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Plan***

***Yucca Mountain Site, Nevada Research
and Development Area, Nevada***

Volume II

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DOCUMENT ORGANIZATION

Introduction

Part A: Description of the mined geologic disposal system

Introduction

Chapter 1--Geology

Chapter 2--Geoengineering

Chapter 3--Hydrology

Chapter 4--Geochemistry

Chapter 5--Climatology and meteorology

Chapter 6--Conceptual design of a repository

Chapter 7--Waste package

Part B: Site characterization program (Chapter 8)

8.0 Introduction

8.1 Rationale

8.2 Issues

8.3 Planned tests, analyses, and studies

8.3.1 Site program

8.3.1.1 Site overview

8.3.1.2 Geohydrology

8.3.1.3 Geochemistry

8.3.1.4 Rock characteristics

8.3.1.5 Climate

8.3.1.6 Erosion

8.3.1.7 Rock dissolution

8.3.1.8 Postclosure tectonics

8.3.1.9 Human interference

8.3.1.10 Population density and distribution

8.3.1.11 Land ownership and mineral rights

8.3.1.12 Meteorology

8.3.1.13 Offsite installations

8.3.1.14 Surface characteristics

8.3.1.15 Thermal and mechanical properties

8.3.1.16 Preclosure hydrology

8.3.1.17 Preclosure tectonics

8.3.2 Repository program

8.3.2.1 Repository overview

8.3.2.2 Configuration of underground facilities (postclosure)

8.3.2.3 Repository design criteria for radiological safety

8.3.2.4 Nonradiological health and safety

8.3.2.5 Preclosure design and technical feasibility

8.3.3 Seal program

8.3.3.1 Seal overview

8.3.3.2 Seal characteristics

CONSULTATION DRAFT

DOCUMENT ORGANIZATION (continued)

- 8.3.4 Waste package
 - 8.3.4.1 Waste package overview
 - 8.3.4.2 Waste package characteristics (postclosure)
 - 8.3.4.3 Waste package characteristics (preclosure)
 - 8.3.4.4 Waste package production technologies
- 8.3.5 Performance assessment program
 - 8.3.5.1 Strategy for preclosure performance assessment
 - 8.3.5.2 Waste retrievability
 - 8.3.5.3 Public radiological exposures--normal conditions
 - 8.3.5.4 Worker radiological safety--normal conditions
 - 8.3.5.5 Accidental radiological releases
 - 8.3.5.6 Higher-level findings--preclosure radiological safety
 - 8.3.5.7 Higher-level findings--ease and cost of construction
 - 8.3.5.8 Strategy for postclosure performance assessment
 - 8.3.5.9 Containment by waste package
 - 8.3.5.10 Engineered barrier system release rates
 - 8.3.5.11 Seal performance
 - 8.3.5.12 Ground-water travel time
 - 8.3.5.13 Total system performance
 - 8.3.5.14 Individual protection
 - 8.3.5.15 Ground-water protection
 - 8.3.5.16 Performance confirmation
 - 8.3.5.17 NRC siting criteria
 - 8.3.5.18 Higher level findings--postclosure system and technical guidelines
 - 8.3.5.19 Completed analytical techniques
 - 8.3.5.20 Analytical techniques requiring development
- 8.4 Planned site preparation activities
- 8.5 Milestones, decision points, and schedule
- 8.6 Quality assurance program
- 8.7 Decontamination and decommissioning

Glossary and acronyms

TABLE OF CONTENTS

	<u>Page</u>
CHAPTER 3 HYDROLOGY	3- 1
Introduction	3- 1
3.1 Description of surface hydrology	3- 8
3.2 Floods	3- 13
3.2.1 Flood history and potential for future flooding	3- 14
3.2.1.1 Ongoing and future studies of flood and debris hazard potentials	3- 22
3.2.2 Flood protection	3- 22
3.3 Locations and distances to points of surface-water use	3- 23
3.3.1 Present quantity and quality of surface water extracted	3- 24
3.3.2 Projected surface-water uses	3- 24
3.4 Chemical composition of adjacent watercourses	3- 25
3.5 Points of ground-water discharge	3- 28
3.5.1 Springs, seeps, and phreatophyte areas	3- 34
3.5.2 Potential for contamination of surface waters and ground waters	3- 49
3.6 Regional hydrogeologic reconnaissance of candidate area and site	3- 50
3.6.1 Hydrogeologic units	3- 50
3.6.1.1 Definition	3- 51
3.6.1.2 Selection of units	3- 51
3.6.1.3 Valley fill aquifer	3- 60
3.6.1.4 Volcanic rock aquifers and aquitards	3- 60
3.6.1.5 Upper carbonate aquifer	3- 60
3.6.1.6 Upper clastic aquitard	3- 60
3.6.1.7 Lower carbonate aquifer	3- 60
3.6.1.8 Lower clastic aquitard	3- 61
3.6.1.9 Spatial relationships of units	3- 61
3.6.2 Relationship among hydrogeologic units	3- 61
3.6.3 Potentiometric levels	3- 62
3.6.4 Hydraulic characteristics of principal hydrogeologic units	3- 64
3.7 Regional ground-water flow system	3- 69
3.7.1 Identification of recharge and discharge areas	3- 70
3.7.1.1 Location of ground-water recharge and discharge areas	3- 70
3.7.1.2 Model of recharge and discharge	3- 70
3.7.1.3 Residence times of the ground water	3- 72
3.7.1.4 Bulk rates of ground-water flow	3- 74
3.7.1.5 Surface-water ground-water interrelations	3- 74
3.7.2 Principal ground-water flow paths	3- 74
3.7.2.1 Oasis Valley subbasin	3- 76
3.7.2.2 Ash Meadows subbasin	3- 77
3.7.2.3 Alkali Flat-Furnace Creek Ranch subbasin	3- 79
3.7.3 Isotopic and regional hydrochemistry	3- 79
3.7.3.1 Regional hydrochemistry	3- 80
3.7.3.1.1 Chemical composition of ground water	3- 80
3.7.3.1.2 Recharge and discharge mechanisms	3- 87
3.7.3.1.3 Hydrochemical facies	3- 94
3.7.3.1.4 Summary	3- 95
3.7.3.2 Regional isotope hydrology	3- 96

CONSULTATION DRAFT

TABLE OF CONTENTS (continued)

	<u>Page</u>
3.7.3.2.1 Tuff aquifer	3- 96
3.7.3.2.2 Carbonate aquifer	3-102
3.7.4 Paleohydrology	3-103
3.7.4.1 Lowering of ground-water levels during the Quaternary Period	3-103
3.7.4.2 Ground-water flow paths during the Quaternary Period . .	3-107
3.7.4.3 Late Wisconsin lakes, marsh deposits, and ground-water levels	3-109
3.7.4.4 Estimation of future ground-water levels	3-110
3.7.4.5 Ground-water recharge during late Wisconsin time: carbon-14 evidence	3-111
3.7.4.6 Conclusions	3-112
3.8 Ground-water uses	3-113
3.8.1 Regional aquifers used for human activities	3-113
3.8.1.1 Alkali Flat-Furnace Creek Ranch subbasin	3-118
3.8.1.2 Ash Meadows subbasin	3-121
3.8.1.3 Oasis Valley subbasin	3-125
3.8.1.4 Proposed water use at Yucca Mountain	3-128
3.8.1.5 Water use for energy development	3-130
3.8.2 Regional ground-water management plans	3-133
3.8.2.1 Ground-water management in Nevada	3-133
3.8.2.2 Local ground-water supply problems	3-135
3.9 Site hydrogeologic system	3-136
3.9.1 Baseline monitoring	3-146
3.9.1.1 Monitoring networks	3-149
3.9.1.1.1 Unsaturated zone	3-149
3.9.1.1.2 Saturated zone	3-153
3.9.1.2 Potentiometric levels	3-159
3.9.1.2.1 Unsaturated zone	3-160
3.9.1.2.2 Saturated zone	3-163
3.9.1.3 Hydrochemistry	3-164
3.9.2 Hydraulic characteristics	3-168
3.9.2.1 Hydraulic characteristics of the unsaturated zone	3-168
3.9.2.2 Hydraulic characteristics of the saturated zone	3-178
3.9.2.2.1 Hydraulic conductivity and fractures	3-178
3.9.2.2.2 Transmissivity and hydraulic conductivity	3-186
3.9.2.2.3 Porosity and storage coefficients	3-191
3.9.3 Ground-water flow system conceptual model	3-192
3.9.3.1 Accessible environment and credible pathways	3-198
3.9.3.2 Potentiometric levels and head relationships	3-198
3.9.3.2.1 Unsaturated zone	3-199
3.9.3.2.2 Saturated zone	3-200
3.9.3.3 Recharge-discharge and leakage	3-202
3.9.3.4 Unsaturated-zone relationships	3-207
3.9.4 Ground-water velocity and travel time	3-213
3.9.4.1 Ground-water travel time in the unsaturated zone	3-214
3.9.4.2 Ground-water travel time in the saturated zone	3-218
3.9.4.3 Retardation and thermal effects	3-220
3.9.5 Hydrochemical confirmation of ground-water behavior	3-221

TABLE OF CONTENTS (continued)

	<u>Page</u>
3.9.5.1 Unsaturated zone	3-221
3.9.5.2 Saturated zone	3-222
3.9.6 Monitoring and verification	3-223
3.9.7 Local ground-water users	3-224
3.9.8 Paleohydrology	3-226
3.10 Summary	3-230
3.10.1 Summary of significant results	3-230
3.10.2 Relation to design	3-236
3.10.3 Identification of information needs	3-237
3.10.4 Relation to Regulatory Guide 4.17	3-240
References for Chapter 3	3R- 1
 CHAPTER 4 GEOCHEMISTRY	 4- 1
Introduction	4- 1
Issues	4- 1
Past data collection	4- 3
Concepts of the site that influence geochemical data collection	4- 3
Present state of data and models	4- 5
4.1 Geochemistry of the host rock and surrounding units	4- 6
4.1.1 Mineralogy and petrology	4- 6
4.1.1.1 General description of the host rock and surrounding units	4- 7
4.1.1.2 Analytical techniques	4- 8
4.1.1.3 Mineralogical, petrologic, and chemical composition of the host rock and surrounding units	4- 9
4.1.1.3.1 The potential host rock	4- 12
4.1.1.3.1.1 Mineralogy of the host rock matrix	4- 12
4.1.1.3.1.2 Mineralogy of fractures in the host rock	4- 12
4.1.1.3.2 Surrounding units	4- 13
4.1.1.4 Mineral stability	4- 36
4.1.2 Ground-water chemistry	4- 39
4.1.2.1 General description of the hydrochemistry	4- 39
4.1.2.2 Major inorganic content	4- 39
4.1.2.3 Trace elements	4- 45
4.1.2.4 Organic content	4- 45
4.1.2.5 Dissolved gas	4- 46
4.1.2.6 Background radioactivity	4- 46
4.1.2.7 Particulates and colloids	4- 46
4.1.2.8 Temperature and pressure	4- 47
4.1.2.9 Mineralogical controls on water composition	4- 49
4.1.2.10 Reference ground-water composition	4- 50
4.1.3 Geochemical retardation processes	4- 51
4.1.3.1 General description of geochemical retardation	4- 52
4.1.3.1.1 Key radionuclides	4- 54
4.1.3.2 Analytical techniques	4- 59
4.1.3.2.1 Preparation of ground-water samples for use in laboratory experiments	4- 61
4.1.3.2.2 Batch sorption methods	4- 62
4.1.3.2.3 Crushed-tuff column method	4- 63

CONSULTATION DRAFT

TABLE OF CONTENTS (continued)

	<u>Page</u>
4.1.3.2.4 Circulating system method	4- 64
4.1.3.2.5 Solubility	4- 64
4.1.3.3 Sorption	4- 64
4.1.3.3.1 Sorption data for tuff	4- 65
4.1.3.3.2 Sorption data from batch experiments	4- 65
4.1.3.3.3 Sorption data from experiments with crushed-tuff columns	4- 82
4.1.3.3.4 Sorption data from circulating-system experiments . .	4- 82
4.1.3.3.5 Comparison of sorption ratios from batch, circulating system, and column measurements	4- 82
4.1.3.3.6 Sorptive behavior as a function of stratigraphic position and mineralogy	4- 82
4.1.3.3.7 Sorptive behavior as a function of ground-water composition	4- 87
4.1.3.4 Processes affecting radionuclide concentrations and speciation in solution	4- 88
4.1.3.4.1 Processes affecting waste element concentrations in solution	4- 91
4.1.3.4.1.1 Speciation	4- 91
4.1.3.4.1.2 Precipitation	4- 91
4.1.3.4.1.3 Natural-colloid formation	4- 92
4.1.3.4.1.4 Radiolysis	4- 92
4.1.3.4.2 Solubility and speciation data	4- 93
4.1.3.4.2.1 Solubility of americium, plutonium, and neptunium in well J-13 water	4- 93
4.1.3.4.2.2 Solid waste dissolution experiments	4- 95
4.1.3.4.2.3 Literature data for waste elements	4- 95
4.1.3.4.3 Solubility modeling	4- 98
4.1.3.4.4 Solubilities of waste elements in Yucca Mountain water	4- 99
4.1.3.5 Matrix diffusion	4- 99
4.1.3.5.1 Diffusion into the rock matrix	4-101
4.1.3.5.2 Retardation by matrix diffusion during fracture flow	4-102
4.1.3.6 Radionuclide transport	4-105
4.1.3.6.1 Transport of suspended solids	4-106
4.1.3.6.2 Gaseous transport	4-108
4.1.3.7 Geochemical retardation in the host rock and surrounding units--anticipated conditions	4-109
4.1.3.8 Geochemical retardation in the host rock and surrounding units--unanticipated conditions	4-114
4.2 Geochemical effects of waste emplacement	4-116
4.2.1 Anticipated thermal conditions resulting from waste emplacement	4-116
4.2.2 Hydrothermal alteration due to the thermal pulse	4-119
4.2.2.1 Hydrothermal alteration of zeolites	4-119
4.2.2.2 Hydrothermal alteration of smectites	4-121
4.2.2.3 Hydrothermal alteration of rhyolitic glasses	4-121
4.2.3 Changes in water chemistry due to the thermal pulse	4-122

TABLE OF CONTENTS (continued)

	<u>Page</u>
4.2.4 Effects of the thermal pulse on radionuclide migration	4-123
4.3 Natural analogs and related field tests	4-126
4.3.1 Natural analogs	4-129
4.3.1.1 Warm and hot springs	4-129
4.3.1.2 Uranium and thorium ore deposits and their host rocks . .	4-130
4.3.1.2.1 The Oklo uranium deposit	4-131
4.3.1.2.2 Other uranium and thorium deposits	4-132
4.3.2 Related field tests	4-134
4.3.2.1 Cambrian site	4-134
4.3.2.2 Los Alamos DP site	4-138
4.4 Geochemical stability	4-141
4.4.1 Man-induced effects	4-141
4.4.2 Potential effects of natural changes	4-143
4.5 Summary	4-145
4.5.1 Summary of significant results	4-145
4.5.2 Relation to design	4-146
4.5.3 Identification of investigations	4-146
4.5.4 Relation to Regulatory Guide 4.17	4-152
References for Chapter 4	4R- 1
 CHAPTER 5 CLIMATOLOGY AND METEOROLOGY	 5- 1
Introduction	5- 1
5.1 Recent climate and meteorology	5- 8
5.1.1 Climate	5- 9
5.1.1.1 Temperature	5- 14
5.1.1.2 Precipitation	5- 16
5.1.1.3 Atmospheric moisture	5- 20
5.1.1.4 Wind speed and direction	5- 21
5.1.1.5 Upper air data	5- 25
5.1.1.6 Severe weather and obstructions to visibility	5- 25
5.1.2 Local and regional meteorology	5- 30
5.1.3 Site meteorological measurement program	5- 36
5.2 Long-term climatic assessment	5- 37
5.2.1 Paleoclimatology	5- 39
5.2.1.1 Quaternary global paleoclimate	5- 39
5.2.1.2 Quaternary regional paleoclimate	5- 41
5.2.1.2.1 Historical lake, climatic, and hydrologic data . . .	5- 42
5.2.1.2.2 Prehistoric lake-level fluctuations	5- 56
5.2.1.2.3 Time series of prehistoric vegetation change	5- 71
5.2.1.2.4 Glacial and periglacial records and their relation to lake levels	5- 89
5.2.1.2.5 Regional paleoclimatic hypotheses	5- 90
5.2.2 Future climatic variation	5- 91
5.2.2.1 Components of the climate system	5- 92
5.2.2.2 Climatic variation	5- 93
5.2.2.3 Climate prediction methods	5- 95
5.2.2.3.1 Prediction of boundary conditions	5- 95
5.2.2.3.2 The prediction of slowly varying components	5- 96
5.2.2.3.3 Prediction of fast-response components	5- 97

CONSULTATION DRAFT

TABLE OF CONTENTS (continued)

	<u>Page</u>
5.2.2.3.4 Spatial disaggregation of predictions	5- 98
5.2.2.3.5 Model validation	5- 99
5.2.3 Site paleoclimatic investigations	5-101
5.2.3.1 Synoptic characterizations of regional climate	5-101
5.2.3.2 Collection and analysis of pollen and pack rat midden macrofossil assemblages from the Great Basin . . .	5-101
5.2.3.3 Coring and paleontological, mineralogical, chemical, and chronological analysis of lacustrine and marsh deposits	5-101
5.2.3.4 Development, validation, and application of a generalized model of late-Quaternary climatic variations to predict future climate change at the Yucca Mountain site	5-102
5.3 Summary	5-102
5.3.1 Summary of significant results	5-102
5.3.2 Relation to design	5-105
5.3.3 Identification of investigations and information needs . . .	5-105
5.3.4 Relation to Regulatory Guide 4.17	5-106
References for Chapter 5	5R- 1

LIST OF FIGURES

<u>Figure</u>	<u>Title</u>	<u>Page</u>
3-1	Hydrographic study area, showing the eight hydrographic areas and major stream channels	3- 2
3-2	Hydrogeologic study area, showing three groundwater subbasins	3- 3
3-3	Locations of crest-stage sites in the hydrographic study area and adjacent areas	3- 12
3-4	Yucca Mountain area and current streamflow measurement sites	3- 17
3-5	Flood-prone areas in the vicinity of Fortymile Wash . .	3- 19
3-6	Surface-water sample sites listed in Table 3.3	3- 26
3-7	Hydrogeologic study area, showing precipitation, recharge areas, and discharge areas	3- 48
3-8	Generalized distribution of hydrogeologic units in the saturated zone of the region	3- 59
3-9	Potentiometric surface of the hydrogeologic study area (based on composite water levels)	3- 63
3-10	Regional ground-water flow paths	3- 75
3-11	Location map showing regional physiographic features . .	3- 81
3-12	Piper diagram showing the regional sodium bicarbonate character of the tuff aquifer	3- 84
3-13	Piper diagram showing the regional mixed character of the lower carbonate aquifer	3- 86
3-14	Piper diagram showing the regional character of the valley fill aquifer	3- 88
3-15	General location of Oasis Valley, Pahute Mesa, and Gold Flat ground-water sampling areas	3- 89
3-16	Concentration trends of sodium, bicarbonate, and chloride	3- 90
3-17	Location of Amargosa Valley (AM) ground-water sample wells	3- 98

CONSULTATION DRAFT

LIST OF FIGURES (continued)

<u>Figure</u>	<u>Title</u>	<u>Page</u>
3-18	$\delta D - \delta^{18}O$ plot of waters from Yucca Mountain area, with radiocarbon content (in percent modern carbon (pmc) units) shown at the corresponding $\delta D - \delta^{18}O$ coordinate	3-100
3-19	Map of south-central Great Basin showing location of uranium-series dated veins	3-104
3-20	Diagrammatic section illustrating effects of possible past or future pluvial-related water-table rise on length of ground-water flow path from Frenchman Flat to points of natural discharge at Ash Meadows in the Amargosa Desert	3-108
3-21	Map showing areas of heavy withdrawals near Yucca Mountain	3-115
3-22	Index map showing water well locations and area boundaries at the Nevada Test Site	3-116
3-23	Location of Yucca Mountain with respect to the relevant hydrographic areas of Death Valley ground-water system and the hydrogeologic study area . .	3-117
3-24	Potentiometric maps of the Amargosa Desert (valley fill aquifer) based on the well data from (1962) (A) and (1984) (B)	3-132
3-25	Simplified stratigraphy of section across Yucca Mountain showing stratigraphic relationships (see Figure 3-28 for location of section)	3-138
3-26	Conceptual east-west hydrogeologic section through the unsaturated zone at Yucca Mountain	3-144
3-27	Locations of drilled monitoring boreholes and core sampling boreholes in the unsaturated zone at Yucca Mountain	3-147
3-28	Preliminary composite potentiometric-surface map of the unsaturated zone, Yucca Mountain	3-148
3-29a	Instrumentation of monitoring borehole USW UZ-1 (0- to 623-feet depth)	3-151
3-29b	Instrumentation of monitoring borehole USW UZ-1 (623-1217.5 feet depth)	3-152

CONSULTATION DRAFT

LIST OF FIGURES (continued)

<u>Figure</u>	<u>Title</u>	<u>Page</u>
3-30	Matric-potential profiles for selected dates at monitoring borehole USW UZ-1 based on data from heat-dissipation probes	3-161
3-31	Matric-potential profiles for selected dates at monitoring borehole USW UZ-1 based on data from thermocouple psychrometers	3-162
3-32	Geochemical facies in the tuff formation in the immediate area around Yucca Mountain, with Stiff diagrams indicating the composition of six samples . . .	3-165
3-33	Moisture-retention curves for the hydrogeologic units within the unsaturated zone at Yucca Mountain	3-171
3-34	Idealized hydraulic-conductivity characteristic curve for a composite (fracture-matrix) porous medium	3-174
3-35	Moisture content of drill cuttings from borehole USW UZ-1	3-175
3-36	Moisture content of drill cuttings from borehole USW UZ-6	3-177
3-37	Computed capacitance coefficient versus pressure head for the Topopah Spring welded (TSw) hydrogeologic unit	3-179
3-38	Borehole-flow survey of drillhole UE-25b#1 (when hole was at total depth of 1,220 m)	3-185
3-39	Vertical distribution of permeable zones in selected wells in the Yucca Mountain area	3-190
3-40	Generalized east-west section through Yucca Mountain showing conceptual moisture-flow system under natural conditions	3-193
3-41	Isopach contour maps within the repository block	3-197
3-42	Standard theoretical flow pattern and boundaries between different flow systems	3-201
3-43	Location of regional and subregional modeled areas, with generalized ground-water flow directions	3-205
3-44	Locations and estimates of ground-water flow into and out of the Alkali Flat-Furnace Creek Ranch subbasin . .	3-206

CONSULTATION DRAFT

LIST OF FIGURES (continued)

<u>Figure</u>	<u>Title</u>	<u>Page</u>
3-45	Numerical grid used in the two-dimensional flow simulations	3-210
3-46	Conceptual hydrogeologic section from Solitario Canyon northwest of the site, to well J-13 in Jackass Flats	3-215
3-47	Water levels in well J-13 since completion	3-227
4-1	Abundance of minerals as a function of depth from surface; determined by x-ray diffraction for core samples from drillhole USW G-2	4- 10
4-2	Abundance of minerals as a function of depth from surface; determined by x-ray diffraction for core samples from drillholes USW GU-3 and USW G-3	4- 11
4-3	Sequence of fraction coatings in the unsaturated zone of drillhole USW G-4	4- 14
4-4	Location map for the zeolite cross sections in Figures 4-5, 4-6, and 4-7	4- 22
4-5	Major zeolite distributions north-south cross section, shown as a dark pattern overlying the stratigraphic cross section of Scott and Bonk (1984)	4- 24
4-6	Major clinoptilolite-mordenite distributions, northwest-southeast cross section	4- 25
4-7	Major zeolite distributions, center to southeast cross section	4- 26
4-8	Histograms comparing the silicon-to-aluminum ratios for heulandites and calcic clinoptilolites from Yucca Mountain, Nevada	4- 31
4-9	Triangular diagram showing the exchangeable-cation ratios in calcic-clinoptilolite and heulandite from zeolitized interval 1 at Yucca Mountain	4- 33
4-10	Triangular diagrams showing ratios of exchangeable cations in clinoptilolites at Yucca Mountain	4- 34
4-11	Selected drillhole and well locations on and near Yucca Mountain exploration block	4- 42
4-12	Relative sodium-potassium-calcium concentration in Yucca Mountain water	4- 44

LIST OF FIGURES (continued)

<u>Figure</u>	<u>Title</u>	<u>Page</u>
4-13	Plot of temperature versus depth in drillholes deeper than 600 m for Yucca Mountain and nearby areas	4- 48
4-14	Variations in the sorptive behavior of strontium and cesium with the abundance of clinoptilolite and stratigraphic depth	4- 84
4-15	Cesium diffusion with slow sorption (tuff sample number G1-1883)	4-103
4-16a	The movement of a 10,000-yr pulse of technetium-99 for $V = 4.5\text{mm/yr}$	4-110
4-16b	The movement of a 10,000-yr pulse of technetium-99 for $V = 0.5\text{ mm/yr}$	4-111
4-17a	Uranium-238 normalized mass flux entering water table for $V = 4.5\text{ mm/yr}$	4-112
4-17b	Uranium-238 normalized mass flux entering water table for $V = 0.5\text{ mm/yr}$	4-113
4-18	Fluid temperature profile showing temperature contours 50 yr after emplacement of a 50 kw/acre heat load using the WAFE code	4-118
4-19	Calculated dissolved silicon dioxide distribution at selected times	4-124
4-20	Calculated total silicon dioxide (dissolved and precipitated) at selected times	4-125
4-21	Calculated chloride distribution at selected times . . .	4-127
4-22	Calculated permeability assuming relationships between pore volume and permeability	4-128
4-23	Schematic of Cambrian explosion cavity and drillholes RNM-1 and RNM-2S	4-135
4-24	Chlorine-36 and tritium concentrations in well RNM-2S water	4-137
4-25	Concentration of plutonium as a function of sampling depth for absorption beds 1 and 2 in 1978	4-139
4-26	Concentration of americium-241 as a function of sampling depth for absorption beds 1 and 2 in 1978 . . .	4-140

CONSULTATION DRAFT

LIST OF FIGURES (continued)

<u>Figure</u>	<u>Title</u>	<u>Page</u>
4-27	The measured and simulated plutonium concentration profiles for the region below bed 1 at the DP West sites	4-142
5-1	Meteorological monitoring stations in the vicinity of Yucca Mountain	5- 3
5-2	Monthly average precipitation (in inches) for stations in the vicinity of Yucca Mountain	5- 19
5-3	Seasonal and annual surface wind distributions for Yucca Flat (1961-1978)	5- 23
5-4	Seasonal and annual wind distributions at 5,000 ft (1,524 m) above mean sea level (328 m above ground level) for Yucca Flat (1957 to 1964)	5- 27
5-5	Seasonal and annual wind distributions at 6,000 ft (1,829 m) above mean sea level (633 m above ground level) for Yucca Flat (1957 to 1964)	5- 29
5-6	Distributions for Pasquill stability classes A, B, and C for Yucca Flat (1961 to 1978)	5- 33
5-7	Distributions for Pasquill stability classes D, E, and F for Yucca Flat (1961 to 1978)	5- 34
5-8	Time lines indicating geologic time periods discussed in text	5- 38
5-9	Great Basin lakes of presumed Pleistocene age	5- 43
5-10	Historical and pristine (corrected) records of fluctuations in the level of Great Salt Lake, Utah	5- 46
5-11	Active river streamflow-gaging and precipitation collection stations located in Pleistocene Lake Lahonton	5- 47
5-12	Precipitation records for Fallon, Nevada and Tahoe City, California, compared with Carson River discharge, 1942 to 1983	5- 48
5-13	Relation of mean monthly evaporation and temperature in the western United States	5- 53
5-14	Historical fluctuations in the levels of Great Salt, Walker, Pyramid, Mono, and Winnemucca Dry lakes	5- 55

LIST OF FIGURES (continued)

<u>Figure</u>	<u>Title</u>	<u>Page</u>
5-15	Lake Bonneville subbasins	5- 59
5-16	The last cycle in the Lake Bonneville Basin as interpreted by (A) Currey and Oviatt (1985), (B) Scott et al. (1983), (C) Spencer et al. (1984), and (D) Morrison (1965)	5- 60
5-17	Surface extent of Lake Lahontan and geography of its basin 14,000 to 12,500 yr before present	5- 61
5-18	Lake Lahontan chronology inferred from data of Born (1972), Benson (1981), and Thompson et al. (1986)	5- 63
5-19	Location of lakes in the Owens River system during pluvial periods in the late pleistocene	5- 64
5-20	Overflow sequence of lakes in the Owens River system . .	5- 65
5-21	Chronologies of lakes in the Russell (Mono) and Searles basins	5- 66
5-22	Summary of carbon-14 dates on subsurface samples composed of organic carbon and wood from Searles Lake	5- 68
5-23	Generalized relationships among elevation, precipita- tion, and the distribution of plant communities for the Great Basin portion of Nevada	5- 74
5-24	Locations of study sites for pollen and pack rat middens listed in Table 5-13	5- 76

CONSULTATION DRAFT

LIST OF TABLES

<u>Table</u>	<u>Title</u>	<u>Page</u>
3-1	Hydrographic areas in the hydrographic study area . . .	3- 9
3-2	Summary of peak streamflow data for selected crest-stage sites in hydrographic study area and adjacent areas	3- 10
3-3	Chemical composition of watercourses adjacent to Yucca Mountain	3- 27
3-4	Ground-water discharge from the Amargosa Desert and Oasis Valley	3- 30
3-5	Springs, seeps, and phreatophyte areas in the hydrogeologic study area	3- 36
3-6	Magnitude of springs in the hydrogeologic study area based on Meinzer's classification of spring discharge	3- 49
3-7	Relation of stratigraphic units to hydrogeologic units in the hydrogeologic study area	3- 52
3-8	Pumping-test data for aquifers in Nevada Test Site and vicinity	3- 65
3-9	Major ground-water discharge in the hydrogeologic study area	3- 72
3-10	Summary of fluxes into and out of the Alkali Flat-Furnace Creek Ranch subbasin	3- 80
3-11	Environmental isotope data for ground-water samples from the samples from the tuff and tuffaceous valley fill aquifers in the region near Yucca Mountain	3- 97
3-12	Environmental isotope data for ground-water samples from the carbonate aquifer and from mixed carbonate-tuff sources	3-101
3-13	Public water suppliers in the community of Amargosa Valley	3-119
3-14	Summary of wells drilled in the Amargosa Desert (hydrographic area 230) according to defined use	3-119
3-15	Nevada Test Site water wells located in the Alkali Flat-Furnace Creek Ranch subbasin	3-120

CONSULTATION DRAFT

LIST OF TABLES (continued)

<u>Table</u>	<u>Title</u>	<u>Page</u>
3-16	Perennial yields, total appropriations, and actual water use for 1985 in the hydrographic areas making up the Alkali Flat-Furnace Creek Ranch Basin	3-122
3-17	Nevada Test Site water wells located in the Ash Meadows subbasin	3-124
3-18	Perennial yields, total appropriations and actual water use for 1985 in the hydrographic areas making up the Ash Meadows subbasin	3-126
3-19	Perennial yields, total appropriations and actual water use for 1985 in the hydrographic areas making up the Oasis Valley basin	3-129
3-20	Estimated water use for exploratory shaft facilities at Yucca Mountain, Nevada	3-131
3-21	Existing total water use in the communities located in the hydrogeologic study area	3-134
3-22	Definition of unsaturated-zone hydrogeologic units and correlation with rock-stratigraphic units	3-139
3-23	Completion records of unsaturated-zone boreholes	3-154
3-24	Ground-water levels, Yucca Mountain area	3-156
3-25	Environmental isotope data for ground-water samples from the tuff aquifers beneath Yucca Mountain	3-167
3-26	Summary of compilations of hydrogeologic properties of hydrogeologic units within the unsaturated zone, Yucca Mountain	3-169
3-27	Preliminary summary of hydrologic characteristics of major stratigraphic units in the vicinity of Yucca Mountain	3-181
3-28	Transmissivity and average hydraulic conductivity at well USW H-1, based on pumping tests, borehole-flow surveys, and slug-injection tests	3-187
3-29	Comparison of hydraulic-conductivity values from in situ tests with core-sample analyses.	3-188
3-30	Parameters used in travel-time calculations for the unsaturated zone	3-216

CONSULTATION DRAFT

LIST OF TABLES (continued)

<u>Table</u>	<u>Title</u>	<u>Page</u>
3-31	Summary of unsaturated-zone travel time for vertical flux of 0.5 mm/yr	3-219
3-32	Estimates for ground-water travel times through the saturated zone	3-220
3-33	Pumping test and water-level data for wells J-12 and J-13 from 1960 through 1969	3-225
3-34	Water production from wells J-12 and J-13 from 1983 through 1985	3-225
4-1	Symbols used for petrologic units, rock types, alteration, and minerals listed in Table 4-2	4- 16
4-2	Summary of petrography for core samples from drillhole USW G-1, determined by point count of glass-covered thin sections	4- 19
4-3	Statistical modes for compositional parameters of phenocrysts for the lower Topopah Spring Member and for underlying units	4- 21
4-4	Zeolitized intervals for Yucca Mountain	4- 28
4-5	Representative clinoptilolite analyses for Yucca Mountain	4- 29
4-6	Element concentrations in ground water from the vicinity of Yucca Mountain	4- 40
4-7	Anion concentrations and other measurements for ground water from the vicinity of Yucca Mountain	4- 41
4-8	Background radioactivity	4- 47
4-9	Radionuclides ordered by ratio of inventory to U.S. Environmental Protection Agency limit for PWR spent fuel for various decay times	4- 55
4-10	Radionuclides ordered by ratio of inventory to the U.S. Environmental Protection Agency limit for high-level waste from pressurized water reactors for various decay times	4- 56
4-11	Radionuclides ordered by ratio of inventory to the U.S. Environmental Protection Agency limit for defense high-level waste for various decay times	4- 57

CONSULTATION DRAFT

LIST OF TABLES (continued)

<u>Table</u>	<u>Title</u>	<u>Page</u>
4-12a	Average sorption ratios from batch sorption experiments on crushed tuff for strontium, cesium, barium, radium, cerium, and europium	4- 66
4-12b	Average sorption ratios from batch sorption experiments on crushed tuff for americium, plutonium, uranium, selenium, technetium, and neptunium	4- 68
4-13a	Average sorption ratios from batch desorption experiments on crushed tuff for strontium, cesium, barium, cerium, and europium	4- 71
4-13b	Average sorption ratios from batch desorption experiments on crushed tuff for americium, plutonium, uranium, technetium, and neptunium	4- 73
4-14	Petrologic characterization of tuff samples from drillholes J-13, UE-25a#1, USW G-1, USW G-2, and USW GU-3	4- 76
4-15	Average sorption ratios from batch, circulating-system, and crushed-tuff column sorption measurements at room temperature	4- 83
4-16a	Sorption ratios for cesium and strontium in different waters	4- 89
4-16b	Sorption ratios for barium, europium, and tin	4- 90
4-17	Analytical results of solutions in equilibrium with their solid phase in 0.1 M NaClO ₄ and well J-13 ground water at pH = 7.0 ± 0.1, 25 ± 1°C	4- 94
4-18	Solubility of waste elements in well J-13 water	4-100
4-19	Sorption rate constants for tuff from drillhole USW G-1	4-104
4-20	Behavior of fission products in industrial reactors and in the Oklo natural reactors	4-132
5-1	Information on meteorological monitoring stations in the vicinity of Yucca Mountain	5- 4
5-2	Climatological summary for Yucca Flat, Nevada, 1962 to 1971	5- 12

CONSULTATION DRAFT

LIST OF TABLES (continued)

<u>Table</u>	<u>Title</u>	<u>Page</u>
5-3	Yucca Flat and Beatty temperature data	5- 15
5-4	Monthly and annual average and maximum precipitation for sites in the vicinity of Yucca Mountain	5- 17
5-5	Wet-bulb depression values calculated for Yucca Flat	5- 22
5-6	Yucca Flat upper air data for 5,000 ft (1,524 m) above mean sea level (328 m above ground level)	5- 26
5-7	Yucca Flat upper air data for 6,000 ft (1,829 m) above mean sea level (633 m above ground level)	5- 28
5-8	Annual extreme wind speed at 30 ft (9.1 m) above ground level and return period for Yucca Flat, Nevada	5- 31
5-9	Extreme maximum and minimum temperatures and probability of occurrence for Beatty, Nevada	5- 31
5-10	Maximum 1- and 24-hr precipitation and probability of occurrence for Yucca Flat	5- 32
5-11	Yucca Flat Pasquill stability class distributions for the period 1961 to 1978	5- 35
5-12	Annual equivalent lake surface evaporation data (calculated from pan evaporation data) in centimeters for three Nevada locations	5- 52
5-13	Summary of palynological and pack rat middens data from the Great Basin, Mojave Desert, and Sierra Nevada with inferred climatic changes	5- 78
5-14	Vegetative cover and inferred climatic regime of the Yucca Mountain region	5- 86

Nuclear Waste Policy Act
(Section 113)

HYDROLOGY

Consultation Draft

***Site Characterization
Plan***

***Yucca Mountain Site, Nevada Research
and Development Area, Nevada***

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Chapter 3

HYDROLOGY

This chapter summarizes present knowledge of the regional and site hydrologic systems. The purpose of the information presented is to (1) describe the hydrology based on available literature and preliminary site-exploration activities that have been or are being performed and (2) provide information to be used to develop the hydrologic aspects of the planned site characterization program.

INTRODUCTION

An understanding of the hydrologic regime is required for many of the issues in the Nevada Nuclear Waste Storage Investigations (NNWSI) Project Issues Hierarchy (Section 8.2). In particular, the nine Performance Issues (1.1 through 1.9) of Key Issue 1 (related to the postclosure environment); the Characterization Programs 8.3.1.2 (geohydrology), 8.3.1.3 (geochemistry), and 8.3.1.9 (water resources); the Performance Issue (4.1) of Key Issue 4 (preclosure); and the Characterization Program 8.3.1.16 (hydrology) all pertain directly to hydrology. In addition, other programs are indirectly related to hydrology, (e.g., Programs 8.3.1.5, 8.3.1.6, and 8.3.1.8 question the effects that climatic change, erosional processes, and tectonic processes exert on the hydrologic characteristics at the site). Furthermore, issues related to design (both in the preclosure and postclosure time frames) require a knowledge of the emplacement environment, a major component of which is moisture. Therefore, an understanding of the hydrologic system in and around Yucca Mountain is paramount to satisfying many of the issues in the issues hierarchy.

One of the primary advantages of the proposed Yucca Mountain repository site is that it is intended to be located in the unsaturated zone, approximately 300 m above the saturated zone. This chapter discusses in detail the present understanding of hydrologic conditions within the unsaturated zone, as well as conditions in the saturated zone and in the surface hydrologic systems; the last two act as boundaries to the unsaturated zone.

For convenience in describing the hydrologic regime, two types of study areas are defined in this chapter: (1) a hydrographic study area (Figure 3-1) that delineates the regional surface water system that encompasses Yucca Mountain and (2) a hydrogeologic study area (Figure 3-2), the boundaries of which approximate the regional ground-water flow system surrounding Yucca Mountain. These study areas represent the areas from which published data were compiled to describe the hydrology of the region and site; additional investigations are planned for parts of these study areas, as described in Section 8.3.1.2.

The hydrographic study area is composed of eight smaller hydrographic areas as described by Waddell et al. (1984) (Figure 3-1). Yucca Mountain lies on the boundary of the Crater Flat and Fortymile Canyon-Jackass Flats hydrographic areas. Detailed discussions of the surface-water system within

CONSULTATION DRAFT

HYDROGRAPHIC AREAS

- 1 DEATH VALLEY AND LOWER AMARGOSA AREA
- 2 AMARGOSA DESERT AND UPPER AMARGOSA AREA
- 3 CRATER FLAT

- 4 FORTY MILE CANYON, JACKASS FLATS
- 5 FORTY MILE CANYON, BUCKBOARD MESA
- 6 MERCURY VALLEY
- 7 OASIS VALLEY
- 8 ROCK VALLEY

— BOUNDARY OF HYDROGRAPHIC STUDY AREA

----- MAJOR STREAM CHANNELS

★ YUCCA MOUNTAIN SITE

Figure 3-1. Hydrographic study area, showing the eight hydrographic areas and major stream channels. Modified from Waddell, et al. (1984).

CONSULTATION DRAFT

GENERAL DIRECTION OF REGIONAL GROUND-WATER FLOW
(QUESTION MARK INDICATES UNCERTAINTY)

- A. OASIS VALLEY SUBBASIN
- B. ALKALI FLAT-FURNACE CREEK RANCH SUBBASIN
- C. ASH MEADOWS SUBBASIN

Figure 3-2. Hydrogeologic study area, showing three ground-water subbasins. Modified from Rush (1970), Blankennagel and Weir (1973), Winograd and Thordarson (1975), Dudley and Larsen (1976), Waddell (1982), and Waddell, et al. (1984).

CONSULTATION DRAFT

the hydrographic study area are presented in Section 3.1 (description of surface hydrology), 3.2 (floods), 3.3 (locations and distances to points of surface-water use), and 3.4 (chemical composition of adjacent watercourses). As these sections discuss, the occurrence of surface-water bodies is extremely limited in the vicinity of Yucca Mountain; their primary importance is in their relationship to the ground-water system, as discussed in Section 3.5 (points of ground-water discharge).

The hydrogeologic study area (Figure 3-2) consists of three ground-water subbasins that together form a part of the Death Valley ground-water basin (Waddell et al., 1984). These subbasins are the Oasis Valley subbasin, Alkali Flat-Furnace Creek Ranch subbasin, and the Ash Meadows subbasin. The division into and the approximate boundaries of these ground-water subbasins have been estimated from potentiometric levels, geologic controls of subsurface flow, discharge areas, and inferred flow paths (Rush, 1970; Blankennagel and Weir, 1973; Winograd and Thordarson, 1975; Dudley and Larson, 1976; Waddell, 1982; Waddell et al., 1984). In some areas, the boundaries are uncertain due to the lack of potentiometric data, the complexity of geologic structures, or the occurrence of interbasin flow of ground water through the lower carbonate aquifer (Winograd and Thordarson, 1975). Regional flow paths in the hydrogeologic study area vary in direction from southerly to southwesterly (Winograd and Thordarson, 1975; Waddell, 1982). Detailed discussions of the regional hydrogeology and regional flow systems are included in Sections 3.6 and 3.7. Detailed discussions of the site (Yucca Mountain) hydrogeology are included in Section 3.9. The discussions include aquifer-hydraulics information, water-balance calculations, and water-chemistry data for both the saturated and the unsaturated zones.

Most of the information on the regional hydrology is available in published reports from general ground-water studies made during the 1960s and 1970s by the Nevada Department of Conservation and Natural Resources or by the Department in cooperation with the U.S. Geological Survey (USGS). The studies were used principally for general water-resource appraisals of parts of Nevada. A few studies of surface hydrology were also made, but because the region is generally arid and because streamflow is essentially ephemeral, very few data have been collected.

In 1957 the U.S. Atomic Energy Commission (later the Energy Research and Development Administration and presently the Department of Energy) began underground testing of nuclear devices at the Nevada Test Site (NTS), and therefore needed an evaluation of the occurrence and movement of ground water within and adjacent to the NTS. There was a need to determine both the potential for radionuclide contamination of ground water and location of water supplies. Thus, a program of hydrologic and geologic studies was established between the Atomic Energy Commission and the USGS and has continued since the late 1950s to the present. In a paper published in 1975, Winograd and Thordarson (1975) summarized the current understanding of the regional hydrogeology and flow systems within the hydrogeologic study area. This report is the basis for the more recent hydrologic studies, such as a regional ground-water flow modeling effort by Waddell (1982). Regional ground-water studies for the NNWSI Project have continued because they provide an understanding of the ground-water flow system in the vicinity of Yucca Mountain. Furthermore, regional studies provide a basis for evaluating

future changes in climate and hydrology. Such possible future effects may be important to isolation of waste at Yucca Mountain.

In general, hydrogeologic data for the region are believed to be adequate for regional quantitative ground-water flow modeling, although there remain some hydrologic features that are not completely understood for the purposes of site characterization. Where it is important to resolve such questions (such as the potential for recharge near Fortymile Wash), plans have been developed for additional field studies as described in Section 8.3.1.2.1. Various hypotheses about the ground-water flow systems and probable present and future fluxes are continuing to be evaluated to improve general understanding and to determine probable error bounds for calculated quantities such as ground-water velocities and future water levels. These hypotheses are discussed in Sections 3.6, 3.7, and 3.9.

Most of the hydrologic information about the Yucca Mountain site has been obtained from studies conducted since 1978. During the first few years of these studies, the emphasis was on describing the hydrogeology of the saturated zone. As additional data became available, consideration was given to locating the repository within the unsaturated zone. The unsaturated zone at Yucca Mountain is thick (500 to 750 m at the proposed repository) and offers several possible advantages over the saturated zone, including longer ground-water travel times from the repository to the accessible environment, the existence of a zeolitic tuff along the flow path that may retard radionuclide transport, and the avoidance of difficulties associated with constructing and operating a repository within the saturated zone. Because of the recent shift of investigative emphasis to the unsaturated zone and the lack of previous investigations of thick unsaturated zones, less is known about it than about the saturated zone. However, present and planned studies are designed to develop a comprehensive understanding of the hydrology of the unsaturated zone at Yucca Mountain (Section 8.3.1.2.2).

Beginning in 1981, hydrologic test holes, some as deep as 1.8 km, were drilled into the saturated zone at Yucca Mountain. The holes were logged to determine lithology and stratigraphy of the rocks penetrated and tested to determine such hydrologic parameters as depth to water, total water yield, water yield as a function of stratigraphic horizon, hydraulic conductivity, transmissivity, and water chemistry, including apparent carbon-14 ages of some of the waters. The upper kilometer or more of the saturated zone penetrated by the wells consists of extensively fractured volcanic tuffs that appear to derive most of their permeability from the fractures, rather than from the porosity of a granular matrix.

Hydraulic tests of the saturated zone in these drillholes were evaluated mostly by porous media methods. Depending on the scale of study, such methods may or may not be applicable for evaluating the results of aquifer tests conducted in fractured media. Investigations of the hydraulic behavior of fractured-rock aquifers are still considered developmental; indeed, most of the studies have been initiated in direct response to the needs of this and other radioactive waste repository projects. Three approaches are commonly taken: (1) the continuum approach in which the aquifer is modeled, both with respect to transport of water and solutes, as an equivalent porous medium; (2) the dual porosity approach in which the rock matrix and fractures are treated as coupled flow systems of highly differing properties; and

CONSULTATION DRAFT

(3) the discrete fracture approach in which flow through a regular geometric network of parallel-plate fractures is simulated using fluid-dynamics equations. Newer approaches involve the application of near-field stochastic analyses to predict far-field hydraulic and transport properties. Alternative approaches will be examined as the project develops and as the state of the modeling art is refined (Section 3.9).

In addition, multiple-well tests, some of which involve tagging ground water with tracers to evaluate the tests, will be conducted within the saturated zone. The purpose of these tests is to determine the nature and extent of the permeability contributed by fractures and to determine effective porosity.

Beginning in 1983, test holes were drilled to depths greater than 300 m within the unsaturated zone. The purposes of these boreholes were to obtain rock samples for the determination of hydrologic properties and to monitor ambient water saturation, potential, and flux in rocks above, below, and within the horizon proposed for the repository. Rock samples have been tested to determine effective porosities, ambient saturations and matric potentials, saturated matrix hydraulic conductivities, and water retention characteristics. Ambient water potentials have been monitored in one test well since 1983. The data from the rock samples show a wide variation in hydrologic property values between the various hydrogeologic units within the unsaturated zone. The in situ water potential data show considerable variability; however, the data are consistent with the supposition that the hydrogeologic units are vertically homogeneous and that steady-state vertical moisture flow occurs under unit vertical hydraulic gradients. The mechanisms of moisture flow and storage (as liquid water, water vapor, or both) at any location within the system depend largely upon the amount of water entering the system as net infiltration. The actual rate and distribution of net infiltration across the unsaturated zone is not known, although likely upper bounds have been established. Because little is known about ground-water occurrence and movement within deep zones of unsaturated, fractured tuffs, an extensive program of field investigations and theoretically based studies is planned to provide a comprehensive understanding of the hydrology of the unsaturated zone at Yucca Mountain (Section 8.3.1.2.2).

Uncertainty associated with the data presented in Chapter 3 is primarily related to the lack of a sufficient number of samples or tests for statistical reliability and also to measurement errors that are inherent in the investigations. As with all scientific investigations, however, certain levels of uncertainty associated with the data are generally acceptable. The level of uncertainty that can be considered acceptable is determined by the level to which the parameter must be known based on the sensitivity of the system to that parameter.

The evaluation of uncertainty associated with measured parameters has been addressed, where possible, by planning testing and sampling programs that are structured so that experimental uncertainty and sampling uncertainty are independently or jointly characterized. These programs for site characterization are described in Section 8.3.1.2. Specific levels and types of existing uncertainties or data needs are discussed throughout Chapter 3.

In very general terms, the regional hydrology and hydrogeology are fairly well understood and their levels of uncertainty are relatively low. However, the unsaturated zone at the site has only recently come under investigation and, therefore, has produced a relatively meager data base. There is also a considerable uncertainty regarding the degree to which the limited data base represents the actual range of conditions at the site.

Examples of uncertainties associated with the hydrologic and hydrogeologic data presented in Chapter 3 are as follows:

1. The quantity of streamflow and flood records are insufficient for accurate predictions of future magnitudes and recurrence intervals because of the relatively short period of record.
2. Several portions of the regional flow system are uncertain, including quantities of evapotranspiration in the Amargosa Desert, vertical components of flow within the saturated zone, and the nature of hydrologic boundary conditions at the site (to be provided by regional models). These factors are insufficiently known due to the lack of data collected to this point.
3. Uncertainties exist concerning the site saturated flow system. Included among these are the reasons for the existence of a locally steep hydraulic gradient (upgradient from the site) and the associated difficulties related to characterization of saturated-zone flux within and around this zone. These uncertainties arise from an insufficient data base for proper quantification of ground-water behavior in the area.
4. Knowledge of the nature and quantities of present and future infiltration is insufficient for use in unsaturated-zone flux or travel-time calculations, due to the lack of existing infiltration measurements and the lack of hydrologic data that could be used to estimate infiltration characteristics.
5. Because of the limited number of studies available, projections of future hydrologic conditions are difficult to make based on paleoclimatic and paleohydrologic data.
6. The quantity of water chemistry data, especially from the unsaturated zone, is not sufficient for verification of conceptual-flow models.
7. Ambient hydrologic conditions within the fractured rocks of the unsaturated zone are not well known because of the very limited number of tests and measurements performed to date.
8. Knowledge of the behavior of water and gases within the fractured terrane at Yucca Mountain and the interaction of the matrix and fracture systems are not sufficient due to the limited amount of data collected to this point and the limited amount of relevant information available in the open literature.

9. The conceptual model of ground-water flow through the unsaturated zone at the site has not been developed to a high confidence level as yet because of the limitations of the supporting data.

These and other areas of uncertainty will be significantly reduced through further data collection and through the application of multiple approaches that may produce corroborative results. These programs will be carried out during site characterization and are described in Section 8.3.1.2.

3.1 DESCRIPTION OF SURFACE HYDROLOGY

Characterization of surface hydrology is required to provide baseline data necessary to resolve several issues. Knowledge gained through the surface hydrology studies will be used to (1) evaluate the potential for erosion in the site area and any hazards that this may cause, (2) evaluate the potential for floods and the potential hazards to the surface facilities due to floods (Section 3.2), and (3) help evaluate the amount of infiltration and establish the relationship between surface runoff and ground-water recharge that occurs in the study area.

The hydrographic study area (Figure 3-1) is the same as the Death Valley Basin Hydrographic Region of California and Nevada, as defined by Rush (1968) and the USGS (1978). This study area consists of eight hydrographic areas. The boundary lines of the eight hydrographic areas within the hydrographic study area are drawn principally along topographic divides. These hydrographic areas serve as the basic morphologic unit of the surface-water hydrologic system. A listing of these areas is shown in Table 3-1.

The general configuration of surface drainage of the hydrographic areas is shown in Figure 3-1. The Yucca Mountain site proposed for waste disposal lies on the boundary between the Crater Flat and Fortymile Canyon-Jackass Flats hydrographic areas (Figure 3-1).

The eastern slopes of Yucca Mountain drain to Fortymile Wash and the northern slopes drain to Beatty Wash; these washes are major tributaries to the Amargosa River. The southern and western slopes of Yucca Mountain drain to the Amargosa River through a smaller unnamed drainage system.

There are no perennial streams in or near the Yucca Mountain area. However, the many ephemeral stream channels, including the large drainage systems of Fortymile Wash and the Amargosa River, flow following significant regional or local storms. Although the region that includes Yucca Mountain has a generally arid to semiarid climate that includes high annual average potential evaporation (about 1,500 to 1,700 mm/yr; Kohler et al., 1959), low average annual precipitation (about 150 mm; Quiring, 1983), and infrequent storms (Section 5.1.1.6), surface runoff does occur. Runoff results from regional storms that occur most commonly in autumn, winter, and spring, and from localized thunderstorms that occur mostly during the summer. Rugged relief, abundant bedrock exposed at the land surface, and sparse vegetal cover promote runoff, particularly during intense rainstorms. The annual precipitation pattern usually follows a bimodal distribution, with greatest

Table 3-1. Hydrographic areas in the hydrographic study area

Area number on Figure 3-1	Hydrographic area	State ^{a,b}	<u>Approximate area</u>	
			(mi ²)	(km ²)
1	Death Valley and Lower Amargosa area	Nevada California	344 5,019	891 13,000
2	Amargosa Desert and Upper Amargosa area	Nevada California	896 1,122	2,321 2,906
3	Crater Flat	Nevada	182	471
4	Fortymile Canyon, Jackass Flat	Nevada	279	723
5	Fortymile Canyon, Buckboard Mesa	Nevada	240	622
6	Mercury Valley	Nevada	110	285
7	Oasis Valley	Nevada	460	1,191
8	Rock Valley	Nevada	82	212

^aData for Nevada modified from Rush (1968).

^bData for California from U.S. Geological Survey (1978).

average amounts occurring during the winter storms and less during the summer (Quiring, 1983). Although runoff can result from severe winter storms, the scanty data available for the region suggests that greatest runoff magnitudes commonly result from summer storms (Table 3-2). These few streamflow data available for the general study area were collected to document flooding. Locations of the gaging sites are shown in Figure 3-3, and data collected at these sites summarized in Table 3-2. (These sites and data are also discussed later in Section 3.2.) No data are presently available that relate runoff to recharge (see studies planned for Fortymile Wash area, Section 8.3.1.2.1.3), and the data are insufficient to develop meaningful flood recurrence intervals.

Quantitative data on rainfall, runoff, and evaporation for the area are not yet adequate to determine rainfall-runoff-recharge relations for individual storms, seasons, or years. Therefore, only general knowledge of runoff parameters is available. Numerical simulations of rainfall-runoff relations are possible, but these models cannot be calibrated until more field data become available. See Section 3.2.1.1 for a discussion of investigations

Table 3-2. Summary of peak streamflow data for selected crest-stage sites in hydrographic study area and adjacent areas^a (page 1 of 2)

Site number ^b	Station number	Station name	Drainage area		Period of record	Peak discharge (m ³ /s)	Date of peak discharge	Discharge per unit area ((m ³ /s)/km ²)
			(mi ²)	(km ²)				
1	10247860	Penoyer Valley tributary near Tempiute, Nevada	1.48	3.83	1964-80	3.68	8/06/68	0.96
2	10248490	Indian Springs Valley tributary near Indian Springs, Nevada	29.0	75.1	1964-80	14.1	8/14/72	0.19
3	10251270	Amargosa River tributary near Mercury, Nevada	110.0	284.9	1963-80	97.1	8/04/68	0.34
4	10251271	Amargosa River tributary No. 1 near Johnnie, Nevada	2.21	5.72	1967-80	9.9	8/04/70	1.73
5	10251272	Amargosa River tributary No. 2 near Johnnie, Nevada	2.49	6.45	1968-80	3.54	8/01/68	0.55
6	10251220	Amargosa River near Beatty, Nevada	470.0	1,217.0	1964-79	453.0	2/24/69	0.37

Table 3-2. Summary of peak streamflow data for selected crest-stage sites in hydrographic study area and adjacent areas^a (page 2 of 2)

Site number ^b	Station number	Station name	Drainage area		Period of record	Peak discharge (m ³ /s)	Date of peak discharge	Discharge per unit area ((m ³ /s)/km ²)
			(mi ²)	(km ²)				
7	10249050	Sarcobatus Flat tributary near Springdale, Nevada	37.1	96.1	1961-80	1.78	9/09/80	0.02
8	10249850	Palmetto Wash tributary near Lida, Nevada	4.73	12.25	1967-80	5.46	7/07/69	0.45
9	10248970	Stonewall Flat tributary near Goldfield, Nevada	0.53	1.37	1964-79	4.25	6/16/69	3.10
10	10249680	Big Smoky Valley tributary near Blair Junction, Nevada	11.4	29.5	1961-79	4.81	10/02/76	0.16
11	10249135	San Antonio Wash tributary near Tonopah, Nevada	3.42	8.86	1965-80	18.7	8/13/72	2.10
12	10249180	Salsbury Wash Tonopah, Nevada	56.0	145.0	1962-80	9.62	3/27/69	0.07

^aSource: Squires and Young (1984), and Waddell et al. (1984).

^bLocation of sites is shown in Figure 3-3.

CONSULTATION DRAFT

Figure 3-3. Locations of crest-stage sites in the hydrographic study area and adjacent areas. Modified from Squires and Young (1984).

currently underway to improve the surface-water hydrologic data base at Yucca Mountain and surrounding areas. Plans for future investigations and analysis of surface-water flow in terms of precipitation, runoff, infiltration, evaporation and transpiration, and floods are given in Section 8.3.1.2.1.

Throughout the hydrographic study area, perennial surface water comes only from springs, and it is restricted to some short reaches of the Amargosa River, to source pools at some large springs, and to some marshes around the edge of the salt pan in Death Valley (Hunt et al., 1966). One small lake, locally known as Crystal Reservoir, with a storage capacity of $2.27 \times 10^6 \text{ m}^3$ (Giampaoli, 1986), occurs in the Ash Meadows part of the Upper Amargosa hydrographic area (Figure 3-1). The water for the reservoir is supplied via a concrete flume from Crystal Pool (Giampaoli, 1986).

The Amargosa River originates in Oasis Valley and continues southeastward through the Amargosa Desert past Death Valley Junction then southward another 75 km, where it turns northwestward and terminates in Death Valley (Figure 3-1). The river carries floodwaters following cloudbursts or intense storms but is normally dry, except for a few short reaches that contain water from springs (Walker and Eakin, 1963), in Oasis Valley between Springdale and Beatty (Malmberg and Eakin, 1962), in Ash Meadows northeast of Death Valley Junction, and near Shoshone, about 40 km south of Death Valley Junction. Springs are discussed in greater detail in Sections 3.3 and 3.5. Base flow to these segments of the river is maintained by ground-water discharge during the winter, when evapotranspiration is at a minimum. During the summer, discharge from the springs is almost entirely lost by evapotranspiration. During winter, ground water discharges into the river south of Alkali Flat near Eagle Mountain about 10 km south of Death Valley Junction (Walker and Eakin, 1963).

Because of the ephemeral character of streamflow, specifically in the Yucca Mountain area and generally throughout the Death Valley regional drainage systems, almost no streamflow data have been collected. The quantity of average annual runoff within the eight hydrographic areas in the Death Valley Basin in Nye County, Nevada, was estimated at less than $620,000 \text{ m}^3$ for each separate area (Scott et al., 1971). However, the magnitudes, frequencies, and durations of flows that comprise these average estimates are unknown. A discussion of the total water budget is given in Section 3.5. Plans for collection of streamflow data are discussed in Section 8.3.1.2.1.

Dry washes provide channels that concentrate runoff and may thus be the principal sources of potential modern ground-water recharge to the area. The implications and evidence of possible recharge by surface water are discussed in Section 3.9.3.3.

There are no manmade water-control structures that could influence conditions at the site, and none is known to be proposed.

3.2 FLOODS

Because of limited streamflow data in the hydrographic study area, the flood history of individual drainages tributary to Death Valley is not well

known. Analysis of the general character of flooding is based on scattered data from peak-flow stations located around the Yucca Mountain area and from quantitative measurements of flood-flow peaks in a five-state region that comprises Arizona, California, Nevada, New Mexico, and Utah.

Flood analyses at Yucca Mountain are needed to provide flood data for design and performance considerations. Flood data will also be used to help evaluate infiltration rates and long term erosion rates. Two hydraulic engineering studies of flood-prone areas at and near Yucca Mountain (Christensen and Spahr, 1980; and Squires and Young, 1984) provided a basis for estimating the magnitudes of future floods with various recurrence intervals. The results of these investigations are presented in Section 3.2.1. Analytical methods other than those presented in ANSI/ANS 2.8-1981 (American Nuclear Standards Institute/American Nuclear Society) for estimating the potential for flooding of the site were used in these introductory studies; the reasons for applying alternate methods and a description of the procedures used are given in Section 3.2.1.

The measures that will be used to protect the site from floods are described in Section 3.2.2.

3.2.1 FLOOD HISTORY AND POTENTIAL FOR FUTURE FLOODING

Moderate to large floods in low-lying areas along the major drainages (drainage areas of several hundred square kilometers), such as the Amargosa River and Fortymile Wash, usually are the result of regional storm systems that most commonly occur during the winter and occasionally occur during autumn and spring. These extensive storm systems sometimes include areally restricted cells that discharge intense precipitation.

Flash floods of similar intensity and areal extent commonly occur as the result of summer thunderstorms. These summer floods usually do not cumulate to cause regional floods, but their intensive character renders them potentially destructive over limited areas. The summer storms are commonly the products of monsoonal air masses that invade southern Nevada and California from the general vicinity of the Gulf of California. The areally restricted storm cells are triggered by local convective lifting of the moist air or by cooler frontal systems that move through the region and intersect with the warmer monsoonal air mass. A detailed discussion on climatic conditions including precipitation, temperature, wind, and severe weather phenomena is given in Section 5.1.

In summary, although flooding can occur over an extensive area, intense floods are generally restricted to relatively small areas and occur as flash floods of short duration.

Flash floods constitute a hazard throughout the Great Basin (Chapter 1) and specifically in southern Nevada. These floods and associated debris flows are among the most important geomorphic processes currently active in the region and local area. They play a major role in the development of alluvial fans, denudation of mountainous landscapes, and the evolution of drainage-channel morphology. Flash-flood discharges range in character from

water-dominated mixtures of sediments and water to debris flows. The hazards posed by debris movement associated with flash floods may be of equal or greater importance with regard to destructive potential than that of the mobilizing water. The intensive rainfall and runoff of flash flooding commonly promotes debris avalanching on steep slopes. Erosion scars caused by these debris avalanches are not uncommon in the Yucca Mountain area.

Data have been collected since the early 1960s on flooding and stream-flow in ephemeral stream channels throughout Nevada. A network of crest-stage gages has been operated for almost two decades. Crest-stage gages are simple devices that record evidence of the peak-flow stage of a stream at the specific site where they are located. They do not record the duration of flow associated with the peak stage; also, they are poor indicators of multiple-peak stages that can occur between visits to the site. Also, because the network of gages was not dense, only a small percentage of ephemeral stream channels throughout Nevada has been monitored. Individual gages were visited monthly during this data collection program to ascertain recent occurrences of stream flow at the gage sites. If evidence of recent streamflow was noted during a visit, an indirect measurement of the peak-flow discharge was made and recorded along with the stage of the peak flow. Additional site visits were made when knowledge of recent flooding or major flooding was available. The resultant data base is a reasonably reliable record of the magnitude and frequencies of peak flows at the crest-stage gaging sites during the tenure of operation of the gages.

Twelve crest-stage gaging sites that were part of the monitoring network are located in the general area of Yucca Mountain. Locations of these sites are shown in Figure 3-3, and data collected at the sites are summarized in Table 3-2. Data collection at these sites was discontinued in 1980 when the statewide network was reduced. Stations 2 through 6 (Figure 3-3) have been reactivated as part of ongoing studies (Section 3.2.1.1).

Table 3-2 indicates that maximum unit peak-flood discharges (peak discharge divided by upstream drainage areas) measured at the 12 gaging stations during about a decade and a half ranged from approximately $0.07 \text{ (m}^3/\text{s)/km}^2$ for drainage areas greater than 100 km^2 to more than $3.0 \text{ (m}^3/\text{s)/km}^2$ for drainage areas less than 10 km^2 . This evidence of recent flooding indicates that occasional locally intense runoff probably will occur in the future within southern Nevada, and may occur in washes at Yucca Mountain.

Several major floods that occurred in some of the washes from 1964 through 1980 are shown in Table 3-2. The most notable flood was associated with a large winter storm that occurred over the upper Amargosa River drainage basin (drainage area, $1,217 \text{ km}^2$) in February 1969. This storm caused an estimated peak flow of $453 \text{ m}^3/\text{s}$ near Beatty with a peak discharge per unit area of $0.37 \text{ (m}^3/\text{s)/km}^2$ (estimated from records for site No. 6). The upper Amargosa drainage basin is immediately west of Fortymile Wash basin, and the two basins have similar physiographic terrains. A peak flow of $97.1 \text{ m}^3/\text{s}$ with a peak discharge per unit area of $0.34 \text{ (m}^3/\text{s)/km}^2$ occurred in August 1968 at site No. 3 in an unnamed tributary to the Amargosa River near Mercury. This drainage basin has an area of 284.9 km^2 ; the basin terrain is also similar to that of Fortymile Wash.

CONSULTATION DRAFT

Long-range flood predictions are difficult to make, even for drainages that have stream-flow records as long as 100 yr. Predictions are especially difficult for drainages with minimal stream-flow records, such as those in the hydrographic study area. Current flood-prediction methods for this area generally involve some form of statistical evaluation of available regional stream flow, precipitation, modeling, and channel morphology data. These evaluations will be supplemented by studies of paleoflood sediments in order to provide a history of major paleofloods in the area (Section 8.3.1.5.2).

Analyses of the effects of probable maximum floods near the surface sites of proposed repository facilities have been initiated. An analysis was made of the flood plain of Fortymile Wash and its southwestern tributaries that include ephemeral stream channels from the east flank of Yucca Mountain (Figure 3-4) (Squires and Young, 1984). An analysis also was made of flood plains of Topopah Wash, a sizeable drainage from the Shoshone Mountain area into Jackass Flats, about 10 to 15 km east of Yucca Mountain (Christensen and Spahr, 1980). These studies of flooding potential of major desert washes are pertinent introductory investigations of flood potential at Yucca Mountain because of their close geographic and physiographic relations to that area.

Squires and Young (1984) estimate the magnitudes of the 100- and 500-yr flood peaks and the regional maximum peaks. The standard errors for the estimates of the 100- and 500-yr floods are relatively large. These relatively large errors result primarily from the short period of record (15 to 20 yr) and the extreme areal variability of floodflows in arid climates. However, the regression approach used by Squires and Young (1984), which is based on data from nearby streams, is believed to be the best available method for peak-discharge determinations. Other methods, such as rainfall-runoff modeling, may give results that are not qualified as to their statistical reliability. These other methods are not believed inherently better than the method used in the study of Squires and Young (1984), which allows a reliability evaluation and is based on nearby flood data. The estimated discharges for the 100- and 500-yr floods on Fortymile Wash and the three southwestern tributaries are considered accurate to no more than two significant figures (Squires and Young, 1984).

The estimates of regional maximum floods made by Squires and Young (1984) are based on a graphical boundary curve developed by Crippen and Bue (1977). The graph defines a boundary curve of maximum discharges that have occurred in drainages of varying sizes and is based on quantitative measurements of flood-flow peaks in a five-state region that includes Arizona, California, Nevada, New Mexico, and Utah. That graph is reproduced and modified by Squires and Young (1984) and includes data from southern Nevada.

The technique for calculating the probable maximum flood is based on a determination of maximum probable precipitation over the drainage or drainages being considered. It also requires detailed knowledge of the drainage basin to effect reasonable and accurate predictions of probable maximum peak flows. The sparse streamflow records, the availability of only minimal precipitation and storm data, and the absence of data on infiltration-runoff characteristics for drainage basins in the Yucca Mountain area requires that many speculations and assumptions would be needed to calculate the magnitudes of probable maximum floods in complex drainages the size of Fortymile Wash and Topopah Wash. Also, the lack of storm and runoff data throughout the

Figure 3-4. Yucca Mountain area and current streamflow measurement sites. Modified from Squires and Young (1984).

CONSULTATION DRAFT

hydrographic study area prevents checking the validity of the various assumptions used.

Because of data limitations just discussed, the methods for predicting extreme discharges of unusually large floods as used by Christensen and Spahr (1980) and Squires and Young (1984) were considered preferable to the methods presented in ANSI/ANS 2.8-1981 at the time the flood studies were conducted for Topopah and Fortymile washes. A preliminary study has been done to evaluate the clear-water probable maximum flood discharge for the Yucca Mountain site (Bullard, 1986). Results of this study will be analyzed to evaluate the flood potential of the site. Further studies are planned and discussed in Section 8.3.1.16.1.

Squires and Young (1984) determined flood magnitudes and frequencies less than the regional maximum by statistically manipulating the measured data collected at the 12 sites of Table 3-2. The study by Christensen and Spahr (1980) defined flood-prone areas of 100-yr, 500-yr, and maximum potential floods for Topopah Wash and its tributaries in the eastern part of Jackass Flats (Figure 3-4). Their maximum potential flood is essentially equivalent to the regional maximum flood of Squires and Young (1984). Christensen and Spahr (1980) reproduced the curve developed by Crippen and Bue (1977) for the region including Yucca Mountain as their basis for determination of the magnitude of their maximum potential flood.

From their investigation of Topopah Wash, Christensen and Spahr (1980) concluded the following:

1. The areas prone to 100-yr floods closely parallel most main channels, with few occurrences of out-of-bank flooding of the areas between the main channel and adjacent secondary channels. Out-of-bank flooding would result in a water depth of less than 0.6 m, with a mean velocity as high as 2 m/s occurring on the steeper slopes. Flood-water depth in the stream channels would range from 0.3 to 2.7 m, with mean velocities of 0.9 to 2.7 m/s.
2. The 500-yr flood would exceed the discharge capacity of all stream channels except Topopah Wash and some upstream reaches of a few tributaries. Out-of-bank flooding of areas between the adjacent channels would result in water depths as much as 0.9 m, with mean velocities greater than 2 m/s. Flood-water depth in the stream channels would range from 0.3 to 3.7 m, with mean velocities ranging from 0.9 to 4 m/s.
3. The maximum potential flood would inundate most of Jackass Flats. Out-of-bank flows in the areas between adjacent channels would have a depth as much as 1.5 m, with a mean velocity as high as 4 m/s. Flood water in the stream channels would have depths of 0.6 to 7 m, with velocities of 1.2 to 7.9 m/s.

Squires and Young (1984) studied the downstream part of Fortymile Wash. Within this area, Fortymile Wash has three tributaries that are informally designated from south to north as Busted Butte Wash, Drill Hole Wash, and Yucca Wash. Approximate flood-prone areas in these washes are shown on Figure 3-5. Squires and Young (1984) conclude the following:

Figure 3-5. Flood-prone areas in the vicinity of Fortymile Wash. Modified from Squires and Young (1984).

CONSULTATION DRAFT

1. Fortymile Wash, within the flood-study area, is a well-defined incised channel, with a cross section of 15 to 21 m depth and 300 to 450 m width. The estimated values of the 100-yr, 500-yr, and regional maximum floods indicate that the flow would stay within the confines of the wash. Estimated depths of flood water in the stream channel would range from 0.9 to 2.4 m for the 100-yr flood, from 1.8 to 3.3 m for the 500-yr flood, and 6.4 to 8.8 m for the regional maximum flood; corresponding mean velocities would be from 1.8 to 2.7 m/s for the 100-yr flood, 3.3 to 4.3 m/s for the 500-yr flood, and from 7.0 to 8.5 m/s for the regional maximum flood.
2. The drainage basin of Busted Butte Wash varies from a shallow valley with meandering ephemeral streams to a deeply incised canyon in the upstream reaches. Drill Hole Wash is characterized by deep canyons extending from Yucca Mountain to its mid-drainage area. Both washes would have estimated flood-water depths of from 0.3 to 1.2 m in the stream channel during the 100-yr flood, and the corresponding mean velocities would range from 1.2 to 2.4 m/s. The 500-yr flood would exceed bank capacities at several reaches of the washes. Depths and mean velocities would range from 0.9 to 3.0 m and 1.5 to 3.3 m/s. The regional maximum flood would inundate all central flat-fan areas in these two watersheds. Flood-water depths in the stream channels would range from 1.5 to 3.7 m, with mean velocities varying from 2.1 to 4.9 m/s.
3. Yucca Wash is contained within an incised channel that is about 14 m deep and 240 m wide at its confluence with Fortymile Wash. The 100-yr, 500-yr, and regional maximum floods would stay within the steep-side-slope stream banks that contain the flood plain. Flood-water depths in the stream channel would range from 0.9 to 1.5 m for the 100-yr flood, from 1.5 to 2.7 m for the 500-yr flood, and from 2.7 to 7 m for the regional maximum flood; corresponding mean velocities would vary from 1.5 to 2.7 m/s for the 100-yr flood, from 2.4 to 3.7 m/s for the 500-yr flood, and from 2.7 to 6.7 m/s for the regional maximum flood.

According to Squires and Young (1984), "Geomorphic studies on the Nevada Test Site have indicated that some of the alluvial surfaces along Fortymile Wash are thousands of years old. Such ages might imply that the surfaces have not been flooded since they were formed several thousand years ago. However, distinct high-water marks are observed along Fortymile Wash in the vicinity of cross-section FM-4" (Figure 3-5), indicating that the alluvial surfaces along Fortymile Wash were inundated. Survival of these alluvial surfaces may be explained by a previous observation in Colorado that a fine-grained alluvial surface overtopped by a flash flood was virtually unaffected (Squires and Young, 1984). They continue, "From these marks and from data on the cross-sectional area and channel slope, a peak flow of about 20,000 ft³/s (570 m³/s) is estimated. Documentation of similar flooding in nearby washes indicates that this flood peak probably occurred during February 1969." A discharge of 20,000 ft³/s (570 m³/s) for this section of stream channel is greater than the estimated discharge for the 100-yr flood (12,000 ft³/s (340 m³/s)), but less than the estimated discharge for the 500-yr flood (57,000 ft³/s (1,600 m³/s)) (Squires and Young, 1984). Additional studies are planned to evaluate paleoflooding in the Yucca Mountain area through the

use of geomorphic and geologic evidence. These studies are discussed in Section 8.3.1.5.2.

The extent of erosion and sediment movement caused by flood flow in Fortymile Wash and its tributaries that drain Yucca Mountain is not known quantitatively. Qualitatively, however, on the basis of knowledge of record major floods elsewhere in southern Nevada (Glancy and Harmsen, 1975; Katzer et al., 1976), erosion of or deposition in channels and flood plains probably would occur during the 100-yr, 500-yr, and regional maximum floods. While erosion or deposition due to an individual flood event cannot be quantified, average rates of stream incision over long periods of time can be evaluated. Section 1.1.3.3.2 discusses surficial processes, and Table 1-3 lists average rates of stream incision in the Yucca Mountain area.

Evidence of erosion and deposition was observed in some channels during field surveys. Channel erosion or aggradation in the existing streambeds could alter the flood-flow characteristics of cross-sectional area, width, mean velocity and maximum depth listed in the report by Squires and Young (1984). The effect of erosion or deposition on flood-flow characteristics would vary from place to place. Because velocities for the 100-yr, 500-yr, and regional maximum flood peaks are high, channel erosion and aggradation appear likely.

Because most of Yucca Mountain is well above expected flood levels of Fortymile Wash and its major tributaries, as assessed by Squires and Young (1984), numerous small ephemeral stream channels from the mountain to the major drainages were not included in their study. Flow in the ephemeral stream channels that may affect the Yucca Mountain site will be assessed in future studies (Section 8.3.1.2.1).

The sparseness of the historic data base on surface-water hydrology, including the movement of both water and debris inhibits accurate prediction of flood and debris hazards for the immediate future. Data needs, plans for future data collection, and data-collection methodologies are given in Section 8.3.1.16.1. Likewise, a deficient understanding of paleoclimates (Section 5.2.1) and past geomorphic processes (Section 1.1.3) limits the ability to predict climatic changes and their probable effects on flood-and-debris-hazards potential over the next several thousands of years. Plans for studies of future climates are discussed in Section 8.3.1.5.1.

Hydrologic conditions that have occurred during the Quaternary that have differed from present conditions, and the likelihood of recurrence over the next 10,000 yr, are described in Sections 3.7.4 and 3.9.8. Evidence for long term changes in hydrometeorology are discussed in Section 3.7.4. A more detailed discussion on past influences of glacial environments in the western United States and their effects on climate is given in Section 5.2. Hoover et al. (1981) describes a system for correlating and mapping surficial deposits in the Yucca Mountain area. The stratigraphic sequence of units, soils, and unconformities that has been observed indicates fluctuating climatic conditions.

3.2.1.1 Ongoing and future studies of flood and debris hazard potentials

Studies are planned to improve the surface-water hydrologic data base at Yucca Mountain and surrounding areas (Section 8.3.1.2.1). The minimal data on stream flow and insufficient knowledge of geomorphic parameters make predictions of flood and debris hazards very speculative. To rectify this deficiency, the NNWSI Project plans to expand the streamflow (movement of water and debris) data base (Section 8.3.1.2.1). A network of stream-gaging stations near Yucca Mountain has been established to monitor and record modern stream flows. Four continuously recording stream gages and five peak-flow gages were installed. Locations of these gages are shown in Figure 3-4. Precipitation gages also installed at these stream gaging sites collect rainfall data. Also, five peak-flow gages at sites 2 through 6 (Table 3-2 and Figure 3-3) have been reactivated. Investigations of flooding also provide stream flow data at other miscellaneous ungaged sites when floods occur near Yucca Mountain or the surrounding region. These investigations also yield data on debris transport, particularly where debris avalanches or debris flows are part of the flooding process.

Investigations of paleofloods, including debrisflows, have been initiated. Paleoflood studies involve analyses of sediment deposits of Quaternary floods and collection of evidence for mass movement of debris by these floods. These investigations are being conducted to improve understanding of Quaternary runoff events, develop a chronology of flooding, provide data on past geomorphic processes for comparison with modern processes, and aid in interpretation of paleoclimates. Plans for future investigations of, and data collection for potential flood and debris hazards, are described in Section 8.3.1.5.2.

Inasmuch as the Yucca Mountain area is located inland and has no significant surface-water bodies or water-control structures located near the site, there is no potential for events such as surges, seiches, tsunamis, dam failures, or ice jams that could affect the site, nor is there any potential for future dam development. The Yucca Mountain area was not glaciated during the Pleistocene (Section 5.2). No evidence for pluvial lakes has been found in the Yucca Mountain area (Mifflin and Wheat, 1979). This indicates that the potential for floods during maximum glaciation is minimal. Additional studies to evaluate the potential for climatic change and the effects of any change on the proposed site are discussed in Section 8.3.1.5.2. No evidence for past flooding induced by landslides in the vicinity of the site has been reported (USGS, 1984, and Section 1.1.3, geomorphic processes).

3.2.2 FLOOD PROTECTION

The rugged terrain at the Yucca Mountain site and local convective storms can cause intense flooding to occur periodically in the normally dry washes that drain Yucca Mountain ridge. Protection from this flooding was considered in choosing the proposed locations for the surface facilities, shafts, and ramps in the current conceptual design. Topographic maps with 20-ft contour intervals on a 1:24,000 scale were used, together with direct observations, to choose the proposed locations. These maps, however, do not provide the detail necessary for final design. New topographic maps have

been compiled (Wu, 1985) of the Yucca Mountain area at a scale of 1:5,000 and a contour interval of 2 m (with 1-m supplemental contours). These will be used for some aspects of final design.

As described in Section 6.2.4.2 (flood protection) a preliminary analysis of the probable maximum flood (PMF), based on a study by Bullard (1986), has been performed for the men-and-materials shaft area because of the local rugged terrain and the proximity of the shaft area to the confluence of two washes that are tributary to the larger Drill Hole Wash. The purpose of the preliminary analysis was to evaluate the feasibility of locating the shaft and its supporting surface complex in such a rugged area. In the analysis, the PMF flows and levels were conservatively estimated and flooding protection provisions were incorporated into the design. The men-and-materials shaft area is benched, with the shaft entrance designed to be above the PMF levels in the diversion channels, located on the north and south sides, as well as the PMF level in Drill Hole Wash. From this preliminary conservative analysis it was concluded that the men-and-materials shaft can be adequately protected from PMF levels at the proposed location.

Further surface facility design will be based on PMF flows and levels, determined in accordance with ANSI/ANS 2.8-1981. This site information will be acquired in accordance with site characterization plans described in Section 8.3.1.14.1, topographic characteristics of potential locations of surface facilities, and Section 8.3.1.16.1, flood-recurrence intervals and levels at potential locations of surface facilities. This design effort will be finalized when definitive flooding data become available, as described in Section 8.3.2.5.7, design analyses including those addressing impacts of surface conditions, rock characteristics, hydrology, and tectonic activity.

3.3 LOCATIONS AND DISTANCES TO POINTS OF SURFACE-WATER USE

The purposes of Section 3.3 are to (1) identify the locations and quantities of the surface water currently used for supply, and (2) project the amounts of surface water that will be used during the next 50- and 100-yr periods. The general absence of surface water that is suitable for supply purposes, precludes addressing those aspects of surface water that are directly related to water supply for populations (i.e., extracted quantity of water, extraction locations, size of population served, projected populations and corresponding water-use figures, etc.).

The hydrographic study area (Figure 3-1) includes all occurrences of surface water that may be affected by or affect site characterization and repository construction and operations. This section presents all known sources of surface water within the hydrographic study area, the quantity and quality of available surface water, and current and projected water needs of industrial, commercial, municipal, agricultural, and domestic water users.

3.3.1 PRESENT QUANTITY AND QUALITY OF SURFACE WATER EXTRACTED

Very little surface water exists in the vicinity of Yucca Mountain. The climate in the region is arid; no perennial streams are present except short reaches of the Amargosa River which are fed by springs (Figure 3-1) (Waddell et al., 1984). Other perennial sources of surface water are the small spring-fed ponds in Oasis Valley, the Amargosa Desert, and Death Valley (French et al., 1984; Waddell et al., 1984). Spring locations and their water chemistries are presented in Sections 3.5.1 and 3.7.3, respectively. Spring water use is discussed in Section 3.8.1.

Several small, man-made reservoirs used for agricultural, milling, and mining purposes are scattered throughout the southern portion of the hydrologic study area. The largest, a man-made reservoir, known locally as Crystal Reservoir, is located in the Ash Meadows area 52 km (32 mi) southeast of Yucca Mountain (Figure 3-1). This reservoir receives its water from Crystal Pool spring, located approximately 0.8 km (0.5 mi) to the north. Crystal Reservoir, which has a capacity of 1,840 acre-feet ($2.27 \times 10^6 \text{ m}^3$) is used for recreational purposes only. Another small reservoir (Clay Camp at NE 1/4, Section 1, T.18S., R.49E.) with a capacity of 43 acre-feet ($53,000 \text{ m}^3$), is used for milling. This and many other reservoirs are fed by wells (Section 3.8.1). Other small reservoirs with capacities much less than 40 acre-feet ($49,300 \text{ m}^3$) exist near Point of Rocks and Death Valley Junction (Giampaoli, 1986).

The Amargosa River (Section 3.1) is not used for water supply. The flow regime has been intermittent to ephemeral throughout historic times, except for the spring-fed reaches. As such, sufficient supplies for domestic, municipal, commercial, agricultural, or industrial supply have not been available (Walker and Eakin, 1963). The chemical composition of Amargosa River water is discussed in Section 3.4.

Hydroelectric power is generated at Hoover Dam near Boulder City, Nevada. Although Hoover Dam is located only 180 km southeast of Yucca Mountain, the hydrologic settings of the two areas are very different. The absence of a perennial river within the boundaries of the hydrographic study area precludes the development of hydroelectric power.

No other sources of surface water exist within the hydrographic study area. Frenchman Lake and Yucca Lake are playas that contain water only after rainstorms. Runoff to these playas serves to recharge the valley-fill aquifer (Claassen, 1985; Section 3.7.1). However, because of the sporadic precipitation and runoff and the excessive amounts of dissolved solids, playas are not considered as possible sources of water supply (Waddell et al., 1984). The virtual absence of perennial surface water has precluded its application for beneficial use in the hydrographic study area.

3.3.2 PROJECTED SURFACE-WATER USES

Arid conditions of the region and the absence of perennial streams and lakes make surface water an unlikely future water supply. Within the hydrographic study area, surface water serves two principal functions:

(1) recharge to the valley-fill aquifer and (2) maintenance of habitats for aquatic species. Surface-water use by man in the study area would probably increase only in the event of a change in climate to a wetter regime. Studies done to date indicate that such an event is possible, but not in the period of repository construction, operation, closure, and decommissioning (DOE, 1986) (see Section 5.2 for a complete discussion of climatic modeling and Section 8.3.1.5.1 for a discussion of planned studies). Consequently, there are no known plans to construct dams or reservoirs in the vicinity of Yucca Mountain site.

3.4 CHEMICAL COMPOSITION OF ADJACENT WATERCOURSES

This section addresses the chemical composition of surface water at the Yucca Mountain site and vicinity. Classically, the section would include discussions of (1) seasonal cycles of physical and chemical limnological parameters, (2) bottom and shoreline configuration, (3) sedimentation rates, (4) sedimentation gradation analysis, and (5) sorption properties. However, because of the arid to semiarid climate of the region, these topics are not applicable. This section presents data required to address questions of the adequacy of the description of the present and expected hydrogeologic characteristics in order to provide the information required by the design and performance issues and the formation of a sufficient baseline against which to assess potential impacts.

The historical records and onsite monitoring of surface water composition is quite limited. Present data consist of two historic and six recent chemical analyses. The analyses are presented here along with discussions on trends and observations.

Within the hydrographic study area, the Amargosa River is the major surface-water source to the regional drainage sink of Death Valley. As described in Section 3.1, the Amargosa River is normally dry, except for short reaches that receive water discharging from the underlying ground-water flow system. Because of the ephemeral character of the Amargosa River only two water samples have been collected; one from a reach near Eagle Mountain and the other from Carson Slough, a major tributary that joins the Amargosa River near Alkali Flat (Figure 3-6). From the chemical analyses presented in Table 3-3, it can be seen that sodium and bicarbonate are the primary constituents. The high sulfate and chloride is indicative of ground-water interaction with playa deposits. The chemical composition of these two surface water samples is quite similar to that of the water from the tuffaceous valley fill aquifer (Section 3.7.3). This similarity suggests that the source of the water sampled is the tuffaceous valley fill. Hunt et al. (1966) indicate that ponding occurs where the shallow ground-water table in southern Amargosa Desert encounters a structural barrier north of Eagle Mountain.

Samples of surface water from the Yucca Mountain area were collected during period of runoff and flooding in July and August of 1984. Water samples were collected from the main stream channel of Fortymile Wash and from two of its principal tributaries, Drill Hole and Busted Butte washes (Figure 3-6). Results of analyses are shown in Table 3-3. All the samples

CONSULTATION DRAFT

Figure 3-6. Surface-water sample sites listed in Table 3.3.

Table 3-3. Chemical composition of watercourses adjacent to Yucca Mountain^a

Parameter	1 Flood sample Fortymile Wash at road M ^b	2 Flood sample Fortymile Wash at road W ^b	3 Flood sample Fortymile Wash above Drill Hole Wash ^b	4 Flood sample Drill Hole Wash at mouth ^b	5 Flood sample Fortymile Wash at J-12 ^b	6 Flood sample Busted Butte Wash ^b	7 Stream sample Carson slough ^c	8 Stream sample Amargosa River near Eagle Mountain ^c
Laboratory sample number	4212602	4248608	4248604	4248605	4248607	428606	3060	3062
Latitude	36°49'07"	36°49'04"	36°49'08"	36°49'11"	36°49'51"	36°47'49"	≈36°25'	≈36°13'
Longitude	116°23'47"	116°23'47"	116°23'46"	116°23'52"	116°23'51"	116°23'51"	≈116°21'	116°22'50"
Site number on Figure 3-6	1	2	3	4	5	6	7	8
Date	7/22/84	8/15/84	8/14/84	8/14/84	8/14/84	8/14/84	1958	1958?
Specific conductance, field (microsiemens per cm at 25°C)	-- ^d	170	70	100	59	120	937	1,860
Specific conductance, laboratory (microsiemens per cm at 25°C)	201	859	198	218	100	217	--	--
pH, field	--	8	8.4	8.3	8.2	8.3	8.5	8.8
pH, laboratory	7.4	7.4	7.5	7.8	7	7.8	--	--
Temperature (°C)	--	21.5	--	--	--	--	10	4.4
Calcium (Ca)	24	31	8.1	9.5	6.7	12	40	24
Magnesium (Mg)	3.3	2.9	0.9	1.3	0.7	1.8	26	29
Sodium (Na)	8.1	8.2	4.1	8.6	2.4	7	125	344
Potassium (K)	7.8	9.1	5.6	7.4	6.3	8.1	16	40
Bicarbonate (HCO ₃)	--	--	--	--	--	--	362	542
Carbonate (CO ₃)	--	--	--	--	--	--	10	33
Alkalinity as CaCO ₃ , lab	73	75	36	42	26	47	--	--
Chloride (Cl)	3.7	1.4	1.3	2.2	2	1.7	40	123
Sulfate (SO ₄)	10	10	6.2	13	6.3	7.9	122	277
Fluoride (F)	0.2	0.2	<.1	0.3	<.1	0.3	2	2.8
Silica (SiO ₂)	25	24	8.7	20	4.5	23	28	26
Arsenic (μg/L as As)	2	2	<1	2	<1	3	0.0	20
Iron (μg/L as Fe)	110	77	18	100	28	200	210	450
Manganese (μg/L as Mn)	3.3	5	11	6	22	10	0.0	0.0
Strontium (μg/L as Sr)	100	100	34	66	31	86	1,800	800
Lithium (μg/L as Li)	6	7	6	14	5	17	--	--
Iodide (I)	0.009	0.005	0.003	0.002	0.004	0.003	--	--
Bromide (Br)	0.049	<0.01	<0.01	<0.01	<0.01	<0.01	--	--
Boron (B)	--	--	--	--	--	--	0.68	2.1
Dissolved solids (sum)	127	122	57	92	45	100	566	1,140

^aValues for chemical constituents are in milligrams per liter unless otherwise indicated. Analyses by U.S. Geological Survey, Denver, Colorado.

^bFlood sample analyses (USGS) from WATSTORE files.

^cStream sample analyses from Hunt et al. (1986).

^d-- indicates no data.

are dilute and are significantly lower in solutes than ground-water samples of the area (Section 3.7.3). However, samples 1 and 2 are noticeably more concentrated in calcium and dissolved solids than are samples 3 through 6. This difference is most likely attributed to the fact that samples 1 and 2 were collected on different days than samples 3 through 6, which means that they are not only from different locations but also from different rainfall events. Even with the last four samples showing remarkably similar compositions, any quantitative interpretation is unwarranted unless the following are known:

1. The compositional variability of storms throughout the year.
2. The compositional variability within a given storm.
3. The lithologic composition of the source areas for the drainage systems.

However, the data do provide information about the base level composition of recharge to the valley-fill aquifer along Fortymile Wash (Section 3.7.3).

Waddell et al. (1984) cite water chemistry data that indicate that washes may be the principal source of ground-water recharge beneath Yucca Mountain. Periodic temperature measurements made in wells in Drill Hole Wash during 1981 to 1983 revealed a pronounced increase in the temperature profile above a depth of 150 m in drillhole UE-25a#7. This occurrence of warmer water may be related to recharge from a major storm that occurred early in March, 1983 (Waddell et al., 1984). Furthermore, as discussed in Section 3.7.3.2, the high tritium concentration of water from well UE-29a#1 in Fortymile Wash (62 tritium units) indicates that precipitation containing bomb-test tritium has recharged the aquifer since the mid-1960 possibly correlating with the major flood in the winter of 1969 (Section 3.2.1). This evidence clearly establishes that modern rainfall and runoff contribute recharge to the shallow ground-water system in the tuffaceous valley-fill aquifer. The extent of this contribution, which probably is small, cannot be further quantified with the existing data. Stable isotopic composition of surface water may also be used to interpret times and conditions of recharge. Planned studies for characterization of the regional surface water are discussed in Section 8.3.1.2.1.

3.5 POINTS OF GROUND-WATER DISCHARGE

Ground-water discharge from basins within the hydrogeologic study area occurs as spring flow, evapotranspiration, ground-water withdrawal, and outflow to other ground-water basins. This section discusses these various forms of discharge in relation to the overall hydrologic budget and identifies naturally occurring springs, seeps, and areas of phreatophytic growth that are located within the hydrogeologic study area. An inventory of water-supply wells located within the study area is presented in Section 3.8.1, as are discussions of spring and well water use. The discharge rates of specific springs are available in the spring inventory, presented later in this

section. Ground-water underflow figures and the associated data from which they were calculated are included in Section 3.7. Ground-water withdrawals from specific wells are listed in Section 3.8.1. A discussion of the potential for contamination of ground-water and surface-water supplies that might result from site characterization activities is also included.

Throughout this section, the term ground water refers to one of several classifications of water lying beneath the surface, including deep and shallow saturated zones, unsaturated zones, soil water and perched water. The particular usage depends on the topic of discussion and may be derived from the context.

The data presented in this section generally address the question of the isolation of radioactive waste from the accessible environment after closure of the mined geologic disposal system at Yucca Mountain (Key Issue 1). More specifically, data concerning the location and nature of points of ground-water discharge may be used to assist in the identification of paths of likely radionuclide travel and in the calculation of ground-water travel time along those paths. Further, this information contributes to the description and understanding of the hydrologic system (both the unsaturated and saturated zones) and to the prediction of potential impacts to the quality and availability of water resources in the vicinity of Yucca Mountain.

In general, definitive discharge data for the hydrogeologic study area are scanty. Walker and Eakin (1963) estimated discharge in the Amargosa Desert; Malmberg and Eakin (1962) made similar estimates for Oasis Valley. These estimates are summarized in Table 3-4. But, because in this instance, it is not possible to separate all the various forms of discharge, the figures presented are intended for comparison purposes only and may not represent the actual conditions present in the two areas.

Evapotranspiration in the Amargosa Desert is approximately 1.4 times greater than spring discharge and 47 times greater than ground-water underflow. In Oasis Valley, evapotranspiration exceeds spring discharge and underflow by factors of 10 and 5, respectively. It is evident from these estimates that evapotranspiration is the primary mechanism for discharge of ground water in these areas and probably throughout the hydrogeologic study area. Surface discharge is not considered significant, because it is ephemeral (Malmberg and Eakin, 1962; Walker and Eakin, 1963). A further discussion is given in Section 3.1. However, the existing data are insufficient to quantify properly the surface-water component of the total discharge. To address this deficiency, site characterization activities will include measurements of overall runoff and the estimation of peak runoff, flood magnitudes, and recurrence intervals. The scope of work and timetable for these activities are discussed in Section 8.3.1.2.1. These plans allow for the ephemeral nature of desert runoff, including the possibility of no runoff events in any single year.

CONSULTATION DRAFT

Table 3-4. Ground-water discharge from the Amargosa Desert and Oasis Valley

Discharge category	Amargosa Desert ^a		Oasis Valley ^b	
	acre-ft/yr	m ³ /s	acre-ft/yr	m ³ /s
Average annual consumptive use by native vegetation	11,500	0.45 ^c	ND ^d	ND
Evaporation from soil	12,000	0.47	ND	ND
Total evapotranspiration	23,500	0.92	2,000	0.08
Surface-water runoff discharging from basin	ND	ND	ND	ND
Spring discharge	17,000 ^c	0.66	200 ^e	0.008
Ground-water underflow	500	0.02	200	0.016

^aSource: Walker and Eakin (1963).^bSource: Malmberg and Eakin (1962).^cTotal spring discharge, includes water becoming underflow, evapotranspiration, etc.^dND = no data.^eDischarge from one spring that lies near the discharge point of the basin.

Several authors have used a variety of techniques to determine the average annual recharge rate for the area encompassing Yucca Mountain. Montazer and Wilson (1984) estimate the average annual recharge rate for Yucca Mountain to be between 0.5 and 4.5 mm/yr on the basis of various published estimates for the region. Using the analyses of Rush (1970), Czarnecki (1985) estimated recharge to be 0.7 mm/yr in a precipitation zone that includes Yucca Mountain. Czarnecki (1985) used a value of 0.5 mm/yr for modeling baseline conditions. Wilson (1985) concludes that 0.5 mm/yr is a reasonable and conservatively large value for flux below the proposed repository horizon at Yucca Mountain. This estimate is based on regional estimates, ground-water model studies, site data, and comparisons with published estimates for other arid and semiarid sites throughout the world. Nichols (1986) determined that recharge near Beatty may have occurred as three discrete pulses during the period 1961 to 1976. By implication, Nichols (1986) acknowledges the high probability that infiltration may not

occur a majority of the time. Nichols (1986) estimated pulses of 2.2, 0.5, and 2.6 cm by comparing available precipitation, evaporation, and soil-moisture content data. The discrete pulses were expected to be dampened out with depth. Rice (1984) used a hydrologic budget technique to estimate recharge for a hydrologic model of the NTS and vicinity. Her calculations estimated recharge to the saturated zone at less than 2.5 mm annually. However, Rice's (1984) model lacked sufficient detail to be used as an accurate representation of the conditions present at Yucca Mountain.

An evaluation of all water inputs and outputs in the hydrogeologic study area aids in the estimation of infiltration to the unsaturated zone at the Yucca Mountain site. A regional hydrologic budget is one method of estimating the amount of water that enters, percolates through, and recharges the saturated zone beneath Yucca Mountain. Recharge volumes represent only a small percentage of the total precipitation. Thus, when estimates of recharge are based on precipitation data, high accuracy of the data is required to keep the recharge estimates from being masked by the uncertainty in the precipitation data. Collecting highly accurate precipitation data is difficult. Also, maintaining a high level of accuracy for other elements of the water budget (e.g., ground-water underflow or total evapotranspiration) probably is unrealistic. This method will be used to the extent possible using both available data and that obtained during site characterization. Determination of recharge to the saturated zone may be accomplished by a variety of other methods, some of which are discussed in Sections 3.7 and 3.9.

The hydrologic budget for recharge within the study area, under steady state conditions, may be defined as follows (based on Fetter, 1980):

$$R_G = \text{Ppt} + I - Et - Q_G - Q_S - Q_W \pm \Delta S_S \quad (3-1)$$

where

- R_G = ground-water recharge
- Ppt = precipitation
- I = surface- and ground-water inflow
- Et = total evapotranspiration
- Q_G = ground-water outflow
- Q_S = surface-water runoff
- Q_W = ground-water withdrawal
- ΔS_S = changes in surface-water storage.

Under equilibrium conditions ground-water withdrawal is not considered and changes in the surface-water storage are considered to be zero.

As indicated, precipitation and inflow (surface and ground water) are the input factors to the total budget. Precipitation records, along with

CONSULTATION DRAFT

other climatological data, are available for the study area and are discussed in Section 5.1. Surface inflow from outside the boundaries of the hydrographic study area does not occur (Section 3.1). Ground-water inflow, however, has not been adequately quantified for the study area and will be studied during site characterization. Estimation of this component of the hydrologic budget will be accomplished through hydrologic modeling as discussed in Section 8.3.1.2.1.

Although discharge through surface-water runoff and ground-water outflow are smaller components of the total discharge, they are equally important for proper quantification of the total discharge. However, the data for the study area are not sufficient to quantify inflow, outflow, and runoff, or to relate these to recharge. Net subsurface flux (outflow minus inflow) must be quantified for proper comparison of recharge versus outflow from the basin. Plans for these studies are discussed in Section 8.3.1.2.1. Surface runoff and ground-water outflow are discussed further in Section 3.1, 3.2, and 3.7.

Total evapotranspiration is the sum of several processes (Robinson, 1957).

$$Et = E_s + E_i + E_x + E_p + E_c + E_{sw} \quad (3-2)$$

where

- Et = total evapotranspiration
- E_s = evaporation from soil
- E_i = evaporation of precipitation intercepted by foliage
- E_x = transpiration by xerophytes
- E_p = transpiration by phreatophytes
- E_c = evaporation from capillary fringe
- E_{sw} = evaporation from surface water.

Although evaporation from soil has not been sufficiently quantified in the study area, available studies indicate that losses due to evaporation from soil may be an important factor in the water budget. In a study by Nichols (1986), soil evaporation near Beatty was estimated to average 97 percent of the annual precipitation during the period 1961 to 1976. Further, Nichols (1986) stated that the high evaporation rate may have allowed for only three recharge events during the 16-yr study period.

The discharge of ground water through bare-soil evaporation can be significant, especially in areas with a shallow water table. Such conditions do not exist in the immediate vicinity of Yucca Mountain. However, shallow ground water does occur within the hydrogeologic study area. Ripple et al. (1972) present a method for estimating evaporation from bare soil areas with a near-surface water table. The procedure is potentially useful for estimating evaporation from areas such as Alkali Flat. An important factor in assessing evaporation from soil is the availability of detailed soil-type and

surficial-deposits maps. These are, for the most part, unavailable for the study area. However, Swadley et al. (1984) describe the surficial deposits near Yucca Mountain. Additional assessment of soil-moisture losses and mapping of surficial deposits will be included in site characterization activities as discussed in Section 8.3.1.5.1.

Data concerning evaporation of intercepted precipitation are not available for the study area. Investigations of this process in Utah and Idaho (Hull and Klomp, 1974; West and Gifford, 1976) report average water-holding capacities for individual sagebrush plants of 1.5 mm. Within the study area, the limited plant density (average shrub coverage ranging from 16 to 29 percent depending on elevation (O'Farrell and Emery, 1976)) and small crown diameters indicate that this element of the water budget is relatively minor, but not insignificant, particularly compared with the average annual recharge estimate of 0.5 mm/yr made by Wilson (1985) for Yucca Mountain.

Data concerning transpiration by xerophytes is lacking for the area of study. Sammis and Gay (1979) estimated the transpiration by a large creosote bush to be 5 to 8 percent of precipitation. Since these plants do not draw water from the water table, it seems that their primary effect is to reduce infiltration from the soil zone to the unsaturated zone. This may be an important factor in the budget.

Phreatophytic transpiration, on the other hand, has been relatively well studied, and some data exist for the study area that can be used for quantification of this parameter; for example, some areas of phreatophytic growth are identified in the spring inventory in Section 3.5.1 and in Section 3.6. Included in the studies of phreatophytes are several attempts at formulating a general method for estimating transpiration based on empirical data (Decker and Wein, 1957; Hughes and McDonald, 1966; Robinson, 1970) and several methods for calculating evapotranspiration using meteorological or hydrologic data (Rantz, 1968; Hanson and Dawdy, 1976). Proper estimation of the rate of transpiration from the various plant types will depend on the availability of information detailing areal distributions and densities of vegetation throughout the study area. Several studies have surveyed the biota in the area of the NTS (O'Farrell and Emery, 1976; Collins et al., 1982; O'Farrell and Collins, 1983; O'Farrell and Collins, 1984; Collins and O'Farrell, 1985); however, additional information will be obtained during site characterization (Section 8.3.1.2.1).

The great depth to the water table in most of the study area all but eliminates the importance of evaporation from the capillary fringe over large portions of the area. This form of discharge is different from evaporation from soil in that moisture is lost from the transition zone between the saturated and unsaturated zones (i.e., the capillary fringe) rather than directly from the zone of saturation or from the upper soil zone that may have been wetted by recent precipitation. Both forms of discharge may occur in areas where the water table is near the surface, such as Alkali Flat and possibly areas of perched water, where it could be a significant factor. Ripple et al. (1972) include a discussion of this process and a method for estimation of its rate.

The presence of a well-developed calcic horizon may act as an infiltration barrier and may retard percolation below the horizon. Infiltration

could accumulate above such a horizon, thereby allowing evaporation from the soil and capillary fringe and transpiration by plants to discharge nearly all of the precipitation entering that area. Such calcic horizons occur commonly throughout the region as well as at Yucca Mountain (Swadley and Hoover, 1983; Swadley et al., 1984; Taylor and Shroba, 1986; Rosholt et al., 1985). (A further discussion is given in Section 1.2.) The influence of the calcic horizon will be evaluated in the infiltration studies discussed in Section 8.3.1.2.2.

Evaporation from those surface-water bodies identified in Section 3.1 can be estimated on the basis of their surface area and pan evaporation data for the area. Pan evaporation for the area is estimated to be over 250 cm/yr according to Nichols (1986) and Dudley and Larson (1976). However, this figure is the potential evaporation. The volume of actual evaporation is expected to be relatively low due to the very limited availability of surface water bodies in the area. Nonetheless, this form of discharge is believed to be an important factor in the water budget. Information concerning the surface areas of these bodies is lacking and plans to obtain this information during site characterization are described in Section 8.3.1.2. Estimates of evaporation during periods of inundation due to flooding could also be made based on the area and period of inundation and pan evaporation. Much of the data necessary for these calculations are available in the form of topographic maps and previous and planned flood studies. Plans for these activities are discussed in Section 8.3.1.2.1.

Planned studies at Franklin Lake playa (Section 8.3.1.2.1) will provide data related to total evapotranspiration. Several techniques will be used to provide information on discharge rates and quantities, soil-moisture, phreatophyte distribution, evapotranspiration rates, and other parameters.

In addition to evapotranspiration from natural sources, quantification of discharge through human-related activities is planned (Section 8.3.1.16). Water use projections are available for agriculture, livestock, and domestic users in publications such as Water for Nevada (Office of the State Engineer, 1974). Discussions of ground-water use are included in Section 3.8.

In summary, studies that have been performed to date indicate that all of the components of ground-water discharge are significant when compared with the estimated recharge rates. In general, there is a paucity of site specific data, and meaningful estimates of these components cannot be made at this time. Additional studies, which will provide the necessary data to quantify these parameters, are included in the site characterization plans discussed in Section 8.3.1.2.

3.5.1 SPRINGS, SEEPS, AND PHREATOPHYTE AREAS

Springs can be classified on the basis of their ground-water source, (i.e., water-table springs and perched springs). Water-table springs discharge where the land surface intersects the water table. Perched springs, however, flow from the intersection of the land surface with a local ground-water body that is separated from the main saturated zone below by a zone of relatively lower permeability and an unsaturated zone.

Seeps are springs with very low discharge rates, commonly so low as to preclude measurement. Similar to springs, seeps may be classified according to their ground-water source.

An inventory of springs, seeps, and phreatophyte areas in the hydrogeologic study area has been compiled and is presented in Table 3-5. This list is based on data from Thordarson and Robinson (1971), Malmberg and Eakin (1962), Pistrang and Kunkel (1964), Schoff and Moore (1964), Miller (1977), Winograd and Thordarson (1975), Dudley and Larson (1976), White (1979), and Waddell et al. (1984). The springs are grouped according to basin boundaries (Sections 3.7.1 and 3.7.2). Springs are located using latitude and longitude coordinates to the nearest second, when available, and township and range designation. Perched springs are noted in the comments column. Hydrostratigraphic unit information, when available, is derived from the reference(s) listed in the reference column.

Also included in the spring inventory are some areas of phreatophytic growth. These areas are responsible for considerable amounts of ground-water discharge through evapotranspiration. Playas and alkali flats have not been included in the spring inventory because of the lack of data concerning quantities of discharge and the locations of many of these areas. As stated previously, such hydrologic conditions do not occur within the immediate vicinity of Yucca Mountain. However, they are believed to be important sites for ground-water discharge and are planned to be identified and mapped during site characterization (Section 8.3.1.2.1).

Areas of spring discharge are grouped into several distinct locations within the hydrogeologic study area. The primary areas are southern Oasis Valley (north of Beatty), Ash Meadows, Furnace Creek Wash (at Death Valley National Monument), and an area of minor ground-water discharge north and west of Yucca Flat. (A further discussion is given in Section 3.7.1.) The first three locations are situated at the discharge areas of the Oasis Valley, Ash Meadows, and Alkali Flat-Furnace Creek Ranch subbasins, respectively, shown on Figure 3-7 (3.7.1.1). The springs located within these three areas are primarily water-table springs and are associated with the regional ground-water flow (Sections 3.7.1 and 3.7.2). Several springs in the Oasis Valley discharge area, however, may flow from a perched water table located near the contact of the Tertiary volcanic beds and the Quaternary alluvium of the area (Waddell et al., 1984). Most of the springs that flow in the area north and west of Yucca Flat are perched springs (Schoff and Moore, 1964; Winograd and Thordarson, 1975). Table 3-6 groups the springs listed in the inventory that include an individual discharge rate, by magnitude of their discharge. Section 3.7 describes the regional flow system and includes information regarding the springs and areas of discharge discussed above.

Table 3-5. Springs, seeps, and phreatophyte areas in the hydrogeologic study area (page 1 of 12)

Latitude	Longitude	Township and range, spring location number	Name ^a	Aquifer ^b	Elevation		Discharge		Temperature		Use	Reference ^c	Comments
					ft	m	gpm	L/s	°F	°C			
ALKALI FLAT-FURNACE CREEK RANCH SUBBASIN (CALIFORNIA)													
36 26 40	116 49 50	27N/01E-23R1	Travertine		400	122	305	19.2	92	33	ND	T&R, 71 P&K, 64	
36 26 30	116 49 40	27N/01E-26A2	Travertine		320	98	220	13.9	92	33	Irrigation	T&R, 71 P&K, 64	
36 26 30	116 49 50	27N/01E-26A5	Travertine		320	98	ND ^d	ND	95	35	Irrigation	T&R, 71 P&K, 64	
36 26 30	116 49 50	27N/01E-26A6	Travertine		330	100	270	17.0	94	34	Irrigation	T&R, 71 P&K, 64	
36 26 30	116 49 50	27N/01E-26A7	Travertine		330	100	ND	ND	85	29	Irrigation	T&R, 71 P&K, 64	
36 26 30	116 49 40	27N/01E-26A1	Travertine		330	100	103	6.5	82	28	Irrigation	T&R, 71 P&K, 64	Discharge value includes discharge from the two springs listed below
36 26 30	116 49 40	27N/01E-25D2	Travertine		400	122	ND	ND	92	33	Irrigation	T&R, 71 P&K, 64	Seep
36 26 30	116 49 40	27N/01E-25D1	Travertine		400	122	ND	ND	92	33	Irrigation	T&R, 71 P&K, 64	
36 26 30	116 49 50	27N/01E-26A3	Travertine		320	98	0.4	0.025	89	32	Irrigation	T&R, 71 P&K, 64	
36 26 30	116 49 50	27N/01E-26A4	Travertine		320	98	4	0.25	94	34	Irrigation	T&R, 71 P&K, 64	
36 30 30	116 49 10	28N/01E-36K1	Nevares		920	280	ND	ND	69	21	Domestic Public supply	M, 77 P&K, 64	Seep
36 30 30	116 49 40	28N/01E-36M2	Nevares		745	227	ND	ND	84	29	Domestic Public supply	M, 77 P&K, 64	
36 30 30	116 49 40	28N/01E-36M1	Nevares		720	220	31	1.96	78	26	Domestic Public supply	M, 77 P&K, 64	
36 30 40	116 49 10	28N/01E-36G2	Nevares		896	273	22	1.39	102	39	Domestic Public supply	M, 77 P&K, 64	

Table 3-5. Springs, seeps, and phreatophyte areas in the hydrogeologic study area (page 2 of 12)

Latitude	Longitude	Township and range, spring location number	Name ^a	Aquifer ^b	Elevation		Discharge		Temperature		Use	Reference ^c	Comments
					ft	m	gpm	L/s	°F	°C			
ALKALI FLAT-FURNACE CREEK RANCH SUBBASIN (CALIFORNIA) (continued)													
ND	ND	28N/01E-36FS1			ND	ND	40	2.5	ND	ND	ND	M, 77 P.59	
36 30 40	116 49 10	28N/01E-36GS1	Nevares		937	286	289	16.97	104	40	Domestic Public supply	M, 77 P&K, 64	
36 29 40	116 51 20	27N/01E-03P1	Salt		0	0	4	0.25	73	23	Unused	P&K, 64	
36 29 40	116 51 20	27N/01E-03K1	Salt		100	30	ND	ND	ND	ND	NA ^e	P&K, 64	
36 27 10	116 51 00	27N/01E-22H1	Furnace Creek Inn Tunnel		50	15	148	9.34	92	33	Irrigation	P&K, 64	
36 26 30	116 49 40	27N/01E-26B1	South Travertine (Sump in Furnace Creek Wash)		280	85	568	35.7	92	33	Irrigation	M, 77 P&K, 64	
36 26 30	116 49 40	27N/01E-26B2	Buried tile in Furnace Creek Wash		250	76	200	12.62	ND	ND	Irrigation	P&K, 64	
36 26 40	116 49 50	27N/01E-23B1	Texas Spring (Tunnel)		380	116	224	14.1	91	33	Irrigation and public supply	P&K, 64	
		27N/01E-14N1			170	52	NA	NA	ND	ND	NA	P&K, 64	Phreatophyte area
		27N/01E-14P1			240	73	NA	NA	ND	ND	NA	P&K, 64	Phreatophyte area
		27N/01E-14Q1			400	122	NA	NA	ND	ND	NA	P&K, 64	Phreatophyte area
		27N/01E-23B2			400	122	NA	NA	ND	ND	NA	P&K, 64	Phreatophyte area
		27N/01E-23B3			400	122	NA	NA	ND	ND	NA	P&K, 64	Phreatophyte area
		27N/01E-23B4			400	122	NA	NA	ND	ND	NA	P&K, 64	Phreatophyte area
36 27 10	116 50 20	27N/01E-23P1			160	49	4	0.25	80	27	ND	P&K, 64	
		27N/01E-23G1			400	122	NA	NA	ND	ND	NA	P&K, 64	Phreatophyte area
		27N/01E-23G2			400	122	NA	NA	ND	ND	NA	P&K, 64	Phreatophyte area

3-37

CONSULTATION DRAFT

Table 3-5. Springs, seeps, and phreatophyte areas in the hydrogeologic study area (page 3 of 12)

Latitude	Longitude	Township and range, spring location number	Name ^a	Aquifer ^b	Elevation		Discharge		Temperature		Use	Reference ^c	Comments
					ft	m	gpm	L/s	°F	°C			
ALKALI FLAT-FURNACE CREEK RANCH SUBBASIN (CALIFORNIA) (continued)													
		27N/01E-23J1			410	125	NA	NA	ND	ND	NA	P&K, 64	Phreatophyte area
		27N/01E-23J2			410	125	NA	NA	ND	ND	NA	P&K, 64	Phreatophyte area
36 27 00	116 50 10	27N/01E-23K1			410	125	NA	NA	ND	ND	NA	P&K, 64	Seep
36 27 00	116 50 10	27N/01E-23K2			400	122	4	0.25	ND	ND	ND	P&K, 64	
36 27 00	116 50 20	27N/01E-23L1			160	49	NA	NA	ND	ND	NA	P&K, 64	Seep
36 27 00	116 50 20	27N/01E-23L2			160	49	NA	NA	ND	ND	NA	P&K, 64	Phreatophyte area
36 27 00	116 50 20	27N/01E-23L3			160	49	NA	NA	ND	ND	NA	P&K, 64	Seep
36 26 40	116 50 10	27N/01E-23Q1			320	98	NA	NA	ND	ND	NA	P&K, 64	Seep
36 26 40	116 50 10	27N/01E-23Q2			320	98	NA	NA	ND	ND	NA	P&K, 64	Seep
36 26 40	116 50 10	27N/01E-23Q3			320	98	4	0.25	72	22	ND	P&K, 64	
36 26 40	116 50 10	27N/01E-23Q4			320	98	NA	NA	ND	ND	NA	P&K, 64	Seep
36 26 40	116 50 10	27N/01E-23Q5			320	98	NA	NA	ND	ND	NA	P&K, 64	Seep
36 26 40	116 50 10	27N/01E-23Q6			320	98	NA	NA	ND	ND	NA	P&K, 64	Seep
36 26 40	116 49 50	27N/01E-23R2			410	125	NA	NA	ND	ND	NA	P&K, 64	Phreatophyte area
36 26 40	116 49 50	27N/01E-23R3			410	125	NA	NA	ND	ND	NA	P&K, 64	Phreatophyte area
36 26 40	116 49 40	27N/01E-24N1			400	122	NA	NA	ND	ND	NA	P&K, 64	Seep
36 26 30	116 50 10	27N/01E-26B4			320	98	NA	NA	ND	ND	NA	P&K, 64	Seep
36 26 30	116 50 10	27N/01E-26B5			320	98	NA	NA	ND	ND	NA	P&K, 64	Seep
36 30 00	116 51 00	27N/01E-3A1	Cow		200	61	18	1.14	81	27	Unused	P&K, 64	
		27N/01E-3B1			240	73	NA	NA	ND	ND	NA	P&K, 64	Phreatophyte area
		28N/01E-34M1			0	0	NA	NA	ND	ND	NA	P&K, 64	Phreatophyte area

Table 3-5. Springs, seeps, and phreatophyte areas in the hydrogeologic study area (page 4 of 12)

Latitude	Longitude	Township and range, spring location number	Name ^a	Aquifer ^b	Elevation		Discharge		Temperature		Use	Reference ^c	Comments
					ft	m	gpm	L/s	°F	°C			
ALKALI FLAT-FURNACE CREEK RANCH SUBBASIN (CALIFORNIA) (continued)													
36 30 10	116 51 50	28N/01E-34N1			10	3	NA	NA	74	23	NA	P&K, 64	Seep
36 30 10	116 51 50	28N/01E-34N2			80	24	NA	NA	ND	ND	NA	P&K, 64	Phreatophyte area
		28N/01E-34P1			100	30	NA	NA	ND	ND	NA	P&K, 64	Phreatophyte area
		28N/01E-35E1			380	116	NA	NA	ND	ND	NA	P&K, 64	Phreatophyte area
		28N/01E-35G1			500	152	NA	NA	ND	ND	NA	P&K, 64	Phreatophyte area
36 30 30	116 50 20	28N/01E-35K1			520	158	NA	NA	76	24	Unused	P&K, 64	Seep
		28N/01E-35N1			380	115	4	0.25	78	26	NA	P&K, 64	
		25N/02E-13GS1	Lemonade	Volcanic Rock	3,800	1,160	<1	<0.06	54	12	ND	USGS Map 1:250,000 Death Valley M, 77	
		26N/02E-13FS1	Navel	Fanglomerate	2,080	634	ND	ND	73	23	ND	USGS Map 1:250,000 Death Valley M, 77	
36°39'	116° 51'	29N/01E-15			1,800	360	ND	ND	ND	ND	ND	USGS Map 1:250,000 Death Valley	
36°37'	116°48'	29N/02E-30			2,000	610	ND	ND	ND	ND	ND	USGS Map 1:250,000 Death Valley	
ALKALI FLAT-FURNACE CREEK RANCH SUBBASIN (NEVADA)													
36 25 40	116 24 50	17S/49E-35d1	Ash Tree	ND	2,175	663	9	0.57	74	23	ND	T&R, 71 W&T, 75	
36 48 07	116 05 13	ND	Cane	Tertiary igneous	4,060	1,286	2	0.13	64	18	Unused	T&R, 71 W&T, 75	Perched (Wahmonie Formation) (W&T, 75) ^c
36 53 00	116 45 00	12S/47E-20bb1		ND	3,200	975	ND	ND	ND	ND	Irrigation	T&R, 71	
36 53 10	116 45 00	12S/47E-20bbb		ND	3,200	975	100	6.31	71	22	Irrigation	T&R, 71	

Table 3-5. Springs, seeps, and phreatophyte areas in the hydrogeologic study area (page 5 of 12)

Latitude	Longitude	Township and range, spring location number	Name ^a	Aquifer ^b	Elevation		Discharge		Temperature		Use	Reference ^c	Comments
					ft	m	gpm	L/s	°F	°C			
ALKALI FLAT-FURNACE CREEK RANCH SUBBASIN (NEVADA) (continued)													
36 56 21	116 16 14	ND	Topopah	Tertiary igneous	6,800	2,072	0.1	0.006	53	12	Unused	T&R, 71	Perched (S&M, 64) ^c
37 02 36	116 12 26	ND	Tippipah	Tertiary igneous	5,240	1,597	8	0.5	54	12	Unused	T&R, 71	Perched (S&M, 64) ^c
37 11 22	116 11 43	ND	Rainier	Tertiary igneous	6,240	1,902	0.1	0.006	67	19	Unused	T&R, 71	
37 26 30	116 06 00	06S/52E-10ad	Indian	Tertiary- Quaternary alluvium	6,685	2,038	2	0.13	49	9	Unused	T&R, 71	
37 30 40	116 05 20	05S/52E-14db	Cliff	Tertiary igneous	6,940	2,115	0.4	0.025	ND	ND	Unused	T&R, 71	
37 37 00	116 20 00	04S/50E-01dd		Tertiary igneous	6,550	1,996	0.1	0.006	ND	ND	Unused	T&R, 71	
36 45 30	116 15	ND	Pavits	Tertiary igneous	3,940	1,201	ND	ND	ND	ND	ND	T&R, 71	Perched (Piapi Canyon Group Wahmonie Forma- tion) (W&T, 75)
36 18 50	116 18 40	19S/50E-02	Grapevine	ND	2,280	695	ND	ND	ND	ND	ND	T&R, 71 W&T, 75	
36 52 50	116 39 40	12S/48E-30	Specie	ND	4,400	1,341	ND	ND	ND	ND	ND	USGS Topographic Map, Death Valley 1:250,000 Quad	
OASIS VALLEY SUBBASIN													
36 55 10	116 44 40	12S/47E-05cda	Beatty	Tertiary volcanic	3,370	1,027	100	6.31	75	24	Public use	T&R, 71 W, 79	Not used (Gram, 1985)
36 56 30	116 43 40	11S/47E-33bac		ND	3,480	1,061	25	1.58	93	34	Domestic	T&R, 71	
36 56 40	116 47 40	11S/46E-26dcc	Lower Indian	Tertiary igneous	4,000	1,219	8	0.5	70	21	Stock	T&R, 71 M&E, 62	Perched (W et al., 84) ^c ;

Table 3-5. Springs, seeps, and phreatophyte areas in the hydrogeologic study area (page 6 of 12)

Latitude	Longitude	Township and range, spring location number	Name ^a	Aquifer ^b	Elevation		Discharge		Temperature		Use	Reference ^c	Comments
					ft	m	gpm	L/s	°F	°C			
OASIS VALLEY SUBBASIN (continued)													
36 57 00	116 47 50	11S/46E- 26ca-1	Middle Indian	Tertiary igneous	4,150	1,265	ND	ND	60	16	Municipal	T&R, 71 M&E, 62	Perched (W et al., 84) ^c ; Beatty Muni- cipal Supply (Gram, 1985)
36 57 10	116 48 20	11S/46E- 26cbb-1	Upper Indian	Tertiary igneous	4,200	1,280	5	0.32	80	27	Municipal	T&R, 71 M&E, 62	Perched (W et al., 84) ^c ; Beatty Muni- cipal Supply (Gram, 1985)
36 57 00	116 43 00	11S/47E- 28dac-1		Tertiary- Quaternary alluvium	3,480	1,061	35	2.21	70	21	Irrigation	T&R, 71	
36 57 30	116 43 10	11S/47E- 28aa-2	Ute	Quaternary alluvium	3,540	1,109	25	1.58	70	21	Irrigation	T&R, 71 W, 79	
36 57 50	116 43 20	11S/47E- 21dbb-2		Tertiary- Quaternary alluvium	3,550	1,082	37	2.33	79	26	Domestic	T&R, 71	
36 58 20	116 43 10	11S/47E- 21aba-2	(Hicks?) (Bailey?)	Tertiary alluvium	3,600	1,097	ND	ND	106	41	Public	T&R, 71	
36 58 30	116 43 20	11S/47E- 16dcd-2	Burro Hot	Tertiary igneous	3,600	1,097	5	0.32	98	37	Public	T&R, 71 M&E, 62	
36 58 30	116 43 20	11S/47E- 16dcd-1	Burro Hot	Tertiary igneous	3,600	1,097	ND	ND	98	37	Domestic	T&R, 71	
36 59 10	116 45 30	11S/47E-18acb	Crystal	Tertiary igneous	3,980	1,207	2	0.13	75	24	Domestic	T&R, 71 M&E, 62	
36 59 30	116 42 50	11S/47E-10ccb		Tertiary- Quaternary alluvium	3,650	1,113	ND	ND	70	21	Stock	T&R, 71	
36 59 30	116 42 50	11S/47E-10ccb		Tertiary igneous	3,680	1,122	ND	ND	65	18	Domestic	T&R, 71	
36 59 40	116 42 30	11S/47E-10bdd		Tertiary igneous	3,800	1,158	49	3.09	75	24	Irrigation	T&R, 71	
37 00 00	116 42 20	11S/47E- 10ab-1	Goss	Tertiary igneous	3,800	1,158	50	3.15	71	22	Irrigation	T&R, 71	

Table 3-5. Springs, seeps, and phreatophyte areas in the hydrogeologic study area (page 7 of 12)

Latitude	Longitude	Township and range, spring location number	Name ^a	Aquifer ^b	Elevation		Discharge		Temperature		Use	Reference ^c	Comments
					ft	m	gpm	L/s	°F	°C			
OASIS VALLEY SUBBASIN (continued)													
37 00 20	116 42 30	11S/47E-03cdb-1		Tertiary igneous	3,840	1,170	40	2.52	73	23	Irrigation	T&R, 71	
37 00 30	116 42 30	11S/47E-04cad		Tertiary-Quaternary alluvium	3,680	1,122	10	0.63	70	21	Irrigation	T&R, 71	
37 00 50	116 43 50	11S/47E-04bb-1		ND	3,720	1,134	7	0.44	65	18	Domestic	T&R, 71	
37 01 40	116 43 20	10S/47E-33abc		Tertiary igneous	3,680	1,122	255	16.09	72	22	Irrigation	T&R, 71	
37 01 50	116 45 10	10S/47E-31aab-1		Tertiary-Quaternary alluvium	3,840	1,170	11	0.69	67	19	Irrigation	T&R, 71	
37 02 00	116 45 20	10S/47E-30d-1		ND	3,850	1,173	25	1.58	58	14	Domestic	T&R, 71	
37 04 30	116 41 30	10S/47E-14bab		Tertiary-Quaternary alluvium	3,182	970	100	6.31	84	29	Irrigation	T&R, 71	
37 37 10	116 43 40	04S/47E-04ca	Antelope	Tertiary alluvium	6,220	1,896	0.4	0.025	53	12	Unused	T&R, 71	
36 59 40	116 51 20	11S/46E-08bdc	Mud	Tertiary-Quaternary alluvium	4,244	1,294	ND	ND	67	19	Stock	T&R, 71	
		12S/47E-20bb1		ND	ND	ND	ND	ND	ND	ND	Irrigation, domestic	MAE, 62	
		11S/47E-7dc1		ND	ND	ND	ND	ND	ND	ND	ND	MAE, 62	
		10S/47E-30d1		ND	ND	ND	25	1.58	58	14	Domestic, stock	MAE, 62	
		10S/47E-32dda		Quaternary alluvium	ND	ND	225	14.19	72	22	ND	W, 79	
ASH MEADOWS SUBBASIN													
36 26 00	116 18 30	17S/50E-35a1		Lower carbonate aquifer	2,328	710	140	8.83	92	33	ND	T&R, 71 W&T, 75	Ash Meadows

Table 3-5. Springs, seeps, and phreatophyte areas in the hydrogeologic study area (page 8 of 12)

Latitude	Longitude	Township and range, spring location number	Name ^a	Aquifer ^b	Elevation		Discharge		Temperature		Use	Reference ^c	Comments
					ft	m	gpm	L/s	°F	°C			
ASH MEADOWS SUBBASIN (continued)													
36 26 10	116 18 50	17S/50E-35b1		Lower carbonate aquifer	2,305	703	17	1.07	83	28	ND	T&R, 71 W&T, 75	Ash Meadows
36 27 50	116 19 00	17S/50E-23b1	(Five Springs Area ?)	Lower carbonate aquifer	2,345	715	193	12.18	94	34	ND	T&R, 71 D&L, 76 W&T, 75	Ash Meadows
36 28 00	116 19 30	17S/50E-22a1	Longstreet	Lower carbonate aquifer	2,305	703	1,042	65.74	82	28	ND	T&R, 71 W&T, 75	Ash Meadows
36 28 50	116 19 30	17S/50E-15a1	Rogers	Lower carbonate aquifer	2,270	692	736	46.43	92	33	ND	T&R, 71 W&T, 75	Ash Meadows
36 29 20	116 20 10	17S/50E-10c1	Bell (Soda)	Lower carbonate aquifer	2,272	693	79	4.98	73	23	ND	T&R, 71 W&T, 75	Ash Meadows
36 29 20	116 20 30	17S/50E-09a1	Fairbanks	Lower carbonate aquifer	2,280	695	1,715	108.2	81	27	ND	T&R, 71 W&T, 75	Ash Meadows
		17S/50E-22ac	McGillivray	Lower carbonate aquifer	-2,300	701	See comments		ND	ND	ND	D&L, 76	Ash Meadows; flow reported in 1986 @ 155gpm; no flow observed in 1971; water level 5 ft below outlet
		17S/50E-35acc	Scruggs	Lower carbonate aquifer	-2,360	719	60	3.78	91	33	ND	D&L, 76	Ash Meadows
		17S/50E-35d1	School	Lower carbonate aquifer	-2,360	719	6	0.38	94	34	ND	D&L, 76	Ash Meadows
		18S/50E-01ca	Collins	Lower carbonate aquifer	-2,360	719	10	0.63	78	26	ND	D&L, 76	Ash Meadows

Table 3-5. Springs, seeps, and phreatophyte areas in the hydrogeologic study area (page 9 of 12)

Latitude	Longitude	Township and range, spring location number	Name ^a	Aquifer ^b	Elevation		Discharge		Temperature		Use	Reference ^c	Comments
					ft	m	gpm	l/s	°F	°C			
ASH MEADOWS SUBBASIN (continued)													
		18S/50E-12dc	Sink	Lower carbonate aquifer	-2,260	689	See comments		ND	ND	ND	D&L, 76	Ash Meadows; flow observed 0 25 gpm in 1986; no flow observed in 1971; water level 3 ft below outlet
ND	ND	17S/50E-21ac	Cold	Lower carbonate aquifer	ND	ND	73	4.60	67	20	ND	D&L, 76	Ash Meadows
36 25 10	116 19 20	18S/50E-03a	Crystal Pool	Lower carbonate aquifer	2,197	670	2,820	177.01	91	33	ND	W&T, 75 T&R, 71	Ash Meadows
36 24 50	115 44 20	18S/55E-01d	Cold Creek	Tertiary- Quaternary alluvium	6,220	1,898	690	43.53	ND	ND	Domestic	T&R, 71	
36 25 00	115 45 50	18S/55E-02a	Willow	Tertiary- Quaternary alluvium	5,990	1,826	340	21.45	ND	ND	ND	T&R, 71	
36 26 40	115 55 40	17S/54E-29d	Big Timber	Cambrian sedimentary	6,720	2,048	ND	ND	ND	ND	ND	T&R, 71	
36 18 30	115 41 10	19S/56E-10c	Three	Cambrian limestone	8,700	2,652	21	1.32	ND	ND	ND	T&R, 71	Spring Mountains
36 19 10	115 40 40	19S/56E-03c	Scout Canyon	Ordovician limestone	8,470	2,582	11	0.69	ND	ND	ND	T&R, 71	Spring Mountains
36 21 30	116 16 20	18S/51E-30d	Last Chance	Lower carbonate aquifer	2,253	687	1	0.063	68	20	ND	T&R, 71 W&T, 75	Ash Meadows
36 21 50	116 15 40	18S/51E-29b		Lower carbonate aquifer	2,275	693	1	0.063	72	22	ND	T&R, 71 W&T, 75	Ash Meadows
36 21 50	116 16 10	18S/51E-30a1	Bole	Lower carbonate aquifer	2,245	684	12	0.76	72	22	ND	T&R, 71 W&T, 75	Ash Meadows

Table 3-5. Springs, seeps, and phreatophyte areas in the hydrogeologic study area (page 10 of 12)

Latitude	Longitude	Township and range, spring location number	Name ^a	Aquifer ^b	Elevation		Discharge		Temperature		Use	Reference ^c	Comments
					ft	m	gpm	L/s	°F	°C			
ASH MEADOWS SUBBASIN (continued)													
36 22 30	116 16 20	18S/51E-19a1	Big (Deep) (Ash Meadows)	Lower carbonate aquifer	2,239	682	1,036	65.36	83	28	ND	T&R, 71 W&T, 75	Ash Meadows
36 23 20	116 16 40	18S/51E-18b1	Jack Rabbit	Lower carbonate aquifer	2,270	692	587	37.03	82	28	ND	T&R, 71 W&T, 75	Ash Meadows
36 24 00	116 18 00	18S/50E-12c1		Lower carbonate aquifer	2,245	684	11	0.69	80	27	ND	T&R, 71 W&T, 75	Ash Meadows
36 24 00	116 18 10	18S/50E-11d3	Davis Ranch (Bradford)	Lower carbonate aquifer	2,242	683	30	1.89	72	22	ND	T&R, 71 W&T, 75	Ash Meadows
36 24 10	116 18 10	18S/50E-11d2	Davis Ranch (Bradford)	Lower carbonate aquifer	2,242	683	5	0.32	74	23	ND	T&R, 71 W&T, 75	Ash Meadows
36 24 00	116 18 10	18S/50E-11d1	Davis Ranch (Bradford)	Lower carbonate aquifer	2,242	683	397	25.05	77	25	ND	T&R, 71 W&T, 75	Ash Meadows
36 24 10	116 16 10	18S/51E-07d1	King Pool (Point of Rock)	Lower carbonate aquifer	2,325	709	1,078	68.01	90	32	ND	T&R, 71 W&T, 75	Ash Meadows
36 24 10	116 16 10	18S/51E-07d4	(Point of Rock)	Lower carbonate aquifer	2,325	709	19	1.20	93	34	ND	T&R, 71 W&T, 75	Ash Meadows
36 24 10	116 16 10	18S/51E-07d3	Indian Rock (Point of Rock)	Lower carbonate aquifer	2,325	709	379	23.91	92	33	ND	T&R, 71 W&T, 75	Ash Meadows
36 24 10	116 16 10	18S/51E-07d2	Indian Rock (Point of Rock)	Lower carbonate aquifer	2,325	709	22	1.83	92	33	ND	T&R, 71 W&T, 75	Ash Meadows
36 24 10	116 16 20	18S/51E-07d5	(Point of Rock)	Lower carbonate aquifer	2,310	704	2	0.13	93	34	ND	T&R, 71 W&T, 75	Ash Meadows

Table 3-5. Springs, seeps, and phreatophyte areas in the hydrogeologic study area (page 11 of 12)

Latitude	Longitude	Township and range, spring location number	Name ^a	Aquifer ^b	Elevation		Discharge		Temperature		Use	Reference ^c	Comments
					ft	m	gpm	L/s	°F	°C			
ASH MEADOWS SUBBASIN (continued)													
36 25 30	116 17 30	17S/50E-36d1	Devils Hole	Lower carbonate aquifer	2,361	720	See comments		93	34	See comments	T&R, 71 D&L, 76 W&T, 75	Ash Meadows. No discharge as flow. <u>Cyprinodon</u> <u>diabolis</u> habitat
36 33 40	115 39 50	16S/56E-16b1	Indian	Tertiary- Quaternary alluvium	3,175	968	430	27.13	78	26	Domestic	T&R, 71	
36 34 30	115 43 40	16S/55E-11a	Cactus	Tertiary- Quaternary alluvium	3,238	987	0.5	0.031	ND	ND	Domestic	T&R, 71	
36 59 10	116 45 30	11S/57E	Quartz	ND	ND	ND	ND	ND	ND	ND	ND	T&R, 71	
37 10 09	116 10 07	ND	Captain Jack	Tertiary igneous	5,490	1,673	0.2	0.013	56	13	Unused	T&R, 71	Perched (SAM, 64) ^c
37 12 13	116 07 54	ND	Whiterock	Tertiary igneous	5,050	1,539	1.0	0.063	48	9	Unused	T&R, 71	Perched (SAM, 64) ^c
37 14 23	116 02 30	ND	Tubb	Tertiary igneous	5,190	1,582	ND	ND	52	11	Unused	T&R, 71	Perched (SAM, 64) ^c
37 14 41	116 04 24	ND	Oak	Tertiary igneous	5,800	1,768	0.1	0.0063	55	13	Unused	T&R, 71	Perched (SAM, 64) ^c
37 31 40	115 56 00	05S/54E-08bc	White Blotch	Tertiary igneous	5,960	1,817	0.2	0.013	42	6	Unused	T&R, 71	
36 27 50	115 57 40	17S/53E-24a	Gold	Cambrian sedimentary	6,780	2,067	ND	ND	ND	ND	ND	T&R, 71	
36 29 10	115 58 20	17S/53E-12c	Rock	Cambrian sedimentary	5,850	1,783	ND	ND	ND	ND	ND	T&R, 71	
36 29 10	115 59 00	17S/53E-11d	Jaybird	Cambrian sedimentary	6,280	1,914	ND	ND	ND	ND	ND	T&R, 71	
36 38 00	115 12 30	ND	Wiregrass	Ordovician dolomite	7,990	2,435	0.5	0.031	ND	ND	ND	T&R, 71	
36 38 50	115 13 50	ND	Pine	Ordovician dolomite	7,360	2,243	ND	ND	ND	ND	ND	T&R, 71	

Table 3-5. Springs, seeps, and phreatophyte areas in the hydrogeologic study area (page 12 of 12)

Latitude	Longitude	Township and range, spring location number	Name ^a	Aquifer ^b	Elevation		Discharge		Temperature		Use	Reference ^c	Comments
					ft	m	gpm	L/s	°F	°C			
ASH MEADOWS SUBBASIN (continued)													
36 39 30	115 13 10	ND	Canyon	Ordovician dolomite	8,020	2,444	ND	ND	ND	ND	ND	T&R, 71	
36 40 50	115 10 40	ND	Sawmill dolomite	Ordovician			8,140		2,481	ND	ND	T&R, 71	
36 41 30	115 12 30	ND	Basin dolomite	Ordovician			7,950		2,423	ND	ND	T&R, 71	
36 42 20	115 00 50	ND	Perkins	Ordovician dolomite			7,900		2,408	ND	ND	T&R, 71	
36 42 30	115 11 00	ND	Yellow Jacket	Ordovician dolomite			7,750		2,362	ND	ND	T&R, 71	
36 42 30	115 14 20	ND	Whiterock dolomite	Ordovician			5,930		1,807	ND	ND	T&R, 71	
36 42 50	115 10 50	ND	Shalecut dolomite	Ordovician			7,400		2,255	ND	ND	T&R, 71	
36 43 00	115 10 50	ND	Bootleg dolomite	Ordovician			7,200		2,194	ND	ND	T&R, 71	

^aBlank means the area is not named.

^bWhen the discharging aquifer could be determined, it is listed; otherwise only the lithology is shown. The main source of the ground water discharging in the Furnace Creek Ranch area is believed to be the lower carbonate aquifer (Waddell et al., 1984).

^cP&K, 64 = Pistrang and Kunkel, 1984; M&E, 62 = Malmberg and Eakin, 1982; T&R, 71 = Thordarson and Robinson, 1971; W&T, 75 = Winograd and Thordarson, 1975; D&L, 76 = Dudley and Larson, 1976; M, 77 = Miller, 1977; W, 79 = White, 1979; W et al., 84 = Waddell et al., 1984; S&M, 64 = Schoff and Moore, 1984.

^dND = No data.

^eNA = Not applicable.

CONSULTATION DRAFT

Figure 3-7. Hydrogeologic study area, showing precipitation, recharge areas, and discharge areas. Modified from Winograd and Thordarson (1975), Waddell (1982), and Waddell et al. (1984).

Table 3-6. Magnitude of springs in the hydrogeologic study area, based on Meinzer's^a classification of spring discharge

Magnitude	Volume of discharge		Number of springs in hydrogeologic study area
	English units	Metric units (L/s)	
1	>100 ft ³ /s	>2830	0
2	10-100 ft ³ /s	283-2830	0
3	1-10 ft ³ /s	28.3-283	9
4	100 gal/min to 1 ft ³ /s	6.31-28.3	18
5	10-100 gal/min	0.631-6.31	27
6	1-10 gal/min	0.0631-0.631	21
7	1 pt/min to 1 gal/min	0.0079-0.0631	8
8	1 pt/min	<0.0079	4

^aAdapted from Meinzer (1923).

3.5.2 POTENTIAL FOR CONTAMINATION OF SURFACE WATERS AND GROUND WATERS

No perennial surface waters exist on or near Yucca Mountain and very few areas of surface water are present within the hydrographic study area (Section 3.1). Ephemeral flow may occur as a result of high-intensity precipitation (Section 5.1). This storm runoff is commonly heavily laden with sediment and debris and may be used by vegetation and, to a minor extent, animals. However, this runoff is not used by humans for any purpose (DOE 1986; Section 3.2.3.3). The limited availability of surface water restricts the extent to which either plants or animals would be affected. Storm runoff would, most probably, only be affected in the immediate vicinity of the site. For this reason, modification of surface runoff, either in quantity or in chemical quality, is expected to have minimal, if any, impact on vegetation or wildlife. Other potential sources of surface runoff, such as dust-control spraying, are not expected to contribute to either surface or ground waters (DOE, 1986).

Ground water in the hydrogeologic study area is not expected to be contaminated or affected during site characterization activities. Controls over site characterization activities discussed below are considered sufficient to minimize the potential for any contamination of the ground water (DOE, 1986).

No contact is to be made with the water table at Yucca Mountain during site characterization except through exploratory boreholes. All water used for construction of the exploratory shaft will be tagged with a suitable

tracer for identification purposes (Section 8.4.2) (Whitfield, 1985). Tracers will be used according to accepted procedures (Bedmar, 1983; Rao, 1983). Waste water from surface facilities and the exploratory shaft will be disposed of away from the repository block at either the hypalon-lined waste rock pile, lined evaporative ponds, or through septic tanks (Section 8.4.1). The exploratory shaft will be constructed using established mining techniques and a minimum of tagged water to prevent significant alteration of subsurface conditions (Section 8.4.2). Subsurface construction techniques will be designed to reduce the effects on the hydraulic characteristics of the rock (Sections 2.8.3 and 3.6.4). It is recognized that drilling fluids used in exploratory boreholes will have an effect on the chemical and hydraulic nature of the subsurface environment, including any water samples taken from that hole. These effects have been studied and can be quantified and controlled to some degree (NRC, 1983; Brobst and Buszka, 1986). If drilling mud is used in the drilling of boreholes, waste fluids would be disposed of on the rock pile or in the evaporative ponds (Section 8.4.1.2.5). Drilling mud may not be used where vacuum drilling is possible (Whitfield, 1985). This process will not only prevent contamination by drilling fluids but also allow for collection of uncontaminated rock and water samples and identification of perched-water zones. Exploratory boreholes will be sealed after the completion of down-hole testing and monitoring to minimize the potential for ground-water contamination through the boreholes (Sections 6.2.8.3 and 6.2.8.4).

3.6 REGIONAL HYDROGEOLOGIC RECONNAISSANCE OF CANDIDATE AREA AND SITE

The occurrence and movement of ground water in the vicinity of Yucca Mountain is, in part, controlled by the regional hydrogeologic system. The hydrogeologic units within this system have varied hydraulic characteristics and interrelationships and, hence, exert varying controls on ground-water flow rates and directions. In this section, the general characteristics of these units, their lateral extents, and their interrelationships are described.

The hydrogeologic study area is shown in Figure 3-2. The boundaries of this area are based upon a synthesis of many published interpretations and supporting geologic and climatologic studies as referenced in Chapters 1 and 5. The hydrogeologic study area includes three ground-water subbasins: Oasis Valley, Alkali Flat-Furnace Creek Ranch, and Ash Meadows, which are all part of the larger Death Valley ground-water basin (Waddell et al., 1984).

3.6.1 HYDROGEOLOGIC UNITS

The hydrogeologic study area falls within the Alluvial Basins Ground-water Region as described by Heath (1984). This region is characterized by alternating basins (valleys) and mountain ranges with unconsolidated alluvial sediments filling in the basins and with consolidated rock cropping out in the mountains and underlying the alluvial deposits in the basins. Individual basins may be hydraulically linked into regional flow systems or may be hydraulically isolated. The degree of hydraulic linkage or isolation of the

hydrogeologic study area with adjacent subbasins or regional ground-water basins or flow systems, and the hydraulic interrelationships of individual subbasins within the study area itself are important considerations in defining and characterizing ground-water flow paths and velocities at Yucca Mountain. The areal distribution of the hydrogeologic units and their lithology and stratigraphy discussed in this section help to define the hydraulic communication on a regional scale and identify areas where additional information will clarify the overall characterization. Much of this information is summarized in Table 3-7. More detailed information about the geologic history and mode of occurrence of these units is provided in Chapter 1.

The existing data base on the areal extent, thickness, and interrelationships among the various geologic units within the hydrogeologic study area is sufficient to define regional hydrogeologic units and their general properties. No additional activities are designed to specifically refine the existing definition. However, the results of many of the planned activities will generate data that will help to further refine the level of knowledge about the regional hydrogeologic system.

3.6.1.1 Definition

A hydrogeologic unit (also referred to as a geohydrologic unit), as defined by Bates and Jackson (1980), is a geologic unit with consistent hydraulic properties such that the unit may be classified as an aquifer, confining layer (aquiclude or aquitard), or a combination of both, making up a framework for a reasonably distinct hydraulic system. The definition of a hydrogeologic unit is usually based on stratigraphic or lithologic characteristics, but as pointed out by Montazer and Wilson (1984), it may also be based upon structural considerations. In this section, the hydrogeologic units that occur in the saturated zone are defined and discussed. The hydrogeologic units recognized in the unsaturated zone at Yucca Mountain are described and discussed in Section 3.9.2.

3.6.1.2 Selection of units

The studies conducted by Winograd and Thordarson (1975) and additional assessments summarized by Waddell et al. (1984) identify the regional hydrogeologic units discussed in the following sections. These discussions are limited to those units that represent distinct hydraulic systems on a regional scale and do not necessarily correspond on a one-to-one basis with units discussed in Section 3.9.2 for the Yucca Mountain area. A generalized map showing the distribution of these units in the hydrogeologic study area is shown in Figure 3-8. As shown in Table 3-7, the regional hydrogeologic units in descending stratigraphic order are (1) the valley fill aquifer, (2) volcanic rock aquifers and aquitards, (3) the upper carbonate aquifer, (4) the upper clastic aquitard, (5) the lower carbonate aquifer, and (6) the lower clastic aquitard.

Table 3-7. Relation of stratigraphic units to hydrogeologic units in the hydrogeologic study area^a
(page 1 of 7)

System	Series	Stratigraphic unit	Major lithology	Maximum thickness (m)	Hydrogeologic unit (underlined) and hydrologic characteristics
Quaternary and Tertiary	Holocene, Pleistocene, and Pliocene deposits	Valley fill	Alluvial fan, fluvial, fanglomerate, lakebed, and mudflow ranges from 0.21 to 2.9 m/d	600	<u>VALLEY FILL AQUIFER</u> Transmissivity ranges from 10 to 400 m ² /d average coefficient of interstitial permeability ranges from 0.21 to 2.9 m/d. Interstitial porosity controls flow of water.
Tertiary	Pliocene	Basalt of Kiwi Mesa	Basalt flows, dense and vesicular	75	<u>LAVA-FLOW AQUIFER</u> Water movement controlled by primary and secondary fractures and possibly by rubble between intercrystalline porosity and conductivity negligible; estimated transmissivity ranges from 4.2 to 1,200 m ² /d; saturated only beneath east-central Jackass Flats.
		Rhyolite of Shoshone Mountain	Rhyolite flows	600	
		Basalt of Skull Mountain	Basalt flows	75	
	? ^b	Thirsty Canyon Tuff	Ash-flow tuff, partially to densely welded; trachytic lava flows	230	No corresponding hydrologic unit. Generally unsaturated; present beneath Black Mountain, north-western part of the basin.
	Miocene ^c	Piapi Canyon Group	Ammonia Tank Member	75	<u>WELDED AND BEDDED TUFF AQUIFER</u> Water movement controlled by primary and secondary joints in densely welded part of ash-flow tuff; transmissivity ranges from 1 to 1,200 m ² /d; intercrystalline porosity and conductivity negligible; nonwelded part of ash-flow tuff, where present, has relatively high
		Timber Mountain Tuff	Rainier Mesa Member	175	

Table 3-7. Relation of stratigraphic units to hydrogeologic units in the hydrogeologic study area^a
(page 2 of 7)

System	Series	Stratigraphic unit		Major lithology	Maximum thickness (m)	Hydrogeologic unit (underlined) and hydrologic characteristics
Tertiary	Miocene	Piapi Canyon Group	Tiva Canyon Member	Ash-flow tuff, nonwelded to densely welded; thin ash-fall tuff near here	90-100	interstitial porosity (35 to 50%) and modest permeability (0.08 m/d) and may act as a leaky aquitard; saturated only beneath deeper parts of Yucca, Frenchman, and Jackass flats.
			Topopah Spring Member	Ash-flow tuff, nonwelded to densely welded; thin ash-fall tuff near base	275	
			Bedded tuff (informal unit)	Ash-flow tuff and fluvially reworked tuff	300	
		Paintbrush Tuff				Transmissivity ranges from 2 to 10 m ² /d; saturated only beneath structurally deepest parts of Yucca, Frenchman, and Jackass Flats; occurs locally below ash-flow tuff members of Paintbrush Tuff and below Grouse Canyon Member of Belted Range tuff.
		Wahmonie Formation		Lava-flow and interflow tuff and breccia; locally hydrothermally altered	1,200	<u>LAVA-FLOW AND TUFF AQUITARD</u> Water movement controlled by connected fractures; interstitial thermally altered porosity and permeability negligible; transmissivity estimated less than 6 m ² /d; contains minor perched water beneath foothills between Frenchman and Jackass flats.
				Ash-fall tuff, tuffaceous sandstone, and tuff breccia all interbedded; matrix commonly clayey or zeolitic	500	Transmissivity ranges from 1 to 3 m ² /d; interstitial porosity is as high as 40%, but interstitial permeability is negligible; (3 x 10 ⁻⁴ to 3 x 10 ⁻⁶ m/d); owing to poor hydraulic connection of

Table 3-7. Relation of stratigraphic units to hydrogeologic units in the hydrogeologic study area^a
(page 3 of 7)

System	Series	Stratigraphic unit	Major lithology	Maximum thickness (m)	Hydrogeologic unit (underlined) and hydrologic characteristics
Tertiary	Miocene ^c	Salyer Formation	Breccia flow, lithic breccia, and tuff breccia, all interbedded with ash-fall tuff, sandstone, siltstone, claystone, matrix commonly clayey or zeolitic	600	fractures, interstitial conductivity probably controls regional ground-water movement; <u>perches minor quantities of water beneath foothills flanking valleys</u> ; fully saturated only beneath structurally deepest parts of Yucca, Frenchman, and Jackass flats; Grouse Canyon and Tub Spring members of Belted Range Tuff may locally be aquifers in northern Yucca Flat.
		Belted Range Tuff	Grouse Canyon Member Ash-flow tuff, densely welded	60	
			Tub Spring Member Ash-flow tuff, non-welded to welded	90	
		Local Informal Units	Ash-fall bedded tuff, nonwelded to semiwelded ash-flow tuff, tuffaceous sandstone. Siltstone, and claystone; all massively altered to zeolite or clay minerals; locally minor welded tuff near base; minor rhyolite and basalt	600	

Table 3-7. Relation of stratigraphic units to hydrogeologic units in the hydrogeologic study area^a
(page 4 of 7)

System	Series	Stratigraphic unit	Major lithology	Maximum thickness (m)	Hydrogeologic unit (underlined) and hydrologic characteristics
Tertiary	Miocene ^c	Rhyolite flows and tuffaceous beds of Calico Hills	Rhyolite, nonwelded and welded ash flow, ash-fall tuff, tuff breccia, tuffaceous sandstone; hydrothermally altered at Calico Hills; matrix of tuff and sandstone commonly clayey or zeolitic	600	<u>TUFF AQUIFER/AQUITARD</u> Rhyolite lavas and ash flows may be transmissive. Bedded tuffs may be zeolitized or argillized and less transmissive. Beneath Yucca Mountain the tuffaceous beds of Calico Hills are mostly unsaturated.
		Crater Bullfrog Tuff	Prow Pass Member Ash-flow tuff, nonwelded to moderately welded, interbedded with ash-fall bedded tuff; matrix commonly clayey or	>600	<u>TUFF AQUIFER/AQUITARD</u> Transmissivity ranges from less than 0.1 to several hundred m ² /d. Interstitial hydraulic conductivity is small (8×10^{-4} to 3×10^{-1} m/d). At Yucca Mountain, unit commonly contains the water table.
		Lithic Ridge Tuff	Member Tram Member Ash-flow tuff, partially to densely welded. Commonly argillized	300	<u>TUFF AQUITARD</u> Not well characterized. Transmissivity about 2×10^{-4} m ² /d. Interstitial hydraulic conductivity low (3×10^{-4} to 6×10^{-5} m/d).
	Miocene (?) Oligocene (?)	Older tuffs and lavas beneath Yucca Mountain	Altered rhyolitic and quartz latitic lavas, and altered bedded and ash-flow tuffs	?	
	Miocene and Oligocene	Rocks of Pavits Spring	Tuffaceous sandstone and siltstone, claystone; fresh water limestone and conglomerate; minor gypsum; matrix commonly clayey, zeolitic, or calcerous	425	See Salyer Formation.

Table 3-7. Relation of stratigraphic units to hydrogeologic units in the hydrogeologic study area^a
(page 5 of 7)

System	Series	Stratigraphic unit	Major lithology	Maximum thickness (m)	Hydrogeologic unit (underlined) and hydrologic characteristics
Tertiary (continued)	Oligocene	Horse Spring Formation	Fresh-water limestone, conglomerate, tuff	300	See Salyer Formation
Cretaceous to Permian		Granitic stocks	Granodiorite and quartz monzonite in stocks, dikes and sills	--	<u>(A MINOR PLUTONIC-ROCK AQUITARD)</u> Complexly fractured but nearly impermeable.
Permian and Pennsylvanian		Tippisah Limestone	Limestone	1,100	<u>UPPER CARBONATE AQUIFER</u> Complexly fractured aquifer; transmissivity estimated in range from 10 to 1,250 m ² /d; intercrystalline porosity and permeability negligible, saturated only beneath western one-third of Yucca Flat.
Mississippian and Devonian		Eleana Formation	Argilliate, quartzite, conglomerate, limestone	2,400	<u>UPPER CLASTIC AQUITARD</u> Complexly fractured but nearly impermeable; transmissivity estimated less than 5 m ² /d; interstitial permeability negligible but owing to poor hydraulic connection of fractures probably controls groundwater movement; saturated only beneath western Yucca and Jackass Flats; interstitial porosity ranges from 2.0 to 18%.

Table 3-7. Relation of stratigraphic units to hydrogeologic units in the hydrogeologic study area^a
(page 6 of 7)

System	Series	Stratigraphic unit	Major lithology	Maximum thickness (m)	Hydrogeologic unit (underlined) and hydrologic characteristics
Devonian	Upper	Devils Gate Limestone	Limestone, dolomite, minor quartzite	>425	<u>LOWER CARBONATE AQUIFER</u> Complexly fractured aquifer supplies major springs throughout eastern Nevada; transmissivity ranges from 10 to 10,000 m ² /d; intercrystalline porosity, 0.4 to 12%; intercrystalline hydraulic permeability, 9×10^{-7} to 4×10^{-3} m/d; solution caverns are present locally but regional ground-water movement is controlled by fracture transmissivity; saturated beneath much of study area.
	---?---	Middle	Nevada Formation	Dolomite	
		Undifferentiated	Dolomite	430	
Devonian and Silurian	Upper	Ely Springs Dolomite	Dolomite	90	
		Eureka Quartzite	Quartzite, minor limestone	100	
	Middle	Antelope Valley Limestone	Limestone and silty limestone	460	
Ordovician	---?---				
	Lower	Pogonip Group	Ninemile Formation	Claystone and limestone interbedded	100
			Goodwin Limestone	Limestone	>275
Cambrian	Upper	Nopah Formation			
		Smoky Member	Dolomite, limestone	325	
		Halfpint Member	Limestone, dolomite, silty limestone	220	
		Dunderberg Shale Member	Shale, minor limestone	70	

Table 3-7. Relation of stratigraphic units to hydrogeologic units in the hydrogeologic study area^a
(page 7 of 7)

(underlined) System	Series	unit	Stratigraphic lithology	Major (m)	Maximum thickness and hydrologic characteristics	Hydrogeologic unit
Cambrian	Middle	Bonanza King Formation				
		Banded Mountain Member	Limestone, dolomite, minor siltstone	750		
		Papoose Lake Member	Limestone, dolomite, minor siltstone	650		
		Carrara Formation	Siltstone, limestone, interbedded (upper part predominantly limestone; lower part predominantly siltstone)	320 290		
	Lower					
Precambrian		Zabriskie Quartzite	Quartzite	70		<u>LOWER CLASTIC AQUITARD</u>
		Wood Canyon Formation	Quartzite, siltstone, shale minor dolomite	700		Complexly fractured but nearly impermeable; supplies no major springs; transmissivity less than 10 m/d; interstitial porosity and permeability is negligible but probably controls ground-water movement owing to poor hydraulic connection of fractures; saturated beneath most of area; interstitial porosity 0.2 to 10%; interstitial hydraulic conductivity ranges from 3 x 10 ⁻⁴ to 4 x 10 ⁻³ m/d.
		Stirling Quartzite	Quartzite, siltstone	1025		
		Johnnie Formation	Quartzite, sandstone, siltstone, minor limestone and dolomite	975		
		Noonday(?) Dolomite	Dolomite	--		

^aSource: Waddell et al. (1984), which was modified from Winograd and Thordarson (1975).

^b? indicates uncertainty.

^cThe three Miocene sequences occur in separate parts of the region. Age correlations between sequences are uncertain. They

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GENERAL DIRECTION OF REGIONAL GROUND-WATER FLOW
(QUESTION MARK INDICATES UNCERTAINTY)

- A. VOLCANIC ROCK AQUIFERS AND AQUITARDS OVERLYING
UPPER CLASTIC AQUITARD OR LOWER CARBONATE AQUIFER
- B. VOLCANIC ROCK AQUIFERS AND AQUITARDS OVERLYING
LOWER CARBONATE AQUIFER
- C. LOWER CARBONATE AQUIFER
- D. UPPER CLASTIC AQUITARD
- E. VOLCANIC ROCK AQUIFERS AND AQUITARDS OF CALDERAS

NOTE: FOR CLARITY, VALLEY FILL AQUIFER NOT SHOWN

Figure 3-8. Generalized distribution of hydrogeologic units in the saturated zone of the region.

3.6.1.3 Valley fill aquifer

The valley fill aquifer is of Tertiary and Quaternary age and is composed of alluvial-fan, fluvial, fanglomerate, lakebed, and mudflow deposits (Table 3-7). The thickness and stratigraphic interrelationships of these deposits are highly variable reflecting varying depositional environments. Thicknesses more than 550 m occur in some areas, but because of the depths to water, areas where the alluvium is saturated are limited. Because of the limited thickness of saturated alluvium, this hydrogeologic unit is of minor importance in defining ground-water flow on a regional scale. The exception is beneath most of the Amargosa Desert, where the valley fill is present in saturated thicknesses sufficient to constitute an important aquifer.

3.6.1.4 Volcanic rock aquifers and aquitards

Volcanic rocks that occur in the region are of Tertiary age and consist of nonwelded to welded ash-flow tuffs and basaltic and rhyolitic flows (Table 3-7). Because of the complex stratigraphy of these volcanics, individual rock types are not differentiated on a regional scale. The aggregate thickness of this unit is not known, but in places it exceeds several thousand meters. These volcanics form the uppermost water-bearing unit throughout most of the northwestern part of the hydrogeologic study area (Figure 3-8).

3.6.1.5 Upper carbonate aquifer

The Tippipah Limestone of Pennsylvanian and Permian age constitutes the upper carbonate aquifer (Table 3-7). Although this unit is as much as 1,100 m thick, it is of minor regional hydrologic significance because it probably is saturated only beneath western Yucca Flat.

3.6.1.6 Upper clastic aquitard

The Eleana Formation of Devonian and Mississippian age constitutes the upper clastic aquitard. This unit comprises primarily argillite with minor quartzite and limestone (Table 3-7). The upper clastic aquitard acts as a confining unit separating the upper carbonate aquifer from the lower carbonate aquifer beneath western Yucca Flat and northern Jackass Flats. In these areas, the upper clastic aquitard is more than 2,400 m thick.

3.6.1.7 Lower carbonate aquifer

The lower carbonate aquifer comprises limestone in the upper part of the Cambrian Carrara Formation and the overlying limestones and dolomites of Cambrian through Devonian age (Table 3-7). The limestones of this aquifer occur extensively in the eastern part of the hydrogeologic study area and are

a major regional aquifer. In places the thickness of this aquifer exceeds 4,700 m.

3.6.1.8 Lower clastic aquitard

Siltstone, quartzite, shale, and sandstone of Precambrian through early Cambrian age comprise the lower clastic aquitard (Table 3-7). Thicknesses greater than 3,000 m of these rocks occur within the hydrogeologic study area. The lower clastic aquitard is an important regional unit because it probably significantly affects the distribution of hydraulic potential and the locations of ground-water discharge areas.

3.6.1.9 Spatial relationships of units

The spatial relationships of these units are primarily a result of the depositional history of the sedimentary rocks, the volcanic history of the area, and the complex structural evolution of the region. The rocks of the lower clastic aquitard and lower carbonate aquifer were deposited in a miogeocline in thick accumulations (Stewart and Poole, 1974). In middle Paleozoic time, a sequence composed of generally poor water-transmitting siliceous clastic rocks was thrust eastward about 150 km along the Roberts Mountains thrust. These rocks now overlie a sequence composed mostly of relatively soluble, water-transmitting carbonate rocks. Subsequent deposition of rocks forming the upper clastic aquitard and upper carbonate aquifer occurred, followed by additional structural activity. In the Tertiary, volcanic activity resulted in the formation of the volcanic rock aquifers and aquitards. More recent geomorphic processes resulted in the deposition of the valley fill aquifer and unsaturated valley fill over these rocks.

3.6.2 RELATIONSHIP AMONG HYDROGEOLOGIC UNITS

Because of the complex geologic history, ground-water flow relationships within and between the hydrogeologic units are also complex. Data are sufficient to allow generalizations of these relationships. The valley fill aquifer lacks lateral continuity. Ground water in the valley fill aquifer in the northeastern part of the hydrogeologic study area has a downward component of flow resulting in leakage of ground water into the underlying aquifers (Winograd and Thordarson, 1975).

Where several aquifers and aquitards are present, such as beneath Yucca Flat, hydraulic heads generally decrease with depth indicating downward leakage toward the lower carbonate aquifer (Winograd and Thordarson, 1975). Near Yucca Mountain, on the other hand, the reverse is true, and an upward hydraulic gradient occurs (Waddell et al., 1984). The upward gradient in this area is discussed in Section 3.9.3.

Although both upward and downward components of interunit flow occur, the primary component of ground-water flow, in a regional scale, is lateral. With respect to detailed characterization of the Yucca Mountain area,

knowledge of the distribution of vertical hydraulic gradients is important. However, a detailed definition of localized vertical gradients over the entire hydrogeologic study area is unnecessary in a regional reconnaissance. Because only limited data are presently available, data obtained during more specific studies will be used to help quantify the magnitude and rate of interunit flow (Section 8.3.1.2.1.2).

3.6.3 POTENTIOMETRIC LEVELS

The generalized locations of ground-water monitoring wells and the regional distribution of potentiometric levels referenced to mean sea level (msl) are presented in Figure 3-9. The distribution of wells provides an adequate data base for the general definition of the regional potentiometric surface. Sources of potentiometric data include Eakin et al. (1963), Malmberg (1967), Mifflin (1968), Winograd and Thordarson (1968), Rush (1970), Thordarson and Robinson (1971), Naff et al. (1974), Winograd and Thordarson (1975), Miller (1977), Harrill (1982), Waddell (1982), Czarnecki and Waddell (1984), Robison (1984), and Robison (1986). Specific construction information for these wells is provided in the cited references and are not included here. The equipotential lines shown on Figure 3-9 are based upon composite water levels from several hydrogeologic units and are averages of the hydraulic heads in the boreholes.

The altitude of the potentiometric surface ranges from over 1,900 m above msl in the northernmost part of the hydrogeologic study area to below sea level in Death Valley. Steep hydraulic gradients occur in several areas, including north of Yucca Mountain, near Death Valley, southeast of Mercury, and north of Yucca Flat. These steep gradients may be due to stratigraphic, structural, or hydraulic conditions, or to a combination of these conditions. Thermal conditions could also affect gradients, but preliminary analyses indicate that such effects probably are not significant influences on these steep gradients. The definition of areas of steep gradients and evaluation of the causes of these gradients are important considerations in determining ground-water flow paths and velocities. Additional syntheses and modeling of the regional hydrogeologic system will be performed as part of site characterization activities (Study 8.3.1.2.1.2). Two-dimensional and three-dimensional models of ground-water flow and sensitivity analyses performed during these studies will help to assess the relative importance of these steep gradients in the overall characterization of regional flow conditions.

The potentiometric data needed to establish temporal histories of variations in the potentiometric surface are available in selected areas, including Yucca Mountain, Amargosa Desert, Ash Meadows, and Franklin Lake playa. Collection of additional data is planned as part of the characterization of the regional ground-water flow system (Study 8.3.1.2.1.2). This data collection will include (1) regional potentiometric level studies to identify and compile additional data, (2) the expansion of the existing monitoring well network, and (3) local water-level monitoring.

Figure 3-9. Potentiometric surface of the hydrogeologic study area (based on composite water levels). Modified from: Waddell et al. (1984).

3.6.4 HYDRAULIC CHARACTERISTICS OF PRINCIPAL HYDROGEOLOGIC UNITS

This section provides information on the hydraulic characteristics of the hydrogeologic units. The capability of the various regional hydrogeologic units to store and transmit ground water is a function of their hydraulic properties. Estimates of ground-water flow paths and velocities require that the hydraulic characteristics of each of the important hydrogeologic units be known. The principal hydraulic characteristics discussed are transmissivity, porosity, and both interstitial and fracture hydraulic conductivity.

Information on the hydraulic properties of the principal regional units is listed in Tables 3-7 and 3-8. These data are based on information published by Winograd (1962a,b), Blankennagel and Weir (1973), Winograd and Thordarson (1975), and Lin (1978). Transmissivity estimates are based upon aquifer testing. In many instances, transmissivity was estimated from specific-capacity data from wells. In the absence of other aquifer test data, these estimates are useful to provide minimum values for transmissivity (Winograd and Thordarson, 1975). The porosity and hydraulic conductivity estimates are based upon laboratory studies, including measurements of grain and bulk densities, and mercury-injection and water-saturation methods.

Based upon the results of these tests, the most transmissive unit is the lower carbonate aquifer, which has transmissivity values as large as 900,000 gpd/ft (10,000 m²/d) (Winograd and Thordarson, 1975). All other aquifers have transmissivities that are at least an order of magnitude smaller. By comparison, the aquitards all have listed values less than about 800 gpd/ft (10 m²/d).

The transmissivity of the valley-fill aquifer ranges from about 800 gpd/ft to 34,000 gpd/ft (10 to 400 m²/d) (Winograd and Thordarson, 1975). The saturated thickness data suggest that these transmissivities reflect interstitial permeability ranging from 5 to 70 gpd/ft² (0.21 to 2.9 m/d).

The volcanic rocks function locally as either aquifers or aquitards, depending on the presence or absence of open, unsealed fractures. The character of these units at Yucca Mountain, where most of the more recent hydraulic data have been collected, is described in Section 3.9.2.2. The interstitial hydraulic conductivity of the volcanic units in the vicinity of the NTS ranges from 0.000007 to 4.1 gpd/ft² (2.8×10^{-7} to 1.6×10^{-1} m/d) (Winograd and Thordarson, 1975). The data indicate that the interstitial hydraulic conductivity of the densely welded zones of the volcanic rock aquifers and aquitards is extremely low and probably substantially limits the movement of ground water through these rocks unless they are fractured. Alternatively, the data indicate that the interstitial porosity and hydraulic conductivity of some nonwelded and partially welded tuff units within this hydrogeologic unit may be large enough to allow significant ground-water movement.

Because of the limited areal extent of the upper carbonate aquifer within the hydrogeologic study area, no data are available concerning its hydraulic properties. Similarly, data are limited on the hydraulic properties of the upper clastic aquitard. It has been inferred (Winograd and Thordarson, 1975) that the upper clastic aquitard is analogous to the lower

Table 3-8. Pumping-test data for aquifers in Nevada Test Site and vicinity^a (page 1 of 4)

Well	Stratigraphic unit	Depth interval (m)	Thickness (m)	Estimated penetration of aquifer (%)	Hydraulic setting	Depth to static water level (m)	Specific capacity (L/s per meter of drawdown) ^b	Transmissivity (m ² /d)			Remarks
								Estimated from specific capacity	Calculated from drawdown curve	Calculated from recovery curve	
79-69a	Carrara Formation	469-518	49	-- ^c	Unconfined	469	110	1 x 10 ⁴	--	--	Well 79-69a is 30 m north-west of well 79-69. During two pumping tests neither drawdown nor recovery could be measured owing to very high aquifer transmissivity and low pumping rates (3.7 and 13.4 L/s). Carrara Formation tapped by well is probably part of upper plate of low-angle thrust fault that crops out a few miles west of well.
79-69	Carrara Formation	469-503	34	--	Unconfined	469	1.28	75	(d)	(d)	Step-drawdown analysis indicates specific capacity of 2.4 (L/s)/m of drawdown and water entry chiefly from interval 490-495 m. Carrara Formation tapped by well is probably part of upper plate of low-angle thrust fault that crops out a few miles west of well.
67-73	Bonanza King Formation (Banded Mountain Member)	311-397	86	5	Confined	256	0.99	100	250	660	Step-drawdown analysis indicates specific capacity of 2.3 (L/s)/m of drawdown.
67-68	Bonanza King (?) ^e Formation	406-593	187		Confined						Step-drawdown analysis indicates specific capacity of 3.5 (L/s)/m of drawdown.
	Nopah Formation (?)	239-356	117	20	Unconfined	239	1.24	75	490	1,100	
66-75	Nopah Formation (Smoky Member)	225-454	224	10	Unconfined	225	0.93	50	140	330	Step-drawdown analysis indicates specific capacity of 2.2 (L/s)/m of drawdown.
88-66	Pogonip Group	777-1,042	266	10	Confined	626	0.08	9	16	66	Water yielded principally from interval 968-1,040 m.
75-73	Pogonip Group	336-565	229	10	Unconfined	336	0.14	7	47	--	--

Table 3-8. Pumping-test data for aquifers in Nevada Test Site and vicinity² (page 2 of 4)

Well	Stratigraphic unit	Depth interval (m)	Thickness (m)	Estimated permeability of aquifer (%)	Hydraulic setting	Depth to static water level (m)	Specific capacity (l/s per meter of drawdown)	Transmissivity calculated from drawdown curve (m ² /d)	Remarks
73-66	Silurian (?) dolomite	956-1,036	80	<5	Confined	529	6.21 ^a	740	Density changes in water column, due to anomalously high water temperature, completely masked water-level fluctuations due to pumping. Air-line measurements permitted approximation of specific capacity.
87-62	Devils Gate Limestone and Nevada Formation undifferentiated	1,128-1,280	172	<20	Confined	600	0.17	12	--
84-68d	Devonian (?) dolomite and calcareous quartzite	860-922	62	<5	Confined	596	0.08	9	30
81-67	Bedded tuff (?) of Piapi Canyon Group	514-549	35	--	Confined	478	0.19	12	16
90-74	Bedded tuff (?) of Piapi Canyon Group	149-204	55	--	Unconfined (?)	149	0.08	2	Constant-rate pumping test not made; specific capacity based on measurements made after 80 min pumping; aquifer is probably bedded tuff or non-welded ash-flow tuff.
90-75	Bedded tuff (?) of Piapi Canyon Group	273-333	59	--	Unconfined (?)	273	0.12	5	Constant-rate pumping test not made; specific capacity reported after 30 min of pumping; aquifer is probably bedded tuff or nonwelded ash-flow tuff.

Table 3-8. Pumping-test data for aquifers in Nevada Test Site and vicinity^a (page 3 of 4)

Well	Stratigraphic unit	Depth interval (m)	Thick-ness (m)	Estimated penetration of aquifer (%)	Hydraulic setting	Depth to static water level (m)	Specific capacity (L/s per meter of drawdown) ^b	Transmissivity (m ² /d)			Remarks
								Estimated from specific capacity	Calculated from drawdown curve	Calculated from recovery curve	
WELDED-TUFF AQUIFER											
75-58	Topopah Spring Member of Paintbrush Tuff	226-270	45	40	Confined (?)	226	11.6	1,240	See remarks	See remarks	Drawdown of 2.1 m measured with air line and test pressure gage in first 3 min of pumping test at rate of 24.4 L/s. Additional drawdown not detectable in subsequent 57 min of pumping test.
74-57	Topopah Spring Member of Paintbrush Tuff	283-444	161	100	Confined (?)	283	4.6	500	850	--	Step-drawdown analysis suggests considerable head losses at face of bore; losses are probably due to poor gun perforation of casing.
81-69	Topopah Spring Member of Paintbrush Tuff	459-511	51	100	Confined (?)	457	0.02	2.5	See remarks	0.6	Well tested by bailing.
LAVA-FLOW AND WELDED-TUFF AQUIFERS											
74-61	Basalt of Kiwi Mesa	317-351	34	100	Unconfined	317	0.52	50	350	--	Combined test of lava-flow and welded-tuff aquifers. Measurements made with test pressure gage and air line.
	Topopah Spring Member of Paintbrush Tuff	351-465	55	45	Confined	--	--	--	--	--	
VALLEY-FILL AQUIFER											
74-70b	Valley fill	210-366	156	60	Unconfined	210	0.35	12	30	31	Value of 21 m ² /d from recovery during 133-day shut-down; other values from 48-hour pumping test.

3-67

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Table 3 8. Pumping-test data for aquifers in Nevada Test Site and vicinity^a (page 4 of 4)

Well	Stratigraphic unit	Depth interval (m)	Thick-ness (m)	Estimated penetration of aquifer (%)	Hydraulic setting	Depth to static water level (m)	Specific capacity (l/s per meter of drawdown) ^b	Transmissivity (m ² /d)			Remarks
								Estimated from specific capacity	Calculated from drawdown curve	Calculated from recovery curve	
74-70a	Valley fill	208-274	66	15	Unconfined	208	0.83	37	95	136	--
75-72	Valley fill	218-285	48	--	Unconfined	218	0.27	10	--	--	Specific capacity and static water level reported by driller.
83-68	Valley fill	489-570	80	80-100	Unconfined	185	0.39	12	160	150	--
91-74	Valley fill	33-113	80	100	Unconfined	33	2.48	120	--	400	Specific capacity after about 211 hours of pumping; driller's log indicates mostly clay below 71.9 m.
91-74a	Valley fill	35-165	130	100	Unconfined	35	6.21	370	--	--	Specific capacity and static water level reported by driller; driller's log suggests chief aquifer in depth interval 34.7-61.0 m.

^aModified from Winograd and Thordarson (1975).

^bSpecific capacity computed at 100 min of pumping.

^c indicates no data.

^dTime drawdown curves in Winograd and Thordarson (1975) indicate a positive boundary of very high transmissivity at 35 min; the "zone" of high transmissivity probably is that tapped by adjacent well 76-69a.

^e(?) indicates data uncertainty.

clastic aquitard and probably has a transmissivity less than about 500 gpd/ft ($6.0 \text{ m}^2/\text{d}$). No porosity data are available for either of these units.

Testing of the lower carbonate aquifer indicates that extremely high transmissivities occur. At well 79-69a (Table 3-8), completed in the Carrara Formation, a transmissivity of about 900,000 gpd/ft ($1 \times 10^4 \text{ m}^2/\text{d}$) was estimated on the basis of specific-capacity data. The total porosity of the lower carbonate aquifer ranges from 0.4 to 12 percent (Table 3-7), with an effective porosity ranging from 0.0 to 9.0 percent (Winograd and Thordarson, 1975).

Values of the interstitial porosity and hydraulic conductivity of the lower clastic aquitard given in Table 3-7 are based on analyses of 43 cores. Although restricted to cores from a single test hole, the data are considered representative of the lower clastic aquitard throughout the study area, because the borehole penetrated 1,072 m of the aquitard and the cores resemble examined outcrop specimens. The total porosity of this unit averages about 4 percent and the effective porosity is about 2 percent.

The transmissivity values summarized above are based upon analyses that assume porous media rather than fracture flow conditions and which presume that the ground-water flow in the saturated rocks satisfies Darcy's law (i.e., the flow rate through porous media is proportional to the head loss and inversely proportional to the length of the flow path). For the purposes of regional characterization, such analyses are considered appropriate. However, their utility for calculation of ground-water flow paths and velocities over distances of a few kilometers or less is questionable. To determine the validity of the use of porous media solutions for flow through fractured media in the hydrogeologic study area, additional evaluations will be performed as summarized in Section 8.3.1.2.3.3. These evaluations will be performed as part of the assessment of regional hydrogeologic data needs in the saturated zone.

3.7 REGIONAL GROUND-WATER FLOW SYSTEM

This section describes the general ground-water flow and principal flow paths in the hydrogeologic study area and its component subbasins. The system hydrochemistry is presented, the ages of ground water are estimated, and the system paleohydrology is described. Information about the regional ground-water flow system is required to define ground-water flow paths, to estimate the effects of climatic changes on the water table altitude at the site, and to understand the ground-water flow system at the site.

Hydrologic conditions beneath Yucca Mountain are controlled in part by the regional ground-water flow system. Flow paths and water velocities beneath the site are determined by aquifer properties and hydraulic gradients. Because hydraulic gradients are affected by the regional distribution of permeability, and the locations and amounts of ground-water recharge and discharge, knowledge of the regional system is used for assessment of hydrologic conditions at Yucca Mountain.

3.7.1 IDENTIFICATION OF RECHARGE AND DISCHARGE AREAS

This section presents information on the (1) location of areas of ground-water recharge to and discharge from the ground-water system, (2) modes of recharge and discharge, (3) residence time of ground water, (4) bulk rates of ground-water flow for specific hydrogeologic units, and (5) surface-water ground-water interrelations.

3.7.1.1 Location of ground-water recharge and discharge areas

The recharge and discharge areas in the ground-water basin shown in Figure 3-7 are taken from Winograd and Thordarson (1975), Waddell (1982), and Waddell et al. (1984). The recharge areas shown are those areas of the basin that receive 200 mm or more average annual precipitation (Winograd and Thordarson, 1975) and are estimated to have relatively significant ground-water recharge, minor recharge probably occurs elsewhere in the basin, including Yucca Mountain. Surface-water runoff along major stream channels, such as Fortymile Wash, probably results in significant recharge, under both modern and post-pluvial conditions (Claassen, 1983; Czarnecki, 1985). Section 8.3.1.2.1 presents plans to evaluate the significance of this recharge. In addition, subsurface inflow probably occurs from the northeast, and may represent up to 58 percent of the total flow coming into the hydrogeologic study area (Czarnecki, 1985). The inflow across the northern and north-eastern boundaries results from recharge in the higher areas to the north of the hydrogeologic study area. The northern boundary of the study area was arbitrarily located and is assumed to be a constant flow boundary rather than a no-flow boundary.

The principal areas of discharge are in the southern Amargosa Desert. Smaller, less significant areas are near Beatty, Indian Springs, and in Death Valley (Figure 3-7).

An additional potential source of inflow to the hydrogeologic study area is recharge from precipitation. Deep percolation from runoff originating from precipitation falling on the basin is included in the estimated recharge from precipitation.

3.7.1.2 Model of recharge and discharge

An empirical method of estimating average annual ground-water recharge from precipitation in desert regions was developed by Eakin et al. (1951). Recharge was estimated as a percentage of the average annual precipitation within an area. Geographic zones in which average precipitation ranges between specified limits were delineated on a map and a percentage of precipitation was assigned to each zone; this represented assumed average recharge from average annual precipitation in that zone. The degree of reliability of the estimate so obtained is related to the degree to which the values approximate actual precipitation, and the degree to which the assumed percentage represents actual percentage of recharge. Neither of these factors is known precisely enough to ensure a significant degree of reliability

for any small area. Crosthwaite (1969) and Watson et al. (1976) compared this method to other methods of estimating recharge and concluded that it was suitable for obtaining an approximate estimate of recharge. This method has proved useful for reconnaissance estimates, and experience in using the method throughout Nevada and the desert areas of western Utah indicates that estimates probably are relatively close to actual long-term average annual recharge (Watson et al., 1976) in many large areas in these desert regions. However, the smaller the scale examined, the greater is the uncertainty of the method.

Claassen (1983) suggests that recharge to the valley-fill aquifer in the west-central Amargosa Desert, based on hydrochemical data, resulted primarily from overland snow-melt runoff during late Pleistocene time, in or near present-day stream channels such as Fortymile Wash. Today this area receives an average precipitation of less than 200 mm/yr. Infrequent, small, and localized modern floods still occur in these channels, and some recharge from them probably does occur.

In his estimates of recharge in the Alkali Flat-Furnace Creek Ranch ground-water subbasin, Czarnecki (1985) assumed that recharge associated with annual precipitation rates of less than 76 mm/yr was minor. Further, in his closing remarks, he states that "additional work is needed to document recharge mechanisms and rates, and to establish analytical expressions between precipitation rates and associated ground-water recharge rates." This is particularly true for channels such as Fortymile Wash on the east side of Yucca Mountain, where highly transmissive stream-channel sediments possibly focus recharge during flooding events. Little is known about the rate or distribution of deep infiltration there. Recharge through Fortymile Wash may affect the water-table altitude and gradient and flow-path directions beneath Yucca Mountain. Increased recharge through Fortymile Wash could result in a water table with higher altitudes and in modified gradients and ground-water flow rates (Czarnecki, 1985).

The amount of estimated average annual ground-water recharge from precipitation has been computed by Rush (1970) to be about 4.5×10^4 acre-ft ($5.6 \times 10^6 \text{ m}^3$) for a 16,000 km² area of the geohydrologic study area. This is equivalent to an average recharge rate of 3.5 mm/yr over the total area. Winograd and Friedman (1972) estimate that as much as 35 percent of the discharge at Ash Meadows in the southern Amargosa Desert (Figure 3-7, may be ground-water underflow from Pahrnatag Valley. The springs at Ash Meadows flow about $2 \times 10^6 \text{ m}^3/\text{yr}$ (Walker and Eakin, 1963); therefore, the underflow from Pahrnatag Valley may be on the order of $7 \times 10^6 \text{ m}^3/\text{yr}$.

Although some uncertainty remains on the quantities and distributions of recharge over the region, the level of uncertainty does not appear significant in terms of affecting site specific analyses and interpretations for Yucca Mountain. Nonetheless, additional work is planned through regional modeling and water balance measurements as described in Section 8.3.1.2.1.

Ground-water discharge from the ground-water basin is by (1) spring flow, (2) evapotranspiration from phreatophyte areas where depths to ground water are less than about 15 m, and (3) evaporation from bare soil areas such as playas, where depth to ground water is less than about 5 m (Rush, 1970), and to a lesser extent, well withdrawals. At the major discharge areas in

the basin (Figure 3-7), discharge is by evapotranspiration by phreatophytes and by natural springs that flow at the land-surface contact between transmissive and less-transmissive hydrogeologic units. The less-transmissive units act as barriers to ground-water flow, causing ground water to flow upward to land surface (Winograd and Thordarson, 1975). Major ground-water discharge is summarized in Table 3-9. Modeling studies by Waddell (1982) and Czarnecki and Waddell (1984) have shown recharge to be a highly sensitive element of regional flow models and additional studies are planned to determine discharge rates (Section 8.3.1.2.1).

Table 3-9. Major ground-water discharge in the hydrogeologic study area

Area (see Figure 3-9)	Nature of discharge	Estimated average discharge (m^3/yr)	Reference
Southern Amargosa Desert	Springs, evaporation, and evapotranspiration	3.0×10^7	Rush (1970), Walker and Eakin (1963),
Death Valley	Springs	6.3×10^6	Waddell (1982)
Near Beatty in Oasis Valley	Springs and evapotranspiration	2.7×10^6	Malmberg and Eakin (1962)
Indian Springs Valley	Springs	9.8×10^5	Maxey and Jameson (1948)
Total discharge		4.0×10^7	

Production wells contribute to the total discharge of the system. Data presented in Table 3-9 and tables in Section 3.8 indicate that well production may represent one third of the total discharge (total discharge from wells is $1.9 \times 10^7 \text{ m}^3/\text{yr}$).

The total discharge shown in Table 3-9 does not include diversion from the study area by wells; well discharge is described in Section 3.8. Flow paths of ground water between recharge and discharge areas and the controls on the flow paths are discussed in Section 3.7.2.

3.7.1.3 Residence times of the ground water

Residence time of water in the hydrogeologic units is dependent on two factors: (1) ground-water velocities and (2) travel distances. Each of these factors is controlled by the highly complex hydrogeologic characteristics of the ground-water system.

A commonly used method for determining ground-water age is carbon-14 dating, although the presence of carbonate rocks along the flow path has an affect on the apparent ages determined (Tamers, 1967; Section 3.7.4). Claassen (1985) discusses limitations in determining ground-water ages and suggests various methods for correcting apparent ages of ground water. He also lists apparent ages of ground water near and at Yucca Mountain (9,100 to 17,000 yr before present) and in the Amargosa Desert (7,300 to 21,400 yr before present). Based on these apparent ages and length of flow paths, Claassen (1985) estimated ground-water velocities to be slower than 4 to 7 m/yr in the Amargosa Desert. Apparent carbon-14 ages from the tuff aquifers are probably not subject to much error, but apparent ages of the carbonate aquifers are assumed to require significant correction. Carbon-14 ground-water data and their interpretation for the Yucca Mountain vicinity are presented and discussed in more detail in Section 3.7.3.2, based on Claassen (1985) and Benson and McKinley (1985).

In a discussion of ground-water velocity, Winograd and Thordarson (1975) developed some estimates for residence and travel time for ground water beneath parts of the basin:

1. Beneath Yucca Flat, assuming an average saturated thickness of about 300 m and vertical movement, the time needed for a water particle to move from the top to the bottom of the tuff aquitard is about 6,000 to 2,000,000 yr, corresponding to velocities of 6×10^{-2} m/yr to 1.5×10^{-4} m/yr.
2. A carbon-14 date of water from the valley-fill aquifer beneath Frenchman Flat suggests that the age of the ground water in the tuff aquitard probably is in the range of several tens of thousands of years based on a sample with an apparent age of 13,000 yr. An age of several hundred thousand years is not beyond possibility for waters near the base of the tuff aquitard beneath Yucca or Frenchman flats, because vertical flow velocities through the underlying aquitard are likely to be very small, based on current rock-property data and recharge estimates.
3. Estimates of the velocity in the lower carbonate aquifer beneath central Yucca Flat range from about 6 to 600 mm/d, or from about 2 to 200 m/yr. Estimated average velocities beneath the Specter Range (15 km southwest of Mercury) are about 100 times larger than those for the lower carbonate aquifer beneath central Yucca Flat, or about 200 m to 20 km/yr and probably represent an upper limit of the ground-water velocities in the lower carbonate aquifer. Velocity estimates for the lower carbonate aquifer beneath Yucca Flat and the Specter Range principally reflect differences in the estimates of the volume of ground water flowing through each area used in estimating these ground-water velocities. These velocities were calculated by dividing the estimated flow through the aquifer by the carbonate aquifer cross-sectional area and the estimated effective porosity of 0.01 to 1 percent.

From wells in the area of Yucca Mountain, Waddell et al. (1984) report apparent carbon-14 ages of ground water ranging from 2,280 to 17,000 yr before present. The large variation in age is probably related to the travel

distances from point of recharge. However, the extent to which mixing of differently aged ground waters may affect apparent ages in the hydrogeologic study area is unknown at present. Age of ground water beneath Yucca Mountain is more fully discussed in Section 3.9.1.3.

3.7.1.4 Bulk rates of ground-water flow

Bulk rates of ground-water flow for specific hydrogeologic units have not been determined; however, Waddell (1982), using a two-dimensional flow model, calculated unit fluxes for 12 locations, extending from Beatty to Death Valley; the values range from $7.0 \text{ m}^2/\text{yr}$ to $2.8 \times 10^3 \text{ m}^2/\text{yr}$. Also, Czarnecki and Waddell (1984) used a two-dimensional ground-water flow model of an area approximately equivalent to the Alkali Flat-Furnace Creek Ranch subbasin to calculate vertically integrated specific fluxes ranging from about $3.6 \text{ m}^2/\text{yr}$ to $2.9 \times 10^3 \text{ m}^2/\text{yr}$. Although only limited field data were available to calibrate the model, the calculated flux rates are consistent with total water balance data, gradients, and hydraulic conductivity and transmissivity data.

Additional modeling work is planned to estimate more accurately regional ground-water flux rates, as described in Section 8.3.1.2.1.

3.7.1.5 Surface-water ground-water interrelations

Runoff and the ponding of water generally result only from unusually intense precipitation or from rapid rates of snow melt. Because of rapid evaporation and commonly high potential for infiltration, those bodies of surface water do not persist in large areas for long periods of time. As a result, within most of the study area, perennial streams and lakes are absent or insignificant. The exceptions are the ground-water discharge areas, where springs support small pools of water and short perennial channels of stream-flow. The flow commonly is used by man, evaporates, supports phreatophytes, or infiltrates back to the ground-water system where it continues to migrate in the subsurface. In the discharge areas, surface-water bodies generally are in hydraulic continuity with the saturated ground-water system. In recharge areas surface-water bodies are generally separated from the saturated zone by a thick unsaturated zone, so that a direct, saturated connection is not present.

Surface-water hydrology is discussed more completely in Section 3.1.

3.7.2 PRINCIPAL GROUND-WATER FLOW PATHS

As described in Section 3.7.1, ground-water movement is from recharge areas mostly in the northern and eastern part of the basin to the discharge areas of the western and southern part of the basin (Figure 3-10). Ground water flows generally southward or southwestward through most of the hydrogeologic study area; however, flow is westerly in the southeastern part

Figure 3-10. Regional ground-water flow paths. Modified from Waddell et al. (1984).

(Figure 3-10). Contour lines of the potentiometric surface are used only to determine the general direction of ground-water flow; local flow direction depends not only on the hydraulic gradient but also on anisotropy, of which little is known in the study area.

Temperature profiles in wells may be indicative of the local movement of ground water. Temperature variations with depth have been measured in selected boreholes in the hydrogeologic study area and surrounding area (Sass and Lachenbruch, 1982). Sass and Lachenbruch have interpreted a large area of anomalously low heat flow in eastern Nevada, extending into the north-eastern part of the hydrogeologic study area. The area, called the Eureka Low, is interpreted to possibly have a downward component of ground-water flow on the order of a few millimeters per year, which is consistent with regional interpretations of recharge and discharge zones by Winograd and Thordarson (1975). The heat flow to land surface in the Eureka Low is less than 0.6 W/m^2 . Elsewhere in the region, upward heat flow is higher and more typical of upward heat flow in the Great Basin (0.6 to 1.0 W/m^2 , see Figure 1-52 in Chapter 1). Beneath Pahute Mesa, the reported heat flow to the land surface is unusually low (0.4 W/m^2); this may be the result of heat being transported out of the area by laterally flowing ground water in the upper 3 km of rock (Sass and Lachenbruch, 1982). That interpretation is consistent with interpretations based on geohydrologic data by Winograd and Thordarson (1975) and Blankennagel and Weir (1973). According to Sass and Lachenbruch (1982), based on an analysis of 60 boreholes on the NTS, thermal gradients indicate that the predominant vertical ground-water flow may be downward in the upper 2 to 3 km, but local upwellings may exist. However, only one of the 60 boreholes studied by Sass and Lachenbruch (1982) was completed in the manner required for a confident analysis of thermal effects of natural ground-water flow. This results in uncertainties regarding complications of local vertical flow (Sass and Lachenbruch, 1982). The influence of temperature variations and heat flow on the regional and site ground-water flow systems will be more thoroughly assessed in site characterization and related to other factors that influence ground-water flow.

The hydrogeologic study area includes three subbasins: Oasis Valley on the west, Ash Meadows on the east, and the centrally located Alkali Flat-Furnace Creek Ranch subbasin. The subbasins consist of recharge areas and flow paths to a major discharge area (Figure 3-7). Each is described in the following sections.

3.7.2.1 Oasis Valley subbasin

Ground-water discharges by evapotranspiration and spring flow near Beatty (Figure 3-7) in Oasis Valley (Malmborg and Eakin, 1962). The water flows to the discharge area from the north and northeast, as shown by Mifflin (1968), Rush (1970), Scott et al. (1971), Blankennagel and Weir (1973), Winograd and Thordarson (1975), Waddell (1982), Waddell et al. (1984). The boundary for the subbasin shown in Figure 3-7 is from Waddell et al. (1984). According to Malmborg and Eakin (1962), virtually all the ground-water discharge from Oasis Valley is by evapotranspiration and underflow to the Amargosa Desert. The estimated average annual evapotranspiration, as summarized in Section 3.7.1.2, is about $2.7 \times 10^6 \text{ m}^3/\text{yr}$, derived from both

spring flow and a shallow water table. The reported spring flow was about $2 \times 10^6 \text{ m}^3/\text{yr}$ (Table 3-5); and additional $4.9 \times 10^5 \text{ m}^3/\text{yr}$ was estimated by them to be ground-water outflow to the Amargosa Desert.

Flow paths are from the principal recharge areas, mostly in the central part of the subbasin (Figure 3-7), southward to the discharge area near Beatty. This flow is principally through volcanic-rock aquifers to the valley-fill aquifer underlying the discharge area. It is not known whether the lower carbonate aquifer is present at considerable depth below parts of the subbasin (Figure 3-7). According to Waddell (1982), the valley fill aquifer is underlain by the lower clastic aquitard. Some of the spring discharge is warm, suggesting deep circulation. Ground water not discharged by evapotranspiration flows southward through the valley fill aquifer, past an alluvium-filled narrows, and into the Amargosa Desert.

3.7.2.2 Ash Meadows subbasin

The Ash Meadows subbasin boundaries (Figure 3-7) are associated with mountain ranges and are generally ground-water divides (potentiometric highs) resulting from greater amounts of precipitation associated with higher land surface altitudes. Along the eastern side of the subbasin, ground water probably flows from Pahrnat Valley into Tikapoo Valley (Winograd and Thordarson, 1975). Winograd and Friedman (1972) estimated that 35 percent of the discharge at Ash Meadows is derived from Pahrnat Valley, based on differences in deuterium content of water from Pahrnat Valley, the Spring Mountains and Sheep Range, and Ash Meadows. Cowart (1979) suggested that the Pahrnat Valley and Spring Mountains are the only significant recharge areas for the Ash Meadows subbasin.

Loeltz (1960) speculated that the source of much of the Ash Meadows discharge was precipitation on the western slope of the Spring Mountains (Figure 3-7) following a flow path beneath Pahrump Valley. However, since 1960, more data have become available that are inconsistent with Loeltz's hypothesis. Based on potentiometric data, ground-water flow from Pahrump Valley to Ash Meadows may exist; however, the unnamed hills between these valleys are underlain by rocks of the lower clastic aquitard (Winograd and Thordarson, 1975; Harrill, 1982). Therefore, the amount of flow from Pahrump Valley to Ash Meadows, if any, probably is minor compared with other recharge and discharge fluxes in the subbasin. Naff et al. (1974) suggest that inter-basin flow is improbable because the lower clastic aquitard crops out nearly continuously between Pahrump Valley and the spring discharge area. Head relations between Pahrump Valley and the Amargosa Desert also suggest a damming effect somewhere between.

Based on the above information, the boundary of the subbasin shown on Figure 3-7 is based primarily on the Winograd and Thordarson (1975) interpretation, which excluded the Spring Mountains and Pahrump Valley from the subbasin.

According to Winograd and Thordarson (1975), the flow paths are from the principal recharge area, mostly in the northern, eastern, and southern mountains of the subbasin, mostly westward to Ash Meadows in southern Amargosa

CONSULTATION DRAFT

Desert where most of the ground water is discharged by many large springs. The ground water migrates downward to the lower carbonate aquifer from overlying aquifers, as described in Section 3.6 and earlier parts of Section 3.7. Flow paths are determined by hydraulic head differences in open fractures in the lower carbonate aquifer.

Flow in the Ash Meadows subbasin is primarily in the lower carbonate aquifer. Hydraulic gradients are low (commonly less than 0.0002) because of high transmissivities of the soluble, fractured lower carbonate aquifer (Winograd and Thordarson, 1975). Distribution of head and direction of flow are greatly affected by permeability differences among the hydrogeologic units (Winograd and Thordarson, 1968).

According to Winograd and Thordarson (1975) spring discharge of $8.7 \times 10^5 \text{ m}^3/\text{yr}$ occurs at Indian Springs. It is probably fed by upward leakage of ground water from the lower carbonate aquifer by way of the valley fill aquifer. A major hydraulic barrier probably exists in the vicinity of the spring that causes the ground water to flow upward.

From northern Yucca Flat to Ash Meadows, ground water apparently flows downward through the valley fill aquifer and volcanic-rock hydrogeologic units into the underlying carbonate aquifers beneath Yucca Flat. The potentiometric level is as much as 43 m lower in the carbonate aquifer than in the overlying units (Winograd and Thordarson, 1975; Doty and Thordarson, 1983). Ground water from the overlying aquifers is drained by the more transmissive underlying carbonate aquifers. These interpretations are based primarily on data from two well locations. It may be desirable to confirm the interpretations with one or more additional wells with multidepth head measurements. Plans for collecting additional data on this type are discussed in Section 8.3.1.2.

The lower carbonate aquifer transmits most of the water in the subbasin, but other lithologies are locally important. Northeast of the Ash Meadows spring line, the saturated thickness of the valley fill aquifer probably is more than 100 m. Beneath Frenchman Flat, both valley fill and volcanic aquifers are saturated beneath the structurally deepest parts of the valley. Most of the valley-fill aquifer beneath Yucca Flat is unsaturated (Winograd and Thordarson, 1975).

Discharge from springs at Ash Meadows is estimated to be $2 \times 10^7 \text{ m}^3/\text{yr}$ (Walker and Eakin, 1963; Dudley and Larson, 1976). An additional unknown amount of ground water flows to the Alkali Flat-Furnace Creek Ranch subbasin to the west, some of which discharges by evapotranspiration (Winograd and Thordarson, 1975). A normal fault, downthrown to the southwest (Healey and Miller, 1971), probably juxtaposes a low-permeability lakebed aquitard or eolian deposits on the downthrown side of the fault against the lower carbonate aquifer across the fault, forcing flow upward (Dudley and Larson, 1976). Discharge is from springs in alluvium downgradient (southwest) of the fault. Regional transmissivities of about $40,000 \text{ m}^2/\text{d}$ have been calculated (Winograd and Thordarson, 1975) for the lower carbonate aquifer in the area northeast of Ash Meadows using estimated values for discharge, hydraulic gradient, and width of the aquifer; this figure is six to nine times greater than that determined from aquifer tests (Waddell, 1982).

3.7.2.3 Alkali Flat-Furnace Creek Ranch subbasin

Winograd and Thordarson (1975) suggest evidence that ground water flows from the south-central Amargosa Desert, through the lower carbonate aquifer, discharging at springs and seeps at Furnace Creek Ranch. Evidence supporting this belief consisted of the following: (1) proximity of spring discharge to the lower carbonate aquifer, (2) temperature relations of spring waters that are similar to those at Ash Meadows, and (3) the chemical quality of Travertine, Texas, and Nevares Springs are nearly identical. Ground water in the lower carbonate aquifer beneath the south-central Amargosa Desert may be derived from two sources--direct southwestward underflow from the Ash Meadows ground-water subbasin through the lower carbonate aquifer (assuming that the aquifer is extensive beneath the central Amargosa Desert), and downward leakage from the valley-fill aquifer beneath the central and south-central Amargosa Desert.

Discharge by natural processes in the Amargosa Desert was estimated by Walker and Eakin (1963) to be about $3 \times 10^7 \text{ m}^3/\text{yr}$, based on mass-balance estimates, which they acknowledge are "crude"; ground-water outflow southward from the subbasin was estimated to be about $6.2 \times 10^5 \text{ m}^3/\text{yr}$. Discharge near Furnace Creek Wash and Cottonball Marsh was estimated, based on some direct measurements of spring discharge and mapping of seeps in alluvium, to be about $6.3 \times 10^6 \text{ m}^3/\text{yr}$ (Hunt et al., 1966). Total flux (volumetric flow rate) in the Alkali Flat-Furnace Creek Ranch subbasin, not including the discharge in the Oasis Valley and Ash Meadows subbasins, is estimated to be about $2 \times 10^7 \text{ m}^3/\text{yr}$ (Waddell, 1982). Waddell (1982) used a parameter-estimation model and determined that total flux into and out of the ground-water system is about $4.3 \times 10^7 \text{ m}^3/\text{yr}$. This value is in good agreement with the total ground-water discharge of $4.0 \times 10^7 \text{ m}^3/\text{yr}$ indicated in Table 3-9.

Using a two-dimensional, finite-element, ground-water flow model, flow paths in the Alkali Flat-Furnace Creek Ranch subbasin were simulated by Czarnecki and Waddell (1984). The model was calibrated against measured head and boundary conditions. The results indicate that ground water flows predominantly southward from the Timber Mountain area in the northern section of the subbasin, beneath Crater Flat and the Amargosa Desert, toward the south and southwest, and to discharge areas located at Franklin Lake playa (Alkali Flat) and Furnace Creek Ranch in Death Valley. The model also indicates that additional minor components of ground water flow into the subbasin occur (1) from Calico Hills southward into Jackass Flats, (2) from Rock Valley into Jackass Flats, (3) from the western Amargosa Desert, and (4) from lateral recharge of ground water from the Ash Meadows subbasin. A summary of ground-water fluxes into and out of the hydrogeologic study area is shown in Table 3-10. Location of fluxes are shown in Section 3.9.3.3.

3.7.3 ISOTOPIC AND REGIONAL HYDROCHEMISTRY

This section describes the hydrochemical and isotopic nature of the ground water within the saturated zone in the hydrologic study area. Based on this data, estimates of ground-water age, origins, residence times, and travel times are presented. Discussions are provided regarding (1) the relative degree of circulation within the hydrogeologic units, (2) areas and

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Table 3-10. Summary of fluxes into and out of the Alkali Flat-Furnace Creek Ranch subbasin^{a,b}

Flux location	Rate ^c (m ³ /d)	Percentage
Timber Mountain ^d	+ 31,446	57.3
Fortymile Wash	+ 22,140	40.3
Rock Valley	+ 1,079	2.0
Calico Hills	+ 158	0.3
Ash Meadows	+ 78	0.1
Western Amargosa Desert	+ 19	<0.01
Franklin Lake playa	- 35,600	64.8
Furnace Creek Ranch, Death Valley	- 19,320	35.2

^aSource: Czarnecki and Waddell (1984).

^bFlux calculated as a residual of the mass balance.

^c+ indicates flow is into subbasin; - indicates flow is out of subbasin.

^dTimber Mountain area was simulated as a constant head boundary in the model.

modes of recharge and discharge, and (3) the delineation of regional hydrochemical facies. Figure 1-10 (Section 1.2.1) provides the distribution of rock types in the region, Figure 3-11 shows the location of many of the regional physiographic features, and Figure 3-2 shows the hydrogeologic subbasins.

3.7.3.1 Regional hydrochemistry

The chemical composition of ground water in the region is principally determined by (1) reactions with carbonate and volcanic rocks or rock fragments; (2) concentration of dissolved chemicals by evaporation; (3) formation of smectites, zeolites, and evaporite minerals; and (4) mixing of waters of different compositions. Because of a scarcity of shales, quartzites and granites, reactions with these rocks do not produce significant changes in the chemical composition of water in the region.

3.7.3.1.1 Chemical composition of ground water

The chemical composition of the ground water within the hydrogeologic study area is discussed in terms of the tuffaceous, lower carbonate, and valley fill aquifers.

Figure 3-11. Location map showing regional physiographic features. Modified from Winograd and Thordarson (1975).

Tuffaceous aquifer

The sodium bicarbonate water, which dominates the tuffaceous rocks, evolves principally by dissolution of rhyolitic volcanic glass and subsequent precipitation of various zeolites and smectite clays (Claassen and White, 1979; White, 1979; Claassen, 1985). White (1979) presents chemical analyses for 13 water samples from the tuffaceous aquifer northwest of Yucca Mountain in the Oasis Valley. The major cations are sodium (50 to 200 mg/L), calcium (0 to 36 mg/L), potassium (1.5 to 11 mg/L) and magnesium (0 to 5 mg/L). The dominance of sodium ions can be attributed to the incongruent dissolution and hydrolysis of volcanic glass. White et al. (1980) report that incongruent dissolution in the tuff beds within Rainier Mesa preferentially releases sodium, calcium, and magnesium while retaining potassium. Calcium and magnesium show progressive depletion relative to sodium, not by ion exchange but due to the effective removal of bivalent ions as clinoptilolite and montmorillonite are precipitated. Greater percentages of sodium ion are associated with larger clinoptilolite:montmorillonite precipitation ratios (Claassen, 1985). The 13 water samples taken from the tuffaceous rocks in Oasis Valley show that the dominant anion is bicarbonate (115 to 330 mg/L), with sulfate, chloride, and fluoride present in concentrations of 0 to 217 mg/L, 14 to 3 mg/L, and 0 to 10 mg/L, respectively. Since the occurrence of carbonate minerals in the tuffaceous rocks is minimal, the principal source and control of bicarbonate is the reaction of dissolved soil-zone carbon dioxide with various mineral phases (White, 1979). Chloride is leached preferentially relative to fluoride from the fresh, previously unreacted, glass phase (White and Claassen, 1980). Aqueous silica (SiO_2) concentrations remain constant, showing supersaturation with respect to cristobalite and quartz, and saturation with respect to silica gel (amorphous hydrated SiO_2). Excess silica is removed from the aqueous systems as silica gel, which is observed in fractures in the tuffaceous rocks of Oasis Valley (White, 1979). For the 13 water samples from Oasis Valley, the pH is reported as ranging from 7.6 to 8.7, and temperatures range from 18 to 41°C.

White et al. (1980) report chemical analyses for 55 water samples from the pores and fractures of the saturated tuffs of Rainier Mesa. The data show that the fracture water is predominantly a sodium bicarbonate. The interstitial water, however, shows much higher concentrations of chloride and sulfate relative to bicarbonate. Relative concentrations are similar to those specified in White (1979) for the Oasis Valley. However, actual concentrations are less with the dominant ions sodium, bicarbonate, and chloride having maximum concentrations of 71 mg/L, 220 mg/L, and 62 mg/L, respectively. The pH values reported range from 6.8 to 8.3.

Two wells drilled in rhyolites in Fortymile Canyon produced water containing predominantly sodium (39 to 44 mg/L) and bicarbonate (54.9 to 56.7 mg/L). Calcium and sulfate are the next dominant, with concentrations of 10 to 23 mg/L and 19 to 22 mg/L, respectively (Waddell, 1985). Measured temperatures were 19 and 20°C. Waddell reports that four samples at varying depths in one well have dissolved oxygen contents of 2 to 5 mg/L and pH values of 7.1 to 7.6.

Benson and McKinley (1985) report the chemical composition of 25 water samples obtained from 14 test wells in and near the exploratory block (Section 3.9.1.3 and Chapter 4). Again, sodium and bicarbonate ions

predominate. Temperatures range from 22 to 56°C, and pH values are 6.6 to 9.2. Although plots of relative cation and anion concentrations indicate a general description of sodium bicarbonate water, the ground water at Yucca Mountain is not chemically homogeneous. The differences in ionic composition among the wells is outlined and the authors conclude that ground water from the wells in the Yucca Mountain area has a significant degree of lateral and vertical chemical variability.

Claassen (1973) reports chemical data from eight wells drawing water from the tuff strata in the north, southeast, and southwest areas of the (NTS). The analyses are in good agreement with the previously reported data.

The chemical composition of 15 ground-water samples from Pahute Mesa were reported by Blankennagel and Weir (1973). As with other waters from the tuff aquifer, the ground water of Pahute Mesa is of a sodium potassium bicarbonate type. Sodium plus potassium ranged from 58 to 99 percent of the total cations, with 8 of the samples having the alkali ions present at more than 90 percent. Bicarbonate is the most abundant anion with bicarbonate plus carbonate making up more than 50 percent of the total anions in 12 of the samples. Water from the western side of the mesa has dissolved solids concentrations ranging from 206 to 336 mg/L, and averaging 280 mg/L. In the eastern part of the mesa, the dissolved solids range from 117 to 248 mg/L, and average 200 mg/L. Blankennagel and Weir (1973) consider that the lesser concentrated water in the east may be due to its nearness to areas of recharge.

Chemical analyses for 12 ground-water samples primarily from the northern area of the NTS are presented by Claassen (1985). Claassen (1985) recognized that some of the samples were atypical of the regional tuffaceous aquifer. These samples "all represent very young, generally perched ground water," thus not having evolved chemically to the extent of the waters representative of the regional aquifer. Specifically, the decrease in calcium relative to sodium is presumed to result from greater quantities of clinoptilolite precipitated from a ground-water composition evolved by reaction with vitric tuff and, therefore represents a trend of increasing maturity of ground water.

Figure 3-12 provides a graphical description of water in the tuffaceous aquifer in the form of a Piper diagram.

Lower carbonate aquifer

Schoff and Moore (1964) recognized two varieties of water in the regional carbonate aquifer: a calcium magnesium bicarbonate type and a mixed type. Winograd and Pearson (1976) determined that the variation in composition corresponded to the distance from recharge area. The authors found that ground water in close proximity to the Spring Mountains recharge area was typical of water moving through an upland carbonate; that is, calcium (45 to 69 mg/L), magnesium (10 to 16 mg/L) and bicarbonate (180 to 290 mg/L) occurred as primary constituents, and sodium, potassium, chloride and sulfate were present in negligible quantities. Water that is primarily calcium magnesium bicarbonate is still found at intermediate points along the flow system; however, concentrations of sodium and sulfate are increased. Finally, near the Ash Meadows discharge area sodium, sulfate, and chloride are present as

Figure 3-12. Piper diagram showing the regional sodium bicarbonate character of the tuff aquifer.

major ions with concentrations of 69 to 100 mg/L, 80 to 110 mg/L, and 21 to 27 mg/L, respectively. The mixed waters of Schoff and Moore are characterized by these last two types. Winograd and Thordarson (1975) attribute this increase in sodium, sulfate, and chloride to the downward leakage of sodium- and sulfate-rich water from the tuff aquitard. Temperatures are typically between 26 and 38°C, with outliers at 6.5 and 64°C; pH values vary insignificantly around 7.4.

The chemical composition of water from four wells tapping the carbonate aquifer on the eastern side of the NTS are reported in Claassen (1973). Three of the wells are in an area where the carbonate aquifer is directly overlain by tuff beds. As expected, the sodium concentration in these waters is high (up to 142 mg/L) relative to calcium (80 mg/L maximum concentration) and magnesium (33 mg/L). Bicarbonate and sulfate are present in concentrations up to 589 mg/L and 76 mg/L, respectively. The fourth well is in an area where the producing zone is separated from the tuff by 637 ft (190 m) of limestone and dolomite, thereby making sodium less dominant than calcium and magnesium. With the exception of one measurement of 9.9, pH values are between 7.0 and 8.3. Temperatures vary insignificantly around 34°C.

Water quality data for three samples from the lower carbonate aquifer in the southern portion of Ash Meadows ground-water subbasin (Claassen, 1985) are characteristic of the mixed water previously described.

Water from the lower carbonate aquifer is depicted graphically in Figure 3-13 in a Piper diagram.

Valley fill aquifer

Water in the tuffaceous valley fill has a composition similar to that of water in the tuffaceous aquifer shown in Figure 3-12. However, because of the shallowness of the water table in some areas, evapotranspiration results in an increased concentration. Analyses of 16 water samples from the tuffaceous alluvium in Oasis Valley (White, 1979) show sodium concentrations ranging between 100 and 315 mg/L and bicarbonate measuring between 207 and 512 mg/L. Chloride (68-100 mg/L) and sulfate (53-250 mg/L) also show a general increase. Other ions--calcium, magnesium, potassium and fluoride--are present in quantities less than 40 mg/L. Although some additional reaction may occur between the ground water and the tuffaceous detritus, White (1979) maintains that the principal mechanism for concentrating the sodium and bicarbonate is the decrease in volume of water due to the proximity of the water table to the atmosphere and soil zone. Temperatures vary between 18 and 31.5°C.

Claassen (1985) presented analyses for approximately 55 wells and springs tapping the valley fill in the Amargosa Desert. Many of the samples were differentiated as to tuffaceous or carbonate alluvium. All analyses show a wide range of concentrations for sodium (30 to 250 mg/L), calcium (0 to 66 mg/L), bicarbonate (116 to 437 mg/L) and sulfate (25 to 235 mg/L). The average temperature for water in seven wells completed in the valley fill aquifer was reported as 21°C.

Figure 3-13. Piper diagram showing the regional mixed character of the lower carbonate aquifer.

There are instances where water from each of the above alluvial aquifers shows the effect of local playa deposits. These deposits tend to increase the concentrations of sulfate and chloride in associated ground water. Claassen (1985) reports that six wells in the Amargosa Desert drawing water associated with playas have sulfate and chloride compositions ranging from 89 to 186 mg/L and 24 to 70 mg/L, respectively. By contrast three nearby wells not associated with playa deposits have sulfate and chloride concentrations ranging from 27 to 35 mg/L and 6 to 10 mg/L, respectively.

The chemical composition of water in the valley fill is depicted graphically in Figure 3-14 in a Piper diagram.

3.7.3.1.2 Recharge and discharge mechanisms

Areas and modes of recharge and discharge for the aquifers are discussed in this section, based on the ground-water hydrochemistry.

Recharge to the tuff aquifer

The highlands of Pahute Mesa and Gold Flat serve as probable recharge areas for the tuff aquifer in Oasis Valley. White (1979) shows a linear trend for concentrations of bicarbonate and chloride plotted against sodium for ground water in the tuffaceous aquifers of Oasis Valley, Pahute Mesa and Gold Flat (Figures 3-15 and 3-16). Ground water from the latter two areas plots in the dilute portion of the trend while water from Oasis Valley plots along a more concentrated part of the line. The similarity in composition supports the suggestion of Blankennagel and Weir (1973) that water beneath Pahute Mesa is related to Oasis Valley ground water, and merely represents a less advanced stage in the chemical reaction sequence farther upgradient (White 1979).

Other parts of the tuff aquifer may be recharged differently. Winograd and Thordarson (1975), Claassen and White (1979), and White et al. (1980) discuss recharge at Rainier Mesa. Consideration of altitude and vegetation would include Timber Mountain as a potential recharge area also. One further, and perhaps critically important recharge area, may be Fortymile Canyon. This was first suggested by Winograd and Thordarson (1975); however, Claassen (1985) presents chemical and isotopic evidence to support the hypothesis that major amounts of ground water beneath and in the vicinity of the canyon were recharged by infiltration of surface runoff. This infiltration may have penetrated the tuff aquifer as well as the alluvial aquifers. Plans to evaluate the potential recharge through Fortymile Wash are described in Section 8.3.1.2.1.

Discharge from the tuff aquifer

There are three primary mechanisms for ground-water discharge from the tuffaceous aquifer in the vicinity of the NTS: (1) vertical leakage to the underlying lower carbonate aquifer, (2) subsurface flow infiltrating into the valley-fill aquifer, and (3) spring discharge. Hydrochemical evidence for the first mechanism is presented in the section addressing recharge to the lower carbonate aquifer. The second mechanism is supported by White (1979)

Figure 3-14. Piper diagram showing the regional character of the valley fill aquifer.

Figure 3-15. General location of Oasis Valley, Pahute Mesa, and Gold Flat ground-water sampling areas. Modified from White (1978).

CONSULTATION DRAFT

Figure 3-16. Concentration trends of sodium, bicarbonate, and chloride. Modified from White (1979).

where he showed that, through the effects of evapotranspiration, water in the tuffaceous alluvium could be produced from water in the tuff aquifer. White (1979) included data from the Oasis Valley alluvium on his plot of bicarbonate and chloride concentrations relative to sodium (Figure 3-16). Alluvial water compositions for specific flow paths ranged from the dilute region at the initial stages along the flow paths to considerably higher concentrations on the linear trend farther downgradient. White (1979) concluded that based on the linearity of the data, water in the alluvium could evolve from tuffaceous compositions by concentration increases due to evapotranspiration.

The third mechanism, direct spring discharge, is most likely to occur where fracture systems are saturated, broken by recent faulting, or intersect the land surface. White (1979) notes several locations in Oasis Valley where this is evidenced. Although a spring may be issuing directly from the tuff aquifer, Claassen (1985) found that the cation composition of tuffaceous spring water may not necessarily be consistent with that of the regional aquifer. Claassen (1985) explained that the spring discharge is young and generally perched ground water, and thus, has not had the opportunity to evolve to a similar chemical nature as the water from the regional aquifer. Claassen (1985) concludes that although the spring discharge may not be representative of the regional tuffaceous aquifer, it does represent a point in the evolution of the composition of waters in tuffs of southern Nevada.

Recharge to lower carbonate aquifer

The highlands of the Sheep Range, northwestern Spring Mountains, and southern Pahranaagat Range are the primary source of recharge to the lower carbonate aquifer. To a lesser extent the Pintwater, Desert, and Spotted ranges also contribute to the recharge (Winograd and Thordarson, 1975). These are highly fractured Paleozoic carbonate rocks that yield ground water of the calcium-magnesium bicarbonate type. However, there are relatively few areas where the lower carbonate aquifer actually yields the calcium-magnesium water characteristic of the recharge areas. This is attributed to vertical leakage in the eastern and northeastern valleys; that is, sodium rich water from the tuff aquifer leaking vertically downward to mix with the calcium-magnesium water in the lower carbonate aquifer. On the basis that Frenchman Flat and Ash Meadows ground waters have nearly identical compositions, Winograd and Thordarson (1975) conclude that vertical leakage of high sodium water must be occurring in northern Indian Springs, northern Three Lakes, and eastern Emigrant and Desert valleys. There are, however, two areas where the carbonate water has had little opportunity to come in contact with tuff or tuffaceous alluvium. Water from these areas, the Spring Mountains and the Pahranaagat Valley, show the untainted calcium magnesium chemistry that would result from precipitation in the eastern Paleozoic carbonate highlands.

A second source of recharge to the lower carbonate aquifer is downward leakage of water from the Cenozoic strata. Schoff and Moore (1964) concluded that based on the distribution of sodium, the water in the Paleozoic carbonate rocks underlying the NTS is being recharged by downward percolation through the tuff and tuffaceous alluvium. When added to the calcium magnesium water already in the carbonate rock, the sodium potassium water associated with these tuffaceous rocks yields the water of mixed chemical character, which is generally found downgradient in the Amargosa Desert.

CONSULTATION DRAFT

Naff et al. (1974) used a Piper diagram as chemical evidence supporting the idea of downward leakage from the tuff aquifer. Points on the Piper diagram representing downward leakage through the Tertiary aquitard and points representing water moving through a potentiometric trough in the carbonate aquifer just below Yucca and Frenchman flats virtually surround points representing spring water issuing from the carbonate aquifer in the Ash Meadows discharge area (Devils Hole). From this, the authors conclude that the chemical character of the discharge at Ash Meadows can be accounted for by ground water percolating downward across the Tertiary aquitard and mixing with water in the carbonate aquifer. Winograd and Thordarson (1975) used the differences and variations in chemical quality between the carbonate aquifer and the tuff aquitard to estimate that leakage across the aquitard probably constitutes 1 to 5 percent of the discharge at Ash Meadows but may contribute as much as 20 percent.

A third source of recharge to the lower carbonate aquifer, possibly contributing as much as 35 percent (Winograd and Friedman, 1972), is the underflow (i.e., trans-basin regional ground-water movement) into the basin from the northeast. Winograd and Thordarson (1975) found that the chemical data supported the hypothesis that ground water originating in the Pahranaagat Valley moves southwestward through the lower carbonate aquifer into the Ash Meadows ground-water basin. Spring water in the Pahranaagat Valley was found to have about one-third as much sodium and potassium and about one-half as much sulfate and chloride as that at the NTS downward crossflow from the tuff aquitard in the areas of Amargosa Desert, northern Three Lakes and northern Indian Springs valleys, the calcium magnesium water in the lower carbonate aquifer of Pahranaagat Valley could readily be altered to the calcium magnesium sodium type (mixed) found beneath eastern Frenchman Flat.

Discharge from the lower carbonate aquifer

Ash Meadows is the primary discharge area for the lower carbonate aquifer (Section 3.7.2). It is a fault controlled spring line in the southeastern and east-central part of the Amargosa Desert. Chemical and temperature data clearly show that although the major springs emerge from Quaternary deposits, the water feeding the spring pools is derived by upward leakage from the underlying and flanking lower carbonate aquifer (Winograd and Thordarson, 1975). These authors found that specific conductance measurements of water from 11 major springs emerging from the lake beds at Ash Meadows ranged from 640 to 750 micromho per centimeter at 25°C, with 8 of these springs having measurements not greater than 675 micromho per centimeter at 25°C. The specific conductance for springs issuing directly from the carbonate aquifer (Devils Hole and Point of Rocks) ranged from 645 to 686 micromho per centimeter at 25°C. By contrast, water from wells tapping valley fill had measurements ranging from 300 to more than 1,000 micromho per centimeter at 25°C. Winograd and Thordarson (1975) cite as support for the discharge hypothesis the consistently similar specific conductance measurements between water from the spring pools and water from the lower carbonate aquifer.

Temperature measurements also corroborate the idea that spring discharge is supplied by direct upward leakage from the lower carbonate aquifer. Temperatures for water discharging directly from the lower carbonate aquifer range from 33.5 to 34°C. Water from the lake beds is between 27 and 33°C for

yields more than 1,000 gpm ($0.06 \text{ m}^3/\text{s}$), and between 23 and 34.5°C for lower yielding springs. Measurements from deep wells tapping the valley fill indicate that the water in the valley fill aquifer ranges from 19.5 to 28.5°C but is usually less than 24.5°C . No seasonal or long-term variation in water temperature at the major springs is indicated by periodic temperature measurements (Winograd and Thordarson, 1975). As with the specific conductance measurements, the springs issuing directly from the lower carbonate aquifer show similar temperature measurements with those from the lake bed springs.

Areal variations in temperature add further support. Near the carbonate-rock ridges that border the discharge area on the northeast, the buried carbonate aquifer is shallow. Water traveling along a funnel or fault zone to the surface has little time to reach temperature equilibrium with the surrounding cooler Cenozoic sediments. However, further from the ridges the carbonate aquifer is buried to depths of 1,000 ft (300 m) or more. In this area, the longer travel time of water moving to the surface allows greater loss of heat to the cooler surrounding sediments (Winograd and Thordarson, 1975).

A second mechanism for discharge from the lower carbonate aquifer is direct upward crossflow into the valley fill. Winograd and Thordarson (1975) found that the chemical character of water from two wells tapping valley fill in east central Amargosa Desert was dissimilar to that of water from valley fill wells in other parts of the study area, but resembled quite closely that of water issuing from the springs at Ash Meadows. However, principal ionic concentrations in the valley fill wells (median values) were 10 to 20 percent lower than the median concentrations for the springs at Ash Meadows. The authors suggested that the difference in absolute ionic concentrations was due to the dilution effect of ground water derived from local recharge (having lower dissolved solids content) on water that had leaked up into the valley fill from the lower carbonate aquifer.

Further support for this hypothesis is offered by Claassen (1985). Observations of water quality, temperature, and hydraulic potential indicate that in the eastern part of the Amargosa Desert upward leakage from the lower carbonate aquifer mixes with water that has been recharged directly to the valley fill. Claassen (1985) observes that water quality from four sites associated with playa deposits on the eastern side of the Amargosa Desert does not show the drastic differences with surrounding nonplaya water, as does other playa-affected water in the study area. In addition, the water from these four sites is chemically homogeneous over a large area relative to other playa-associated water. These observations lead Claassen (1985) to the idea that an integrating factor, such as source water leaking up from the carbonates, is in effect.

A temperature of 30.6°C was measured in only one of the aforementioned wells. This temperature is closer to temperatures measured in water from wells in the carbonate aquifer (27 to 30°C) than to temperatures measured in water from wells in the distal valley fills (21°C). Furthermore, at this same location the difference in hydraulic potential between the carbonate aquifer and the valley fill is just 6 m, almost the smallest difference observed in the west central Amargosa Desert. Claassen (1985) explains that the smaller potential difference and the presence of more permeable sand and

CONSULTATION DRAFT

gravel in the valley fill at this site supports the idea of significant upward leakage from the carbonate aquifer.

Recharge to the valley fill aquifer

There are three principal sources of recharge to the valley fill aquifer: (1) upward leakage from the tuff aquifer, (2) upward leakage from the lower carbonate aquifer, and (3) overland flow. All but the third have been previously discussed under sections relating to discharge from the tuff and lower carbonate aquifers. Surface runoff is considered the most likely recharge mechanism for tuffaceous alluvium in the west-central Amargosa Desert. Winograd and Thordarson (1975) suggested that the low dissolved solids content of water from six wells along Fortymile Wash reflected recharge primarily via infiltration along the arroyo rather than underflow from Pahute Mesa and Timber Mountain. Claassen (1985) found that sodium-calcium-magnesium concentrations of water in the tuffaceous valley fill in the Amargosa Desert were generally inconsistent with the composition of water in the bedrock aquifers to the north. The data showed that the sodium concentration for valley fill ground water was consistently less than that for bedrock aquifers, sometimes as much as 15 percent. Claassen (1985) concluded that valley fill recharge was primarily surface runoff from the tuffaceous highlands infiltrating along present day wash bottoms.

Discharge from the valley fill aquifer

The most likely mechanism for discharge from saturated valley fill is evapotranspiration. Because of the close proximity of the water table to the ground surface in some areas, Oasis Valley for instance, direct exchange with the atmosphere as well as transpiration occurs (White, 1979). White found that along a valley fill flow path, upgradient samples contained less than half the dissolved solids of the water in the alluvium at the lower end of the flow path (458 mg/L compared with 1,040 mg/L). White (1979) acknowledged that there may be some additional reaction between the ground water and tuffaceous alluvium but concluded that the principal reason for the increase in dissolved solids was a decrease in the volume of ground water due to direct evaporation and transpiration through the vegetative cover.

3.7.3.1.3 Hydrochemical facies

The major hydrochemical facies of the NTS and vicinity were delineated by Winograd and Thordarson (1975). These authors recognized the three types of water described by Schoff and Moore (1964) and suggested two additional facies. The hydrochemical nature of the NTS and vicinity can thus be characterized as follows:

1. A sodium-potassium-bicarbonate facies is present in western Emigrant Valley, Yucca Flat, Frenchman Flat, Jackass Flat, Pahute Mesa, and Oasis Valley. Water of this facies is found in tuff, rhyolite, and valley fill aquifers rich in volcanic detritus. Subsequent to the Winograd and Thordarson (1975) studies, Benson and McKinley (1985) provide data indicating that the water of Yucca Mountain is also of this facies.

2. A calcium-magnesium-bicarbonate facies occurs in the Spring Mountains, southern Indian Springs, southern Three Lakes Valley, and Pahranaagat Valley. In these areas, wells tapping either the lower carbonate aquifer or the valley fill aquifer rich in carbonate detritus draw water of this facies.
3. The calcium-magnesium-sodium-bicarbonate facies is found in the lower carbonate aquifer between Ash Meadows and the eastern NTS. This facies is the result of mixing the first two facies.
4. The playa facies is recognized to be quite variable, depending to some extent on the depth of the sampling well. Occurrences are restricted to playas where ground water is discharged by evapotranspiration (near Death Valley Junction) or to shallow discharge areas.
5. The sodium-sulfate-bicarbonate facies is found in a few wells in the west-central Amargosa Desert.

3.7.3.1.4 Summary

In summary, the chemistry of ground water in the NTS and vicinity is determined primarily by interactions between water and the reactive components of rock and soil zones through which the water has traveled. Water having moved only through tuff strata or through valley fill rich in tuffaceous detritus has a sodium-bicarbonate nature. Ground water that has traveled only through the lower carbonate aquifer or valley fill rich in carbonate detritus is of calcium-magnesium-bicarbonate-character. In areas of downward crossflow from the Cenozoic strata, the lower carbonate aquifer has water of a mixed nature, that is, a calcium-magnesium-sodium-bicarbonate type.

The regional hydrochemical evidence presented contributes significantly to the understanding of the hydrologic system. Recharge to the tuff aquifer in Oasis Valley is principally by underflow and flow from the northern tuffaceous highlands. The lower carbonate aquifer is recharged by precipitation in the Paleozoic carbonate highlands to the northeast. Other sources of recharge to this aquifer are interbasin underflow from the Pahranaagat Valley and downward leakage from the Cenozoic strata. In addition to upward leakage from underlying Cenozoic and Paleozoic strata, the valley fill alluvium receives recharge from surface runoff infiltrating along present day stream channels.

Discharge from the tuff aquifer occurs as (1) vertical leakage to the underlying lower carbonate aquifer, (2) subsurface flow infiltrating into the valley fill, and (3) direct spring discharge. Ash Meadows, a fault-controlled spring line, is the primary discharge area for the lower carbonate aquifer. In other areas of the Amargosa Desert, water from the lower carbonate aquifer leaks upward into the valley fill. Evapotranspiration is the most likely mechanism for valley fill discharge.

CONSULTATION DRAFT

The data presented are sufficient for a semiquantitative description of the hydrochemical environment of the NTS and vicinity, but additional data are necessary for a fully quantitative understanding. Plans for future hydrochemical studies are described in Section 8.3.1.2.3.2.

3.7.3.2 Regional isotope hydrology

This section discusses the data presently available on regional isotope hydrology for the tuff and the carbonate aquifers.

3.7.3.2.1 Tuff aquifer

The environmental radioisotope content of ground-water samples from the tuff aquifers at Yucca Mountain is discussed in Section 3.9.1.3. Additional data are reported by Waddell (1985) for test wells UE-29a#1 and UE-29a#2 and by Claassen (1985) for the Amargosa Desert and parts of the NTS. These data are also summarized in Section 3.9.1.3.

Radiocarbon dates are available from samples of ground water from the tuffaceous aquifer upgradient from Yucca Mountain. The samples are from wells UE-29a#1 and UE-29a#2, in the bottom of Fortymile Canyon. The major ion composition of the water is very similar to that at Yucca Mountain but it has three-fold the radiocarbon content (Table 3-11). In these two wells the radiocarbon content of water ranges from 75.3 percent modern carbon (pmc) units in the shallower abandoned well UE-29a#1 (which was completed to 65.5 m) to 60.0 pmc in the 86.7 to 213.4-m interval of adjacent well UE29a#2 (Waddell, 1985). This stratification is consistent with occasional rapid wash bottom recharge from ephemeral flow (rain storm water) in Fortymile Canyon. Depth to water here is about 25 m and the upper layers of the section are composed of permeable alluvium and brecciated rhyolite. This recharge hypothesis is confirmed by the high tritium content of the shallower water (62 TU), which could only be derived from precipitation bearing bomb-test tritium recharging the aquifer since the mid-1960s. Since the carbon-14 activity is below 100 pmc, appreciable dilution of these recent recharge waters must have occurred, indicating that the recharging water must have had an even higher tritium concentration. Further indirect support of this mechanism is given by the δ deuterium- δ oxygen-18 (δD - $\delta^{18}O$) data. These data plot on the meteoric-water line significantly above the cold, high altitude recharge waters observed under Yucca Mountain (Section 3.9.1.3). The water shows no sign of evaporitic enrichment and is consistent in isotopic composition with warmer, mid-continental intense rainstorm runoff that could cause Fortymile Wash to flow. Rare, high intensity rainfall events therefore probably currently recharge the shallow tuff aquifer at this location. Note that locations for the Amargosa Valley wells listed in Table 3-11 are shown in Figure 3-17.

Claassen (1985) shows that this tendency for the higher carbon-14 content waters in the tuff to occur along washes also prevails down hydraulic gradient from Yucca Mountain. Wells J-12, J-13, Amargosa-9, and Amargosa-18 are good examples. The recharge through wash bottoms need not be recent, as

CONSULTATION DRAFT

Table 3-11. Environmental isotope data for ground-water samples from the samples from the tuff and tuffaceous valley fill aquifers in the region near Yucca Mountain^a

Well designation ^b	Collection date	δD (o/oo SMOW) ^c	$\delta^{18}O$ (o/oo SMOW) ^c	$\delta^{13}C$ (o/oo PDB) ^d	^{14}C (pmc) ^e	HTO ^f (TU) ^f
UE-29a#1 ^g	01/29/82	-92.0	-12.4	-12.6 ^h	75.3	62.0
UE-29a#2 ^{g,i}	01/08/82	-93.5	-13.0	-12.6 ^h	62.3	11.0
UE-29a#2 ^{g,i}	01/15/82	-93.0	-13.1	-13.1 ^h	60.0	11.0
J-12 ^j	03/26/71	-97.5	-12.8	-7.9	32.2	<68.0
J-13 ^j	03/26/71	-97.5	-13.0	-7.3	29.2	<68.0
UE-25c#1	09/30/83	-102	-13.5	-7.1	15.0	<0.3
UE-25c#2	03/13/84	-100	-13.4	-7.0	16.6	<0.6
UE-25c#3	05/09/84	-103	-13.5	-7.5	15.7	0.6
USW H-6	10/16/82	-106	-13.8	-7.5	16.3	<3.0
USW H-6	06/20/84	-105	-14.0	-7.3	10.0	1.2
USW H-6	07/06/84	-107	-14.0	-7.1	12.4	0.3
USW VH1 ^j	02/11/81	-108	-14.2	-8.5	12.2	6.0
AM-4	03/04/74	-103	-13.2	-7.1 ^k	19.3	--
AM-9	03/01/74	-102	-12.6	--	28.4	--
AM-11	03/05/74	-101	-13.1	--	20.8	--
AM-13	03/05/74	-102	-13.0	--	19.3	--
AM-15	03/05/74	-104	-13.0	--	18.4	--
AM-18 ^l	03/06/74	-102	-13.0	--	27.8	--
AM-20 ^l	03/06/74	-102	-12.4	--	13.8	--
AM-21	06/25/74	-99	-13.2	-8.4	27.4	--
AM-23	03/31/71	-103	-13.4	-7.1	17.1	--
AM-25	03/31/71	-102	-13.4	-5.6	15.6	--
NTS#8 ^m	03/24/71	-104	-13.0	-12.1	25.4	--

^aSources: Data from Benson and McKinley (1985) and as noted in footnotes. AM-samples from Claassen (1985).

^bWell locations for UE-, J- and USW-holes are shown in Figure 3-22. Well locations for AM (Amargosa) wells are shown in Figure 3-16.

^c δ deuterium and δ oxygen-18 are reported in parts per thousand relative to the standard mean ocean water (SMOW) standard.

^d δ carbon-13 is reported in parts per thousand relative to the Pee Dee belemnite (PDB) carbonate standard.

^eCarbon-14 activity is reported as a percent of the modern carbon (pmc) standard.

^fHTO = tritium; reported in tritium units (TU).

^gWaddell (1985).

^hValue reported as positive in reference.

ⁱAlso reported by Benson and McKinley (1985).

^jAlso reported by Claassen (1985).

^k-- indicates no data.

^lCompletion in tuffaceous material unsure.

^mClaassen (1985).

Figure 3-17. Location of Amargosa Valley (AM) ground-water sample wells.

it has been in well UE-29a#2. Claassen (1985) demonstrates that low bicarbonate waters in the tuff having δ carbon-13 ($\delta^{13}\text{C}$) values near atmospheric (-7 parts per thousand with respect to the Peedee belemnite carbonate standard (PDB) have probably had minimal isotopic dilution of their original carbon-14 by carbon-12 from carbonate minerals and have not exchanged with biogenic carbon in the soil root zone. Although generally an upper limit, "uncorrected, apparent radio-carbon ages" (i.e., dates obtained by assuming that the initial radiocarbon content of the recharge water was equal to that of the atmosphere (approximately 100 pmc), without correcting for exchange with soil carbon dioxide or carbonate minerals) may be a realistic estimate, in this case, of the true ground-water residence times. In this case, the drainageways also served for paleorecharge. The waters in these locations are no more than 10,000 yr old while the water away from the drainage have apparent ages more on the order of 15,000 yr before present (B.P.).

Figure 3-18 is a $\delta\text{D}-\delta^{18}\text{O}$ diagram of the analyses on water samples from the tuff aquifers (in Tables 3-11 and 3-12). The radiocarbon content is plotted (in pmc) at each $\delta\text{D}-\delta^{18}\text{O}$ coordinate. The data from Yucca Mountain are circled. The younger, wash-bottom recharge samples fall higher on the plot than does the older, high altitude, cold continental recharge in the same formations under Yucca Mountain (Section 3.9.1.3). This is consistent with rainstorm runoff, rather than snow melt recharge. The low $\delta^{13}\text{C}$ values in both instances suggest that the recharge has always occurred in poorly vegetated areas.

The combined isotopic data is consistent with an arid (poorly vegetated in the recharge zone), continental system recharging through infiltration of surface flow along permeable surface drainages. The recharge water about 15,000 yr ago was typical of cold, high altitude precipitation (or snow melt) and has evolved toward water suggesting warmer rapid recharge from intense storms. This model is consistent with the paleoclimatic interpretation of the area (Sections 3.7.4 and 5.2.1).

Note that even in the most down-gradient (southern) portions of the tuff aquifer system, carbon-14 contents tend not to be less than about 12 pmc (about 17,000 yr B.P. apparent radiocarbon age). Several hypotheses have been advanced concerning the absence of older water (of middle Wisconsin age). Based on ground-water velocity boundary conditions and on paleoclimatic data for the site, Claassen (1985) has proposed that precipitation in the area in the middle Wisconsin may have been inadequate to contribute significant recharge to the tuff aquifer system. Major recharge of the system therefore occurred in the late Pleistocene through the early Holocene. Some minor recharge must still occur today.

The collection of additional ground-water samples, as well as rainfall and snow samples, will contribute significantly to an improved interpretation of this system. Plans for further sample collection are presented in Sections 8.3.1.2.1 and 8.3.1.12.

Figure 3-18. $\delta D - \delta^{18}O$ plot of waters from Yucca Mountain area, with radiocarbon content (in percent modern carbon (pmc) units) shown at the corresponding $\delta D - \delta^{18}O$ coordinate. Values in parenthesis are from the lower carbonate aquifer. Underlined values are from mixed carbonate/tuff origin. The remaining values are from the tuff formations, with the Yucca Mountain samples circled.

CONSULTATION DRAFT

Table 3-12. Environmental isotope data for ground-water samples from the carbonate aquifer and from mixed carbonate-tuff sources^a

Well or spring designation	Collection date	δD o/oo SMOW ^b	$\delta^{18}O$ o/oo SMOW ^b	$\delta^{13}C$ o/oo PDB ^c	^{14}C (pmc) ^d	HTO (TU) ^e
CARBONATE AQUIFER						
UE-25p#1 ^{f,g}	02/09/83	-106	-13.5	-4.2	3.5	<3.1
UE-25p#1 ^{f,h}	05/12/86	-106	-13.8	-2.3	2.3	3.1
16S/51E-23 ⁱ	-- ^j	--	-13.6	-4.6	1.8	--
Fairbanks Spring NE ^k	12/13/74	--	--	-5.2	2.2 ± 0.3	0.1 ± 0.2
Fairbanks Spring SW ^k	--	-103	-13.6	-4.9	1.8 ± 0.2	0.0 ± 0.4
Rogers Spring ^k	--	-102	--	-4.6	1.5 ± 0.3	0.0 ± 0.2
Longstreet Spring ^k	--	-103	--	-4.8	2.7 ± 0.4	0.0 ± 0.2
Scruggs Spring ^k	03/10/75 ^l	-103	--	-4.7	1.1 ± 0.3	0.2 ± 0.2
Crystal Pool ^{k,m}	--	-102	-13.7	-5.0	11.6 ± 0.7	0.6 ± 0.4
Devils Hole ^k	--	--	-13.6	-5.0	2.8 ± 0.4	0.3 ± 0.2
King Spring ^{k,m}	--	-104	--	-4.7	1.7 ± 0.4	0.4 ± 0.2
Big Spring ^{k,m}	03/08/75	-102	--	-4.6	2.9 ± 0.4	0.3 ± 0.1
MIXED SYSTEMS						
AM-3 ⁿ	10/20/72	-102	-12.8	--	15.6	--
AM-5 ⁿ	11/17/72	-99.5	-13.2	-6.8	12.4	--
AM-8 ⁿ	03/01/74	-103	-13.4	-7.3	21.9	--
AM-10 ⁿ	06/26/79	-97.5	-13.2	-5.2	24.8	--
AM-16 ⁿ	03/01/74	-104	-12.7	--	10.0	--
AM-17 ⁿ	03/01/74	-105	-12.8	--	18.9	--
AM-19 ⁿ	03/06/74	--	--	--	40.3	--
AM-27 ^o	08/18/82	-105	-13.8	-3.6	7.0	--
AM-29 ^p	03/31/71	-105	-13.8	-3.4	--	--
AM-30 ^o	06/24/79	-104	-13.7	-4.4	10.3	--
AM-47 ^o	03/31/71	-102	-13.1	-6.2	31.4	--
AM-50 ^o	06/25/79	-104	-13.6	-5.7	--	--
AM-60 ^p	12/16/68	--	--	-5.9	28.8	--

^aData from Claassen (1985) unless otherwise indicated.^b δD and δ oxygen-18 ($\delta^{18}O$) are reported in parts per thousand relative to Standard Mean Ocean Water (SMOW) standard.^cCarbon-13 ($\delta^{13}C$) is reported in parts per thousand relative to the Pee Dee belemnite (PDB) carbonate standard.^dCarbon-14 (^{14}C) activity is reported as a percent of the modern carbon standard (pmc).^eHTO = tritium; T is reported in tritium units (TU).^fBenson and McKinley (1985).^gSampled from the 381 to 1,197 m interval.^hSampled from 1,297 to 1,805 m interval.ⁱClaassen (1985) reports this well as "Amargosa Tracer Well #2."^jData from multiple collection dates.^kWinograd and Pearson (1976).^lRepresentative of multiple samplings.^mSimilar values reported by Riggs (1984).ⁿTuffaceous lithology, possible carbonate influence.^oCarbonate lithology, possible tuff influence.^pIn both tuff and carbonate formations.

3.7.3.2.2 Carbonate aquifer

Environmental isotope data for ground-water samples from the carbonate aquifer and from mixed carbonate-tuff sources are listed in Table 3-12. Benson and McKinley (1985) have reported isotopic data for ground waters from the carbonate aquifer in the area around Yucca Mountain (drillhole UE-25p#1). Claassen (1985) reports two carbonate aquifer analyses and several analyses on mixed carbonate-tuff systems from the Amargosa Desert. Winograd and Pearson (1976) and Riggs (1984) report analyses of spring water from Ash Meadows, which drains primarily the area to the east of Yucca Mountain and the Amargosa Desert.

All samples from the carbonate aquifer show low carbon-14 contents (<4 pmc; except for Crystal Pool, which is discussed in the following paragraphs), as would be expected at the distal end of a regional carbonate aquifer system. The use of "apparent, uncorrected radiocarbon ages," which may be justified in the tuff aquifer is unsupportable in the carbonate system. The modeling of the carbon-14 content of the recharge waters is extremely complicated in a semiconfined carbonate aquifer system. The carbon-13 analyses show that interaction has occurred between the relatively carbon-14-free carbonate minerals of the aquifer and the flowing ground water. Small adjustments of the isotopic balance to account for this dilution process can greatly affect the resulting radiocarbon dates in such systems (Muller and Mayo, 1986). Without further modeling, these waters can only be taken to be "old," with 30,000 yr B.P. as the conservative upper limit. A very simple example can show why this is an upper limit. A pure carbonate water, one which derived half its carbon from active CO_2 gas and half from dead carbonate minerals, is one half-life younger than its apparent age. Thus, a ground water with an apparent age of 30,000 yr B.P. is probably around 24,000 yr old, if no isotopic dilution occurs as a result of precipitation/dissolution (i.e., isotopic exchange with the aquifer matrix).

The carbon-14 content of ground water at the center of a 16-km-long fault-controlled spring line at Ash Meadows is 5 times greater than that in water from other major springs along the lineament. Winograd and Pearson (1976) have examined the radiocarbon anomaly of Crystal Pool. They have shown it to probably be caused by megascale channeling, with water moving to this discharge point at velocities appreciably greater than those to adjacent springs. Such channeling is consistent with the known flow regime of karstic aquifers.

The data from Claassen (1985) for wells penetrating both tuff and carbonate lithologies show a variability in isotopic compositions expected in mixed systems (Table 3-12). Several of these wells are shallow, with waters also probably derived from valley fill sediments. There are insufficient data available to assess the extent and direction of mixing.

Few deuterium and oxygen-18 data are available for the carbonate aquifer (Figure 3-18 and Table 3-12). The waters show no hydrothermal fractionation and do not resemble expected modern precipitation. Additional stable isotopic data from the carbonate aquifer in the vicinity of Yucca Mountain is expected to provide further information on the interconnection of the tuff and carbonate aquifers. Plans for further data collection are presented in Section 8.3.1.2.

3.7.4 PALEOHYDROLOGY

The purpose of paleohydrologic studies is to determine hydrologic conditions during the Quaternary Period that have differed significantly from present conditions. This information will be used to evaluate the likelihood of episodic conditions recurring that may affect the regional ground-water-flow system over the next 100,000 yr. Of specific interest are (1) the maximum altitude of the water table during pluvial periods of the Quaternary Period, (2) the effects of pluvial water-table rises on shortening of ground-water flow paths to discharge areas, and (3) the magnitude of increases in recharge during pluvial periods. With such information, questions such as the following can be addressed: What is the possibility of the repository being flooded by a rising water table during a return of pluvial conditions? How large an infiltration flux might move through the repository in the future?

Evidence for former, higher water tables, changes in length of ground-water flow paths, and the presence and absence of pluvial lakes in the south-central Great Basin are discussed in the following sections. Although the period of major concern is the past 100,000 yr, the discussion will include paleohydrologic information extending into late Pliocene time.

3.7.4.1 Lowering of ground-water levels during the Quaternary Period

There is geologic evidence for former high levels of the water table in the south-central Great Basin (Figure 3-19) during the Quaternary Period. The evidence consists of tufas, spring orifices, calcitic veins, and cylindrical-lined tubes that mark the routes of former ground-water flow to springs; former water levels marked by calcitic cave deposits on the walls of Devils Hole (a fault-controlled collapse feature at Ash Meadows); and widespread marsh deposits (composed of chalk and authigenic clay) precipitated from former ground waters. Most of these features have been described in reports by Winograd and Thordarson (1975), Dudley and Larson (1976), Winograd and Doty (1980), Khoury et al. (1982), Pexton (1984), Winograd et al. (1985), and Hay et al. (1986). The paleohydrologic significance of these deposits is the subject of papers by Winograd and Doty (1980), Pexton (1984), and Winograd and Szabo (1986).

Winograd and Doty (1980) mapped tufa and vein calcite in the Ash Meadows area of the Amargosa Desert, south of Highway 95 and east of Highway 29. They concluded that the potentiometric surface in the lower carbonate aquifer was once apparently as much as 50 m higher than the highest modern water levels in the Ash Meadows discharge area namely that in Devils Hole (altitude 719 m). Uranium disequilibrium dating of several of these calcitic veins (AM-7, AM-10, and DH-1 in Figure 3-19) show that the higher stands of water levels occurred during Pleistocene time, 1,000,000 to 500,000 yr ago (Pexton, 1984; Winograd et al., 1985; and Winograd and Szabo, 1986).

Still higher and older ground-water levels are recorded throughout the central Amargosa Desert by widespread marsh deposits of ground-water origin. These deposits are described by Khoury et al. (1982) and Pexton (1984). The marsh deposits have been referred to as lacustrine or playa deposits (Walker

Figure 3-19. Map of South-Central Great Basin showing location of uranium-series dated veins. Modified from Winograd and Szabo (1986).

and Eakin, 1963; Denny and Drewes, 1965; and Swadley, 1983). Hay et al. (1986) present evidence based on stratigraphy, mineral composition, fossil assemblages, and lithofacies relationships that these sediments were deposited in marshes or ponds fed by ground water rather than in lake or playa environments. In this report, they will be referred to as marsh deposits. The marsh deposits occur at altitudes as high as 790 m (Winograd and Doty, 1980) in Amargosa Flat in the east-central portion of the Amargosa Desert. This is about 70 m above the highest modern water level in Devils Hole (altitude 719 m).

A sample from the south end of Crater Flat described as nodular tufa spring deposits has been dated by the uranium-series method to approximately 30,000 yr B.P. (Szabo et al., 1981). Another sample described as seep-deposited tufa or calcrete from the south of Yucca Mountain was dated to $78,000 \pm 5,000$ yr B.P. (Szabo et al., 1981). These dates may not be correct, however, because this dating method is only considered accurate if the material is from a closed chemical system (Swadley et al., 1984; Bradley, 1985). Szabo et al. (1981) state that these types of deposits may form open systems. Studies to reexamine and redate the deposits are planned and discussed in Section 8.3.1.2. In the Ash Meadows area, these marsh deposits range in age from about 2 to 3 million years (Pexton, 1984) and presumably mark the altitude of regional ground-water discharge during late Pliocene and perhaps earliest Pleistocene time.

The occurrence of the various deposits, marking Quaternary water tables, at altitudes tens to over 100 m above the modern water table probably resulted from a combination of tectonism, water-table lowering due to increasing aridity since Pliocene time, and erosion. Winograd and Szabo (1986) believe that the altitude of the water-table-related deposits at Furnace Creek Wash (Figure 3-19) principally reflect uplift of the terrane, rather than an actual lowering of the water table. However, they infer that regionally the water table in the lower carbonate aquifer is likely to have lowered during the Quaternary Period. The following evidence is cited in support of their hypothesis:

1. The several-thousand-meter topographic relief in Death Valley dates largely from Pliocene and Pleistocene time (Hunt and Mabey, 1966; USGS, 1984), and the movement of the floor of Death Valley probably has been downward relative to sea level and to bordering areas (Hunt and Mabey, 1966).

2. Interbasin flow of ground water through Paleozoic carbonate rock and Tertiary welded-tuff aquifers toward Ash Meadows and Death Valley occurs today (Winograd and Thordarson, 1975). Interbasin flow of ground water toward Death Valley likely also occurred during the Quaternary Period in response to the lowering of ground-water discharge outlets; as a result, Death Valley gradually evolved during the Quaternary into a regional sump for ground water.

3. An increase in aridity due to uplift of the Sierra Nevada and Transverse Ranges would have resulted in a reduction of ground-water recharge.

Another climate-related control on the decline of the regional water table may have been the level of Lake Manly, a Pleistocene lake that filled

Death Valley to about 120 m above sea level (Hunt and Mabey, 1966). With the drying of the lake, ground-water base level in Death Valley would have declined. The exact age of Lake Manly and its duration are not well known. Hunt and Mabey (1966) considered the lake to be no older than the Wisconsin. Hale (1985) suggests that it is much older, perhaps of early- to mid-Pleistocene age. That the lake is either very old, had a short duration, or both, is suggested by the paucity of geomorphic and sedimentologic evidence for its existence (Hunt and Mabey, 1966).

Winograd and Szabo (1986) also suggest that the regional water-table decline postulated for the lower carbonate aquifer during the Quaternary Period was accompanied by a water-level decline in the overlying Cenozoic welded-tuff and valley-fill aquifers. They note that "The suggested progressive lowering of the regional water table throughout the Quaternary does not preclude superimposed and relatively rapid cyclical fluctuations in water level in response to the glacial (i.e., pluvial) and interglacial climates of the Pleistocene." Such fluctuations are discussed in Section 3.7.4.3.

Evidence for increased aridity in the south-central Great Basin is discussed here (also in Sections 5.2.1 to 5.2.3) because of its importance to the above-cited inference of water-table lowering throughout the Quaternary Period. Huber (1981) demonstrated that the central Sierra Nevada has been rising since early Miocene time (23.7 million years ago) and that the rate of uplift has been increasing from approximately 0.03 m/1,000 yr during early Miocene to 0.35 m/1,000 yr at present. In the past 3 million years the Sierra Nevada has risen about 1,000 m. Because the major source of cool-season (November to May) precipitation for the Great Basin is the Pacific Ocean, this major uplift is likely to have significantly reduced precipitation in the region. Warm-season precipitation, which is derived largely from the Gulf of California and the Gulf of Mexico, is not considered important because this moisture is unlikely to contribute much to ground-water recharge (Winograd and Riggs, 1984). G.I. Smith et al. (1983) suggest that when the Sierra Nevada was about 1,000 m lower (3 million years ago), about 50 percent more moisture might have moved into the Great Basin. Based on paleobotanical records, Raven and Axelrod (1978) and Axelrod (1979) believe that aridity increased in the Great Basin, Mojave Desert, and Sonoran Desert during the late Tertiary and the Quaternary periods. They attributed this to uplift of the Sierra Nevada, Transverse Ranges, Peninsular Ranges, and the Mexican Plateau. On the basis of sediment depositional environments, Pexton (1984) believes that the Ash Meadows-Amargosa Desert area became more arid during the Quaternary Period. Winograd et al. (1985) describe a major and progressive depletion in the deuterium content of ground-water recharge in the Spring Mountains (Figure 3-19) during the Quaternary Period, and the most likely explanation for this is a decrease in Pacific moisture due to uplift of the Sierra Nevada and Transverse Ranges.

Lowering of the water table at Ash Meadows might have occurred during the Quaternary Period, even in the absence of vertical tectonics or increased aridity. Winograd and Szabo (1986) cite Winograd and Doty (1980), noting that "the major springs at the Ash Meadows oasis (Figure 3-19), differ in altitude by as much as 35 m and are as much as 50 m lower than the water level in Devils Hole. Thus periodic initiation of discharge from new spring orifices (or an increase in existing discharge) in the lower portions of this oasis due to faulting and extensional fracturing would have resulted in new

and lower base-levels for ground-water discharge." Implicit in this hypothesis is the belief that the faulting would be extensional opening new (or widening old) avenues of discharge from the buried Paleozoic carbonate rock aquifer that underlies eastern Ash Meadows and that feeds all the modern springs (Winograd and Thordarson, 1975). In support of this hypothesis, Winograd and Szabo (1986) note that most of the calcitic veins in Pliocene and younger rocks at Ash Meadows strike $N.40^{\circ} \pm 10^{\circ} E$. This is nearly at right angles to Carr's (1974) estimate of the direction of active extension in the region, namely $N.50^{\circ} W$. to $S.50^{\circ} E$. This mechanism may also have periodically lowered the water table in east-central Death Valley where the difference in altitude between the highest (Nevares) and lowest (Texas) major springs discharging from the regional carbonate aquifer is about 170 m (Winograd and Thordarson, 1975).

In summary, available data from calcitic veins and marsh deposits indicate a 50 to 70 m apparent lowering of the water table in the east-central Amargosa Desert since the middle Pleistocene and an apparent lowering of perhaps as much as 130 m since the end of the Pliocene. A strong inference can be made that the water-table altitude in the south-central Great Basin actually declined regionally throughout the Quaternary due to increasing aridity coupled with an absolute lowering of ground-water base level in Death Valley. This postulated general lowering does not preclude superimposed water-table rises in response to pluvial climates of the Pleistocene.

3.7.4.2 Ground-water flow paths during the Quaternary Period

The distribution of calcitic veins, tufas, and marsh deposits also suggest a lengthening of ground-water flow paths during Quaternary time. Regardless of whether the offset in the paleo-water table marked by these deposits was caused by water-level decline or land-surface rise (Section 3.7.4.1), the occurrence of these deposits kilometers to tens of kilometers upgradient from areas of modern ground-water discharge indicates that flow to points of ground-water discharge were shorter in the past. Variations in flow-path length for the Ash Meadows area in response to different water-table levels are shown diagrammatically in Figure 3-20. Ground water formerly discharged from the lower carbonate aquifer about 14 km northeast of Ash Meadows, the present discharge area. Based on the age of a calcite vein (AM-10 in Figure 3-19), this discharge occurred about 750,000 yr ago (Winograd and Szabo, 1986).

The widespread marsh deposits of the Amargosa Desert also show the location as well as the altitude of former ground-water discharge. In east-central Amargosa Desert (east of Highway 29 and south of Highway 95), these deposits were precipitated from ground water discharging from the lower carbonate aquifer, either directly or indirectly via Tertiary aquitards. Like the calcitic veins, they occur as much as 14 km upgradient from Ash Meadows. West of Highway 29 and both north and south of Highway 95 the marsh deposits may reflect not only former discharge from the lower carbonate aquifer (which crops out in the west-central portion of the valley) but also from the Cenozoic welded-tuff and valley-fill aquifers. The marsh deposits in north-central Amargosa Desert are as much as 70 km from Alkali Flat, the modern discharge area in the southern Amargosa Desert.

Figure 3-20. Diagrammatic section illustrating effects of possible past or future pluvial-related water-table rise on length of ground-water flow path from Frenchman Flat to points of natural discharge at Ash Meadows in the Amargosa Desert. Water-level rises of 40 and 150 m would initiate discharge from lower carbonate aquifer at points A and B, respectively 14 and 21 km northeast of modern spring lineament; arrows depict ground-water flow. Modified from Winograd and Doty (1980).

3.7.4.3 Late Wisconsin lakes, marsh deposits, and ground-water levels

During the late Wisconsin, 21,000 to 10,000 yr ago, average annual precipitation is believed to have been as much as 100 percent greater than modern, with average annual temperature as much as 6°C cooler (Spaulding et al., 1984). Not unexpectedly, changes in climate from late Wisconsin to present conditions have (Section 5.2) produced major changes in the hydrology of the region.

Mifflin and Wheat (1979) discuss pluvial lakes of late Wisconsin age in Nevada. In southern Nevada, they examined Frenchman Flat, Yucca Flat, Gold Flat, Kawich Valley, Sarcobatus Flat, Stonewall Flat, Emigrant (Groom Lake) Valley, Papoose Lake Valley, Pahrump Valley, and Indian Springs Valley, which generally surround the Yucca Mountain area. Of these valleys, they reported shoreline development only in Gold Flat, Kawich Valley, and Emigrant Valley. They report marsh deposits in Indian Springs, Pahrump, and Sarcobatus valleys. Mifflin and Wheat (1979) conclude that "southern Nevada did not contain perennial lakes during the Wisconsinan pluvial," but that "there were likely a number of areas of marsh environments in southern Nevada due to concentrated ground water and spring discharge considerably in excess of present discharge. In south-central and southern Nevada significant differences in the amount and location of ground-water discharge seem apparent."

Hooke (1972) believed that a 90-m deep lake existed in Death Valley about 11,000 yr B.P. Smith and Street-Perrot (1983) agree that there was a lake between 12,000 and 21,000 yr B.P. but conclude that any lake or lakes were small and much shallower.

Quade (1986) reports marsh deposits in southern Indian Springs Valley. These deposits are assumed to be of late Wisconsin age, based on their stratigraphic and topographic similarity to dated deposits in northwestern Las Vegas Valley. They lie at a maximum altitude of 1,037 m, 20 to 50 m higher than the present water table in the valley-fill aquifer. If these deposits are of ground-water origin, as Quade believes, they may record a higher water table during late Wisconsin time. The marsh deposits occur above an inferred major hydraulic barrier (Winograd and Thordarson, 1975), and are 15 to 20 km downgradient from the Spring Mountains, a major recharge area. These factors may bear on the apparent water-table rise.

Mifflin and Wheat (1979) report marsh deposits, presumably of late Wisconsin age in Sarcobatus Flat and Pahrump Valley. They do not provide information on the location of these deposits so that we cannot compare their maximum altitudes with that of the preirrigation water table in these valleys.

Preliminary uranium-disequilibrium dating of calcitic cave deposits in Devils Hole, in central Ash Meadows suggests a 10-m rise in water level in the lower carbonate aquifer about 30,000 yr ago at Ash Meadows (Winograd and Szabo, 1986).

3.7.4.4 Estimation of future ground-water levels

Winograd and Doty (1980) using a variety of boundary conditions, suggest that during future pluvials, water levels in the regional carbonate aquifer at Ash Meadows and areas upgradient might be a few tens of meters higher than, or perhaps even lower than, modern levels. See Section 3.9.8 for estimation of possible future water levels in the Cenozoic volcanic rocks beneath Yucca Mountain and vicinity.

Winograd and Szabo (1986) concluded that there was a progressive lowering of the water table in the south-central Great Basin during the Quaternary. They noted that "a continued decline of the regional water table in the next 100,000 yr (and beyond?)" was likely (Section 3.9.8).

Craig (1984) makes a comprehensive statistical-climatologic analysis of climate conditions affecting the Death Valley drainage systems that existed during the last glacial maximum. In one analysis, Craig modeled full glacial conditions without assuming a major decrease in lake evaporation rates. In this instance, precipitation increased by 90 percent and pluvial lakes formed in some valleys where lakes are known to have existed in the Late Pleistocene, but no lake formed in Death Valley and the lakes that resulted were not of a size "substantial enough to directly influence the regional groundwater potentiometric surface" (Craig, 1984). In an alternate analysis, Craig (1984) assumes a major reduction in lake evaporation rates. He states that, "If lake evaporation rates decrease by 27 percent, a major lake will form in Death Valley. This lake will be of an extent comparable to the highest recorded stand of Lake Manly during the Wisconsin." Under these climatic conditions, he states "recharge rates, and so ground-water flow conditions, will change significantly during a glacial event". Craig's (1984) basis for considering a 27 percent decrease in rate of evaporation from pluvial lakes was paleoclimatic evidence that evaporation was significantly lower during the Late Pleistocene. As validation for this assumption, he notes that the model predicted "a lake in Death Valley as large as is documented in the geologic record of the late Pleistocene." Craig's (1984) model predicts that such climatic conditions would have sustained lakes in Timber Mountain caldera, in Frenchman Flat, at Ash Meadows, and elsewhere. If Lake Manly and other major lakes existed in Wisconsin time, regional ground-water levels would indeed have been different. However, Lake Manly may be of early-to-middle Pleistocene age (Section 3.7.4.1), when the Sierra Nevada was several hundred meters lower and the Transverse Ranges were not a topographic barrier to inflow of Pacific storms. Thus, the age of Lake Manly has bearing on the validity of Craig's (1984) alternate analysis for reconstruction of late Pleistocene paleoclimatology of the Death Valley area, and by extension for prediction of the effects of future pluvials on ground-water levels. Hunt and Mabey (1966) also note that: "The slight erosion and sedimentation record of Lake Manly may mean that the lake was of brief duration and its level may have fluctuated rapidly. Whatever the cause, this California lake left one of the least distinct and most incomplete records of any Pleistocene lake in the Great Basin." Thus, even if Lake Manly is of Wisconsin age, the climatic changes it portends for the ground-water regime may have been short lived.

The paucity of geomorphic and sedimentologic evidence for Lake Manly makes it difficult to date the age of this major ancient lake in Death

Valley. Resolving whether this lake is closer to 100,000 or 1 million years old would be of value to the paleohydrologic and paleoclimatologic analyses of Czarnecki (1985) and Craig (1984). Application of new or established methods, such as dating of desert varnish (Dorn, 1983) and uranium-series disequilibrium dating of lake-water cemented gravels, may hold some promise (Section 8.3.1.2.1). The reconnaissance (air photo) observations of Mifflin and Wheat (1979), regarding the absence of shorelines of late Pleistocene age in Yucca and Frenchman Flats and the presence of such shorelines in Gold Flat, Kawich Valley, and Emigrant Valley will be confirmed by more detailed study (Section 8.3.1.2.1).

3.7.4.5 Ground-water recharge during late Wisconsin time: carbon-14 evidence

Apparent ground-water ages determined from carbon-14 data (Benson and McKinley, 1985; Claassen, 1985) indicate that much of the ground water within the tuffaceous rocks in the vicinity of Yucca Mountain and within the valley fill aquifer beneath the central Amargosa Desert was recharged during late Wisconsin time, about 10,000 to 18,000 yr ago. Holocene-age (less than 10,000 yr) water occurs in Fortymile Canyon at well UE-29a#2 (4,000 yr B.P.) and at well J-12 (9,100 yr B.P.).

According to Claassen (1985), "little or no radiocarbon dilution probably occurred during the evolution of water in tuff or tuffaceous valley fill; therefore, the unadjusted ages are taken as true ages." He concluded that for ground waters that contain as much as 20 percent of carbonate-derived constituents unadjusted ages could be as much as 1,000 yr too old, and that for a higher percentage there could be a serious problem in determining true ages from carbon-14 data. Since expected ground-water flow paths of waters sampled from Cenozoic aquifers in the vicinity of Yucca Mountain involve little or no contact with carbonate rocks, carbon-14 ages probably are near true ages, provided the well sampled did not tap water of more than one age.

Carbon-14 data from well USW H-6 (Benson and McKinley, 1985) show an increase in apparent ground-water age with depth in tuffaceous rocks. The interval 526 to 1,220 m (the entire saturated thickness penetrated) yielded water with an age of 14,600 yr; the interval 608 to 646 m yielded water 16,800 yr old; and the interval 753 to 853 m yielded water 18,500 yr old. A flow-meter survey (Benson et al., 1983) shows significant water yield between the water table (526 m) and a depth of 790 m. Because water pumped from the entire saturated zone (526 to 1,220 m) included older water from the two deeper zones, it is probable that water in the uppermost part of the saturated zone is considerably younger than 14,600 yr.

Ground water with apparent age dates of 10,000 to 17,000 yr B.P. occurs in the valley-fill aquifer of the central Amargosa Desert (Claassen, 1985), 25- to 40-km downgradient from Yucca Mountain. Claassen (1985) believes, as did Winograd and Thordarson (1975), that ground water in the valley fill represents pluvial recharge resulting from infiltration of surface-water runoff along the distributaries of Fortymile Wash. Therefore, ground waters of about the same age as those beneath the upper zone of saturation near Yucca Mountain could have been emplaced 25 to 40 km downgradient. Plans to evaluate recharge along Fortymile Wash are discussed in Section 8.3.1.2.1.

Section 8.3.1 also discusses plans to evaluate the apparent ground-water ages presented by Benson and McKinley (1985).

Most of the wells sampled by Claassen (1985) that tap tuffaceous valley fill yielded water with apparent age less than 15,000 yr. Speculating on the absence of older water in the central Amargosa Desert, he considered that any water older than 17,000 yr B.P. might have moved downgradient, beyond his study area. Claassen (1985) favors an explanation based on climatic studies, concluding that "snowfall earlier than about 20,000 years B.P. was insufficient to result in snowmelt recharge, while subsequent climatic conditions caused such recharge."

Deducing ground-water velocities from the carbon-14 data is difficult. Claassen (1985) estimates a maximum velocity of 7 m/yr for the region between the head of Fortymile Canyon and the central Amargosa Desert. Claassen (1985) believes that actual velocities are probably much slower, but "because recharge may occur anywhere but not necessarily everywhere along surface drainageways, no probable minimum velocity can be calculated." Because recharge probably occurred by widespread infiltration along distributaries that resulted in layering of waters of differing ages, the estimation of ground-water velocities within the aquifer would be difficult even if test holes were available to selectively sample for carbon-14 ages of water with depth. Plans to drill test holes are discussed in Section 8.3.1.2.3. If recharge of the same-age water can occur simultaneously along long reaches of Fortymile Wash and its distributaries in the Amargosa Desert, map distances may be irrelevant for calculating ground-water velocity. However, under such recharge conditions, it might be possible to draw conclusions regarding paleo versus modern recharge rates and the effects of ground-water velocity on the flushing out of old recharge water (Section 8.3.1.2.3).

3.7.4.6 Conclusions

In summary, the following conclusions can be drawn:

1. Evidence from dated calcitic veins and marsh deposits indicates that the regional water table (more correctly the potentiometric surface) in the south-central Great Basin has apparently declined 50 to perhaps 130 m during the Quaternary Period and that this decline has been accompanied by a lengthening of ground-water flow paths to points of discharge by 14 to perhaps 60 to 70 km. Considerations of the neotectonics of Death Valley, regional interbasin flow through the lower carbonate aquifer, and of the probable increasing aridity in the region during the Quaternary, collectively suggest that an actual and progressive water-table decline occurred in the region during this period.
2. Superimposed on the indicated long-term decline of water table are rises reflecting pluvial climates. In Devils Hole, at Ash Meadows, preliminary uranium-series dating of calcitic cave deposits indicates that about 30,000 yr ago the water table in the lower carbonate aquifer stood about 10 m above modern level. In southern Indian Springs Valley, marsh deposits, of presumed ground-water

origin, occur as much as 20 to 50 m above the modern water table in the valley-fill aquifer; these deposits probably are late(?) Wisconsin in age.

3. Reconnaissance observations indicate that pluvial lakes of late Pleistocene age were not present in Yucca Flat or Frenchman Flat, though such lakes apparently existed in Gold Flat, Kawich Valley, and Emigrant Valley, 70 km north and northeast of Yucca Mountain.
4. Ground water of late Wisconsin age is present in the uppermost part of saturated volcanic rocks beneath Yucca Mountain and in the valley fill aquifer in the Amargosa Desert. The valley fill aquifer was recharged by surface-water runoff along the distributaries of Fortymile Wash. A quantitative interpretation of the carbon-14 data to measure ground-water velocity will require test wells that permit water to be sampled selectively from several depths in the zone of saturation (Section 8.3.1.2.3).

3.8 GROUND-WATER USES

This section identifies the principal regional ground-water users in the hydrogeologic study area (refer to the introduction of this chapter for a discussion of the study area boundaries). The data presented include the locations of areas of heavy withdrawal, the estimated rates of these withdrawals, and the respective hydrogeologic unit sources. The discussion of water use is presented according to the basin divisions given in the introduction and is then further broken down into the hydrographic areas that have been established by the Nevada State Engineer. Water withdrawals within each hydrographic area are summarized according to use when known (irrigation, industrial, commercial, municipal and quasimunicipal, domestic stock-watering); domestic water use is estimated based upon current population data and the guidelines presented in Morros (1982a). Preliminary estimates of repository-related water withdrawals are presented and the potential impacts on local water users are identified. Possible stresses to the ground-water system that could result from repository-related water withdrawals are assessed in light of the current knowledge of the hydrogeologic study area.

The remainder of this section addresses the regional and local ground-water management plans that have been enacted or are proposed within the hydrogeologic study area. The relationship of these plans to repository-related activities and plans for compliance with these State and local restrictions is briefly outlined. Plans for regional monitoring of potential ground-water-related problems are presented in Sections 8.3.1.2 and 8.3.1.16.

3.8.1 REGIONAL AQUIFERS USED FOR HUMAN ACTIVITIES

The regional aquifers in the hydrogeologic study area that are used for human activities are the valley fill aquifer and the lower carbonate aquifer. The welded tuff aquifer is locally important; it is developed only in the southwestern areas of the NTS in support of the former Nevada Research and

Development Area (Figures 3-21 and 3-22). The characteristics of these aquifers and their relationships to each other are presented in Sections 3.6 and 3.7, respectively.

For discussing water use in the hydrogeologic study area, the hydrographic regions delineated by the Nevada State Engineer are the most useful and practical units for consideration (Figure 3-23). For each of the hydrographic areas, the Nevada State Engineer has determined or estimated the perennial yield of the reservoir. The maximum amount of ground water that can be appropriated from a given basin in Nevada is limited to its perennial yield. Thus, the relationship between the perennial yield and the amount of water being used forms the basis for ground-water management in Nevada. When these amounts are known, a quantitative statement of ground-water availability within a specified basin can be made.

Perennial yield (also referred to as safe yield; Todd, 1980) is defined as the amount of water that can be withdrawn on an annual basis from a ground-water basin without depleting the reservoir (Scott et al., 1971). When the amount of withdrawals from a basin exceeds the perennial yield, an overdraft is created that can produce undesirable effects (Todd, 1980) since the water must come from storage within the aquifer. The result is known as aquifer mining and is prohibited in Nevada (Morros, 1982a). When aquifer mining (or over appropriation) is identified within a basin, the State Engineer may issue a statement known as a designation order. The designation order is a means of protecting the aquifer(s) from overuse by defining the boundaries of the area of overdraft and restricting the issuance of permits within that area. Since it is difficult to determine the perennial yield of a basin in which no pumping has occurred (Todd, 1980), aquifer overdrafts tend to be a common problem.

The following subsections discuss current water use for each community located within the three hydrogeologic subbasins of the study area, within the framework of the hydrographic areas. The proposed water use at Yucca Mountain is also examined within this framework, and the potential impacts of repository-related withdrawals are identified. The town of Pahrump, which is located outside of the hydrogeologic study area, is also discussed briefly. The discussion of State and local ground-water management plans and their relationship to activities at Yucca Mountain is deferred to Section 3.8.2.

The town of Pahrump is located approximately 100 km southeast of Yucca Mountain. Studies done by Harrill (1982) and Winograd and Thordarson (1975) indicate that the Pahrump Valley is not located within the boundaries of the Alkali Flat-Furnace Creek Ranch subsystem. Large outcroppings of clastic materials form the mountains to the northwest that separate the Pahrump flow system from the adjacent Ash Meadows subbasin (Harrill, 1982) and allow little or no underflow from the Pahrump Valley into the geohydrologic study area. Well development in Pahrump is completely within the valley fill aquifer, which underlies the community. This aquifer locally receives recharge from the Spring Mountains. Information presented in Harrill (1982) and Winograd and Thordarson (1975) suggests that site characterization activities and the possible subsequent development of a geologic repository in the Alkali Flat-Furnace Creek Ranch basin probably would not affect the Pahrump flow system.

Figure 3-21. Map showing areas of heavy withdrawals near Yucca Mountain. Modified from French et al. (1984).

CONSULTATION DRAFT

Figure 3-22. Index map showing water well locations and area boundaries at the Nevada Test Site. Modified from Witherill (1986).

3.8.1.1 Alkali Flat-Furnace Creek Ranch subbasin

The Yucca Mountain site is located within the Alkali Flat-Furnace Creek Ranch subbasin to the north of the Amargosa Desert (Figure 3-23). Very little ground water is withdrawn in the northern and central parts of this subbasin. In addition, very little ground water has been appropriated to the north of Yucca Mountain (upgradient) according to information filed with the Nevada State Engineer's Office.

The major ground-water users in the area, the town of Amargosa Valley and small rural communities of the northeastern Amargosa Desert, are located in the southwestern portion of the Alkali Flat-Furnace Creek Ranch subbasin (Figure 3-23) (French et al., 1984). Most of the water to these users is supplied by wells; however, there has been some spring development. Most residences rely on individual wells, while some trailer parks, public facilities, and commercial establishments are served by small, private water companies. Table 3-13 summarizes the public water suppliers in the area, the type of well used, and the population served. Table 3-14 identifies the total number of known water wells drilled in the Amargosa Desert according to their defined uses. All wells are completed in and produce from the valley-fill aquifer.

The only mineral production operation in the Amargosa Desert is owned by the American Borate Corporation. The operation, located between Amargosa Valley, Nevada, and Death Valley Junction, California (Figure 3-21), was decommissioned in July, 1986. The facility consisted of a large mineral processing plant and a housing development for its employees (French et al., 1984). Water for the community was pumped from a shallow well, having a capacity of 125 gpm (681 m³/day). The water was treated by a reverse osmosis process before distribution to reduce the total dissolved solids content. A complete analysis of water chemistry is not available, but the fluoride content is known to be greater than State drinking water standards allow. Ground water was also used to supply the mineral processing plant and was obtained from three wells, with reported capacities of 260, 100, and 125 gpm (1,417, 549, and 681 m³/d). Well log data indicate that all the wells were completed in, and produced from the valley fill aquifer (well logs are filed with the Office of the State Engineer). The plant closed permanently in July of 1986 (Pahrump Valley Times Star, 1986) and water use has ceased.

A small portion of the NTS receives its water from wells drilled in the Alkali Flat-Furnace Creek Ranch basin. Five wells, located in the western areas, supply water for drilling, construction, fire protection, and consumption uses. Table 3-15 summarizes the relevant well data. Section 3.8.1.2 presents information on NTS wells located in the Ash Meadows subbasin.

In addition to well production, a number of springs supply water to the region. The main concentration of springs is in Death Valley in the vicinity of Furnace Creek Ranch, approximately 50 to 60 km southwest of the Yucca Mountain site. Many points of ground-water discharge have been identified in Death Valley National Monument in California (refer to Section 3.5.1 for a tabulation of springs) (Winograd and Thordarson, 1975). The water supply for the National Park Service facilities is derived principally from three groups of springs: Travertine Springs, Texas Springs, and Nevares Springs (French et al., 1984). Discharge data are summarized in Table 3-5. This water is

CONSULTATIVE DRAFT

Table 3-13. Public water suppliers in the community of Amargosa Valley^{a,b}

Supplier	Type	Population served
American Borate Trailer Park	Community	300
Amargosa Water Company (IMV)	Community	45
Embrey's Trailer Park	Community	45
Mountain View Apartments and Shopping Center	Community	75
Amargosa Elementary School	Single user	(c)
Amargosa Senior Citizens' Center	Single user	(c)
Coach House Bar	Single user	(c)
Roadside Park 801NY	Single user	(c)
Water-N-Hole	Single user	(c)

^aSource: SAIC (1986).

^bWells are located in the southern portion of the Alkali Flat-Furnace Creek Ranch subbasin and in the southwestern portion of the Ash Meadows subbasin.

^cIn general these systems serve a transient population of at least 25 persons per day.

Table 3-14. Summary of wells drilled in the Amargosa Desert (hydrographic area 230)^a according to defined use^b

Type of user	Number of wells
Domestic	199
Commercial	5
Industrial	3
Irrigation	164
Municipal	5
No data	<u>21</u>
Total	397

^aSee Figure 3-23.

^bTabulated from well logs filed with the Office of the Nevada State Engineer.

CONSULTATIVE DRAFT

Table 3-15. Nevada Test Site water wells located in the Alkali Flat-Furnace Creek Ranch subbasin^a.

Area number ^b	Well	Pumping rate (gpm) ^c	Unit source ^d	Treatment required	Total withdrawals for 1985, in gal (acre-feet) ^e
18	8	400	Tuff	None	63,683,000 (195.4)
19	19	360	Tuff (?) ^f	None	114,467,200 (351.2)
20	U20a-2	340	Tuff	None	20,165,900 (61.9)
25	J-12	815	Tuff	None	25,049,800 (76.9)
25	J-13	680	Tuff	None	37,811,000 (116.0)

^aSource: Witherill (1986) unless otherwise noted.

^bRefer to Figure 3-22 for well locations.

^cTo convert to cubic meters per second, multiply by 6.31×10^{-5} .

^dBased on information provided in Claassen (1973).

^eTo convert acre-feet to cubic meters, multiply by 1.23×10^3 .

^f(?) = uncertain. Information regarding specific members is not available.

used both for irrigation and for human consumption. There is no estimate available to indicate the quantity of water required to maintain the National Park Service facilities at their present level of public service (French et al., 1984). The population served by this water supply varies during the year. From October through April, approximately 800 persons live in the area on a semipermanent basis, and an additional 2,000 persons live in the area as visitors. From May through September, the number of semipermanent residents decreases, and there are few visitors (French et al., 1984).

There are three resorts located within the boundaries of the Death Valley National Monument: the Stovepipe Wells Hotel, Furnace Creek Inn, and Furnace Creek Ranch. At Stovepipe Wells Hotel, northwest of Furnace Creek Ranch, potable water was trucked in for many years from a storage tank at Emigrant Ranger Station, 14 km (9 mi) southwest of the hotel. In 1973, the National Park Service constructed an underground storage tank in the alluvial fan south of Stovepipe Wells Hotel and began trucking water to this tank from Nevares Spring. No other improvements to the system have been documented. Pistrang and Kunkel (1964) report that water from an excavated sump lined

with buried drainage tile in the Furnace Creek Wash is conveyed in a buried pipeline to the Furnace Creek Inn and Furnace Creek Ranch resorts. The measured flow rate ranged from 190 to 206 gpm (1,037 to 1,123 m³/d).

Water use within the Alkali Flat-Furnace Creek Ranch subbasin occurs primarily in the Amargosa Desert (hydrographic area 230 in Figure 3-23). The perennial yield to this hydrogeographic area is estimated to be 24,000 acre-feet per year ($3.0 \times 10^7 \text{ m}^3/\text{yr}$) (French et al., 1984). An estimated 17,000 acre-feet per year ($2.1 \times 10^7 \text{ m}^3/\text{yr}$) of this total is naturally discharged from springs and seeps in the Ash Meadows area, and nearly 10,000 acre-feet per year ($1.2 \times 10^7 \text{ m}^3/\text{yr}$) is artificially discharged from wells in the Amargosa Valley. Thus, an overdraft of 3,000 acre-feet per year ($3.7 \times 10^6 \text{ m}^3/\text{yr}$) currently exists. Water levels in wells drilled in the valley fill aquifer declined an average of 12.3 ft (3.7 m) between 1963 and 1984 (Nichols and Akers, 1985). Total appropriations (Table 3-16) in 1985 were over 70,000 acre-feet per year ($8.6 \times 10^7 \text{ m}^3/\text{yr}$); if these rights to appropriate water were exercised, rapid depletion of the valley fill aquifer would result. Given the current and projected population estimates (Smith and Coogan, 1984; DOE, 1986), this scenario is highly unlikely.

Crater Flat (hydrographic area 229 in Figure 3-23) is currently overdrawn because of an appropriation made to Saga Exploration, Inc. for the development of the Panama-Stirling Mine, located on the east side of Bare Mountain (Section 1.7). Although an overdraft exists, no protective measures will be taken (i.e., no designation order will be issued) because the water has been appropriated for mining, which is considered a preferred use under the Nevada Revised Statutes. Under the statutes, overdrafts for mining are allowable for periods not to exceed 5 yr.

Future applications to appropriate ground water in the Amargosa Desert area will be carefully regulated by the Nevada State Engineer; the majority of applications for irrigation permits submitted through 1982 have been denied (Morros, 1982b). Section 3.8.2 discusses the actions that have been taken by the Nevada State Engineer to mitigate overappropriation problems in this area.

In summary, although the Alkali Flat-Furnace Creek Ranch subbasin as a whole is overappropriated, actual annual water use is less than 30 percent of the perennial yield (based on Table 3-16 for 1985). The overappropriation problem exists primarily in the Amargosa Desert (hydrographic area 230 in Figure 3-23). Outside of the Amargosa Desert, a small number of ranches and domestic users withdraw small amounts of ground water. The amount of water withdrawn in hydrographic areas other than the Amargosa Desert represents only a fraction of the perennial yield that is available to those areas (Table 3-16).

3.8.1.2 Ash Meadows subbasin

Several small, unincorporated communities are located within the Ash Meadows subbasin. A large part of the NTS, including the community of Mercury, is also located within this subbasin. Very little water is withdrawn in this subbasin.

Table 3-16. Perennial yields, total appropriations, and actual water use for 1985 in the hydrographic areas making up the Alkali Flat-Furnace Creek Ranch subbasin

Hydrographic area ^a		Perennial yield (AFY) ^b	Total appropriations (AFY) ^c	Water use in 1985 (AF) ^d	Comments
Number	Name				
227A (West)	Fortymile Canyon (Jackass Flats)	2,000	320.0	192.9	Yucca Mountain site
227B	Fortymile Canyon (Buckboard Mesa)	3,600	NA ^e	0	Nevada Test Site (NTS)
228 (East)	Oasis Valley	1,000	1.10	ND ^f	
229	Crater Flat	900	2533.48 ^g	2533.48	
230	Amargosa Desert	24,000	71,613.66	9,672.0 ^h	Overappropriated; designated
147 (East)	Gold Flat	1,900	ND	292.4	NTS
157	Kawich Valley	2,200	ND	0	
173A (South)	Railroad Valley (southern part)	ND	ND	ND	
Subbasin Totals		35,600	74,468.24	12,690.78	

^aHydrographic areas are shown on Figure 3-23.

^bFrom Scott et al., 1971. AFY = Acre-feet per year. To convert to cubic meters per year, multiply by 1.23×10^3 .

^cTabulated from preliminary abstracts filed with the Office of the Nevada State Engineer.

^dData from Giampaoli, 1986. AF = Acre-feet. To convert to cubic meters, multiply by 1.23×10^3 .

^eNA = not applicable.

^fND = no data.

^gAppropriations for mining and milling may exceed the perennial yield because (1) appropriations for mining activities are for short-term use, and (2) mining and milling applications are considered preferred water uses by the State Engineer.

^hData from Coache, ca. 1986.

The Ash Meadows area is located 42 to 60 km (26 to 37.5 mi) southeast of Yucca Mountain (Figure 3-21). Water is pumped from the lower carbonate aquifer (Dudley and Larson, 1976). Virtually all irrigation withdrawals of ground water in this area currently are made by the Preferred Equities Corporation (Giampaoli, 1986). Water from 8 major wells, supplemented by 3 small capacity wells, is used to irrigate more than 12 km² (4.6 mi²) of cropland. Pumping peaked at a rate of approximately 1.2 million cubic meters per month during the years 1970 and 1971. Detailed pumping records were not maintained after 1971, while the land was controlled by the Spring Meadows, Inc. (Dudley and Larson, 1976). In 1971, ground-water withdrawals associated with the planned development of a large agricultural enterprise caused a decline in the water level of the pool at Devils Hole (Dudley and Larson, 1976). This natural pool formed from the collapse of the limestone bedrock, is the only remaining habitat of the endangered species, Devils Hole pupfish, Cyprinodon diabolis.

As a consequence of court action, ground-water withdrawals in this area are now restricted to a degree that is sufficient to maintain the water level in Devils Hole (Dudley and Larson, 1976). The land, which was previously owned by the Preferred Equities Corporation, was turned over to Nature Conservancy in 1986 and then transferred to the Federal Government (Giampaoli, 1986). Other springs in the Ash Meadows area are discussed in Section 3.5.1.

The NTS receives its water from wells operated by Reynolds Electrical and Engineering Company, Incorporated (REECo). The NTS accommodates a worker population of approximately 5,000 individuals, most of whom reside in Las Vegas and other nearby communities; a very small percentage of this workforce resides in Mercury on an intermittent basis.

There are 12 NTS wells that currently withdraw water from the Ash Meadows subbasin for construction, drilling, fire protection, and consumption uses. Some of the water requires treatment before distribution. Table 3-17 summarizes the available information for these 12 wells. Section 3.8.1.1 presents the same type of information for NTS water wells located in the Alkali Flat-Furnace Creek Ranch basin.

The unincorporated community of Indian Springs, Nevada (population 912) is located 80 km (50 mi) east-southeast of Yucca Mountain (Figure 3-21). The community receives its water from the Indian Springs Sewage Company, Inc., which is a public corporation owned by stockholders. As of 1981, there were 53 water customers, including several trailer parks with multiple connections. The company has one well that is completed at the depth of 590 ft (180 m) and is capable of pumping at a rate of 550 gpm (2,998 m³/d). The water from this well meets State drinking water standards and requires no chemical treatment. The water supply system is metered and is a separate system from that which supplies Indian Springs Air Force Base (French et al., 1984).

Approximately 80 other shallow, domestic wells in the Indian Springs area are in use. Information regarding water quality and quantity for these wells is unavailable. All wells in the area are probably completed in the valley-fill aquifer (Giampaoli, 1986), which is thought to be recharged locally by the lower carbonate aquifer (Winograd and Thordarson, 1975).

CONSULTATIVE DRAFT

Table 3-17. Nevada Test Site water wells located in the Ash Meadows subbasin^a

Area number ^b	Well ^b	Pumping rate (gpm) ^c	Unit source ^d	Treatment required	Total withdrawals for 1985, in gal (acre-feet) ^e
1	UE1t	270	ND ^f	None	5,594,200 (17.2)
2	2	165	Lower carbonate aquifer(?) ^g	None	20,630,500 (63.3)
3	A	160	Valley fill	Chlorination	39,182,300 (120.2)
5	5b	240	Valley fill	None	6,107,800 (187.5)
5	5c	325	Valley fill	None	61,170,400 (187.7)
5	UE5C	350	ND	None	4,319,000 (13.3)
6	C	270	Lower carbonate aquifer(?)	None	26,162,900 (80.3)
6	C-1	280	Lower carbonate aquifer(?)	None	26,170,300 (80.3)
6	4	650	ND	None	41,815,300 (128.3)
15	UE15d	270	Tuff	Chlorination	ND
16	UE16d	194	ND	None	15,605,000 (47.9)
22	Army Well-1	530	Lower carbonate aquifer(?)	Chlorination	53,916,700 (165.4)

^aData from Witherill (1986) unless otherwise noted.^bRefer to Figure 3-22 for well locations.^cgpm = gallons per minute. To convert to cubic meters per second, multiply by 6.31×10^{-5} .^dBased on information provided in Claassen (1973), and Winograd and Thordarson (1975, plate 2b).^eTo convert gallons to cubic meters, multiply by 3.785×10^{-3} .^fND = No data.^g(?) = uncertain.

(Refer to Sections 3.6 and 3.7 for discussions of aquifer characteristics and local recharge areas.)

Indian Springs Air Force Base is located adjacent to the community of Indian Springs, Nevada. The population, including military personnel and their dependents, is approximately 900 persons. The water distribution system for this installation is owned and operated by the U.S. Air Force. Three wells, probably completed at shallow depths, supply water to the base. Two of the three wells are used for potable supply; the third well no longer provides potable water due to the presence of organic materials. All water distributed through this metered system is chlorinated to control bacteria (French et al., 1984).

Table 3-18 summarizes known water use in the hydrographic areas of the Ash Meadows subbasin. Very little water is pumped outside of the areas and communities previously discussed. Indian Springs Valley (hydrographic area 161 in Figure 3-23) is currently overappropriated. Actual water use in 1985 exceeded the perennial yield to the basin. Adverse effects resulting from overdraft include a decline in the static water level (SWL) and a reduction in spring discharges (Giampaoli, 1986). Information provided in Table 3-18 suggests that Yucca Flat (hydrographic area 159 in Figure 3-23) is currently overdrafted. Well data in Table 3-17 indicate that these withdrawals may be from the lower carbonate aquifer. On the basis of this information and the fact that there is no detectable decline reported (Witherill, 1986), it is reasonable to conclude that Yucca Flat is not overdrawn. An overdraft also appears to exist for Frenchman Flat (hydrographic area 160 in Figure 3-23). All withdrawals in this area are from the valley-fill aquifer and have been in excess of the perennial yield since 1964 (Claassen, 1973). There has not been any decline in the SWL, which suggests that the estimated perennial yield is too low (Claassen, 1973; Witherill, 1986).

As can be seen from Table 3-18, little of the ground water in the Ash Meadows subbasin is currently privately appropriated or used. Since much of the land within this basin is under the jurisdiction of federal agencies and has been withdrawn from the public domain, future private appropriation of the water is not expected.

3.8.1.3 Oasis Valley subbasin

The Oasis Valley subbasin encompasses the western parts of the Gold Flat and Oasis Valley hydrographic areas, 147 and 228 (Figure 3-23) respectively. Very little ground water is used in this area because of the low population density.

The unincorporated community of Beatty, Nevada (population 925), receives its water from the Beatty Water and Sanitation District. The water is supplied from four wells; three are located in town, and the fourth (Indian Springs Well) is approximately 6.4 km (4 mi) north of the town (Figure 3-21). The system supplies water to approximately 25 customers and the source capacity is about 500,000 gpd ($1.9 \times 10^3 \text{ m}^3/\text{d}$) (Gram, 1985). The water quality of the three wells in town, which produce from the valley fill aquifer (White, 1979), does not meet State drinking water standards because

Table 3-18. Perennial yields, total appropriations and actual water use for 1985 in the hydrographic areas making up the Ash Meadows subbasin (page 1 of 2)

Hydrographic area ^a		Perennial yield ^b (AFY)	Existing appropriations (AFY) ^c	Water use in 1985 (AF) ^d	Comments
Number	Name				
158A	Emigrant Valley (Groom Lake)	2,500	59.8	ND ^e	
158B	Emigrant Valley (Papoose Lake)	<10	0	0	
159	Yucca Flat	350	NA ^f	537.5	Nevada Test Site (NTS)
160	Frenchman Flat	100	NA	388.5	NTS
161	Indian Springs Valley	500	754.98	679.0 ^g	
168	Three Lakes Valley (N)	4,000	11.48	ND	
169A	Tikaboo North	2,600	21.72	ND	
169B	Tikaboo South	4,000	ND	ND	
211	Three Lakes Valley (S)	5,000	ND	100 ^{g, h}	

3-126

CONSULTATION DRAFT

Table 3-18. Perennial yields, total appropriations and actual water use for 1985 in the hydrographic areas making up the Ash Meadows subbasin (page 2 of 2)

Hydrographic area ^a		Perennial yield ^b (AFY)	Existing appropriations (AFY) ^c	Water use in 1985 (AF) ^d	Comments
Number	Name				
225	Mercury Valley	8,000	NA	165.4	NTS
226	Rock Valley	8,000	320.0	ND	
227A (East)	Fortymile Canyon (Jackass Flats)	<u>2,000</u>	<u>NA</u>	<u>0.</u>	NTS
Subbasin totals		37,060	1,117.98	1,870.4	

^aHydrographic areas are shown on Figure 3-23.

^bData from Scott et al. (1971). AFY = Acre-ft per year. To convert to cubic meters per year, multiply by 1.23×10^3 .

^cTabulated from preliminary abstracts filed with the Office of the Nevada State Engineer.

^dAF = Acre-feet.

^eND = no data.

^fNA = not applicable.

^gData from Giampaoli (1986).

^hEstimated use at Indian Springs Correctional Center.

CONSULTATIVE DRAFT

of excessive fluoride (water chemistry data are presented in Section 3.7.3). The EPA recently raised the maximum allowable fluoride content; if the State of Nevada raises its limits, the water from these three wells could meet the new standard with minimal treatment. The fourth well, located north of Beatty at Indian Springs (not to be confused with the community of Indian Springs, Nevada), is of higher quality and does meet the current State drinking water standards. This well is believed to produce from a perched tuff aquifer (White, 1979). Well production capacity from the Indian Springs well has recently decreased from 389 to 86 gpm (2.1×10^3 to 4.7×10^2 m³/d) due to overpumping and subsequent dewatering of the aquifer (Gram, 1985). Five commercial establishments (Bailey's Hot Springs, Fran's Star Ranch, Ray's Last Stand, Scottie's Junction, and Cottontail) have water supplies and are available to the public. These establishments are all single-users and generally serve transient populations of 25 or fewer people per day. Outlying ranches in the area around Beatty have their own wells.

There are two gold mines operating in the Beatty area that require water for production purposes. The mines are owned by E. R. Fegert, Incorporated, and are located near Rhyolite and the boundary of the Death Valley National Monument. Water for production originates from two sources: One of these is a well owned by the company that produces 40 gpm (218 m³/d). There are no data available to indicate the aquifer from which the well produces. An additional 483,000 gal (1,828 m³) of water per month originates from one of the in-town wells operated by Beatty Water and Sanitation District (French et al., 1984). These mines are currently in the developmental phase; when planned future expansion is realized, peak employment will be approximately 100 persons. Another well is planned that will supplement the current water supplies (French et al., 1984).

Table 3-19 summarizes the known water use within the portions of Gold Flat and Oasis Valley basins (hydrographic areas 228 and 147 in Figure 3-23) that are located within the Oasis Valley subbasin. Note that although water use does not currently exceed the perennial yield in either basin, Oasis Valley (hydrographic area 228) has been designated. The designation of this area by the Nevada State Engineer was a protective measure to prevent the overappropriation.

The water-supply problems that the community of Beatty is experiencing are not related to the designation of the basin (Section 3.8.2.2). A solution may now be imminent since the EPA raised the allowable limit of fluoride content in drinking water.

3.8.1.4 Proposed water use at Yucca Mountain

Water-use estimates for site characterization and subsequent repository construction and operation at Yucca Mountain are still in the early stages of development. The water-use figures presented in this section are preliminary and will probably change as more is learned about Yucca Mountain.

Wells J-12 and J-13, located along Fortymile Wash, supply water to the former Nuclear Research and Development Area facilities at the NTS (Young, 1972). Both of these wells are completed in and producing from the welded

Table 3-19. Perennial yields, total appropriations and actual water use for 1985 in the hydrographic areas making up the Oasis Valley subbasin

Hydrographic area ^a		Perennial yield (AFY) ^b	Total appropriations (AFY) ^c	Water use in 1985	Comments
Number	Name				
147(Southwest)	Gold Flat	1,900	NA ^d	ND ^e	Nevada Test Site
228(West)	Oasis Valley	<u>2,000</u>	<u>1528.72</u>	<u>Minor</u> ^f	Designated
Subbasin totals		3,900	1528.72	Minor	

^aHydrographic areas are shown on Figure 3-23.

^bData from Scott et al. (1971). AFY = Acre-feet per year.

^cTabulated from preliminary abstracts filed with the Office of the Nevada State Engineer.

^dNA = Not applicable.

^eND = No data.

^fData from Giampaoli (1986).

tuff aquifer (refer to Section 3.6.4 for a description of aquifer characteristics). Well J-13 was in continuous service from 1962 to 1969 and caused only a slight decline in the water level (Thordarson, 1983; refer to Section 3.9.7). By 1980, the static water level had recovered to within 0.1 m (0.3 ft) of its original elevation as water withdrawals became intermittent. A more detailed discussion of wells J-12 and J-13 appears in Section 3.9.7.

Test results by Thordarson (1983), Claassen (1973), and Young (1972) indicate that the welded tuff aquifer contains adequate water to supply repository construction, operation, and closure (Morales, 1986). The maximum annual water demand for the repository is anticipated to rise to a peak of 120 million gal (454,200 m³) per year by the end of the seventh year of construction, and decrease to about 115 million gal (435,280 m³) per year, and remain at this level for the next 25 yr. The minimum average water demand for the following 23 yr of operation would be approximately 2.5 million gal (9,462 m³) per year (Morales, 1986). Estimates of the quantity of water needed to support site characterization activities are considerably less. Preliminary estimates from Pedalino (1986) indicate that maximum total water use, for all phases of the exploratory shaft construction and testing will be 41,507,100 gal (1.57 x 10⁵ m³) (Table 3-20). Section 8.4 provides plans for construction of the exploratory shaft and surface facilities.

To date, published information on water use in the hydrogeologic study area indicates that siting a geologic repository at Yucca Mountain probably would not adversely impact other local users (DOE, 1986). According to current plans (Morales, 1986), water for site characterization activities would be drawn from wells J-12 and J-13 (welded tuff aquifer) until a distribution system could be constructed.

The local effects of ground-water withdrawals vary between the different areas previously discussed. The most recent potentiometric map of the hydrogeologic study area was constructed by Waddell et al. (1984) (Figure 3-9). Contour lines of the potentiometric surface are drawn at 100 m intervals, and are too general to show drawdown from pumping. Figure 3-24 shows the potentiometric surface in the Amargosa Desert, as measured from wells in 1962 and in 1984. A decline in the water level and one depression cone can be seen. The decline in water levels, and the formation of a depression cone, resulted from withdrawals made by 27 wells in the valley. Pumping rates for domestic wells range from 10 to 40 gpm while industrial wells produce from 100 to 800 gpm (including irrigation and mining wells) (Giampaoli, 1986). No other local potentiometric maps are published that show depression cones in areas of ground-water withdrawals, and no injection wells exist in the hydrogeologic study area. Plans for developing large-scale potentiometric maps of the hydrogeologic study area are presented in Section 8.3.1.2.2.

3.8.1.5 Water use for energy development

This section examines water use for the purpose of energy development in the hydrogeologic study area. Only geothermal resource development is considered, because it is the only form of ground-water-related energy development known to be present in southern Nevada.

CONSULTATIVE DRAFT

Table 3-20. Estimated water use for exploratory shaft facilities at Yucca Mountain, Nevada^{a,b}

Characterization phase	Amount of water needed	
	gal	m ³
PHASE I: SURFACE FACILITY CONSTRUCTION (30 WEEKS)		
Site preparation	15,376,000	5.8×10^4
Facilities construction	1,410,500	5.3×10^3
Pipe cleaning, flushing, testing, and filling	248,500	9.4×10^2
Fire suppression	<u>90,000</u>	<u>3.4×10^2</u>
Subtotal	17,125,000	6.46×10^4
PHASE II: SHAFT SINKING (107 WEEKS)		
Setup, collar, headframe	645,000	2.4×10^3
Shaft sinking and testing	4,962,350	1.9×10^4
Station construction and changeover	2,618,200	9.9×10^3
Raisebore second exploratory shaft	<u>481,050</u>	<u>1.8×10^3</u>
Subtotal	8,706,600	3.31×10^4
PHASE III: TEST CONSTRUCTION AND SUPPORT (174 WEEKS)		
Excavation	10,225,800	3.9×10^4
Construction	835,200	3.2×10^3
Test support	<u>4,614,500</u>	<u>1.7×10^4</u>
Subtotal	15,675,500	5.92×10^4
Total	41,507,100	1.57×10^5

^aSource: Pedalino (1986).^bIncludes water for personal use and consumption.

EXPLANATION

— — — INFERRED FAULT — From Winograd and Thordarson (1975, page C70, plate 1); delineated on basis of gravity survey

GENERALIZED AREA OF SPRING DISCHARGE

GENERALIZED AREA WHERE NET WATER-LEVEL DECLINE BETWEEN 1962 AND 1984 EXCEEDED 10 FEET

— 2270 — WATER-LEVEL CONTOUR — Shows altitude of ground-water level in valley-fill deposits. Dashed where approximately located. Contour interval 10 feet. Datum is sea level. Contours for 1962 from Walker and Eakin (1963, plate 3)

• 72 WELL — Number is water-surface altitude, January 1984, in feet above 2,200 feet (question mark indicates lack of water-level measurement in 1984). Datum is sea level

Figure 3-24. Potentiometric maps of the Amargosa Desert (valley fill aquifer) based on the well data from (1962) (A) and (1984) (B). Modified from Nichols and Akers (1985).

The low-temperature geothermal resources located in the vicinity of Yucca Mountain have little potential for exploitation (refer to Section 1.7 for an evaluation of geothermal resources). The use of these resources for energy development is highly unlikely. Water temperatures measured in test holes in the vicinity of Yucca Mountain are in the range of 50 to 60°C at depths to 1,800 m (5,906 ft) (Craig and Robison, 1984). Current technology requires reservoir temperatures of at least 180°C for commercial power generation (White, 1973). Consequently, it is extremely unlikely that high temperature waters would be present at depths that are economically attractive. Should advances in technology make deeper geothermal resources economically attractive, many areas other than Yucca Mountain could be exploited (Garside and Schilling, 1979).

In summary, there currently is very little potential for geothermal energy resource development within the hydrogeologic study area.

3.8.2 REGIONAL GROUND-WATER MANAGEMENT PLANS

The first part of this section presents an overview of Nevada's philosophy of ground-water management and the measures that are taken to mitigate ground-water supply problems. Agencies involved in the appropriation of ground-water resources and programs or laws that govern ground-water use that might relate to activities at Yucca Mountain are identified. The second part of this section identifies the nearby communities that might be affected as a result of the proposed repository-related withdrawals. Site characterization activities and the construction and operation of a geologic repository at Yucca Mountain would increase the demand for water in the immediate vicinity of the site, as discussed in Section 3.8.1.4. Water use projections for the next 50- and 100-yr periods currently are not available. Data and information that are needed to assess the potential impacts are delineated. Plans for obtaining these data are presented in Section 8.3.1.16.2.

3.8.2.1 Ground-water management in Nevada

Most of the unincorporated communities located within the geohydrologic study area do not have municipal water supply systems. Water use is governed by the appropriate Board of County Commissioners. The State of Nevada, however, has responsibility for protecting its ground-water resources. Areas and communities examined in this section are Pahrump, Beatty, Amargosa Valley, and Area 25 (Figure 3-21) of the NTS; these areas are located proximal to the proposed repository site, or historically have experienced water supply problems. Estimated and projected water use figures for each community located within the hydrogeologic study area are presented in Table 3-21.

Water use in Nevada is governed by the Office of the State Engineer and the Division of Water Resources. Chapter 534 (Title 48-Water) of the Nevada Water Laws outlines and delineates the allowable uses of ground waters. Morros (1982a) presents the laws and statutes relating to all aspects of ground-water and geothermal resources, as described in Chapter 534 and Chapter 534A respectively.

CONSULTATIVE DRAFT

Table 3-21. Existing total water use in the communities located in the hydrogeologic study area

Community	Estimated population 1985 ^a	Estimated existing water use (mg/d) ^b	Maximum population increase due to site characterization activities ^c	Projected increase in water use due to site characterization activities ^c (mg/d)
Indian Springs	912	0.8	85	0.07
Indian Springs Air Force Base	900	0.4	ND ^d	ND
Beatty	925 ^e	0.8	2	0.002
Mercury	300	0.3	ND	ND
Pahrump	5,500	4.7	127	0.1
Ash Meadows	41	0.04	ND	ND
Johnnie	8	0.007	ND	ND
Amargosa Valley	1,905 ^f	1.6	6	0.005
Crystal	65	0.06	ND	ND
Rhyolite	4	0.003	ND	ND
Death Valley Junction	<u>20^g</u>	<u>0.017</u>	<u>ND</u>	<u>ND</u>
Totals	10,580	8.7	220	0.177

^aPopulation data from Smith and Coogan (1984) except where otherwise noted.

^bmg/d = million gallons per day. Maximum permissible water use in southern Nevada, is 1,800 gal per day per residential unit, or almost 2 acre-feet per year (assuming 2.1 residents per housing unit).

^cData from DOE (1986).

^dND = no data.

^eData from Gram (1985).

^fIncludes population of Amargosa Valley, American Borate Mill, American Borate Housing Complex, and the Spring Meadows area.

^gCommunity distribution system. Water quality does not meet U.S. Environmental Protection Agency drinking water standards (French et al., 1984).

The State of Nevada uses a recharge-use philosophy in its management of ground-water resources. As previously discussed in Section 3.8.1, total annual ground-water withdrawals from any given basin may not exceed the perennial yield (Morros, 1982a). Unfortunately there are many difficulties inherent in determining the actual amount of recharge to an undeveloped basin where no pumping has occurred. In many instances, aquifer overdrafts occur before the perennial yield has been firmly established (Todd, 1980).

At the first indication of an aquifer overdraft, measures are taken to carefully regulate ground-water withdrawals. A provision of the Nevada Water Laws grants authority to the State Engineer to designate ground-water basin boundaries in areas where he deems that ground-water resources are being depleted. Such designation orders have been issued in four areas of consideration: (1) the Amargosa Desert ground-water basin, (2) the Pahrump Artesian Basin, (3) Oasis Valley West, and (4) Indian Springs Valley. Designation of a ground-water basin places permits on a preference basis but does not establish which types of permits have priority. Preference is determined on a case-by-case basis by the State Engineer.

3.8.2.2 Local ground-water supply problems

In May of 1979, the State Engineer issued an order designating the Amargosa Desert ground-water basin to prevent further overappropriation of basin ground water. The areas and communities affected by this designation order are Amargosa Valley, rural water users in the Amargosa Desert, and Ash Meadows. Although no order of preference has been established by the State Engineer, permit applications on file for wells located within the Amargosa Desert ground-water basin indicate that irrigation and agricultural use has the lowest preference. New permits for, or changes to domestic and quasi-municipal uses would be favored (Morros, 1982b). Water use at the NTS and the Yucca Mountain site, which are located outside the designated Amargosa Desert ground-water basin boundaries, is not affected by the designation order.

Although water-use in Oasis Valley is currently considered minor (Giampaoli, 1986), the western portion of the Oasis Valley (hydrographic area 228) was designated in 1980 to prevent over appropriation. Recharge to the ground water basin is estimated to be approximately 2,000 acre-feet per year ($2.5 \times 10^6 \text{ m}^3/\text{yr}$).

The community of Beatty may experience water supply shortages in the near future. Gram (1985) reported that the population growth rate from 1980 to 1985 was nearly 3 percent, while the demand for water had increased nearly 10 percent. Although a high-quality water source has not yet been developed, Gram (1985) has presented several recommendations that could alleviate future water supply problems.

The Indian Springs Valley (hydrographic area 161 in Figure 3-23) was designated in 1980. Water-level declines in the Indian Springs community caused a decrease in local spring discharge (Giampaoli, 1986). Although the amount of water appropriated is minor, total water withdrawals and natural

discharge in the form of springs and evapotranspiration are in excess of the perennial yield.

The town of Pahrump, located approximately 100 km southeast of Yucca Mountain, has experienced ground-water overdraft problems in the past. Excessive agricultural withdrawals before 1970 caused a lowering of the water level in the valley fill aquifer. In 1970, the State Engineer ordered a moratorium on the issuance of water permits for irrigation from the Pahrump Artesian basin. Certificated appropriations and development permits for ground water in the Pahrump Valley totaled 90,790 acre-feet ($112 \times 10^6 \text{ m}^3$) in 1970, although in recent years actual water withdrawal has averaged approximately 39,720 acre-feet per year ($49 \times 10^6 \text{ m}^3$) (DOE, 1986). The repository project probably would not affect the water resources in the Pahrump Valley, since the proposed site is within a different subbasin of the Death Valley ground-water system (Figure 3-23) (Winograd and Thordarson, 1975; Harrill, 1983; French et al., 1984; Waddell et al., 1984).

3.9 SITE HYDROGEOLOGIC SYSTEM

In previous sections of Chapter 3, the regional hydrogeologic setting of Yucca Mountain was considered; this section focuses on the site hydrogeologic system in the immediate vicinity of the site at Yucca Mountain (generally within a few kilometers of the outer boundary of the repository). The site at Yucca Mountain is unique among those being considered in that the repository is intended to be situated deep in the unsaturated zone, about 300 m below land surface and 250 m above the water table (Chapter 6).

The general concept of ground-water flow at Yucca Mountain assumes that a small fraction of the local precipitation enters the unsaturated zone as net infiltration below the surficial plant-root zone, to percolate generally downward and past the repository horizon, and eventually to reach the water table as recharge. Preliminary analyses indicate that temporal variations of net infiltration may be damped out within the uppermost few tens of meters within the unsaturated zone (Weeks and Wilson, 1984; Montazer and Wilson, 1984). Below these depths, moisture flow (net flux) may be under effectively steady-state conditions (Weeks and Wilson, 1984; Klavetter and Peters, 1986). Because net infiltration is irregularly distributed over the surface of Yucca Mountain and because the hydrologic properties within the unsaturated zone are heterogeneous, considerable spatial variability in flux probably occurs at any horizon within Yucca Mountain, including the repository (Montazer and Wilson, 1984).

Recharge at Yucca Mountain joins and moves laterally through the saturated-zone ground water, whose likely sources of recharge are the mountainous highlands north of the site (although there may be some recharge that is more local; see Section 8.3.1.2.1 for plans to evaluate possible recharge along Fortymile Wash). Ground water arriving at the water table beneath Yucca Mountain probably mixes with ground water in the saturated zone only within the upper few hundred meters of saturated Tertiary tuffs, regardless of the particular hydrogeologic units along the flow path. This occurs because vertical hydraulic gradients within this interval are small, and potentiometric heads at greater depth in less permeable rocks of Tertiary and

Paleozoic age are significantly greater than the water table. At least within the upper part of the saturated zone, the water moves south or southeast from Yucca Mountain.

The rocks underlying Yucca Mountain through which the ground water moves are mostly fractured tuffs of Tertiary age; the major stratigraphic units are shown in a schematic geologic section in Figure 3-25. Because physical properties of the rocks that are believed to control movement of ground water in the unsaturated zone are not completely consistent with stratigraphic boundaries, hydrogeologic boundaries for the unsaturated zone have been defined and are shown in Table 3-22.

An understanding of the saturated-zone hydrologic system is required for waste isolation because this zone contains substantial portions of many pathways to the accessible environment. An understanding of saturated-zone hydrology is also needed in order to evaluate the hydrologic effects of future climate changes; these effects include potential rises in the water table and changes in gradients and paths of ground-water flow in the saturated zone. The distributions of critical parameters, such as effective porosity, are presently unknown and will be evaluated by future testing programs (Section 8.3.1.2.3).

This section describes monitoring networks, from which the following baseline hydrologic data are being obtained: potentiometric levels; hydraulic characteristics of the formations that may store and transmit water in the unsaturated and saturated zones; conceptual ideas of modes and flow paths of ground water, including recharge and discharge; hydrochemistry, ground-water age, and hydraulic parameters as indicators of ground-water velocity and travel times; local ground-water uses that could affect the natural flow system; and past hydrogeologic conditions, to the extent that an understanding of them may be helpful for prediction of future conditions.

Because of the need to emphasize the unsaturated zone at Yucca Mountain, hydrogeologic characterization of the site poses a dual problem. Specifically, it is necessary (1) to define the macroscopic hydrologic system within the unsaturated zone and (2) to describe this system in terms of the microscopic hydrologic processes that occur within the thick sequence of welded and nonwelded, fractured and unfractured tuffs that make up the unsaturated zone. The fundamental concepts and theory of moisture flow and storage within variably saturated natural media have been developed for specific application to problems of soil physics (Hillel, 1980), petroleum-reservoir engineering (Amyx et al., 1960) and geothermal resource evaluation (Faust and Mercer, 1977). The resultant generally considered theory may not be applicable to indurated tuffs that have low porosity and low permeability and that also may be highly fractured. Consequently, a complete characterization of the site hydrologic system must examine the mechanisms by which moisture can be stored and transported within the fractures and interstices of deeply buried tuffs. Only then can a set of hypotheses be developed to construct a macroscopic conceptualization within which these mechanisms act to determine the present, naturally occurring hydrologic system.

A major part of the site characterization program for the unsaturated zone, as described in Section 8.3.1.2.2, is devoted to field, laboratory, and theoretically based studies to test various concepts and hypotheses for

CONSULTATION DRAFT

Figure 3-25. Simplified stratigraphy of section across Yucca Mountain showing stratigraphic relationships (see Figure 3-28 for location of section).

CONSULTATIVE DRAFT

Table 3-22. Definition of unsaturated-zone hydrogeologic units and correlation with rock-stratigraphic units^a.

Rock-Stratigraphic Unit		Hydrogeologic Unit	Approximate Range of Thickness (m)	Lithology ^b
Alluvium		QAL	0-30	Irregularly distributed surficial deposits of alluvium and colluvium
Paintbrush Tuff	Tiva Canyon Member	TCw	0-150	Moderately to densely welded, devitrified ash-flow tuff
	Yucca Mountain Member	PTn	20-100	Partially welded to nonwelded, vitric and occasionally devitrified tuffs
	Pah Canyon Member			
	Topopah Spring Member	TSw	290-360	Moderately to densely welded, devitrified ash-flow tuffs that are locally lithophysae-rich in the upper part, includes basal vitrophyre
Tuffaceous beds of Calico Hills		CHn	100-400	Nonwelded to partially welded ash-flow tuffs
Crater Flat Tuff	Prow Pass Member	CHnz		Vitric
	Bullfrog Member	CFu	0-200	Zeolitized
				Undifferentiated, welded and nonwelded, vitric, devitrified, and zeolitized ash-flow and air-fall tuffs

^aSources: Montazer and Wilson (1984) and as noted in footnotes.

^bLithology summarized from Ortiz et al. (1985).

QAL - Quaternary Alluvium
 TCw - Tiva Canyon welded unit
 PTn - Paintbrush nonwelded unit
 TSw - Topopah Spring welded unit
 CHn - Calico Hills nonwelded unit
 CHnv - Calico Hills nonwelded vitric unit
 CHnz - Calico Hills nonwelded zeolitized unit
 CFu - Crater Flat undifferentiated unit

moisture flow in fractured tuffs. These studies are to be combined with large-scale field and mathematical-modeling studies to further develop and test conceptualizations of the overall macroscopic hydrologic system. To establish an internally consistent conceptual framework and terminology, the conventional theory of moisture flow and storage in variably saturated porous media is reviewed in the following paragraphs, together with some of its potential limitations and inadequacies with respect to conditions thought to prevail at the Yucca Mountain site.

The term moisture is introduced here in a generic sense to include both liquid water and water vapor. Moisture flow and storage, thus, include the combined flow and storage of liquid water and water vapor. Under partially saturated conditions, water vapor will be present within the pore gas, regarded as a mixture of air and water vapor. Under most natural conditions, water vapor and liquid water will be in local thermodynamic phase equilibrium within the pore and fracture space. In partially saturated porous media, the equilibrium vapor pressure is a function of both temperature and matric potential and at constant temperature decreases with decreasing saturation (Hillel, 1980). Liquid-water movement in a partially saturated porous medium is described usually by Darcy's law, whereas water-vapor movement proceeds either as molecular diffusion under a water-vapor concentration gradient or by advection accompanying bulk pore-gas flow.

Unless specified otherwise, saturation refers to liquid-water saturation, S_l , and is defined to be the fractional pore volume or fracture space occupied by liquid water. Saturation and other bulk hydrologic properties are meaningful only when defined as averages over volumes of rock sufficiently large to enclose many pores or fractures (over which the saturation and other bulk hydrologic properties may be taken to be averaged) but yet must be small compared with the overall size of the macroscopic hydrologic system. Because representative linear pore dimensions are of the order of 1 micrometer (Peters et al., 1984) or less, while representative fracture spacings are of the order of 5 cm or more (Scott et al., 1983), a considerable difference may exist in the scale of (1) the bulk-rock volume appropriate for defining the rock-matrix hydrologic properties, which are averaged over the pore space, and (2) the volume containing many fractures that is required to define the bulk properties for the fractures. Under these conditions, the pores and fractures may effectively define two overlapping continuum systems, each regarded as an equivalent porous medium. Such a double-porosity model may be appropriate, for example, to describe a highly fractured but otherwise homogeneous tuff in which the fractures bound distinct rock-matrix blocks. At the opposite extreme, the fracture density may be so low that bulk fracture properties cannot be defined meaningfully, in which case the fractures that are present would have to be regarded as discrete entities. Such may be the case in an otherwise unfractured hydrogeologic unit that is transected by a fault.

Under partially saturated conditions, liquid water is bound to the solid within the pore and fracture openings either by surface-tension (capillary) forces or, at very low saturations, by physical or chemical adsorption. The strength of the bonding force is measured in terms of an equivalent negative pressure, or pressure head, here designated the matric potential. In a fractured medium, the matric potential within the fractures need not be equal to that within the enclosing rock matrix, although pressure equilibration

will tend to be established over time. Large differences in matric potential between rock matrix and fractures may occur under transient conditions, for example, within and following an advancing infiltration front, and thus may affect significantly moisture movement under such conditions (Klavetter and Peters, 1986).

Matric potential is a function of liquid-water saturation, and an analytic or graphical representation of the functional relation defines the moisture-retention curve for the porous medium. Moisture-retention curves for most media are not unique; they display hysteresis in which the precise relation between matric potential and saturation depends on whether saturation is increasing (wetting curve) or decreasing (drying curve). Standard techniques, using mercury intrusion, pressure-plate apparatus, thermocouple psychrometers, and centrifuges have been developed (Hillel, 1980) by which to measure the moisture-retention curves for small soil or rock samples. The intact-fracture test within the exploratory shaft (Section 8.3.1.2.2) is designed, in part, to investigate the moisture-retention properties appropriate to discrete fractures.

The vector volumetric flux of liquid water moving under isothermal conditions through a partially saturated, natural hydrogeologic unit, regarded as an equivalent porous-medium continuum system, is generally presumed to be determined by the spatial gradients of matric and gravitational potentials and by the hydrologic properties. This functional dependence is expressed mathematically by the Darcy's law for unsaturated liquid-water flow, one version of which may be written as (Hillel, 1980):

$$q_L = -K_L K_R \text{ grad } (\psi_L + z), \quad (3-3)$$

where

q_L = volumetric flux of liquid water (L^3/L^2T)

K_L = saturated (liquid) hydraulic conductivity (L/T)

K_R = relative hydraulic conductivity (dimensionless)

ψ_L = capillary-pressure head (L)

z = vertical coordinate (positive upward) relative to a specified x and y-coordinate plane (L)

$\text{grad} = i \frac{\partial}{\partial x} + j \frac{\partial}{\partial y} + k \frac{\partial}{\partial z}$, where i , j , and k are unit vectors aligned

with the x-, y-, and z-coordinate axes, respectively ($1/L$).

In a fractured medium under the dual-porosity model, identical versions of Equation 3-3 would apply separately to the rock-matrix and fracture systems with q_L , ψ_L , K_L , and K_R defined appropriately for these two distinct systems.

The saturated hydraulic conductivity, K_ℓ , entering Equation 3-3 may be expressed in terms of the intrinsic permeability, k , of the medium as (Hillel, 1980):

$$K_\ell = \frac{k \rho_\ell g}{\mu_\ell}, \quad (3-4)$$

where

k = intrinsic permeability (L^2)

ρ_ℓ = liquid-water mass density (M/L^3)

g = gravitational acceleration (L/T^2)

μ_ℓ = liquid-water viscosity (M/LT).

The medium is (hydrologically) homogeneous if k and K_ℓ are spatially invariant and is isotropic if k and K_ℓ are independent of direction. In general, k and K_ℓ are assumed to vary stochastically within an otherwise homogeneous natural medium with mean values that are spatially invariant to within some statistically definable limits of uncertainty (Dettinger and Wilson, 1981; Travis, 1986). These mean values constitute one basis on which to identify and discriminate between distinct hydrogeologic units. The parameters k and K_ℓ are likely to be anisotropic in layered media, such as bedded tuffs, and in fractured media in which the resultant directional dependence is determined by the geometry of the fracture network.

The relative hydraulic conductivity, K_R , is defined as the ratio of hydraulic conductivity at a given matrix saturation to that at complete matrix saturation (K_ℓ). The parameter, K_R , is of unit magnitude at complete saturation and, usually, decreases rapidly towards zero with decreasing saturation or, equivalently, matric potential. Because the functional dependence of K_R on S_ℓ , or ψ_ℓ , is difficult to determine through direct measurement, procedures have been developed for soils by which to infer closed-form analytic representations for this dependence from the corresponding moisture-retention curve (Brooks and Corey, 1964; Mualem, 1976; van Genuchten, 1980). It is not clear that these approximating representations are appropriate to indurated fractured tuffs, but they have been used in preliminary studies at Yucca Mountain (Klavetter and Peters, 1986; Rulon et al., 1986) in lieu of more fundamental data. That this approach is appropriate to describe the rock-matrix properties is supported indirectly in that the moisture-retention curves measured on individual samples of tuff can be fitted accurately, for example, by the analytic representation developed for soils by van Genuchten (1980). Fractured-rock systems introduce further complications that depend on the mechanisms by which moisture is transported in partially saturated fractures and define an area of current active research (Montazer and Harrold, 1985; Wang and Narasimhan, 1985). The bulk-permeability test to be conducted within the exploratory shaft (Section 8.3.1.2.2) is intended to obtain direct empirical data on the bulk hydrologic properties for the unsaturated, fractured Topopah Spring welded unit.

The moisture-retention curve together with the dependence of relative hydraulic conductivity on saturation or, equivalently, matric potential, constitute the set of moisture-characteristic relations for a particular porous medium under partially saturated conditions. Consequently, these relations must be supplied as part of the suite of hydrologic-property data for each hydrogeologic unit within the unsaturated zone at Yucca Mountain. A hydrogeologic unit, in this usage, is defined to be a functional unit composed of an interval or volume of rocks within which the mean hydrologic properties are effectively spatially invariant. A hydrogeologic unit may include, in whole or in part, one or more rock-stratigraphic units.

In addition to the storage and transport of moisture as liquid water within the unsaturated zone, moisture may be stored and transported as water vapor within the air-filled pore space. The bulk flow of air in pores and fractures probably is usually regarded to be Darcian in which the appropriate potential is the local pore-gas pressure. Gas flow is more complex, however, because the compressibility of the gas must be taken into account through an appropriate equation of state. Furthermore, in pores and fractures whose apertures are a few times the mean free path of pore-gas molecules, the gas will tend to slip past the pore and fracture walls to produce an effective increase of gas permeability k_g according to the law $k_g = k (1 + b/P)$ (Klinkenberg, 1941), where k is the (intrinsic) permeability of the medium, P is the ambient pore-gas pressure, and b is an empirical constant appropriate to the pore-gas system under consideration. For example, Reda (1985a) determined experimentally that the permeability with respect to nitrogen gas of a sample of the TSw unit was a strongly dependent function of pore-gas pressure for which the Klinkenberg constant $b = 0.76$ MPa.

Not only does bulk-gas flow and storage need to be considered in order to assess the efficacy of vapor-phase moisture transport within the unsaturated zone at Yucca Mountain; but also, air is a very convenient fluid medium to use in field determinations of permeability by monitoring both barometrically induced pressure changes with depth in wells (Weeks, 1978) and the response in observation wells due to air injection within packed-off zones in neighboring test wells. The monitoring and mathematical modeling of gas-tracer distribution and movement (using both environmental and introduced tracers) yield information on the overall pore-gas flow system from which inferences pertaining to the liquid-water system may be drawn (Section 3.9.1.3.).

The division of the unsaturated zone at Yucca Mountain into hydrogeologic units is a first step in defining the macroscopic hydrologic system at the site. As described in Section 1.2.2, and depicted in Figure 3-26, Yucca Mountain is composed of a stratified sequence of Tertiary volcanic rocks consisting of ash flow and ash-fall tuffs that, at land surface, are overlain by irregularly distributed deposits of alluvium and colluvium. The division of this sequence into hydrogeologic units as described here is based on the divisions developed by Montazer and Wilson (1984) and by Ortiz et al. (1985) and is shown correlated with formal rock-stratigraphic units (Chapter 1) in Table 3-22. The hydrogeologic units are identified in Table 3-22 by symbols such as TSw, and these symbols are used to designate hydrogeologic units in subsequent discussions of the unsaturated zone to distinguish clearly between hydrogeologic and rock-stratigraphic units. The symbol QAL indicates alluvium, TCw indicates the Tiva Canyon welded (hydrogeologic)

CONSULTATION DRAFT

UNSATURATED-ZONE HYDROGEOLOGIC UNITS

Figure 3-26. Conceptual east-west hydrogeologic section through the unsaturated zone at Yucca Mountain. Modified from Montazer and Wilson (1984).

unit, PTn indicates the Paintbrush nonwelded unit, TSw indicates the Topopah Spring welded unit, CHn indicates the Calico Hills nonwelded unit (with "v" indicating the vitric facies and "z" indicating the zeolitic facies), and CFu indicating the Crater Flat unit.

Each of the individual hydrogeologic units listed in Table 3-22 belongs to one of three broadly based hydrogeologic rock types distinguished qualitatively as follows (the values given for saturated matrix hydraulic conductivity and for fracture density are taken from Section 3.9.2.1):

1. Densely to moderately welded tuffs that are highly fractured. This group includes units TCw and TSw of Table 3-22 and is characterized by low saturated hydraulic conductivity of the rock matrix (about 1×10^{-11} m/s) and relatively high fracture densities (about 10 to 40 fractures/m³). Under fully saturated conditions, the fracture hydraulic conductivity probably exceeds that of the rock matrix by several orders of magnitude. The rocks of this group are characterized by relatively low matrix porosities of about 10 percent.
2. Nonwelded vitric tuffs containing few fractures. This group includes the PTn and CHnv units of Table 3-22 and is characterized by relatively high saturated hydraulic conductivity of the rock matrix (about 10^{-8} to 10^{-6} m/s) and low fracture densities (1 to 3 or fewer fractures/m³). These rocks have relatively high matrix porosities (about 30 to 45 percent).
3. Nonwelded zeolitized tuffs containing few fractures. The prototype example is the CHnz unit of Table 3-22, which has a saturated hydraulic conductivity of about 10^{-11} m/s or less, a fracture density of 1 fracture/m³ or less, and a matrix porosity of about 30 percent. A tuff sample need not be completely zeolitized to belong in this classification. The tuffs in CHnz range from partially to completely zeolitized but are grouped into this one rock type when their hydrologic properties are similar.

The hydrologic properties of the hydrogeologic units are discussed in greater detail in Section 3.9.2.1.

The formulation of a conceptual model of a natural process or phenomenon is the essence of scientific method that according to Russell (1948), "consists in inventing hypotheses which fit the (available) data, which are as simple as is compatible with this requirement, and which make it possible to draw inferences subsequently confirmed by observation." In general, the conceptual model of a natural hydrologic system considers the macroscopic configuration and state of the system together with the microscopic hydrologic processes operating within it.

Generalized conceptual models for the hydrologic system within the unsaturated zone at Yucca Mountain have been presented by Montazer and Wilson (1984) and Sinnock et al. (1986). The geologic framework underlying these conceptualizations is illustrated schematically in Figure 3-26, which depicts a generalized east-west cross section through the Yucca Mountain block at the repository site (Figure 3-25).

As depicted in Figure 3-26, the geologic framework for the hydrologic system within the unsaturated zone is defined by a block of layered, east-dipping hydrogeologic units that is bounded above by land surface, below by the water table, laterally on the west by a west-dipping normal fault, and laterally on the east by one or more west-dipping normal faults. In addition, the interior of the block may be transected by one or more high-angle faults (e.g., the Ghost Dance fault) across which hydrologic properties may change abruptly. In addition, the bounding and internal faults may act preferentially either as conduits for or as barriers against moisture flow. The processes of moisture flow that are envisioned to be occurring or that could occur under natural conditions within this macroscopic system are described and evaluated in terms of currently available data in Section 3.9.3.4.

3.9.1 BASELINE MONITORING

Baseline monitoring is an engineered system for continuing measurement of existing ground-water conditions that will serve as an historical data base for future observational comparison. Long-term monitoring of potentiometric levels and hydrochemistry may demonstrate the degree of stability of the geohydrologic system during the period required for waste isolation. In addition, water-level observations would record any changes of the water table that would affect the overall ground-water travel time in both the unsaturated and saturated zones beneath the repository.

Eight monitoring boreholes have been drilled into the unsaturated zone (Figure 3-27). Of these only one, drillhole USW UZ-1, located north of the Yucca Mountain site, is currently being monitored. Ten additional boreholes will be included in the unsaturated-zone monitoring network (Section 8.3.1.2.2). Monitoring will be accomplished with a variety of instruments permanently emplaced at various levels within these boreholes. These instruments will be connected to an automatic data acquisition system on the ground surface. Data collected during monitoring will include downhole temperature, pneumatic pressure, matric potential, and pore gas sampling for hydrochemical analysis. The instrumentation to be installed within the boreholes is designed to permit monitoring to continue for an indefinite period of time after the site characterization program has been completed, as required for performance confirmation.

Based on observations of instrument performance and analysis of data collected from drillhole USW UZ-1, the concept of long-term monitoring within the unsaturated zone in deep boreholes appears tenable. After more than 2 yr of monitoring, the majority of the instruments installed in drillhole USW UZ-1 were still functioning and producing reasonable data (Montazer et al., 1985). Considerable experience in instrumenting a deep borehole and in assessing ambient conditions within the unsaturated zone at Yucca Mountain was gained from the trial instrumentation of drillhole USW UZ-1. This experience will be applied to the instrumentation of the remaining boreholes (Section 8.3.1.2.2).

A network of 25 test holes exists from which water-level measurements of the saturated zone are obtained (Figure 3-28). Ten of the holes are

Figure 3-27. Locations of drilled monitoring boreholes and core sampling boreholes in the unsaturated zone at Yucca Mountain.

CONSULTATION DRAFT

Figure 3-28. Preliminary composite potentiometric-surface map of the saturated zone, Yucca Mountain. Modified from Robison (1986).

monitored continuously, and the remainder are measured periodically. Those being measured periodically are being converted to continuous monitoring. At least 8 additional water-table drillholes are planned to define further the gradient of the potentiometric surface (Section 8.3.1.2.3.1). Observation of small water-level changes is difficult because depths to water at Yucca Mountain are great, ranging from about 300 to 750 m below land surface (Robison, 1986), and because the changes are small relative to the measuring precision attainable. Based on available data, it can be concluded that seasonal variations in water levels probably are a fraction of a meter, and no long-term trends have been discerned yet.

3.9.1.1 Monitoring networks

This section describes the monitoring wells currently in existence in the unsaturated and the saturated zones at Yucca Mountain.

3.9.1.1.1 Unsaturated zone

Monitoring of ambient hydrologic conditions in the unsaturated zone is an important component of the planned hydrogeologic investigations for site characterization at Yucca Mountain. The monitoring program, both present and planned, is designed to quantify the energy status of the variably saturated rock-water system at Yucca Mountain. Moisture flux, whether under saturated or unsaturated flow conditions, cannot be measured directly and must be calculated from measured values of hydraulic conductivity and total potential gradient along the path of flow. Fluid potentials that govern variably saturated two-phase (liquid and gas) flow are as follow: (1) matric potential for the liquid phase; (2) gravitational potential; (3) gas-phase pressure, or pneumatic potential; (4) osmotic potential; and (5) thermal potential. In situ borehole monitoring at Yucca Mountain is designed to measure directly these potentials (except the gravitational potential, which is given directly by the height of the monitoring point above some arbitrarily selected datum). Laboratory testing of core and in situ pneumatic testing before instrumenting the boreholes are the means by which hydraulic conductivities (under various degrees of saturation) are obtained in order to calculate moisture and gas fluxes once the potential gradients are known. Plans for these testing programs are described in Section 8.3.1.2.2.

Hydrologic conditions within the unsaturated zone at Yucca Mountain are presently being monitored only in drillhole USW UZ-1 (Figure 3-27), which was completed in November 1983. This drillhole was drilled without water using a reverse air vacuum system to minimize disturbing local hydrologic conditions near the borehole (Whitfield, 1985). It was drilled to a total depth of 387 m, whereupon drilling was discontinued because an apparent perched-water zone was encountered. However, because the water was contaminated with the drilling-fluid polymer used to drill drillhole USW G-1, located 305 m from drillhole USW UZ-1, it was speculated that the perched-water horizon was not natural but a result of drilling drillhole USW G-1 (Whitfield, 1985). Following drilling and geophysical logging, drillhole USW UZ-1 was instrumented at 33 depth levels over a total depth of 371 m (Montazer et al., 1985). At

15 of these levels, 3 well screens were embedded in dry, coarse-sand columns that were isolated from each other by thin layers of bentonite, columns of silica flour, and isolation plugs consisting of expansive cement. The upper and lower screens within each of these 15 levels were equipped with access tubes for gas sampling. Thermocouple psychrometers to measure matric potential were emplaced in the upper and middle screens and pressure transducers to measure pore-gas pressure in the middle and lower screens at each level. In addition to these instruments, 18 heat-dissipation probes, which also measure matric potential, were installed within the columns of dry silica flour. Six of the heat-dissipation probe stations were also equipped with thermocouple psychrometers. The configuration of the completed borehole and the location of the 33 instrument stations within the borehole are shown in Figure 3-29a and 3-29b. The operation and calibration of the thermocouple psychrometers and the heat-dissipation probes are described by Thamir and McBride (1985). Thermocouple psychrometers measure matric potentials over the range from -4.5 to -75 bars (-0.5 to -7.5 MPa), whereas the useful range of the heat dissipation probes is from -0.1 to -5 bars (-0.01 to -0.5 MPa) (Thamir and McBride, 1985). All instruments were connected to an automatic data-acquisition system installed at land surface. Data loggers were programmed to take hourly readings of the thermocouple psychrometers and twice daily readings of the heat-dissipation probes during the first 50 days of operation, after which the thermocouple psychrometers were read twice daily and the heat-dissipation probes once daily. The pressure transducers and thermocouples were sampled twice hourly during the first 90 days of operation and, subsequently, have been sampled every 2 h (Montazer et al., 1985).

The results of monitoring in borehole USW UZ-1 are described in a detailed report by Montazer et al. (1985). Preliminary findings indicated that after more than two years of regular monitoring, most of the instruments were still functioning and producing reasonable data.

Several operational problems were identified during the 2-yr trial monitoring period. These included possible long-term drift of the pressure transducers, long equilibration times for the thermocouple psychrometers placed in the dry, silica-flour columns, and frequency of failure of the heat-dissipation probes. Data from drillhole USW UZ-1 indicate that the gas-sampling program has had a positive effect on minimizing the equilibration time for psychrometers in screens equipped with gas sampling access tubes. Apparently, the periodic withdrawal of formation gas across the initially dry, coarse-sand columns decreased the time needed to establish moisture equilibrium between the instrument chamber and host formation. These equilibration problems will be addressed during site characterization (Section 8.3.1.2.2).

In addition to the continued matric-potential, gas-pressure, and gas-sampling monitoring at drillhole USW UZ-1, an expanded network of monitoring boreholes is planned (Section 8.3.1.2.2) to enlarge the areal and vertical coverage of ambient and changing hydrologic conditions within the unsaturated zone. Deep boreholes that penetrate the tuffaceous beds of Calico Hills are to be drilled north, east, south, and west of the repository block. The remaining boreholes are intended to penetrate the uppermost Topopah Spring Member of the Paintbrush Tuff and will be used to assess the rate and spatial distribution of net infiltration. Eight wells of this network have been

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Figure 3-29a. Instrumentation of monitoring borehole USW UZ-1 (0- to 623-feet depth). Modified from Montazer et al. (1985).

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Figure 3-29b. Instrumentation of monitoring borehole USW UZ-1 (623-1217.5 feet depth). Modified from Montazer et al. (1985).

drilled but not instrumented. These wells are listed in Table 3-23 and their locations shown in Figure 3-27.

3.9.1.1.2 Saturated zone

As indicated in Section 3.9.1, water-level measurements presently being made provide a foundation for a monitoring program. The monitored sites (most of those shown on Figure 3-28 are part of the monitoring program) include two geologic test holes, which were drilled as deep as 1,800 m to obtain data on lithology and stratigraphy; seven hydrologic test holes, in which pumping and other tests were performed to determine hydraulic characteristics of the formations; and 14 water-table holes, which were planned to penetrate the water table only a minimum amount. The water-table holes are areally distributed to enable definition of the potentiometric surface at Yucca Mountain so that gradients and probable flow paths can be determined. Additional water-table holes are planned to be drilled and monitored (Section 8.3.1.2.3.). The planned principal purpose of the drillholes now being observed was not for baseline monitoring, and there was no formal selection process for their locations or depths. However, they are situated for general coverage of the upper part of the saturated zone, through which ground-water flow paths from the repository are likely.

The construction of the geologic and hydrologic series of test holes was such that the first early measurements of head were composites, reflecting averages of all the members of the Paintbrush and Crater Flat Tuffs that were penetrated below the water table. Subsequent to initial completion, four piezometers of varying lengths were installed in one drillhole (drillhole USW H-1) in order to measure heads in each of four zones (water levels for the upper three intervals are very similar; water level in the deepest interval is about 54 m higher; refer to Table 3-24 and Robison, 1986). A single semipermanent packer was installed between permeable zones in each of 5 hydrologic holes (drillholes UE-25b#1, USW H-3, USW H-4, USW H-5, and USW H-6), enabling comparison of head in the zones above and below the depth of the packer setting. Potentiometric heads in the Tertiary rocks penetrated in drillhole UE-25p#1 were measured while drilling, but these rocks were sealed off before the well penetrated the underlying Paleozoic rocks, and present measurements reflect head only of the Paleozoic interval (Section 3.9.1.2).

Because the range of water-level altitudes on the eastern side of Yucca Mountain area is small, measurement precision is important for determination of fluctuations, gradients, and probable flow paths. Therefore, a discussion of some of the aspects of water-level measurements is presented below.

Periodic measurements of water levels typically involve lowering, over a pulley with depth counter, a portable cable to which is attached a float switch or pressure transducer with readout to land surface; depth adjustments include a factor for counter error, and conversion is made from transducer output (millivolts) to submergence of the transducer meters or feet. Repeat measurements are made on the same date, and averaged. Because the depths to water are large, probable error for an individual water-level measurement may be as much as 0.1 m.

Table 3-23. Completion records of unsaturated zone boreholes (page 1 of 2)

Borehole designation ^a	Drilling method	Total depth (m)	Diameter (cm)	Depth interval (m)	Casing inside diameter (cm)	Casing depth interval (m)	Rock-stratigraphic unit penetrated at total depth
USW UZ-1	Reverse vacuum	386.8	122 91 61 44.5	0-12.5 12.5-29.6 29.6-30.8 30.8-386.8	41	0-12.0	Topopah Spring Member (Paint-brush Tuff)
UE-25UZ-4	Odex/ cored	111.9	15 10.8	0-68.9 68.9-111.9	13	0-17.7	Topopah Spring Member (Paint-brush Tuff)
UE-25UZ-5	Odex	111.3	15	0-111.3	13	0-5.2	Topopah Spring Member (Paint-brush Tuff)
USW UZ-6	Reverse vacuum	575.2	76 61 44	0-12.2 12.2-103.9 103.9-575.2	66 48	0-12.2 0-98.8	Prow Pass Member (Crater Flat Tuff)
USW UZ-6S	Odex	158.2	22 18	0-150.9 150.9-158.2	10	0-0.91	Topopah Spring Member (Paint-brush Tuff)
USW UZ-7	Odex	63	15.2	0-63	12.7	0-6	Topopah Spring Member (Paint-brush Tuff)

Table 3-23. Completion records of unsaturated zone boreholes (page 2 of 2)

Borehole designation ^a	Drilling method	Total depth (m)	Diameter (cm)	Depth interval (m)	Casing inside diameter (cm)	Casing depth interval (m)	Rock-stratigraphic unit penetrated at total depth
USW UZ-8	Odex	107	21.6	(b)			Topopah Spring Member (Paint-brush Tuff)
USW UZ-13	Odex/ cored	131	15.2 10	0-125 125-131	12.7	0-101	Topopah Spring Member (Paint-brush Tuff)

^aLocation of boreholes shown on Figure 3-27.

^bDrilling incomplete.

Table 3-24. Ground-water levels, Yucca Mountain area^a (page 1 of 3)

Hole or well number	Location ^b		Hole depth ^c (m)	Land-surface altitude ^d (m)	Depth correction ^e (m)	Date measured ^f	Geologic unit ^g	Water level (corrected)		
	North (ft)	East (ft)						Interval ^h (m)	Depth to water ⁱ (m)	Altitude ^j (m)
UE-25b#1	765,243	566,416	1,220	1,200.7	0.25	12/03/83	Th/Tlr	Composite	470.6	730.9
						08/01/83	Th/Tct	471-1199	470.3	730.4
						08/01/83	Tct/Tlr	1199-1220	472.2	728.5
						11/07/83	Tpt/Tct	Composite	400.3	730.3
UE-25c#1	757,095	569,680	914	1,130.6	0.06	11/07/83	Tpt/Tct	Composite	400.3	730.3
UE-25p#1	756,171	571,485	1,805	1,114.2	0.06	02/ /83	Th/Tcp	383-500	383.9	730.1
						11/07/83	Pz/Pz	1297-1805	364.7	749.4
						03/23/82	Tcp/Tof	Composite	571.7	754.2
						09/17/82	Tpt/Tof	Composite	524.9	1,029.0
USW G-1	770,500	561,000	1,829	1,325.9	0.67	11/30/83	Tcb/Tof	Composite	750.3	730.2
USW G-2	778,824	560,504	1,831	1,553.9	0.16	11/30/83	Tcb/Tof	Composite	750.3	730.2
USW G-3	752,780	558,483	1,533	1,480.5	0.57	04/27/83	Tcp/Tct	Composite	539.5	730.0
USW G-4	765,807	563,082	915	1,269.5	1.53	02/25/82	Tcp/Tof	Composite	572.1	730.9
USW H-1	770,254	562,388	1,829	1,303.0	0.19	11/01/83	Tcp/Tcp	572-673	572.4	730.7
						11/01/83	Tcb/Tcb	716-765	572.4	730.7
						11/01/83	Tct/Tct	1097-1123	571.7	731.4
						11/01/83	Tof/Tof	1783-1814	518.2	784.9
						11/19/82	Tcb/Tlr	Composite	570.8	732.4
						11/03/83	Tcb/Tlr	751-1190	750.8	732.3
						11/03/83	Tlr/Tlr	1190-1219	729.0	754.0
						12/30/82	Tcp/Tlr	Composite	518.7	729.8
USW H-4	761,643	563,911	1,219	1,248.5	0.45	06/16/83	Tcp/Tlr	518-1181	518.2	730.3
						06/16/83	Tlr/Tlr	1181-1219	518.1	730.4
						12/22/83	Tcb/Tl	Composite	704.2	774.7
USW H-5	766,634	558,909	1,219	1,478.9	0.08	11/07/83	Tcb/Tl	704-1091	703.8	775.1
						11/07/83	Tl/Tl	1091-1219	703.8	775.1
						12/15/82	Tcp/Tlr	Composite	526.6	775.1
						10/24/83	Tcp/Tcb	526-1187	526.1	775.6
USW H-6	763,299	554,075	1,220	1,301.7	0.05	10/24/83	Tct/Tlr	1187-1220	524.7	777.0
						10/31/83	Th/Tcb	Composite	471.0	730.4
						11/01/83	Tcp/Tcp	Composite	571.0	730.3
USW WT#1	753,941	563,739	515	1,201.4	0.33	10/31/83	Th/Tcb	Composite	471.0	730.4
USW WT#2	760,661	561,924	628	1,301.3	0.53	11/01/83	Tcp/Tcp	Composite	571.0	730.3

Table 3-24. Ground-water levels, Yucca Mountain area^a (page 2 of 3)

Hole or well number	Location ^b		Hole depth ^c (m)	Land-surface altitude ^d (m)	Depth correction ^e (m)	Date measured ^f	Geologic unit ^g	Water level (corrected)		
	North (ft)	East (ft)						Interval ^h (m)	Depth to water ⁱ (m)	Altitude ^j (m)
UE-25 WT#3	745,995	573,384	348	1,030.0	0.27	10/31/83	Tcb/Tcb	Composite	300.5	729.5
UE-25 WT#4	768,512	568,040	482	1,169.2	0.46	11/01/83	Th/Th	Composite	438.9	730.4
UE-25 WT#6	780,576	567,524	383	1,314.8	0.24	10/31/83	Th/Th	Composite	283.8	1,031.3
USW WT#7	755,570	553,891	491	1,196.9	0.03	10/24/83	Tpt/Tcp	Composite	421.2	775.7
USW WT#10	748,771	553,302	430	1,123.4	0.03	10/24/83	Tpt/Tpt	Composite	347.7	775.7
USW WT#11	739,070	558,377	441	1,094.1	0.12	10/24/83	Tpt/Th	Composite	363.9	730.2
UE-25 WT#12	739,726	567,011	399	1,074.7	0.20	10/31/83	Tpt/Th	Composite	345.4	729.3
UE-25 WT#13	756,884	578,843	352	1,032.5	0.01	10/31/83	Tpt/Tpt	Composite	303.3	729.2
UE-25 WT#14	761,651	575,210	399	1,076.4	0.09	11/07/83	Tpt/Th	Composite	346.2	730.2
UE-25 WT#15	766,116	579,806	415	1,083.2	0.19	12/01/83	Tpt/Tpt	Composite	354.2	729.0
UE-25 WT#16	774,420	570,395	519	1,210.9	0.06	12/01/83	Th/Th	Composite	472.7	738.2
UE-25 WT#17	748,420	566,212	443	1,124.0	0.48	11/07/83	Tcp/Tcp	Composite	394.6	729.6
Well J-11	740,968	611,764	405	1,050.0	-- ^k	03/22/73	--/Tpt	Composite	317.4	732.6
Well J-12	733,509	581,011	347	953.5	--	12/05/83	Tpt/Tpt	Composite	226.2	727.3
Well J-13	749,209	579,651	1,063	1,011.3	--	10/31/83	Tpt/Tlr	Composite	283.2	728.1
USW VH-1	743,356	533,626	762	963.5	--	02/12/81	Qa/Tcb	Composite	184.2	779.3
USW VH-2	748,320	526,264	1,219	974.4	--	04/23/83	Qa/Tcb	Composite	164.0	810.4

^aRobison (1986). See text for discussion of methods.

^bLocation: Nevada State Coordinate System Control Zone.

^cHole depth: Total depth drilled.

^dLand-surface altitude: Altitude above sea level at well, as determined by U.S. Geological Survey National Mapping Division, 1984. Altitude at wells J-11, J-12, J-13 and holes USW VH-1, USW VH-2 reported by Holmes and Narver, Inc., contractor to U.S. Department of Energy.

^eDepth correction: Correction needed to adjust measured depth or altitude to true value because of hole deviation from vertical. The correction is computed for a depth approximately equal to the depth of the measured water level; this correction is obtained from a downhole gyroscopic survey.

Table 3-24. Ground-water levels, Yucca Mountain area^a (page 3 of 3)

Footnotes (continued)

^fDate measured: Date of a water-level measurement.

^gGeologic unit: Unit before the virgule is the shallowest geologic unit represented by the water level; unit after virgule is the deepest geologic unit represented by the water level. Units are defined as follows: Qa = alluvium; Tpt = Topopah Spring Member of Paintbrush Tuff; Th = tuffaceous beds of Calico Hills; Tcp = Prow Pass Member of Crater Flat Tuff; Tcb = Bullfrog Member of Crater Flat Tuff; Tct = Tram Member of Crater Flat Tuff; Tl = Lava; Tlr = Lithic Ridge Tuff; Tof = older flows and tuffs; Pz = Paleozoic rocks.

^hInterval: Depth interval of the hole represented by the water-level measurement. Composite levels represent mixed hydraulic heads of the entire interval between the water table or the lower end of the casing and the bottom of the hole. Where a specific interval is indicated, the zone was isolated using a single packer installed to determine hydraulic head differences above and below the packer.

ⁱDepth to water: Depth based on direct measurement of water levels using downhole wireline equipment, adjusted for depth correction, where available. Accuracy of measurements is approximately ± 0.1 m. Depth in USW VH-2 estimated from geophysical logs.

^jAltitude: Computed altitude of water level above sea level, based on land surface altitude and measured depth to water (corrected). Where more than one altitude is reported, the value used on the map (Figure 3-28) is underscored in the table.

^k-- indicates no data.

The apparent range of water-level fluctuations, as indicated by recently-installed downhole pressure transducers in some drillholes, is commonly smaller than the precision from periodic or repetitive measurements made with calibrated tapes or cables, and, therefore, hydrographs of the limited data available might be imprecise and possibly misleading. For some uses a need exists for more accurate and precise measurement of water-level changes than normally has been feasible for deep water levels; the problem is discussed further in the following paragraph and will be addressed in the continuing monitoring program (Section 8.3.1.2.3).

Pressure transducers can sense extremely small changes of water levels, but they require determination of a factor to convert transducer output (millivolts) to depth (meters) of submergence of the transducer below the water level. This factor may change slightly, and it must be determined both before and following the period of transducer measurements, using measurements independent of the transducer. In addition, and perhaps more important, some transducers have exhibited zero-point (zero submergence) drifts equivalent to a meter or more of water-level change. Therefore, apparent trends in water levels (beyond the earth-tide or other cyclic variations that occur within a few days or less) cannot be confirmed until the downhole measuring equipment is removed, checked, and recalibrated; and until prorated, adjusted values of the data have been computed and plotted for the period between calibrations. This process will be part of the continuing monitoring program, as described in Section 8.3.1.2.3.

Ground-water flow direction and travel time are, in part, a function of hydraulic gradient, which is determined from water levels in wells. At Yucca Mountain, high accuracy and precision are needed because in parts of the area (Figure 3-28) the water table is nearly flat and calculation of the gradient is sensitive to small errors in water-level measurements. For comparison of levels among wells, true depths must be calculated, which involves a correction for hole deviation from vertical. To calculate water-level altitudes, highly precise altitudes of the measuring points at land surface are needed.

A further complicating factor that may affect the water levels measured in the saturated zone relates to density effects. However, effects on water levels caused by water-density differences due to differences in ground-water temperatures among wells probably are small. In addition, since the water in most of the wells in the eastern and southeastern areas of Yucca Mountain have similar chemistries and temperatures, density differences probably have little if any effect on the hydraulic gradient within that area.

3.9.1.2 Potentiometric levels

This section provides data on matric potentials obtained from drillhole USW UZ-1 in the unsaturated zone. In addition, data on potentiometric levels from wells in the saturated zone are presented and discussed, as well as an assessment of the possibility of short-term changes in these levels.

3.9.1.2.1 Unsaturated zone

In situ measurements of matric potential within the unsaturated zone at Yucca Mountain have been attempted only at the 33 instrumentation stations that are arrayed vertically within the prototype monitoring borehole USW UZ-1. A preliminary analysis and interpretation of data collected in drillhole usw uz-1 over a 2-yr period following completion of the borehole are presented by Montazer et al. (1985). As described in Section 3.9.1.1.1, matric potentials in the borehole are being measured both by heat-dissipation probes and by thermocouple psychrometers. Profiles of matric potential within the borehole on a sequence of selected dates are shown in Figure 3-30 for the measurements made by heat-dissipation probes, and in Figure 3-31 for the thermocouple-psychrometer data (Montazer et al., 1985). Much of the data shown in Figure 3-31 exceed the useful range of matric potential over which the heat-dissipation probes were calibrated (-0.01 to -0.5 MPa). Some of the higher matric-potential data shown in Figure 3-30 exceed the maximum value, -0.5 MPa for which the thermocouple psychrometers can be calibrated reliably. Nonetheless, the data are useful to indicate the trends. The measurements of matric potential at each level within the borehole exhibited a trend during which, following the disturbance of natural conditions produced by drilling and instrumenting the borehole, the immediate environment of each instrument presumably equilibrated with the ambient hydrologic conditions within the surrounding host rock. In addition, because the heat-dissipation probes and the thermocouple psychrometers were emplaced under differing initial conditions in different backfill materials of different hydrologic properties, there is a considerable range of values of matric potential measured simultaneously by different instruments at the same level within the borehole. By May 1985, however, most of the instruments appeared to have stabilized sufficiently to yield a matric-potential profile that probably is indicative of ambient conditions within the host rock (Montazer et al., 1985).

Presuming conditions within the borehole to have equilibrated with those in the host rock, the data of Figures 3-30 and 3-31 indicate that matric potentials within the PTn hydrogeologic unit at drillhole USW UZ-1 range from -10 to -1 bars (-1 to -0.1 MPa) with a mean value of about -5 bars (-0.5 MPa) over the thickness of the unit. The matric potentials within the TSw unit range from -10 to -1 bars (-1.0 to -0.1 MPa), with an approximately constant mean value of about -3 bars (-0.3 MPa) over the thickness of the unit. Montazer et al. (1985) concluded that downward moisture flux within the TSw unit occurs predominantly as liquid-water flow within the rock matrix of the unit, with little or no liquid-water flow within the fractures. Assuming that the nearly constant matric potential over the depth interval from 122 to 224 m in drillhole USW UZ-1 (Figure 3-31) implied unit vertical hydraulic gradient, Montazer et al. (1985) estimated that the downward liquid-water flux within the TSw unit at drillhole USW UZ-1 probably is within the range from 0.1 to 0.5 mm/yr.

Pneumatic potentials (pore-gas pressures) are being monitored by down-hole pressure transducers emplaced in borehole USW UZ-1 and have been measured within a piezometer nest installed in borehole UE-25a#4, about 1.6 km southeast of drillhole USW UZ-1. Diurnal and barometrically induced fluctuations of gas pressure of about 0.25 kPa (Montazer et al., 1985) have been observed in these boreholes to depths of about 30 m. Such fluctuations are apparently damped out below this depth, although seasonally induced

Figure 3-30. Matric-potential profiles for selected dates at monitoring borehole USW UZ-1 based on data from heat-dissipation probes. Modified from Montazer et al. (1985).

Figure 3-31. Matric-potential profiles for selected dates at monitoring borehole USW UZ-1 based on data from thermocouple psychrometers. Modified from Montazer et al. (1985).

pressure variations are observed to occur at greater depths (Montazer et al., 1985). The pneumatic data were used by Montazer et al. (1985) to make preliminary estimates of air permeability in the uppermost units penetrated by these boreholes but have not been interpreted in terms of the overall bulk-gas flow system within the unsaturated zone.

3.9.1.2.2 Saturated zone

The present monitoring program consists of measuring water levels and observing their variations with time. Selected water levels in the vicinity of Yucca Mountain are shown in Figure 3-28 and Table 3-24. The figure and table are modified from Robison (1986), the most recent source of published data; however, the figure and table contain revisions based on resurveying in 1984 of land-surface altitudes at most of the well sites. The water levels range in altitude from about 1,030 to 730 m above sea level and generally represent water-table or unconfined conditions. Within the level of available precision, potentiometric levels show little variation with depth in the upper part of the saturated zone (Table 3-24). In the deeper part of the saturated zone (drillhole USW H-1, in the older Tertiary rocks; drillhole UE-25p#1, in the Paleozoic carbonate rocks), potentiometric heads are higher than in the overlying rocks (Table 3-24).

Figure 3-25 a simplified section through Yucca Mountain (location shown on Figure 3-28, shows the general relationship of stratigraphic units and the water table. For clarity, only the Solitario Canyon fault is shown, although the section is known to include numerous steeply dipping normal faults.

In the Yucca Mountain area, the potentiometric surface occurs principally in members of the Paintbrush Tuff and underlying Crater Flat Tuff (the geologic units, not the hydrostratigraphic units defined for the unsaturated hydrologic zone). Slope of the potentiometric surface at Yucca Mountain, based on gradients indicated by the contours, is south to southeast, similar to the regional topographic slope.

The gradient in the north is relatively steep compared with that in the south. Water-level altitudes in Figure 3-28 are highest to the north at drillholes USW G-2 and UE-25 WT#6; however, it is not known whether the southward slope toward drillhole USW H-1 is uniform, or if there are abrupt deflections. The cause for this steep slope is not known yet; it may be that southward movement of ground water is inhibited in this area by a low-permeability formation, such as the tuffaceous beds of Calico Hills, or by a fault or other unknown structural control. Test drilling to resolve these, and possible alternate, hypotheses is planned (Section 8.3.1.2.3).

West of the crest of Yucca Mountain, in drillholes USW H-6, USW WT-7, and USW WT-10 and also in USW H-5 on the crest, the water levels are about 775 m above sea level. Water levels are as much as 45 m lower at drillholes USW G-3 and USW H-3, just east of drillhole USW WT-7. A fault in Solitario Canyon may itself be poorly permeable and restrict eastward movement of ground water, or the fault may juxtapose permeable zones against less permeable zones and thereby restrict movement. Pumping tests in existing or

proposed drillholes are expected to be used to determine the hydraulic effects of the fault (Section 8.3.1.2.3.1).

From the eastern edge and southern end of Yucca Mountain to western Jackass Flats, the potentiometric surface ranges from about 728 to 730 m in altitude with a general southeastward slope (Figure 3-28, and Robison, 1986).

Hydrographs of water-level data are not presented, for the reasons given in Section 3.9.1.1.2. In regions where precipitation is greater and water levels are shallower, there is typically a well-defined seasonal or annual correlation between precipitation and ground-water levels. At Yucca Mountain, however, ground-water levels are not expected to show short-term or seasonal fluctuations that can be correlated with precipitation. This is because of extremely low rates of net infiltration and recharge attributable to local precipitation, the great depths to the water table, and ground-water travel times through the unsaturated zone that may be many thousands of years. Long-term precipitation data are not available for the site, although precipitation stations have been established recently in the Yucca Mountain area as part of a study to determine rainfall-infiltration relationships (Section 8.3.1.2.1).

Well J-13, 8 km from the site, and well J-12, 12 km from the site, (Figure 3-22) are in service for water supply; their production is discussed in Section 3.9.7.

3.9.1.3 Hydrochemistry

This section presents the hydrochemical information obtained from the baseline monitoring program of the site and its immediate surroundings. Hydrochemical data have not been collected at the site long enough for reliable time series to be developed. Therefore, temporal trends, if they exist, cannot be directly identified from the available data at this time. Some indirect information about changes in the hydrochemical conditions at the site over time can be obtained from understanding the hydrochemical evolution of the waters interacting with their host-rock environment. Their interaction is discussed in detail in Sections 4.1.1.4, 4.1.2.9, and 4.4.2. On the basis of this discussion, there is no reason to believe that while the existing climatic conditions prevail, the natural hydrochemical conditions at the site will not remain very similar to those currently observed.

Chapter 4 presents in detail the current knowledge of site ground-water temperature (4.1.2.8), major chemical composition (4.1.2.2), dissolved gas concentrations (4.1.2.5), pH and Eh (Tables 4-6 and 4-7 respectively), background radioactivity (4.1.2.6) and organic content (4.1.2.4), which will not be repeated here.

Major-ion data are available from six wells on, or immediately adjacent to the site. These wells tap the tuff formations. The data have been plotted on Stiff diagrams in Figure 3-32, which show that sodium bicarbonate water prevails throughout the area. This is consistent with the regional predominance of that facies as seen in the Piper diagram in Figure 3-12.

Figure 3-32. Geochemical facies in the tuff formation in the immediate area around Yucca Mountain, with Stiff diagrams indicating the composition of six samples

CONSULTATION DRAFT

Total dissolved solids and specific conductance are parameters measured to get a general idea of the salinity of a sample when complete analyses are not available. The total dissolved solids average about 300 mg/L and specific conductance is approximately 300 microsiemens/cm (at 25°C) for the ground water at the site. These parameters are not presented in detail since they contribute little when complete chemical analyses are available such as those in Tables 4-6 and 4-7 of Chapter 4. No density measurements of the local ground water are available; but because of the low solute concentration, values should vary insignificantly from 1 g/cm³ (at 4°C).

Environmental isotopic data are available for ground-water samples from wells in the tuff aquifers beneath Yucca Mountain (Benson et al., 1983; Benson and McKinley, 1985). Both carbon-14 activity and "uncorrected, apparent radiocarbon age" of the samples were reported. In areas where carbonate minerals may contribute significant amounts of carbon, it can be misleading to use these apparent ages in interpreting a flow system. Such apparent ages, which can be off by more than a factor of two, have been omitted in Table 3-25. Since corrected ages (determined by modeling the initial radiocarbon activity of the water; e.g., IAEA, 1983) are not available, only the sample activity measurements were used. The measured carbon-14 activities are relatively homogeneous for an area this size, particularly when accounting for sampling, sample preparation, and analytical uncertainties. Samples from the two southernmost drillholes (USW H-3 and USW H-4) have lower carbon-14 content (mean value 11.1 percent modern carbon) than samples from the other, more northerly drillholes (mean value 20.1 percent modern carbon) (Table 3-25). The significance of this difference cannot be fully assessed without carbon-14 modeling. Nevertheless, ground-water residence times of the southern samples may be on the order of 5,000 yr greater than the northern samples. Because significant amounts of carbonate minerals do not occur in the tuff or in tuffaceous valley fill with which these ground waters have come into contact, minimal radiocarbon dilution may have occurred from such contact and apparent ages may be more realistic than at other sites.

The δ carbon-13 ($\delta^{13}\text{C}$) values may show a spatial variability similar to those of the carbon-14. The values are consistent with recharge under very poorly vegetated conditions. The $\delta^{13}\text{C}$ value reported for sample USW H-3 is above the mean atmospheric value for carbon dioxide. The explanation of this unusual phenomenon could shed light on the differences in radiocarbon content between this sample and the northern ones. The $\delta^{13}\text{C}$ values below about -10 parts per thousand in dilute waters suggest that little carbonate rock/water and soil gas/water exchange has occurred (Section 3.7.3).

The three tritium values of <62 TU in Table 3-25 serve only to indicate that no large amounts of bomb-test tritium have reached the aquifer at those points. The other samples, which were analyzed by a more sensitive method to give a more precise result, are consistent with trace tritium contamination of samples of very old water (very old, in the context of tritium dating implies several hundred years or more). Such contamination is common and not unexpected at Yucca Mountain.

The deuterium and oxygen-18 data in Table 3-25 plot on and below the meteoric line (Section 3.9.1.3), occurring between a deuterium excess of +10 (mean continental meteoric water) to +5. Such a small deviation from the

Table 3-25. Environmental isotope data for ground-water samples from the tuff aquifers under the exploratory block and its immediate area^a

Well designation ^b	Collection date	δD (o/oo SMOW) ^c	$\delta^{18}O$ (o/oo SMOW) ^c	$\delta^{13}C$ (o/oo PDB) ^d	^{14}C (pmc) ^e	HTO (TU) ^f
UE-25b#1	08/07/81	-99.5	-13.4	-10.7	-- ^g	--
UE-25b#1	09/09/81	-101.0	-13.4	-10.7	16.7	<62.0
UE-25b#1	07/20/82	-99.5	-13.5	-8.6	18.9	0.6
USW G-4	12/09/82	-103.0	-13.8	-9.1	22.0	--
USW H-1	10/20/80	-103.0	-13.4	--	19.9	<6.0
USW H-1	12/08/80	-101.0	-13.5	-11.4	23.9	<6.0
USW H-3	03/14/84	-101.0	-13.9	-4.9	10.5	0.6
USW H-4	05/17/82	-104.0	-14.0	-7.4	11.8	<3.0
USW H-5	07/03/82	-102.0	-13.6	-10.3	18.2	<62.0
USW H-5	07/26/82	-102.0	-13.6	-10.3	21.4	<62.0

^aSource: Benson and McKinley (1985).

^bWell locations indicated in Figure 3-32.

^c δD and $\delta^{18}O$ are reported in parts per thousand relative to Standard Mean Ocean Water (SMOW) standard.

^d $\delta^{13}C$ is reported in parts per thousand relative to Pee Dee belemnite carbonate (PDB) standard.

^e ^{14}C activity is reported as a percent of the modern carbon (pmc) standard.

^fT is reported in tritium units (TU).

^g-- indicates no data.

world-wide average meteoric deuterium excess is within expected local continental variation. The water is extremely depleted in heavy isotopes of both oxygen and hydrogen, consistent with high altitude, cold (winter) continental recharge or snow-melt. No evaporative enrichment or thermal alteration is observed. No complete hydrochemical or isotopic analyses are available for water from the unsaturated zone at the site. Methods (triaxial compression, high-speed centrifuge and vacuum distillation) are currently under development for extracting uncontaminated samples upon which to perform these analyses. Yang (1986) has reported some preliminary calcium and sodium concentration data, showing that calcium is elevated in the pore water relative to the ground water, to the extent that it dominates over sodium. Complete chemical and isotopic characterization of the infiltrating pore water is needed (1) to develop an understanding of the hydrochemical nature of the water that may contact a waste package and (2) to isotopically trace the movement of these waters through the unsaturated zone. The plans for collecting these data as a function of depth are presented in Section 8.3.1.2.3.

The carbon-14 and carbon-13 composition of the unsaturated zone $\text{CO}_2(\text{g})$ phase has been determined by Yang et al. (1985a) on samples obtained by the new methods developed by Haas et al. (1983). Radiocarbon from bomb-test fallout was observed to a depth of 12.2 m in borehole UZ-1 on the exploration block. Radiocarbon activity decreased to below 100 percent modern carbon below 18.3 m, which may indicate that downward gas-phase transport from the surface has not occurred beyond this depth since the mid-1960. Improvements in sample collection methods and the interpretation of these results with pore-water hydrochemical data as outlined in Section 8.3.1.2.3 will enhance the understanding of two-phase transport in the unsaturated zone at the site.

3.9.2 HYDRAULIC CHARACTERISTICS

This section presents information on the hydraulic characteristics of the unsaturated and the saturated zones. The hydraulic characteristics of the unsaturated zone include fracture characteristics, porosity, saturated matrix hydraulic conductivity, and moisture characteristic relations. For the saturated zone, values for hydraulic conductivity, transmissivity, porosity, and storage coefficient are presented and discussed.

3.9.2.1 Hydraulic characteristics of the unsaturated zone

Mean values of hydrologic properties for most of the hydrogeologic units defined in Table 3-22 are presented in Table 3-26. The compilation of Montazer and Wilson (1984) included data from several sources. The data of Peters et al. (1984) were based on analyses of core samples from test wells USW G-1, USW G-4, and USW GU-3; the data of Tien et al. (1985) includes data from wells USW H-1, USW G-1, J-13, UE-25a#1, and USW GU-3; and the data reported by Weeks and Wilson (1984) were obtained from core samples from well USW H-1. The range of mean values among the references cited for each property within each hydrogeologic unit reflects the effects of lateral and vertical spatial heterogeneity within each unit. Table 3-26 provides only

Table 3-26. Summary of compilations of hydrogeologic properties of hydrogeologic units within the unsaturated zone, Yucca Mountain

Hydro-geologic unit ^a	Source of data	Range of thickness (m)	Grain density (kg/m ³) ^b	Fracture density (no./m ³) ^c	Porosity ^b	Saturated matrix hydraulic conductivity (m/s)
TCw	(d)	0-150	ND ^f	10-20	0.12	2×10^{-11}
	(e)	ND	2,490	ND	0.08	9.7×10^{-12}
PTn	(d)	20-100	ND	1	0.46	1×10^{-7}
	(e)	ND	2,350	ND	0.40	3.9×10^{-7}
TSw	(d)	290-360	ND	8-40	0.14	3.5×10^{-11}
	(e)	ND	2,580	ND	0.11	1.9×10^{-11}
CHnv	(d)	100-400	ND	2-3	0.37	5×10^{-8}
	(e)	ND	2,370	ND	0.46	2.7×10^{-7}
CHnz	(d)	100-400	ND	2-3	0.31	9×10^{-11}
	(e)	ND	2,230	ND	0.28	2.0×10^{-11}

^aHydrogeologic units are defined on Table 3-22.

^bSee Chapter 2 for more data.

^cScott et al. (1983).

^dMontazer and Wilson (1984).

^ePeters et al. (1984) and Peters et al. (1986).

^fND = no data.

values typical of the samples tested for hydrologic properties. These data are based, in general, on too few samples (several score for the entire unsaturated zone at Yucca Mountain) to permit meaningful statistical analyses to be performed for each hydrogeologic unit. An indication of the variance within and between sample sets is shown by the saturated matrix hydraulic conductivity, whose values listed in Table 3-26 range over two orders of magnitude.

The discrepancy in the values reported in Table 3-26 for saturated matrix hydraulic conductivity for hydrogeologic unit CHnv reflects, in part, the current uncertainty in the position of hydrogeologic unit contacts and, in part, the heterogeneity of the units. Rock-sample collection for the measurement of hydrologic properties is to be done through the program of surface-based borehole drilling and coring and of sampling within the

CONSULTATION DRAFT

exploratory shaft, as described in Section 8.3.1.4. Plans to define the hydrogeologic framework are described in Section 8.3.1.2.

The hydrologic property data of Table 3-26 must be supplemented by developing sets of moisture-characteristic curves, that is the functional dependence of liquid-water saturation and relative hydraulic conductivity on the liquid-water potential within the rock matrix and fractures appropriate for each hydrogeologic unit. In unfractured rocks, these relations refer to the storage and movement of liquid water within and through the interstitial pore space. In fractured rocks, allowance must be made for the storage and movement of water within the interconnected fracture openings as well as for the movement of water between the fracture openings and the rock-matrix pore space. Standard mercury-intrusion (Gregg and Sing, 1967), centrifugal (Russell and Richards, 1938; Hassler and Brunner, 1945), pressure plate (Hillel, 1980) and psychrometric (Papendick and Campbell, 1981) techniques are available for the laboratory determination of the relation between matric potential and saturation in soil and unfractured rock samples. Saturated hydraulic conductivities are measured using standard permeametry techniques (Hillel, 1980). The saturated matrix hydraulic conductivity measurements reported in Table 3-26 were generally made using a constant-head permeametry technique on samples from drillhole core.

Techniques for the measurement of relative hydraulic conductivity values are available (Amyx et al., 1960) but are generally limited in applicability to consolidated rocks with permeabilities much larger than the values for welded tuff or zeolitic tuff samples. Techniques for the measurement of relative hydraulic conductivity on very low permeability tuffs, such as welded tuff matrix material from TSW, are under development, and theoretically based methods have been formulated to estimate relative hydraulic conductivity in unconsolidated porous media from matric potential, saturation, and pore-size-distribution data (Brooks and Corey, 1964; Mualem, 1976).

A set of moisture-retention curves under drainage conditions relating matric-potential and saturation is shown in Figure 3-33 and was developed by Peters et al. (1984) for the matrix properties of most of the hydrogeologic units listed in Table 3-22. These curves were obtained by fitting the van Genuchten (1978) analytic representation to laboratory psychrometric data obtained for unfractured samples extracted from cores from test wells USW G-4 and USW GU-3. Because psychrometric techniques are appropriate only for matric potentials less than about -3 bars (-0.3 MPa), the moisture-retention curves reported by Peters et al. (1984) are not well determined for matric potentials that exceed this value. In lieu of direct measurements, the van Genuchten (1978) representation can also be used to estimate matrix relative hydraulic conductivity for these units. Such estimates must be regarded as highly tentative, however, because it is not known to what extent this analytic representation for relative hydraulic conductivity is appropriate for the tuffs at Yucca Mountain (Section 8.3.1.2.2). Furthermore, the curves in Figure 3-33 are based on laboratory determinations on small sample sets from only two locations, and therefore, the curves may not be representative of the units at Yucca Mountain as a whole.

Standard laboratory methods are not yet available by which to determine the moisture-characteristic relations for fractures and fractured rocks, and reliance must be made on theoretically based models and approximations.

Figure 3-33. Moisture-retention curves for the hydrogeologic units within the unsaturated zone at Yucca Mountain. Modified from Peters et al. (1984).

Liquid-water storage within fractures probably is insignificant, but the flow of liquid water within and across fractures is not yet well understood. Theoretical models for liquid-water flow in single fractures have been developed (Wang and Narasimhan, 1985, and Montazer and Harrold, 1985) but have not been field or laboratory tested. The intact-fracture test to be performed in the exploratory shaft at Yucca Mountain (Section 8.3.1.2.2.4) is designed to investigate in detail the mechanics of flow in a single fracture under variable conditions of saturation and stress. Fractures may or may not impede liquid-water flow at low matrix saturations, but longitudinal flow within the fractures may dominate liquid-water flow above some critical matrix saturation (Montazer and Wilson, 1984; and Klavetter and Peters, 1986). Consequently, at high matrix saturations the fracture systems within the densely welded, fractured hydrogeologic units (e.g., the TCw and TSw units) and within the fault zones may become highly efficient pathways for liquid-water flow. Liquid-water flow within the fractures may be Darcian with an appropriately defined potential and hydraulic conductivity (for example, under the parallel-plate, cubic-law approximation for flow in single fractures), but, as pointed out previously, the potential in the fractures need not be equal to the matrix potential in the rock matrix, especially under highly transient conditions (Klavetter and Peters, 1986).

Considerable attention must be given to the problem of how best to represent the hydrologic properties and hydrologic response of a highly fractured porous medium. Separate treatment of the matrix and the individual fractures is not practical. In the first place, no way is known to generate a complete set of fracture location and geometry data. Secondly, if the Topopah Spring welded unit has a mean fracture density of 20 fractures/m³ and has a mean thickness of 300 m over the approximately 7 x 10⁶ m² area of the central Yucca Mountain block, then one would have to consider flow in approximately 4 x 10¹⁰ discrete fractures. Consequently, the practical alternative is to regard the matrix and fractures as constituting either separate but overlapping continuum systems or as a single composite continuum system. These approaches assume that the matrix and fracture properties can be represented as spatial averages over rock mass volumes whose linear dimensions are very much smaller than the thickness of the hydrogeologic unit but sufficiently large to include a representative, statistical sample of hydraulically connected fractures. The rock mass volume over which the averaging is performed is commonly designated as a representative elementary volume (REV) (Bear, 1972). The macroscopic continuum approach for fractures and rock matrix has been examined theoretically by Klavetter and Peters (1986), who conclude that it appears to be applicable to the unsaturated, fractured tuffs at Yucca Mountain. Specifically, they consider a "composite-porosity" model which describes fluid movement in a single continuum composed of both matrix material and fractures. This approach would not be applicable to sparsely fractured media or to media in which the mean fracture apertures exceeded a few millimeters and, thus, for which the medium would cease to be approximated by a bundle of capillary tubes. One principal objective of the bulk-permeability test, which is to be performed within the exploratory shaft as described in Section 8.3.1.2.2.4, is to field-test the REV hypothesis as it may apply to the fractured Topopah Spring welded unit at the potential repository horizon.

Preliminary results based on a dual-porosity, porous-medium-equivalent representation for fractured hydrogeologic units have been reported by

Montazer and Wilson (1984), Wang and Narasimham (1985), and Klavetter and Peters (1986). The dependence of relative hydraulic conductivity on liquid-water potential in a fractured porous medium under the composite dual-porosity continuum hypothesis may be expected to exhibit the qualitative appearance shown in Figure 3-34. At low matrix saturations, little or no water moves longitudinally within the fracture openings, and the effective hydraulic conductivity is controlled by that of the fracture-bounded matrix blocks. As the matrix approaches complete saturation, however, the movement of water within and along the fracture aperture rapidly becomes more efficient so that at complete saturation the fractures may be dominant contributors to the net hydraulic conductivity. The relative contributions of fractures and matrix to the net effective hydraulic conductivity depend on the fracture frequency, aperture-size distribution, and degree of interconnectivity. The hydrogeologic unit TSw exemplifies the extremes that may be encountered. This unit has a matrix saturated hydraulic conductivity of about 10^{-11} m/s and a fracture density of 8 to 40 fractures/m³ (Table 3-26). In well J-13, the Topopah Spring unit is fully saturated. It is the major source of water to the well and has an estimated net bulk, presumably fracture-controlled, hydraulic conductivity of about 8×10^{-6} m/s (Thordarson, 1983).

A further complication arises inasmuch as the hydrogeologic units may be anisotropic with respect to intrinsic permeability and thus to hydraulic conductivity. Data are presently insufficient to perform an adequate assessment. Because chemical alteration can be expected to destroy preferred orientations of rock properties, the matrix properties of the altered, zeolitized CHnz unit probably are largely isotropic. The fracture and fault systems within the densely welded units, however, probably introduce an inherent anisotropy wherever they are present and contribute significantly to moisture or pore-gas flow. Two principal fracture sets have been identified within the Yucca Mountain block (Scott et al., 1983). One set strikes north-northwest and the other strikes north-northeast, and both fracture sets exhibit steep to vertical dips. Most of the faults bounding and within the Yucca Mountain block are typical Basin and Range style high-angle normal faults that dip to the west, strike to the north, and exhibit small individual displacements (2 to 5 m). The Solitario Canyon fault, which bounds Yucca Mountain on the west, is a northward-striking high-angle normal fault that dips to the west and has a displacement ranging from 20 to 200 m along its trace. The Ghost Dance fault within the Yucca Mountain block is likewise a west-dipping, north-striking normal fault with a displacement of about 25 m. These faults and their associated fracture zones in the more competent hydrogeologic units probably introduce a fundamental preferential directional control on moisture movement, whether these fault zones act as conduits for or barriers to flow. Quantitative data by which to characterize rock-matrix and fracture-induced anisotropy is currently lacking but will be examined by field testing in surface-based boreholes and in the exploratory shaft and by laboratory measurements on cores and on rock samples obtained during excavation of the exploratory shaft facility, as described in Section 8.3.1.2.

Available data on the magnitude and distribution of moisture content within the unsaturated zone are scanty and incomplete. Drill cuttings were collected continually and their gravimetric moisture contents measured during the drilling of test boreholes USW UZ-1 and USW UZ-6 (Whitfield, 1985). The resulting moisture-content profiles are shown in Figure 3-35 for drillhole

Figure 3-34. Idealized hydraulic-conductivity characteristic curve for a composite (fracture-matrix) porous medium. Modified from Montazer and Wilson (1984).

Figure 3-35. Moisture content of drill cuttings from borehole USW UZ-1. Modified from Whitfield (1985).

USW UZ-1 and in Figure 3-36 for drillhole USW UZ-6. The two profiles are similar in that they indicate relatively high moisture contents in the non-welded tuff intervals and lower moisture contents in the Tiva Canyon and Topopah Springs welded units. Both profiles exhibit some degree of irregularity or scatter. The scatter within a given profile may be due to (1) inherent variations in moisture content or (2) undetermined experimental errors. Differences between the two profiles may also be attributed to these two causes. The data indicate, for example, that the moisture content within the Paintbrush nonwelded (PTn) hydrogeologic unit has a representative mean value in the two wells of about 0.2 kg/kg, which, assuming a mean porosity of 0.43 and a grain density of about 2,400 kg/m³ (Table 3-26, Section 3.9.2.1), yields a mean saturation of about 0.6 for this unit. The interval of the Tiva Canyon welded (TCw) unit penetrated in drillhole USW UZ-6 has a mean moisture content of about 0.025 kg/kg, which, with a mean porosity of 0.1 and a mean grain density of about 2,500 kg/m³ (Table 3-26), yields a mean saturation for this unit of about 0.6. The Topopah Spring welded (TSw) hydrogeologic unit has a mean moisture content of about 0.025 kg/kg in drillhole USW UZ-1 and about 0.02 kg/kg in drillhole USW UZ-6. Using a mean porosity of 0.12 and a grain density of about 2,600 kg/m³ for this unit (Table 3-26), a mean saturation is obtained for the TSw unit of about 0.5 in drillhole USW UZ-1 and about 0.4 in drillhole USW UZ-6. These very tentative results generally are consistent with the ambient mean saturations of 0.61 for the PTn unit, 0.67 for the TCw unit, and 0.65 for the TSw unit reported by Montazer and Wilson (1984).

Using the moisture-retention curves shown in Figure 3-33, these estimates of in situ saturations may be converted into corresponding matric potentials. For example, if the saturation in the PTn unit is 0.6, the curve in Figure 3-33 for this unit implies a matric potential of about -0.75 Mpa, a value that is compatible with the range of matric potentials for this unit implied by the matric-potential data obtained to date from drillhole USW UZ-1 (Figures 3-30 and 3-31). However, if the saturation in the TSw unit is 0.5, the curve for this unit in Figure 3-33 implies a corresponding matric potential of about -7 MPa, a value quite different from the mean of about -0.3 MPa for the TSw unit implied by from the matric-potential data in drillhole USW UZ-1 (Figures 3-30 and 3-31). This apparent discrepancy emphasizes the need for continued monitoring of ambient hydrologic conditions within the unsaturated zone at Yucca Mountain, as well as for the direct laboratory measurement of the hydrologic properties for statistically large samples of cores, drill cuttings, and bulk-rock samples taken from within each of the hydrogeologic units at numerous locations within Yucca Mountain.

Under transient moisture-flow conditions, moisture may be taken into or released from storage within the system, and parameters describing the moisture-storage properties of the system must be supplied as part of the hydrologic properties. Storativity is defined by Bear (1972) to be the change in volume of liquid water storage per unit volume of bulk rock mass upon unit change of liquid-water potential. The rate at which the quantity of water held in storage changes with time is equal to the product of storativity and the time rate of change of liquid-water potential. In partially saturated media, storativity depends upon (1) the relations between ambient matrix and fracture saturations and the liquid-water potential in the matrix and fractures; (2) the liquid-water compressibility; and (3) the effective bulk compressibility appropriate to the rock matrix and the fractures.

Figure 3-36. Moisture content of drill cuttings from borehole USW UZ-6. Modified from Whitfield (1985).

Consequently, specification of moisture storage and movement within the unsaturated zone requires the availability of hydro-mechanical-thermal properties and their spatial variation for each hydrogeologic unit. These data are to be determined through field, laboratory, and theoretical studies described in Sections 8.3.1.2 and 8.3.1.15. Thermomechanical data presently available for the rocks composing the hydrogeologic units at Yucca Mountain have been compiled and summarized in Section 2.4.

Using approximate values for the rock-matrix and fracture mechanical properties, Klavetter and Peters (1986) have performed a preliminary analysis of capacitance (conceptually similar to storativity but with different dimensions) for the hydrogeologic units in the unsaturated zone at Yucca Mountain. They regarded each fractured unit to be represented as a composite porous medium in which the matric potential in the fractures is equal to that in the rock matrix. Their results, calculated for the TSw unit, are shown in Figure 3-37, which shows the contributions to the total capacitance due to a number of independent storage mechanisms. Under the assumed conditions, the results plotted in Figure 3-37, which are typical of those for the other hydrogeologic units, indicate that changing fracture and rock-matrix saturations dominate storage capacity with decreasing matric potential while the bulk rock-matrix compressibility becomes the dominant contributor near complete saturation.

Fundamental hydrologic-property data for the bulk-gas flow system within the unsaturated zone are presently lacking. Highly preliminary estimates of in situ air permeability in the TCw and PIn units at boreholes USW UZ-1 and USW UE-25a#4 have been reported by Montazer et al. (1985). Downhole pore-gas pressures are being monitored at regular intervals in borehole USW UZ-1 (Montazer et al., 1985; see also Section 3.9.1.1.1), and the results of pore-gas sampling and chemical analyses have been reported by Yang et al. (1985a) (Section 3.9.1.3). An extensive program of gas-phase monitoring, chemical sampling, gas-tracer studies and pneumatic-property determinations is planned as part of the surface-based, exploratory shaft, and laboratory investigations of the unsaturated zone and are described in Section 8.3.1.2.2.

3.9.2.2 Hydraulic characteristics of the saturated zone

This section presents values for and discusses the following hydraulic characteristics for each hydrogeologic unit within the saturated zone: hydraulic conductivity, transmissivity, porosity, and storage coefficient.

3.9.2.2.1 Hydraulic conductivity and fractures

Analytical approaches to the determination of hydraulic conductivity of rocks in the saturated zone for the Yucca Mountain site have involved variations of Theis equations, which are based on Darcian flow, and assumptions of nonsteady discharge of water from wells. The specific or detailed nature of data from each test hole has varied; pumping responses have varied because of variations of hydraulic conditions among well sites; and conceptual ideas of the nature of the systems have evolved as experience

Figure 3-37. Computed capacitance coefficient versus pressure head for the Topopah Spring welded (TSw) hydrogeologic unit. Modified from Klavetter and Peters (1986).

CONSULTATION DRAFT

has been gained, although analysis of the data for each well site has been based on assumptions that are generally common to each:

1. The tested rock has primary matrix porosity and is homogeneous and isotropic.
2. The secondary porosity is controlled by fractures. The fractures generally are vertical or very steep within the saturated zone. The volume of water stored in fractures is relatively small in comparison to that stored in matrix pores.
3. The flow directly into the well is through fractures only, but within the formation there is flow from matrix pores to the fractures. Hydraulic conductivity of fractures is probably several orders of magnitude greater than the conductivity of the matrix.
4. A sufficient number of fractures has been intercepted by the pumped well to yield representative responses.
5. The distances between fractures are small in comparison with the dimensions of the ground-water system being considered.
6. The orientation of fractures is assumed to be random, on a large scale, and the system appears to be laterally isotropic (Craig and Robison, 1984; Lahoud et al., 1984; Rush et al., 1984).

The above is a double-porosity concept; however, the parameters necessary to analyze fracture aperture, spacing, and continuity were not available, and porous-media solutions were used in calculations of bulk values of in situ transmissivity and hydraulic conductivity shown in Table 3-27 (in which transmissivity for individual stratigraphic units ranges from 0.001 to 276 m²/day, and hydraulic conductivity ranges from 2×10^{-6} to 2.3 m/day). Where appropriate, data will be reevaluated by alternative approaches, such as those based on fracture-flow concepts (Section 8.3.1.2.3).

In situ values of parameters such as transmissivity or hydraulic conductivity were determined from single-hole tests, the limitations of which are recognized; observation holes were not used because none that were deep enough were also close enough that they could be expected to show responses that would be separable from normal background variations due to earth tides and barometric affects. However, Moench (1984) (1) evaluated responses from an observation hole, 100 m from UE-25b#1 (he used a double-porosity conceptual model), (2) concluded that fracture skin may be important, and (3) calculated fracture system hydraulic conductivity to be about 10^{-5} m/s and block system hydraulic conductivity to be about 2×10^{-6} m/s). Also, as the precision of water-level measurements is increased (Section 3.9.1.1.2), aquifer tests that include responses in observation holes at substantial distances from the pumped well may be feasible (Section 8.3.1.2.3).

In the single-well tests the general approach to data collection was the same for each hole. It included measurement of drawdown and recovery of water levels due to pumping from the entire hole (or a major interval in it); a borehole-flow survey while pumping, to define individual productive zones;

Table 3-27. Preliminary^a summary of hydrologic characteristics of major stratigraphic units in the vicinity of Yucca Mountain (page 1 of 2)

Stratigraphic unit	Typical character	In situ (field) analyses					Laboratory analyses (cores)						
		Saturated thickness (m)	Transmissivity ^b (m ² /d)	Average hydraulic conductivity ^c (m/d)	Well or hole tested	Reference ^d	Saturated matrix hydraulic conductivity		Matrix porosity ^e		Well or hole analyzed	Reference ^d	
							m/d	No. of samples	%	No. of samples			
Topopah Spring Member	Moderately to densely welded tuff	167	120	0.7	J-13	7	3x10 ⁻⁷ to 7x10 ⁻⁷ to 2x10 ⁻⁴ to 5x10 ⁻⁴	5 to 18 to 1	4 - 33 to 6 - 30 to 12	5 to 24 to 1	J-13 UE-25a#1 UE-25b#1	7 to 1 to 4	
Tuffaceous beds of Calico Hills	Zeolitized, nonwelded tuff, vitric tuff	148	(82)	0.5	UE-25b#1	4	4x10 ⁻⁸ to 3x10 ⁻⁴	6	20 - 34	7	UE-25a#1	1	
Prow Pass Member of Crater Flat Tuff	Nonwelded to moderately welded tuff	116 135 174 150 111	167 150 36 - 142 ^f (85) 14	1.44 1.1 0.2-0.8 ^f 0.4 0.1	USW H-1 USW H-1 USW H-4 UE-25b#1 UE-25p#1	2 6 9 4 3	6x10 ⁻⁵ to 2x10 ⁻⁵ to 1x10 ⁻⁴ to 1x10 ⁻³ to 6x10 ⁻⁷ to 1x10 ⁻³	3 to 8 to 5	28 - 29 to 10 - 25 to 17 - 30	3 to 12 to 18	USW H-1 UE-25a#1 USW G-4	6 to 1 to 5	
Bullfrog Member of Crater Flat Tuff	Nonwelded to densely welded tuff	125 119 159 132	0.8 70 - 276 ^f (85) (7)	0.006 ^f 0.6-2.3 ^f 0.4 0.05	USW H-1 USW H-4 UE-25b#1 UE-25p#1	6 9 4 3	3x10 ⁻⁵ to 2x10 ⁻⁴ to 1x10 ⁻³ to 1x10 ⁻³ to 2x10 ⁻⁴ to 5x10 ⁻⁴	10 to 3 to 2	19 - 34 to 17 - 34 to 24 - 27	9 to 3 to 6	USW H-1 UE-25a#1 USW G-4	6 to 1 to 5	
Tram Member of Crater Flat Tuff	Nonwelded to moderately welded ash-flow and bedded tuffs	284 354 352 183	2x10 ⁻³ 0.7 70 - 276 ^f (3.3)	7x10 ⁻⁶ 0.002 ^f 0.2-0.8 ^f 0.02	USW H-1 USW H-3 USW H-4 UE-25p#1	6 8 9 3	4x10 ⁻⁶ to 4x10 ⁻⁴	9	18 - 26	9	USW H-1	6	
Lithic Ridge Tuff and older tuffs	Partially welded ash-fall tuffs	594 110 371	0.001 0.1 >10	2x10 ⁻⁶ 1x10 ⁻³ >0.03	USW H-1 USW H-3 UE-25p#1	6 8 3	6x10 ⁻⁵ to 3x10 ⁻⁴	2	9 - 17	2	USW H-1	6	

Table 3-27. Preliminary^a summary of hydrologic characteristics of major stratigraphic units in the vicinity of Yucca Mountain (page 2 of 2)

Stratigraphic unit	Typical character	In situ (field) analyses					Laboratory analyses (cores)					Refer- ence ^d
		Saturated thickness (m)	Transmis- sivity ^b (m ² /d)	Average hydraulic conductivity ^c (m/d)	Well or hole tested	Refer- ence ^d	Saturated matrix hydraulic conductivity		Matrix porosity ^e		Well or hole anal- yzed	
							m/d	No. of samples	%	No. of samples		
Lone Moun- tain Dolo- mite and Roberts Mountain Formation	Carbonate rocks Paleo- zoic age (lower) carbonate aquifer of Winograd and Thordarson, 1975)	>561	108	0.2	UE-25p#1	3						

^aInterpretive analyses from in situ testing at some drillholes are not completed yet, including those from drillholes USW G-4, USW H-6, and UE-25c#1,2,3.

^bDetermined from pumping tests, borehole-flow surveys, and slug-injection tests; parentheses indicate approximate value because reported values reflect more than one stratigraphic unit. See references cited for details of individual drillholes tested.

^cObtained by dividing transmissivity by saturated thickness, which in some cases may not agree with values reported in the cited reference. Productive zones are typically thin, fractured intervals rather than a generally-uniform rock matrix; therefore, the porous-media concept of hydraulic conductivity is not necessarily appropriate.

^d1, Anderson (1981); 2, Barr (1985); 3, Craig and Robison (1984); 4, Lahoud et al. (1984); 5, Peters et al. (1984); 6, Rush et al. (1984); 7, Thordarson (1983); 8, Thordarson et al. (1984); 9, Whitfield et al. (1985).

^eChapter 2 - Geoengineering.

^fLower value based on straight-line method of pumping-test analysis; higher value based on Theis-recovery method (see Whitfield et al. (1985)).

and measurement of head changes of selected (straddled) intervals due to removal (by swabbing) or injection (falling head) of water.

Claassen and White (1979) studied the possible extent of fracture control on the Rainier Mesa ground-water system in the northern part of the NTS; they concluded that the hydraulic characteristics of the Paintbrush Tuff are controlled by fracture porosity. Blankennagel and Weir (1973) observed, through extensive hydrologic testing of Pahute Mesa, that welded tuff and rhyolite have larger bulk hydraulic conductivities than nonwelded or bedded tuff because the former are brittle and fracture readily.

Interconnection of some fractures in welded tuffs beneath Pahute Mesa and Yucca Mountain was demonstrated by hydrologic tests in holes that included slug-injection or withdrawals of water of isolated depth intervals, pumping tests, and borehole-flow surveys to determine intervals that yield water. The volume of water removed during pumping tests is many times greater than that expected from fractures near the drill holes (Blankennagel and Weir, 1973; Rush et al., 1984), which may indicate that these fractures are connected to more distant ones, although there may also be some contribution from the rock matrix.

Hydraulic conductivity of tuffs may decrease with increasing depth; welded tuffs are generally less conductive than rhyolites at the same depth (Waddell et al., 1984). At greater depths, formation of minerals in fractures (commonly clays, silica, calcite, zeolites, or iron and manganese oxides) or closing of fractures because of greater lithostatic stresses tend to decrease the aperture of fractures. Rush et al. (1984) show that, in drillhole USW H-1, bulk hydraulic conductivity (determined from pumping tests) is substantially greater than matrix hydraulic conductivity (determined from cores) for the shallower rocks, but at greater depths the bulk and matrix hydraulic conductivity are approximately equivalent. They concluded that there may be no fracture flow below a depth of 790 m in rocks penetrated by drillhole USW H-1 because the greater overburden pressure decreases fracture porosity. It is not known if these conditions prevail throughout the Yucca Mountain area.

Controls on fracture density are not clear. Besides being a function of rock type, fracture density is a function of structural setting. A zone of increased fracturing and faulting on the southeastern and eastern sides of Yucca Mountain has been mapped (Scott and Bonk, 1984). Another feature evident from surface mapping is that fracture density decreases in the northern part of Yucca Mountain, where displacement and number of faults are less than in the southern part. Fracture density is presumed to increase near faults of large displacement, and a corresponding increase in local bulk hydraulic conductivity could be expected. Plans to examine this are presented in Section 8.3.1.2.3.

The probable importance of flow in fractures in the saturated zone is indicated by two lines of evidence. First, hydraulic conductivity of the matrix, as measured on cores and by slug-injection tests of nonproductive parts of drillholes, is much lower than in situ conductivity measurements in productive parts of drillholes (Rush et al., 1984). Second, productive zones, as determined by borehole-flow and temperature surveys performed during pumping tests, correlate with fracture zones determined from televi-

CONSULTATION DRAFT

and caliper logs (Lahoud et al., 1984; Rush et al., 1984). Preliminary indications from studies in progress at test holes UE-25c#1, 2, and 3 indicate that hydraulic conductivities of fractures are several orders of magnitude greater than matrix conductivities (Waddell et al., 1984; SCP Section 8.3.1.2.3).

Two sets of faults and fractures are present at Yucca Mountain (Scott et al., 1983). The first set strikes north-northwest and dips steeply to the west. The second strikes north-northeast and also dips steeply to the west. The minimum compressive stress is oriented N. 50 to 60 degrees W., which is coincident with the regional direction of tectonic extension (Carr, 1974). Fractures with these orientations may tend to be closed by tectonic stresses, whereas fractures with orientations of N. 30 to 40 degrees E. may tend to be more open. Therefore, the system may not be hydraulically isotropic because the north-northeast-striking fracture set may be more transmissive. Multiple-well hydraulic tests in the vicinity of drillhole UE-25c#1 are planned to test the hypothesis that fracture orientation may tend to control the direction or rate of ground-water flow (Section 8.3.1.2.3).

Using fracture frequency and orientation data from drillhole USW H-4, Erickson and Waddell (1985) calculated the probable shape and orientation of the hydraulic-conductivity ellipsoid at that site; they determined that the plane containing the two larger principal axes strikes northeast and is nearly vertical. The hydraulic conductivity in this plane is 5 to 7 times that of the smallest principal hydraulic conductivity.

Most fractures are steeply dipping, and vertical permeability within a single cooling unit may be approximately the same as maximum horizontal permeability within a single cooling unit. However, because of nearly horizontal layering in ash-flow tuffs, due to emplacement and cooling mechanics, and because of the presence of bedded, nonwelded tuffs, vertical variations in fracture frequency probably occur (Waddell et al., 1984). Some indication of vertical hydraulic conductivity may be obtainable from the fracture-orientation testing in the vicinity of drillhole UE-25c#1 (Section 8.3.1.2.3).

Vertical and horizontal interconnectivity of fractures was demonstrated by a pumping test of drillhole UE-25b#1, which was pumped for 29 days at about 12 L/s from the interval 853 to 914 m (lower Bullfrog to upper Tram Member). The results of a similar borehole-flow survey are shown on Figure 3-38. A sodium bromide tracer was placed in drillhole UE-25a#1, which is 107 m away from drillhole UE-25b#1 and is open to the formation from the water level (470 m) to total depth of 762 m (tuffaceous beds of Calico Hills to upper Bullfrog Member). Breakthrough of the tracer in UE-25b#1 occurred in two days, and above-background concentrations continued for the next 26 days of pumping (Waddell, 1984; Waddell et al., 1984).

Knowledge of the role of fractures in controlling paths and velocities of ground-water flow at Yucca Mountain is insufficient to characterize the flow system adequately. An understanding is also needed of the applicability of the concepts of porous-media equivalents for describing the flow system. Improved understanding of these factors is expected to come from planned tracer tests at Yucca Mountain (Section 8.3.1.2.3.1). In addition, other methods of data analysis will be evaluated (Section 8.3.1.2.3.1).

Figure 3-38. Borehole-flow survey of drillhole UE-25b#1 (when hole was at total depth of 1,220 m). Modified from Lahoud et al. (1984).

3.9.2.2.2 Transmissivity and hydraulic conductivity

Beneath Yucca Mountain, the Topopah Spring Member is unsaturated, but at well J-13 it is saturated (Figure 3-25), and most water production from the well is from the Topopah Spring Member (Young, 1972). Thordarson (1983) analyzed pumping tests of well J-13 and concluded that the Topopah Spring Member has a transmissivity of $120 \text{ m}^2/\text{day}$ and a hydraulic conductivity of 1.0 m/day .

Drillhole USW H-3 is located on the western edge of the repository block (Figure 3-28) in an area relatively undisturbed by fracturing and faulting, and where the Topopah Spring Member, tuffaceous beds of Calico Hills, Prow Pass Member, and most of the Bullfrog Member are unsaturated. Total transmissivity at the site is only about $1 \text{ m}^2/\text{d}$, virtually all of it from the Tram Member (Thordarson et al., 1985).

Borehole-flow surveys were made in each hole in which pumping tests were conducted to determine the depth intervals that are productive. Figure 3-38 is an example of a borehole-flow survey, this one from drillhole UE-25b#1, in which drawdown curves due to pumping were used to calculate a transmissivity of $340 \text{ m}^2/\text{d}$ (Lahoud et al., 1984). Surveys from other holes include drillhole USW H-1, (Rush et al., 1984); drillholes USW H-3, (Thordarson et al., 1984); USW H-4, (Whitfield et al., 1985); USW H-5, (Bentley et al., 1983); and UE-25p#1, (Craig and Robison, 1985).

Data from in situ tests can be compared with those from laboratory tests. For example, Table 3-28, which shows data from drillhole USW H-1 (Rush et al., 1984), includes transmissivity and hydraulic conductivity, based on pumping tests of major intervals, borehole-flow surveys, and slug-injection tests of selected intervals (straddles). In Table 3-29 these values are compared with matrix hydraulic conductivity of cores from the saturated zone in drillhole USW H-1, as determined in the laboratory (Rush et al., 1984). That in situ hydraulic conductivity is generally greater in the shallower units, which are observed to be more highly fractured suggests that fractures transmit more water than does the rock matrix.

Stratigraphic controls on the distribution of permeability cannot be readily established or predicted with confidence for untested areas; this heterogeneity can be seen in Figure 3-39, which shows permeable zones for selected wells. Fractures contribute more to the permeability of test holes than does the rock matrix; therefore, productive intervals are determined by the depths at which holes intercept transmissive fractures. The dip of most fractures is very steep, and drillholes are nearly vertical; therefore, a statistically significant number of fractures is not likely to be intercepted by a vertical drilling program of reasonable design. This bias will be partially ameliorated through data gathered from horizontal boreholes extending off the exploratory shaft.

In the following several paragraphs, the saturated hydraulic characteristics of each major stratigraphic unit beneath the Yucca Mountain site is discussed.

Table 3-28. Transmissivity and average hydraulic conductivity at well USW H-1, based on pumping tests, borehole-flow surveys, and slug-injection tests^a

Depth interval (m)	Transmissivity (m^2/d)	Horizontal average hydraulic conductivity (m/d)
PROW PASS MEMBER		
572-597	74	3
597-616	18	1
616-652	40	1
652-653	18	18
653-688	≤ 1	$\leq 3 \times 10^{-2}$
687-694	1×10^{-1}	1×10^{-2}
687-697	1×10^{-1}	1×10^{-2}
PROW PASS AND BULLFROG MEMBERS		
694-736	1×10^{-2}	$\leq 2 \times 10^{-4}$
BULLFROG MEMBER		
736-741	6×10^{-1}	1×10^{-1}
741-758	2×10^{-1}	1×10^{-2}
758-792	1×10^{-2}	3×10^{-4}
792-811	$\leq 3 \times 10^{-3}$	$\leq 2 \times 10^{-4}$
BULLFROG AND TRAM MEMBERS		
811-926 ^b	5×10^{-3}	4×10^{-5}
TRAM MEMBER AND FLOW BRECCIA		
926-1,200 ^b	$\leq 2 \times 10^{-3}$	$\leq 7 \times 10^{-6}$

CONSULTATION DRAFT

Table 3-28. Transmissivity and average hydraulic conductivity at well USW H-1, based on pumping tests, borehole-flow surveys, and slug-injection tests^a (continued)

Depth interval (m)	Transmissivity (m ² /d)	Horizontal average hydraulic conductivity (m/d)
FLOW BRECCIA AND LITHIC RIDGE TUFF		
1,200-1,407 ^b	$\leq 2 \times 10^{-3}$	$\leq 1 \times 10^{-5}$
LITHIC RIDGE TUFF AND OLDER TUFFS		
1,407-1,621 ^b	8×10^{-4}	4×10^{-6}
OLDER TUFFS		
1,621-1,829 ^b	2×10^{-4}	1×10^{-6}

^aSource: Rush et al. (1984).

^bValues for interval computed by difference from overlapping injection-test intervals.

Table 3-29. Comparison of hydraulic-conductivity values from in situ tests with core-sample analyses

Stratigraphic unit	Selected depth interval from Table 3-28 (m)	Average hydraulic conductivity (m/d)	
		Total, from in situ tests (Table 3-28)	Matrix, from core analyses ^a
Prow Pass Member	616-652	1	^b 8×10^{-5}
Prow Pass and Bullfrog members	694-736	$< 2 \times 10^{-4}$	^c 3×10^{-4}
Bullfrog Member	758-792	3×10^{-4}	^d 4×10^{-4}

Table 3-29. Comparison of hydraulic-conductivity values from in situ tests with core-sample analyses (continued)

Stratigraphic unit	Selected depth interval from Table 3-28 (m)	Average hydraulic conductivity (m/d)	
		Total, from in situ tests (Table 3-28)	Matrix, from core analyses ^a
Bullfrog and Tram members	811-926	4×10^{-5}	^c 4×10^{-5}
Tram Member and flow breccia	926-1200	$<7 \times 10^{-6}$	^e 2×10^{-4}

^aCore-sample analyses from Rush et al. (1984).

^bAverage for three core-sample analyses.

^cAverage for four core-sample analyses.

^dAverage for five core-sample analyses.

^eAverage for six core-sample analyses.

The Topopah Spring Member of the Paintbrush Tuff which has saturation only east of the repository block (Figure 3-25) has a high hydraulic conductivity. Many fractures have been observed in cores and in television surveys; lithophysal cavities are common in some zones (Scott et al., 1983). Problems related to loss of drilling fluids while penetrating the Topopah Spring Member are common. Matrix hydraulic conductivity of the unit is small. Thordarson (1983) reported hydraulic conductivities from cores at well J-13 that range from 2×10^{-4} to 8×10^{-7} m/d; however, the Topopah Spring Member is fractured, has an average in situ hydraulic conductivity of 0.7 m/d, and is the most productive unit penetrated by the well.

The tuffaceous beds of Calico Hills are above the water table in most hydrologic test holes near Yucca Mountain. In drillhole UE-25p#1 the unit is mostly saturated but poorly permeable; during a pumping test, a borehole-flow survey indicated the unit yielded less than 2 percent of the total amount pumped. Hydraulic conductivity of the unit was not separately determined by straddle tests; however, Craig and Robison (1984) concluded that the 41 m of saturated Calico Hills penetrated by UE-25p#1 has a transmissivity of about $0.5 \text{ m}^2/\text{day}$. In contrast in drillhole UE-25b#1, the tuffaceous beds of Calico Hills yielded 32 percent of the production (Figure 3-38) during pumping of the entire hole (Lahoud et al., 1984). In the vicinity of UE-25b#1, the unit is zeolitized and would have low matrix hydraulic conductivity; therefore, the relatively high values determined from in situ tests (Table 3-27) probably are due to nearby normal faults or shear fractures (Waddell et al., 1984).

FORMATION	WELL NUMBER						
	USW H-6	USW H-5	USW H-1	USW G-4	USW H-4	UE- 25b#1	J- 13
TOPOPAH SPRING MEMBER OF PAINTBRUSH TUFF							WT
TUFFACEOUS BEDS OF CALICO HILLS			WT	WT		WT
PROW PASS MEMBER OF CRATER FLAT TUFF	WT	WT			WT
BULLFROG MEMBER OF CRATER FLAT TUFF
TRAM MEMBER OF CRATER FLAT TUFF				TD		TD
LAVA	TD	TD
TUFF OF LITHIC RIDGE			TD ↓		TD		TD
WT = WATER TABLE TD = TOTAL DEPTH PERMEABLE ZONE							

Figure 3-39. Vertical distribution of permeable zones in selected wells in the Yucca Mountain area. Modified from Benson et al (1983).

The Prow Pass and Bullfrog members of the Crater Flat Tuff are generally similar in hydrologic properties and are commonly the most productive ones penetrated by drillholes in the vicinity of Yucca Mountain (Figures 3-38 and 3-39). Their in situ bulk hydraulic conductivity is controlled largely by fractures (Waddell et al., 1984). Results from in situ tests and core analyses for drillhole USW H-1 are shown in Table 3-29 (Rush et al., 1984).

The Tram Member of the Crater Flat Tuff generally is much less permeable than any of the overlying saturated units; at drill hole UE-25b#1, for example, the Tram yielded no measurable production during pumping of the entire hole (Figure 3-38; Lahoud et al., 1984). An exception occurs at drill hole USW G-4, where about 98 percent of the water production was from the Tram Member, although the overlying Bullfrog and Prow Pass Members are saturated; yield probably is from shear fractures noted in the lithologic log (Bentley, 1984). Table 3-27 shows that hydraulic conductivity for the Tram Member is as low as 7×10^{-6} m/d (drillhole USW H-1), and as high as 0.2 to 0.8 m/d (drillhole USW H-4).

Hydraulic conductivity of Tertiary rocks older than the Crater Flat Tuff ranges from about 2×10^{-6} to 3×10^{-2} m/d (Table 3-27).

Pre-Tertiary age rocks have been penetrated in only one drillhole in the vicinity of Yucca Mountain, at drillhole UE-25p#1 (Figures 3-28 and 3-25), where Paleozoic carbonate rocks of the Lone Mountain Dolomite and Roberts Mountains Formation were observed (Craig and Robison, 1984). They reported an average hydraulic conductivity value of 0.2 m/day.

3.9.2.2.3 Porosity and storage coefficients

Laboratory results of total porosity of cores ranging from 4 to 34 percent are shown in Table 3-27 and discussed in Chapter 2. Porosities reported by Anderson (1981) and by Thordarson (1983) were determined by water-saturation techniques, as described by Anderson (1981). The values reported for Lahoud et al. (1984) and Rush et al. (1984) were calculated by helium pycnometer; they reported generally similar values for porosity calculated from dry-bulk and grain densities.

Waddell (1984) estimated effective porosity between two drill sites to be on the order of 1×10^{-4} to 1×10^{-3} , based on breakthrough of a tracer placed in drillhole UE-25a#1 while drillhole UE-25b#1 was being pumped.

Using a method that involves calculating equivalent fracture apertures, Erickson and Waddell (1985) determined that fracture porosity at drillhole USW H-4 is 1×10^{-4} to 1×10^{-3} . They concluded that, if most water moves through fractures rather than the rock matrix, fracture porosity approximates effective porosity. Although results from single-hole tests are not necessarily reliable, Rush et al. (1984) used slug-injection tests to estimate storage coefficients that range from 2×10^{-5} to 6×10^{-6} among the stratigraphic units penetrated at drillhole USW H-1. Using the same data from drillhole USW H-1, Barr (1985) calculated values for storativity (storage coefficient) that are about an order of magnitude lower.

CONSULTATION DRAFT

Various hydraulic tests are being performed in the vicinity of drill-holes UE-25c#1, UE-25c#2, and UE-25c#3 to obtain more values for effective porosity and storage coefficients that will give a range of values that may be expected in the vicinity of the Yucca Mountain site (Section 8.3.1.2.3.1).

3.9.3 GROUND-WATER FLOW SYSTEM CONCEPTUAL MODEL

As indicated in the remarks prefacing Section 3.9, a conceptual model of a ground-water flow system uses available hydrogeologic data for the system to synthesize a general understanding of the overall (macroscopic) hydrologic system in terms of the specific (microscopic) hydrologic processes that are expected to operate within it. The macroscopic elements of the conceptual model include the hydrogeologic framework, a set of hydrologic boundary conditions (including the direction and magnitude of moisture or ground-water flux across the boundaries), and a knowledge of the present (initial) hydrologic state of the system, that is, the spatial distribution of hydraulic head in the saturated zone and of matric potential or saturation in the unsaturated zone. The microscopic elements are supplied by or developed from the fundamental concepts describing liquid and gaseous flow and storage in permeable media. Because it is usually abstracted from an incomplete set of data, a conceptual model generally consists of a collection of hypotheses that must be tested not only against existing data but also continually against new data as they become available. Consequently, one principal utility of a conceptual model is to guide data collection to permit adequate testing and evaluation of hypotheses. The hypotheses, thus, may be accepted, rejected, refined, or replaced by alternatives in the light of new data and improved understanding of the flow system.

The unsaturated zone beneath Yucca Mountain consists of a stratified block of east-dipping hydrogeologic units composed of welded and nonwelded tuffs of contrasting hydraulic properties. Some of these units are highly fractured, and the fractures may either enhance or retard the flow of moisture. The Yucca Mountain block contains and is bounded by west-dipping high-angle normal faults that, depending on location and ambient hydrologic conditions, may be pathways for or barriers to moisture flow. The mechanisms of moisture flow and storage (as liquid water, water vapor, or both) at any location within the system depend on the amount of water entering the system as net infiltration. The actual rate and distribution of net infiltration over the surface of Yucca Mountain is not presently known, although likely upper bounds have been established.

Generalized conceptual models for moisture flow within the unsaturated zone beneath Yucca Mountain have been developed by Montazer and Wilson (1984) and Sinnock et al. (1986). In addition, specific hydraulic problems were conceptualized by these authors, Klavetter and Peters (1986), and Wang and Narasimhan (1985, 1986). The essential features of these models define the hydrogeologic framework and identify the possible processes of moisture flow and storage that occur under the constraints imposed by the hydrogeologic framework. One conceptual model is illustrated schematically in Figure 3-40, which depicts a generalized east-west cross section through Yucca Mountain. The hydrogeologic framework consists of the eastward-tilted Yucca Mountain block, composed of the stratified sequence of hydrogeologic units listed in

Figure 3-40. Generalized east-west section through Yucca Mountain showing conceptual moisture-flow system under natural conditions. Modified from Montazer and Wilson (1984).

Table 3-22. The block is bounded on the west by a westward-dipping high-angle normal fault (the Solitario Canyon fault), is transected internally by westward-dipping high-angle faults (for example, the Ghost Dance fault), and is bounded on the east by a set (or zone) of imbricate westward-dipping high-angle normal faults. The upper hydrologic boundary consists of the land surface, across which water enters the unsaturated zone as infiltration directly from precipitation or from runoff. The lower hydrologic boundary is the water table, whose configuration is presumed to be known. Steady-state moisture flow is presumed to be established in all but a thin, near-surface interval of the unsaturated zone. In this instance, steady-state flow may be assumed if the time required to change conditions within hydrogeologic units below the near-surface interval are much longer than the period of interest.

The principal qualitative aspects of moisture flow within the unsaturated zone beneath Yucca Mountain, as conceptualized in Figure 3-40, are summarized by the following hypotheses, as summarized from Montazer and Wilson (1984):

1. Moisture enters the system as net infiltration below the plant-root zone principally as liquid-water flow into and within the fractures of the surficial Tiva Canyon welded unit (TCw) with subsequent uptake under capillary forces into the TCw matrix. It is expected that at sufficiently low rates of net infiltration, all the water entering the TCw unit may be drawn into the matrix before the full thickness of the TCw unit is traversed.
2. Hydraulic gradients directed parallel to the dip may induce lateral flow either under saturated or unsaturated conditions at the contact between the TCw unit and the underlying Paintbrush nonwelded unit (PTn) as a consequence of (1) efficient fracture-dominated flow in the TCw unit at high infiltration rates or (2) capillary barrier effects. Capillary barriers may inhibit water movement from a unit of low matrix or fracture conductivity into a unit of higher conductivity by capillary forces in the low-conductivity unit.
3. Both vertical and lateral flow may occur in the relatively high-conductivity PTn unit. Lateral flow may occur because of the intrinsic anisotropy of this unit as well as because the low matrix conductivity of the underlying Topopah Spring welded unit (TSw) may impede the vertical movement of water from the PTn unit into the TSw unit.
4. Flow in the TSw unit is expected to be essentially vertical and under steady-state conditions to occur as flow within the matrix for fluxes less than some critical value of flux related to the saturated matrix hydraulic conductivity, and predominantly as fracture flow at fluxes higher than the critical value.
5. Lateral flow may be induced in the TSw unit at its contact with the underlying Calico Hills nonwelded unit (CHn). The circumstances under which this may occur depend on the magnitude of the flux in the TSw unit and whether this unit is underlain by the low-conductivity zeolitic facies (CHnz) or the relatively higher-conductivity vitric facies (CHnv) of the CHn unit. At low fluxes

- within the TSw unit, lateral flow may be produced by capillary-barrier effects within the matrix of the TSw unit where it overlies the CHnv unit. At high fluxes, efficient fracture flow in the TSw unit may produce lateral flow as well as vertical flow where the low-conductivity CHn unit underlies the TSw unit.
6. Flow in both the CHnv and CHnz units is predominantly vertically through the matrix (although a lateral component may occur parallel to the bedding within the vitric CHnv unit) and continues directly to the water table wherever the latter transects the CHn unit. Where the CHn unit lies above the water table, flow is presumed to proceed vertically downward to the water table through the Crater Flat undifferentiated unit (CFu).
 7. The nearly vertically oriented fault zones and their associated fracturing may be highly effective pathways for vertical moisture flow, especially in the competent TCw and TSw units. But faults may impede lateral flow and may thus produce perched-water bodies where the faults transect zones or horizons of significant lateral flow.
 8. Temperature-driven moisture transport may occur, especially within the highly fractured TSw unit. This could be expected to occur by molecular diffusion if local thermodynamic phase equilibrium is maintained between liquid water and water vapor within the system, which would produce a water-vapor concentration gradient along the natural geothermal gradient. Of greater importance may be the advective transport of water vapor accompanying thermally or barometrically driven upward bulk-gas flow within the fractures of the TSw unit. Under steady-state conditions, the upward movement of water vapor in the air-filled fracture openings would be compensated by downward return flow of liquid water within the rock matrix.
 9. Moisture flow within the deep unsaturated zone at Yucca Mountain may be occurring under essentially steady-state conditions. The steady-state hypothesis implies that moisture flow within the natural system is occurring predominantly as vertically downward liquid-water flow within the rock matrix of the hydrogeologic units with possible water-vapor movement within the air-filled pore and fracture space. Significant liquid-water flow within the fractures may occur but only as near-surface, transient, nonequilibrium events that are followed by eventual uptake by the rock matrix of water descending through the fractures. Equilibrium would prevail between the liquid-water matric potentials in the fractures and the unsaturated rock matrix at depth. Liquid water movement could occur in fractures as well as the matrix. Although the amount of fracture flow probably would be small, additional understanding of the factors controlling fracture flow is needed to assess this phenomenon. Under these conditions, the liquid-water flux through the matrix of the TSw unit would be expected to be less than the mean matrix saturated hydraulic conductivity of about 1 mm/yr, as determined for this unit by Peters et. al. (1984).

This set of qualitative hypotheses is evaluated in terms of presently available data and uncertainties in Section 3.9.3.4, unsaturated zone relationships.

A general geohydrologic conceptual model for saturated ground-water flow near Yucca Mountain is based upon division of the Tertiary stratigraphic units into (1) moderately and densely welded tuff and (2) nonwelded and bedded tuffs, as a result of the different hydrologic properties of these materials. The moderately and densely welded tuffs are characterized by relatively low porosities, abundant fractures, and low matrix permeability. In nonwelded and bedded tuffs, porosities are greater and fractures are fewer, although fractures do occur. Unless the rock contains clays or zeolites, matrix conductivities are generally larger than in the welded tuffs (Waddell et al., 1984).

This model provides a basis for predicting saturated hydraulic properties near Yucca Mountain, provided that the welding characteristics and extent of secondary mineralization of units are known. The Paintbrush Tuff is mostly unsaturated; the Topopah Spring Member, which is unsaturated in the repository block, but saturated east of Yucca Mountain (Figures 3-28 and 3-25), consists mostly of moderately to densely welded tuffs. The Crater Flat Tuff (Prow Pass, Bullfrog, and Tram members) consists of partially to moderately welded tuffs. The tuffaceous beds of Calico Hills and Lithic Ridge Tuff are predominantly nonwelded and bedded tuffs. In addition, bedded tuffs commonly separate the major ash-flow tuff units.

Partial alteration of vitric tuffs to zeolites or clays reduces their permeability substantially. Thickness of the zeolitic facies of the tuffaceous beds of Calico Hills increases from southwest to northeast (Figure 3-41), and the entire unit is altered beneath the northern and northeastern parts of the Yucca Mountain repository block (Montazer and Wilson, 1984) (see Section 3.9.2.1).

All the Tertiary stratigraphic units just discussed are within the regional "tuff aquitard" considered by Winograd and Thordarson (1975). Beneath Yucca Mountain are Paleozoic carbonate rocks, equivalent to the regional "lower carbonate aquifer" of Winograd and Thordarson (1975). These rocks occur at a depth of 1,200 m in drill hole UE-25p#1 and probably much deeper elsewhere (Craig and Robison, 1984). These carbonate rocks are permeable, but at Yucca Mountain the potentiometric head, as measured at drill hole UE-25p#1 (which is the only drillhole penetrating the Paleozoic rocks), is greater in the Paleozoic rocks than in overlying rocks. These rocks, in ascending order include Lithic Ridge Tuff and other tuffs of low permeability, Crater Flat Tuff, and Paintbrush Tuff (Craig and Robison, 1984; Robison, 1984). Therefore, ground water moving through the shallow part of the saturated zone in the vicinity of the Yucca Mountain repository block probably does not have significant local interaction with water from the Paleozoic rocks.

Figure 3-41. Isopach contour maps within the repository block: (A) Total thickness from disturbed zone to the water table; (B) Thickness of undisturbed Topopah Spring welded unit TSw; (C) Thickness of the Calico Hills nonwelded vitric unit, CHnv; (D) Thickness of the Calico Hills nonwelded zeolitic unit CHnz; (E) Thickness of the Prow Pass welded unit, PPw; (F) Thickness of the Prow Pass nonwelded unit, PPn; (G) Thickness of the Bullfrog welded unit. Modified from Sinnock et al. (1986).

3.9.3.1 Accessible environment and credible pathways

The proposed repository horizon at Yucca Mountain is within the unsaturated TSw unit, about 250 m above the water table and ranging from about 80 m to 140 m above the underlying CHn unit. The most credible pathway for liquid-water movement from the repository horizon enroute to the accessible environment depends to a large extent on the flow mechanisms with the TSw unit. These, in turn, depend on the saturation as well as on the flux and its time rate of change. If the matrix and fracture saturation are sufficiently low, liquid-water movement is expected to be restricted to the matrix of the TSw unit and will tend to be directed vertically downward continuing through the CHn unit to the water table.

In the unsaturated zone, matrix saturations within the fractures of the TSw unit may exceed some critical values and cause significant liquid-water flow. The fractures, depending on local conditions, may be capable of delivering relatively large volumes of water from the repository horizon rapidly to the contact between the TSw and CHn units. Adequate data are presently lacking to evaluate these hypotheses quantitatively. Ambient saturation within the TSw unit at the repository horizon is inferred, from data obtained from test wells USW H-1, USW UZ-1, and UZ-6, to be in the range 0.4 to 0.6 (Sections 3.9.2.1 and 3.9.3.4), and the vertical flux at this horizon has been estimated by Wilson (1985) to be 0.5 mm/yr or less. This value is less than the mean value for the saturated matrix hydraulic conductivity of about 1 mm/yr for the TSw unit (3×10^{-11} m/s in Table 3-26), and implies that the flux probably occurs principally in the matrix and not in fractures of the TSw unit. However, Montazer and Wilson (1984) indicate that fracture flow may occur at saturations less than 100 percent. In addition, according to Montazer and Wilson (1984), net infiltration may be considerably greater than the average flux in the TSw unit. Tests are planned to evaluate the conditions under which flow in fractures and faults may occur (Section 8.3.1.2.3), thus aiding in the definition of flow paths in the unsaturated zone.

In the saturated zone, differences in potentiometric levels indicate general directions of ground-water flow. In general, ground-water flow beneath Yucca Mountain probably is southward through the site, southeastward away from the site into the Fortymile Wash area. However, because of the nearly flat potentiometric surface under parts of Yucca Mountain, specific flowpath directions are currently difficult to define. Furthermore, the degree of anisotropy has not been evaluated. Additional water-table holes and extensive multiple-well and single-well tracer tests may help define anisotropy, hydraulic connections, and probable flow paths in the saturated zone (refer to Section 8.3.1.2.3.1).

3.9.3.2 Potentiometric levels and head relationships

This section discusses the relationships between matrix potentials for hydrogeologic units within the unsaturated zone and between potentiometric levels for units within the saturated zone.

3.9.3.2.1 Unsaturated zone

Little direct information on matric potentials and moisture fluxes is presently available for the unsaturated zone beneath Yucca Mountain. Weeks and Wilson (1984) report the results of measuring ambient rock-matrix saturations on 19 samples from cores obtained in test well USW H-1 (Figure 3-28). The well was drilled using an air-foam detergent mixture as drilling fluid (Rush et al., 1983), which minimized possible moisture uptake by the core samples. Mercury intrusion techniques were used to develop moisture-retention curves for each of the samples from which ambient matric potentials were inferred. The resulting set of moisture-retention curves differs considerably from the analytic representations of Peters et al. (1984) shown in Figure 3-33, which were based on psychrometric measurements made over a range of matric potentials much less than the minimum value of about -0.20 MPa that could be measured by Weeks and Wilson (1984). Only one sample was available from the PTn hydrogeologic unit at a depth of 33.5 m below land surface; it had a matric potential of -0.15 MPa at an ambient saturation of 0.51. Sixteen samples were available for the TSw unit spanning a vertical interval of 278 m within the unit. The lowest saturation measured within this sample set was 0.47, the highest was 0.89, and the mean was 0.65. There was a poorly defined indication within the sample set for saturation to increase with depth. The matric potentials inferred from the measured saturations for the samples were highly variable, ranging from -0.07 to more than -0.25 MPa with a mean of -0.18 MPa and a standard deviation of -0.06 MPa. Two samples were available from the CHn unit at a depth of 530 m and both indicated complete matrix saturations. Because of the considerable scatter within the set of data, the matric-potential and saturation values for the samples from the TSw unit do not define a unique, representative moisture-retention curve for this unit. It is not known to what extent this scatter may be attributed to intrinsic heterogeneity of properties and conditions within the TSw unit or is due to undetermined experimental errors of measurement. Consequently, these data from drillhole USW H-1 probably are best interpreted as indications only of the hydrologic conditions that may be expected to be encountered within the hydrogeologic units of the unsaturated zone at Yucca Mountain.

As described in Section 3.9.1.1.1, test well USW UZ-1 (Figure 3-27) has been drilled to a total depth of 387 m within the unsaturated zone at Yucca Mountain and instrumented with thermocouple psychrometers and heat-dissipation probes to monitor in situ matric potentials at 33 selected depths within the TCw, PTn, and TSw hydrogeologic units (Montazer et al., 1985). Data from two years of continued monitoring indicate that hydrologic conditions within the borehole may be in approximate equilibrium with ambient conditions in the surrounding units. The data available to date are shown in Figures 3-30 and 3-31. Although the data show considerable variability, there is no well-defined systematic variation of matric potential with depth within hydrogeologic units, which is consistent with the supposition that the individual hydrogeologic units are relatively homogeneous in the vertical direction and that steady-state vertical moisture flow occurs under unit vertical hydraulic gradient. The trend with time, if interpreted as the approach to equilibrium between the well and its surroundings, indicates that present ambient matric potentials generally are greater than -1 MPa, and that the mean matric potential over the total depth of the borehole is about -0.3 MPa. These data from borehole USW UZ-1 are consistent with the range of

CONSULTATION DRAFT

values of matric potential inferred by Weeks and Wilson (1984) to be representative of conditions within the interval of the unsaturated zone penetrated by borehole USW H-1. As described in Section 8.3.1.2.2, additional test holes in the unsaturated zone will be drilled and instrumented to better define the distribution of matric potentials and fluxes.

3.9.3.2.2 Saturated zone

Measured head values used to construct the saturated-zone potentiometric map (Figure 3-28) are mostly composite heads, reflecting heads similar to those in the zones of higher transmissivity.

The hydraulic gradient is low in western Jackass Flats (Fortymile Wash area) and in the Amargosa Desert; the gradient is high in volcanic rocks north of Yucca Mountain and across northern Yucca Mountain. Low-permeability rocks likely occur where the gradients are steep, but the steep gradients may be caused by poorly-permeable faults. In areas where it may be important to know the cause of steep gradients, additional drilling or hydrologic testing is planned (Section 8.3.1.2.3).

Significant vertical hydraulic gradients have been observed in only a few drillholes: in drillhole UE-25p#1, the head in Paleozoic carbonate rocks that occur below a depth of 1.2 km is about 19 m higher than in the shallow zone; in drillhole USW H-1 the head is about 54 m higher in the older tuffs of Tertiary age, at a depth of 1.8 km; and preliminary data from drillhole USW H-3 suggest that the head below 0.8 km (Tram Member of Crater Flat Tuff) is about 40 m higher. These higher heads occur only within or below intervals of low permeability. Semipermanent packers are installed in lower sections of several drillholes (UE-25b#1, USW H-3, USW H-4, USW H-5, and USW H-6) to enable comparison of water levels above and below the packers. Except in drillhole USW H-3, vertical differences of head within the Paintbrush and Crater Flat tuffs have been less than a meter.

The upward component of ground-water flow that is indicated by higher heads in the deeper part of the saturated zone may be the result of the position of Yucca Mountain within the regional ground-water flow system. Simple models of ground-water basins consist of an area where recharge occurs (vertical flow is downward), and where discharge occurs (vertical flow is upward). In intermediate areas, flow is transitional from downward, to strictly horizontal, to upward (Figure 3-42). Although the Alkali Flat-Furnace Creek Ranch subbasin (of which Yucca Mountain is a part) is not simple, these concepts generally apply. Thus, Yucca Mountain is in an area where local vertical flow may be either up or down (Waddell et al., 1984). Erickson and Waddell (1985) observed slight upward and downward movement among different intervals of the same drillhole (USW H-4). This upward and downward movement may also be due to differences in horizontal and vertical hydraulic conductivity beneath Yucca Mountain.

Figure 3-42. Standard theoretical flow pattern and boundaries between different flow systems. Modified from Toth (1963).

3.9.3.3 Recharge-discharge and leakage

The ultimate source of moisture in the unsaturated zone at Yucca Mountain is by net infiltration from precipitation on the mountain, and the quantity of precipitation is used by some techniques to estimate net infiltration and, subsequently, recharge. Net infiltration refers to the rate and quantity of water entering the unsaturated zone below the plant-root zone and represents moisture that is effectively inaccessible for direct return to the atmosphere by evapotranspiration. Recharge is the rate and quantity of water entering the saturated zone from the unsaturated zone (the term specifically excludes underflow into or out of the region of interest; recharge to adjacent areas subsequently may enter the region of interest as underflow). Direct measurements of precipitation at Yucca Mountain have been initiated, but too recently to provide reliable annual average values (Section 5.1). However, precipitation at the site has been estimated from the regional distribution of precipitation or from relationships established between altitude and precipitation. From a regional map of precipitation presented by Winograd and Thordarson (1975), precipitation at Yucca Mountain is estimated to be about 100 to 150 mm/yr (Section 5.1). Quiring (1983) established local relationships between altitude and precipitation for 1964-81 at the NTS. Quiring's (1983) data included limited data outside the NTS and data from 13 stations within the NTS that range in altitude from 3,000 to 7,490 ft (914 to 2,283 m). From a map showing precipitation-to-altitude ratios (Quiring, 1983), precipitation for the approximate range of altitudes at Yucca Mountain (about 1,220 to 1,465 m) is estimated to be 138 to 166 mm/yr, or an areal average of about 150 mm/yr (Montazer and Wilson, 1984). Quiring's (1983) data indicate that about 73 percent of this quantity falls during October through April (Montazer and Wilson, 1984).

Neither the spatial distribution nor the temporal variation of net infiltration is known at Yucca Mountain. However, an intensive field and experimental investigation of this process has been planned (Section 8.3.1.2.1). Spatial variations of net infiltration are mostly dependent on the variations in properties of the surficial units, topography and the vegetative cover. The surficial materials consist of alluvial and colluvial deposits and fractured welded tuffs, principally the Tiva Canyon welded unit. The area underlain by outcrops of nonwelded units is small. Alluvial and colluvial deposits fill the valleys and washes that form the main channels for surface runoff. The Tiva Canyon welded unit crops out throughout most of the repository block, and this unit has the most exposure to precipitation (Montazer and Wilson, 1984). Vegetation is sparse and consists mostly of xerophytes that delimit the depth of the plant-root zone.

Because of the great thickness of the unsaturated zone, the temporal variation and spatial distribution of recharge beneath Yucca Mountain are not expected to be equal to the temporal variation and surficial distribution of net infiltration. Temporal variations of net infiltration are expected to be damped rapidly with depth in the uppermost few tens of meters of the unsaturated zone (Weeks and Wilson, 1984) to produce approximately steady-state moisture flow at greater depths. Slow temporal variation of the moisture flux at any horizon is expected to occur generally only in response to long-term climatic changes. The spatial distribution of recharge beneath Yucca Mountain depends complexly on the hydrogeologic framework, the hydrologic properties, and the flow processes within the unsaturated zone and is not

amenable to direct determination. Because the ground-water travel time through the unsaturated zone is likely to be of the order of several tens of thousands of years and because precipitation is decreasing regionally (Section 5.2.1), the present rate of net infiltration at Yucca Mountain probably is somewhat less than the present rate of recharge. Consequently, the present rate of net infiltration over the surface of Yucca Mountain probably is not a reliable indicator of either the net recharge beneath Yucca Mountain nor the net moisture flux, at or below the repository horizon.

Recharge to the saturated zone and the deep aquifers has been estimated within the region surrounding Yucca Mountain. For example, recharge to the regional carbonate aquifer, which underlies much of the NTS and vicinity, was estimated to be 3 percent of the precipitation falling on upland outcrops of this aquifer on the basis of measured spring discharge at Ash Meadows (Winograd and Thordarson, 1975). Applying this percentage to average precipitation at Yucca Mountain (150 mm/yr) would imply a recharge rate of about 4.5 mm/yr. In addition, Waddell et al. (1984), using data from Winograd (1981), estimated a moisture flux of about 0.5 mm/yr through the alluvium in Yucca Flat, which, however, being 40 km northeast of Yucca Mountain, probably is not typical of the Yucca Mountain site. Also, Case et al. (1984) determined a typical value of pore velocity of "about one-third centimeter per year" (under unsaturated conditions) in Frenchman Flat, 40 km east of Yucca Mountain. This is equivalent to a flux of 1 mm/yr if an effective porosity of 30 percent is assumed for the soil at Frenchman Flat. None of these studies provides a reliable basis for estimating recharge at Yucca Mountain. As at all sites, the actual value of recharge at Yucca Mountain depends on site-specific, local microclimatic, soil, vegetative, topographic, and hydrogeologic conditions.

Methods for estimating recharge for ground-water basins in Nevada were developed by Maxey and Eakin (1949) and refined by Eakin et al. (1951) and Malmberg and Eakin (1962). The resulting Maxey-Eakin method uses regional relationships among recharge, altitude zones, and precipitation but it does not account for a number of factors, such as runoff during intense storms, local rainfall distribution, temperature, and vegetative cover. Rush (1970) used the method to estimate recharge in the vicinity of the NTS; Czarnecki (1985) used Rush's estimates as a basis for input to a two-dimensional ground-water flow model to simulate present hydrologic conditions over an area whose boundaries are approximately the same as the Alkali Flat-Furnace Creek Ranch subbasin of Waddell (1982). Czarnecki (1985) used Rush's (1970) results and estimated recharge to be about 0.5 mm/yr in a precipitation zone that includes Yucca Mountain as well as parts of Jackass Flats and Crater Flat.

Geothermal heat flux can be used for indirect estimation of ground-water percolation rates, where the conductive and convective components of the heat flux can be separated (Sass et al., 1980). Sass and Lachenbruch (1982) estimated a vertical water flux of 1 to 10 mm/yr in wells near Yucca Mountain that are deeper than 1 km. The geothermal method can only be expected to yield percolation or recharge values within an order of magnitude because, among other reasons, it assumes one-dimensional ground-water flow, without accounting for lateral flow. Montazer et al. (1985) used the method developed by Bredehoeft and Papadopoulos (1965) to calculate the water flux in the

unsaturated zone. Montazer et al. (1985) estimated 0.02 to 0.05 mm/yr of upward flux in the TSw unit.

No springs exist near Yucca Mountain, and no discharge from the saturated ground-water system occurs by evapotranspiration because the water table, which is 300 to 750 m below land surface (Robison, 1986), is much too deep to be affected.

Wilson (1985) reviewed available site and regional hydrogeologic data in order to set conservative upper limits on the present, net vertically downward moisture flux below the repository horizon at Yucca Mountain and on the present rate of net recharge to the saturated zone in the vicinity of Yucca Mountain. Wilson (1985) concludes (1) that the liquid-water percolation flux, directed vertically downward in the matrix of the TSw unit below the repository horizon, probably is less than 0.2 mm/yr, and (2) that the areally averaged rate of net recharge to the saturated zone in the vicinity of Yucca Mountain probably is less than 0.5 mm/yr. Although Wilson (1985) considered a number of processes, such as upward water-vapor flow in the fractures of the TSw unit at the repository horizon, these upper bounds on percolation and recharge fluxes must be regarded as preliminary estimates that have as-yet-unknown limits of uncertainty. These estimates are expected to be refined using data obtained from borehole monitoring of ambient hydrologic conditions within the unsaturated zone, from the surface-based infiltration experiments, and from the in situ measurements and experiments within the exploratory shaft, described in Sections 8.3.1.2.1 and 8.3.1.2.2.

Ground-water flow modeling studies of the saturated zone at Yucca Mountain and vicinity have been conducted by Waddell (1982), Czarnecki and Waddell (1984), Rice (1984), and Czarnecki (1985). These two-dimensional models consider areal variation in hydrologic properties, but vertically combine the properties of hydrogeologic units. Results from a three-dimensional steady-state ground-water flow model (Sections 8.3.1.2 and 8.3.1.2.3) will be used to gain a better understanding of the flow system.

The regional model (18,000 km²) of Waddell (1982), Figure 3-43, of which the Yucca Mountain area is only a small part, specified constant ground-water flux over a model basin in the vicinity of Pahranaagat Valley and Pahrump Valley, and constant flux out of the basin at Furnace Creek Ranch in Death Valley. No-flow boundaries coincide approximately with the other boundaries of the regional ground-water basin (Figure 3-7). Constant-head nodes were used at Alkali Flat to simulate ground-water discharge. This model calculated vertically-integrated lateral fluxes of 0.04 to 0.2 m²/d at or near the Yucca Mountain repository block, based on assumed transmissivities of about 7 m²/d and 750 m²/d. The model and the results of Rice (1984) are generally similar to those of Waddell (1982).

The subregional model of Czarnecki and Waddell (1984) (Figure 3-43), covering 6,000 km², is smaller than that of Waddell (1982). Specific fluxes into the flow system were applied along the northern boundary of Jackass Flats, along Rock Valley, the western edge of the Amargosa Desert, the western edge of Ash Meadows, and along Fortymile Canyon (Figure 3-44). Fluxes out of the system were simulated by specifying discharge at Alkali Flat (Franklin Lake Playa) and Furnace Creek Ranch. Other external boundaries were specified as no-flow. Hydraulic heads along the northern

Modified from Czarniecki and Waddell (1984).

CONSULTATION DRAFT

Figure 3-44. Locations and estimates of ground-water flow into and out of the Alkali Flat-Furnace Creek Ranch subbasin. Modified from Czarnecki and Waddell (1984).

boundary, near Timber Mountain, were specified as constant. Vectors of simulated vertically-integrated ground-water flux vary in direction from southeast to south, through and away from the Yucca Mountain repository block, and have a magnitude of about 2 to 3 m²/d. These model-derived fluxes result from a parameter-estimation approach, in which an estimated transmissivity of 3,340 m²/d was applied; although this value of transmissivity is somewhat greater than obtained from pumping tests, it is not outside the expected range of agreement.

3.9.3.4 Unsaturated-zone relationships

The factors controlling land-surface infiltration and the subsequent delivery of net infiltration as recharge to the water table are discussed in Section 3.9.3.3. The spatial distribution of net infiltration is one factor that controls moisture fluxes within the deep unsaturated zone beneath Yucca Mountain. No direct measurements of net infiltration have been made at Yucca Mountain and thus neither its magnitude nor its spatial and temporal distribution over the surface of Yucca Mountain are known. Montazer and Wilson (1984) estimate that the present average rate of net infiltration over the Yucca Mountain block may be expected to range from about 0.5 to no more than 4.5 mm/yr. As discussed in Section 3.9.3.3, Wilson (1985), in reviewing and analyzing available data, concluded that a value of 0.5 mm/yr may be a conservative upper bound for the mean vertical-downward percolation flux of water across the repository horizon. Net infiltration is expected to be episodic and unevenly distributed over the surface of Yucca Mountain. These irregularities, however, probably are smoothed rapidly with depth (Weeks and Wilson, 1984) so that the mean percolation rate at depth within Yucca Mountain is affected significantly only by large-scale climatic changes. If these changes occur sufficiently slowly in time, then the hydrologic system within the unsaturated zone can be presumed to adjust continuously to these changes so that an overall steady-state flow system may approximately exist at any point in time.

If, as is implied by the data obtained from test wells USW H-1, USW UZ-1, and USW UZ-6, the ambient liquid-water saturation within the repository TSw unit is within the range from about 0.5 to 0.7, then liquid-water percolation through the unit probably is matrix dominated (Klavetter and Peters, 1986; Peters et al., 1986) with little or no flow of water within fractures. Laboratory data (Table 3-26) indicate that the saturated hydraulic conductivity for the matrix of the TSw unit is about 3.5×10^{-11} m/s which, under a unit vertical hydraulic gradient, corresponds to a maximum vertical liquid-water flux through the matrix of about 0.8 mm/yr. At matrix saturations less than 1.0, this maximum flux within the TSw matrix would be reduced by the relative hydraulic conductivity appropriate to the matrix pore geometry and the ambient saturation. The relation between relative hydraulic conductivity and saturation for the TSw unit matrix material is not known. The van Genuchten (1980) representation used by Peters et al. (1984), however, would suggest a relative hydraulic conductivity factor of the order of 10^{-2} to 10^{-3} to be appropriate. Hence, the vertical percolation flux through the TSw unit, and by implication across the repository horizon within the unit, may well be about or much less than the 0.5 mm/yr maximum value estimated by Wilson (1985). This finding is consistent with the range of values from 0.1

CONSULTATION DRAFT

to 0.5 mm/yr estimated by Montazer et al. (1985) from data obtained from borehole USW UZ-1 (Section 3.9.1.2.1). It must be emphasized that these analyses are very preliminary and approximate and presume that the TSw unit is laterally and vertically homogeneous and isotropic and that steady-state flow conditions prevail.

As described in Section 8.3.1.2.2, hydrologic data from the unsaturated zone at Yucca Mountain are to be obtained directly only from the exploratory shaft test facility and from a relatively small number of surface-based boreholes. Consequently, it will be necessary to rely on indirect methods, such as numerical flow and solute-transport modeling and geostatistical techniques, to infer the state of the presently existing natural hydrologic system by interpolating within and extrapolating from a somewhat incomplete set of field data (Section 8.3.1.2.2). Using data that are presently available for the site, several investigators have used these methods to perform rudimentary analyses and develop preliminary interpretations. These investigations have been directed in part towards examining some of the hypotheses underlying the conceptual model presented in Section 3.9.3 and contribute especially to identifying specific problems that will need to be resolved as part of the site-characterization program. These initial assessments, therefore, not only provide insight into the conceptualization of unsaturated-zone relationships at Yucca Mountain but also define specific data needs to be met by subsequent data acquisition at the site.

Peters et al. (1986) used a one-dimensional model to simulate liquid-water flow in an idealized but representative vertical column through the unsaturated zone at Yucca Mountain. They investigated the mode of liquid-water flow for a set of prescribed percolation fluxes within the column. The column was simulated to be 530.2 m in height and consisted of a basal CHnz or CHnv unit overlain by an upward succession of the units TSw, PTn, and TCw. The mean hydrologic properties for these units were taken from Peters et al. (1984). The results of these simulations indicate that virtually completely saturated conditions would be expected to occur within the TSw and TCw units for percolation fluxes exceeding about 0.5 mm/yr. This conclusion results from the values of the saturated matrix hydraulic conductivity for the TSw and TCw units that were used in the numerical simulations. The mean matrix conductivity values used for those two units were about 0.5 mm/yr. In a steady-state simulation, the value of percolation flux which induces saturated-matrix conditions in a particular hydrogeologic unit is directly proportional to its value of saturated matrix hydraulic conductivity. The results of the numerical simulations also predict that the time for the unsaturated system to return to steady state after a small change (a few tenths of a millimeter per year) in the steady percolation flux is on the order of tens or hundreds of thousands of years. These results suggest that the response time of the percolation flux at depth to changes in the net infiltration rate is very long. Greater percolation fluxes under this model would be expected to induce increasing volumes of flow within the fractures of the TSw and TCw units. However, the applicability of these results to Yucca Mountain is uncertain, because of (1) the limitations of one-dimensional modeling, which does not account for lateral heterogeneity, boundary conditions, and flow, and (2) the uncertainty in the representativeness of the hydrologic properties. None the less, these results, together with the moisture-content data from boreholes USW UZ-1, UZ-6, and H-1, which indicate that the rock matrix of the penetrated units is not saturated,

support the empirically based estimates of Wilson (1985) and Montazer et al. (1985) that the average percolation flux within much of the TSw unit probably is less than about 0.5 mm/yr.

Sinnock et al. (1986) used one-dimensional modeling techniques to calculate probable ground-water travel times from the repository horizon to the water table (Section 3.9.4). These calculations assumed that steady-state moisture-flow conditions prevail within the unsaturated zone at Yucca Mountain and considered steady-state percolation fluxes of 0.1, 0.5, and 1.0 mm/yr. By varying flow-model parameters and using a Monte Carlo approach, probabilistic distributions for the ground-water travel time were developed. Although these calculations demonstrate the utility of the methodology, the quantitative results were sensitive to hydrologic-property input data that are as yet highly uncertain. Sinnock et al. (1986) emphasize the need for obtaining reliable field and laboratory hydrologic-property data for all of the hydrogeologic units that affect the travel-time calculations. In addition, they recommend that, to the extent that the sparse distribution of data-gathering sites will allow, geostatistical techniques be used to estimate the spatial correlation of hydrologic properties within and between hydrogeologic units. Geostatistical methods are also essential for the description of the unsaturated zone and the estimation of its hydraulic properties from the limited empirical data set that will be available from the Yucca Mountain site characterization studies.

Rulon et al. (1986) modeled multidimensional liquid-water flow within the unsaturated zone at Yucca Mountain by constructing a two-dimensional, vertical cross-sectional flow model to investigate: (1) the difference in steady-state flow patterns and moisture distributions that could result under the extremes of matrix-dominated or fracture-dominated liquid-water flow within the TCw and TSw units; and (2) the likelihood for the occurrence of down-dip lateral flow within hydrogeologic units or at the contacts between units in the presence of a highly-transmissive bounding fault-zone that could readily divert all lateral influx directly to the water table. The model numerically approximated the equations describing variably saturated ground-water flow within the unsaturated zone on the basis of the rectangular grid shown in Figure 3-45. The stair-stepping hydrogeologic-contact boundaries depicted in Figure 3-45 were introduced to simulate approximately the eastward dip of the hydrogeologic units within the Yucca Mountain block. The boundary conditions imposed on the model included a basal static water table, an upper land-surface infiltration boundary, and lateral boundaries consisting of a vertical impermeable boundary on the west and a highly transmissive vertical fault zone on the east. The bounding fault zone was introduced as an artifice to receive lateral flow from the units abutting the zone, however, the hydrologic characteristics of the vertical faults at Yucca Mountain are not known. Any correlation between this model and the system in place at Yucca Mountain may be premature. Further, any assumptions involving the effects of these faults on ground-water flow, travel times, etc., should be considered preliminary. Plans to investigate the hydrologic characteristics of the Ghost Dance Fault are presented 8.3.1.2.2.6.

Three sets of steady-state simulations were performed in which liquid-water flow was assumed to occur (1) as matrix flow in all of the hydrogeologic units, (2) only in the fractures of the TCw and TSw units but in the matrix of the other units, and (3) in either or both the matrix and the

CONSULTATION DRAFT

TCw TIVA CANYON WELDED UNIT

PTn PAINTBRUSH NONWELDED UNIT

TSw TOPOPAH SPRING WELDED UNIT

CHnv CALICO HILLS NONWELDED VITRIC UNIT

CHnz CALICO HILLS NONWELDED ZEOLITIC UNIT

Figure 3-45. Numerical grid used in the two-dimensional flow simulations. Modified from Rulon et al. (1986).

fractures of the TCw and TSw units. Constant values of land-surface net infiltration rate, spanning the range from 0.015 to 4.5 mm/yr, were used as parameters within each of the three sets of simulations.

Because the moisture-characteristic relations for each of the hydro-geologic units considered by Rulon et al. (1986) differed considerably from those used by Peters et al. (1984) for the same units, the steady-state vertical profiles of matric potential and saturation computed by Rulon et al. (1986) for a specified net-infiltration rate (or percolation flux) likewise differed quantitatively from those computed by Peters et al. (1984). Both sets of model calculations, however, predicted that steady net infiltration rates (or percolation fluxes) exceeding values in the range from 0.5 to 1.0 mm/yr would produce complete matrix saturation in the TSw and TCw units after the mountain equilibrates to the change in flux. These conditions could lead to liquid-water flow within the fractures of these units. Both sets of calculations are consistent with the hypothesis that the present net percolation flux in the TSw unit is less than about 0.5 mm/yr (Wilson, 1985; Montazer et al., 1985), and that the flux is transmitted predominantly within the rock matrix of the TSw unit under conditions of less than complete matrix saturation.

Although the representation depicted in Figure 3-45 of inclined hydro-geologic units abutting a vertical fault zone is crude, the model results of Rulon et al. (1986) suggest that there is the potential for significant lateral flow above the repository TSw unit. At a high rate of net infiltration (4.5 mm/yr), significant lateral flow was predicted to occur within fractures of the uppermost TCw unit. At lower infiltration rates, significant lateral diversion of vertical flux was predicted to occur within the high matrix-conductivity PTn unit above its contact with the underlying TSw unit. Because these model calculations, as well as those of Peters et al. (1986), are based on highly idealized representations of the physical system, use a coarse grid size, and use hydrologic-property data that are preliminary and with unknown limits of uncertainty, the model results may not be quantitatively reliable. One of the major tasks to be accomplished through the data-acquisition program at the site, as detailed in Section 8.3.1.2, is to collect sufficiently large sample sets to delimit the uncertainties associated with the quantitative predictions of the numerical models.

This discussion of the preliminary numerical models and their results presumes that natural moisture flow within the unsaturated zone at Yucca Mountain occurs solely as liquid-water movement through rock-matrix pores and fractures under isothermal, steady-state conditions. It will be difficult to assess the steady-state flow hypothesis during the short time span available to the site-characterization program. However, some indication of the degree to which steady-state flow is realized may be inferred from surface-based infiltration measurements and experiments (Section 8.3.1.2.1). If steady-state flow prevails, then the present rate and distribution of net infiltration over the surface of Yucca Mountain should be demonstrably consistent with the magnitude and distribution of moisture flux within the interior of the Yucca Mountain block, whether inferred from in situ measurements of hydrologic conditions or predicted through numerical simulations. Because of their expected low magnitudes, fluxes at depth probably will not be directly measurable within the unsaturated zone but will have to be inferred from measured potential distributions and hydrologic properties. In addition,

numerical simulations may be needed to interpret these data and to estimate the limiting conditions under which these data support or refute the steady-state hypothesis. The analysis and interpretation of the Yucca Mountain hydrologic system will be complicated greatly if the steady-state approximation is invalid. The appropriateness of this approximation will be evaluated during site characterization through monitoring of the ambient in situ water potential and numerical analyses (Section 8.3.1.2).

The numerical modeling results reported here are preliminary but are indicative of the kinds of analyses that can be performed through numerical simulations. Mathematical modeling is essential to gaining an understanding of the overall hydrologic system, and, as described in Section 8.3.1.2, considerable reliance is to be placed on these techniques. All the hydrologic data obtained during site characterization are ultimately to be synthesized in a comprehensive model for the flow of moisture within the unsaturated zone at Yucca Mountain.

Because of the presence of the natural geothermal gradient, moisture flow within the unsaturated zone at Yucca Mountain is not strictly isothermal. Thermal logging of drillhole USW G-1 yielded a mean gradient of $1.35^{\circ}\text{F}/100\text{ ft}$ ($24.6^{\circ}\text{C}/\text{km}$) (Tien et al., 1985) which suggests a temperature difference of about 13°C between the water table and land surface. A temperature difference of this magnitude probably has very little effect on the mechanics of liquid-water movement. Reda (1985b), for example, found experimentally that the intrinsic permeability of a sample of the Topopah Spring welded unit remained constant at $3.0 \times 10^{-19}\text{ m}^2$ to within 10 percent over the temperature range from 25 to 90°C . The largest effect probably would be that of an increase in effective hydraulic conductivity with depth due to the reduction of the viscosity of water with increasing temperature. The change in water viscosity due to its temperature dependence over the thickness of Yucca Mountain could be expected to produce a maximum fractional change of hydraulic conductivity of about 30 percent. However, this effect probably is small compared to the variability in hydraulic conductivity arising from rock-matrix heterogeneity and inherent errors of measurement. The thermal expansivity of water would produce a change in water density of no more than about 0.5 percent and can thus be regarded to be a negligible effect.

A more likely but untested effect of the geothermal gradient within the unsaturated zone would be to induce an upward migration of moisture as water vapor, compensated by a downward return flow of liquid water. As described briefly in Section 3.9.3, the physics of the phenomenon is based on the presumption that liquid water and water vapor are in local thermodynamic phase equilibrium within the pore and fracture space and, thus, that the dependence of temperature on depth will produce a water-vapor concentration gradient aligned with the geothermal gradient. Consequently, a diffusive flux of water vapor would be induced upwards parallel to the geothermal gradient. Ross (1984) analyzes the efficacy of such a process within a porous medium having properties and under conditions similar to those expected for the Yucca Mountain site. He concludes that unless the net infiltration rate is less than 0.03 mm/yr the effects of upward water vapor diffusion are likely to be negligible compared to the matric-potential driven movement of liquid water. However, this transport mechanism may be considerably more pronounced in highly fractured unsaturated media in which the

upward movement of water vapor could occur through the air-filled fracture openings, especially if driven additionally by bulk-gas flow (Montazer and Wilson, 1984; Montazer et al., 1985). At Yucca Mountain this effect could be of significance within the thick, fractured, repository TSw unit at the low rates of liquid water percolation (0.5 mm/yr) that may be occurring with this unit. Using temperature data obtained from test well USW UZ-1, Montazer et al. (1985) conclude that the ambient temperature gradient could produce an upward bulk-gas flux of 1.25 to 2.5 m/yr within the unsaturated zone. If the gas were saturated with respect to water vapor, such upward flow would produce an upward advective moisture flux of about 0.025 to 0.05 mm/yr under present conditions. Gas-pressure monitoring in test well USW UZ-1 shows a well-defined barometric response. It is unknown whether gas pressure variations at depths appropriate to the repository TSw unit would be sufficient to induce significant bulk gas flow. Nevertheless, the analysis of the hydrologic system within the unsaturated zone at Yucca Mountain, especially under nonisothermal, barometrically varying conditions, must consider the simultaneous, coupled flow of gas and moisture, the latter occurring both as liquid water and water vapor. Field measurements in surface-based boreholes, experiments in the exploratory shaft, and theoretically based modeling studies, as described in Section 8.3.1.2.2, are intended to address these issues.

3.9.4 GROUND-WATER VELOCITY AND TRAVEL TIME

Issue 1.6 of the issues hierarchy (discussed in the introduction to this document) addresses the performance objective defined in 10 CFR 60.113(a)(2): "The geologic repository shall be located so that pre-waste-emplacement ground-water travel time along the fastest path of likely radionuclide travel from the disturbed zone to the accessible environment shall be at least 1,000 years...". The following discussion addresses this issue, focusing on that portion of the hydrogeologic system extending from the disturbed zone to the accessible environment. The components of this system which must be considered include: (1) the hydraulic properties of the formations through which the water will flow; (2) the present hydrologic conditions in the vicinity of the proposed repository; and (3) the locations and lengths of the flow paths between the disturbed zone and the accessible environment. The data and information required to assess the influences of these components on ground-water velocity and travel times are presently limited, and will be supplemented by the hydrologic investigations discussed in Section 8.3.1.2. The preliminary estimates of ground-water velocities and travel times presented here are based on sparse data and qualitative observations of hydrologic processes from limited field investigations.

Issue 1.6 addresses ground-water flow from the edge of the disturbed zone to the accessible environment. Based on the NRC Draft Generic Technical Position Paper on the Disturbed Zone (NRC, 1986b) the edge of the disturbed zone will be assumed conservatively to be 50 m below the repository level for this preliminary analysis. The actual extent of the disturbed zone will be determined during site characterization (Section 8.3.5.12.5). Based on the proposed amendments (NRC, 1986a) to 10 CFR Part 60, the accessible environment is located beyond 5 km from the outer boundary of the repository underground facility.

CONSULTATION DRAFT

Section 3.9.3 presents the current conceptual model of the ground-water flow system at Yucca Mountain. This model (Figure 3-46) considers water moving downward beneath the repository in the unsaturated zone until it reaches the water table. Flow in the unsaturated zone is predominantly in the vertical direction, although lateral flow is a possibility. Once it reaches the water table, the flow in the saturated zone is generally horizontal to the accessible environment in a direction controlled by both the hydraulic gradient and the properties of the fractured medium. It is expected that flow to the accessible environment will be generally down gradient, with some directional variability induced by the anisotropy of the fractured rock matrix. The travel time estimates made below are based on this model, because it appears to be the most reasonable based on present understanding of the flow system and the present limited data base. As more data become available and our understanding of the system improves, the model will be modified, and the resulting travel time estimates adjusted accordingly.

3.9.4.1 Ground-water travel time in the unsaturated zone

Estimates of pre-waste emplacement ground-water travel times through the unsaturated zone have been made using the model described in Sinnock et al. (1986). This model makes the following assumptions:

1. Moisture flow in the unsaturated zone occurs under steady state conditions in which the vertical flux of moisture within the deep unsaturated zone is equal to the ambient hydraulic conductivity.
2. The unsaturated-zone flux below the disturbed zone is vertically downward and uniformly distributed in time and space.
3. The ambient (effective) hydraulic conductivity through the matrix varies spatially as a function of saturation. Under conditions of a unit vertical hydraulic gradient (Weeks and Wilson, 1984), the vertical volumetric flux through the matrix becomes numerically equivalent to the unsaturated matrix hydraulic conductivity at the existing saturation level. As the saturation reaches 100 percent, the matrix is assumed to conduct water at a rate numerically equivalent to the saturated conductivity. Any remaining flux is assumed to travel through fractures, even though fracture flow may be initiated at lower saturation values.
4. Water does not move rapidly through fractures that are not connected to the surface until fluxes approach the saturated matrix hydraulic conductivity. This is due to the strong negative capillary pressures exerted by the pores of the matrix, which draw water away from the fractures (Sinnock et al., 1984; Wang and Narasimhan, 1984; Montazer and Wilson, 1984).

Parameters used in the travel-time calculations are shown in Table 3-30 (Sinnock et al., 1986). In addition, the initial flux was taken to be 0.5 mm/yr. This figure is based on Wilson (1985) using "two lines of evidence:

Figure 3-46. Conceptual hydrogeologic section from Solitario Canyon northwest of the site, to well J-13 in Jackass Flats. Modified from Scott and Bonk (1984).

Table 3-30. Parameters used in travel-time calculations for the unsaturated zone^a (page 1 of 2)

Parameter	Hydrogeologic unit ^b							Remarks
	TSw	CHnv	CHnz	PPw	PPn	BFw	BFn ^c	
Vertical hydraulic gradient, i	1.0	1.0	1.0	1.0	1.0	1.0	1.0	$i = \text{grad} (\psi + z)$, where ψ is the pressure head and z the elevation head. If $ \text{grad } \psi \ll 1$, $i_x \approx i_y \approx 0$ and $i_z \approx 1$. This implies ^y vertical gravity flow.
Geometric mean saturated-matrix hydraulic conductivity $\bar{K}_s$ (mm/yr) ^d	0.72 (31)	107 (8)	0.54 (31)	88 (10)	22 (7)	118 (2)	22 (NA) ^e	$K_s = \exp (\text{mean} [\ln (K_s)])$ Values in parentheses are the number of measurements.
$\bar{K}_s \exp (-\sigma [\ln K_s])$	0.12	1.9	0.037	29	3.3	58	3.3	Values of K_s that correspond to a ± 1 standard deviation of $\ln(k_s)$ ($\pm \sigma [\ln K_s]$) around the arithmetic mean of $\ln (K_s)$.
$\bar{K}_s \exp (+\sigma [\ln K_s])$	4.1	6,090	7.6	261	142	240	142	
Mean effective porosity $\bar{n}_e \pm 1\sigma [n_e]$	0.11 $\pm .05$ (138,12)	0.32 $\pm .09$ (23,6)	0.27 $\pm .05$ (65,10)	0.24 $\pm .06$ (27,4)	0.25 $\pm .06$ (75,2)	0.22 $\pm .09$ (120,2)	0.25 $\pm .06$ (NA)	$\bar{n}_e = \bar{n}_b (1 - \bar{S}_r)$, where n_b is the mean bulk, dry porosity and $\bar{S}_r$ is the mean residual saturation. The standard deviation $\sigma [n_e] = (1 - \bar{S}_r) \sigma [n_b]$. Ordered pairs in parentheses are number of measurements of n_b and S_r , respectively.
Range of thicknesses	0-72	0-135	0-133	0-44	0-122	0-91	0-55	Thicknesses between disturbed zone and water table for area

Table 3-30. Parameters used in travel-time calculations for the unsaturated zone^a (page 2 of 2)

Parameter	Hydrogeologic unit ^b							Remarks
	TSw	CHnv	CHnz	PPw	PPn	BFw	BFn ^c	
Range of thicknesses (continued)								within the design repository boundaries.
(m) ^f	(98.5)	(95.3)	(94.5)	(83.2)	(63.1)	(25.6)	(7.5)	Values in parentheses are percentages of total repository area underlain by the units.
Empirical constant ^e	5.9	4.2	7.0	4.0	5.2	4.6	5.2	Empirical constant that represents the effects of the relationship between pore-size distribution and saturation on the amount of the effective porosity, n_e , available for flow; the effect of the empirical constant is to reduce flow area and thus increase particle velocity relative to values calculated using q/n_e .

^aSource: Sinnock et al. (1986).

^bTSw = Topopah Spring welded unit; CHnv = Calico Hills vitric unit; CHnz = Calico Hills zeolitic unit; PPw = Prow Pass welded unit; PPn = Prow Pass nonwelded unit; BFw = Bullfrog welded unit; BFn = Bullfrog nonwelded unit.

^cAssumed to be hydrologically identical to PPn.

^dSaturated conductivity and effective-porosity data are from Sandia National Laboratories Tuff Data Base (SNL, 1985).

^fRange of thickness, Sandia National Laboratories Interactive Graphics Information System (IGIS) (SNL, 1985).

^eNA = Not available.

CONSULTATION DRAFT

(1) calculations of flux in the proposed host rock, based on field and laboratory evidence; and (2) an estimate of the recharge rate beneath the Yucca Mountain area, based on regional relationships developed among precipitation, altitude, and recharge rate."

Table 3-31 provides a summary of the travel times for the 0.5 mm/yr flux (Sinnock et al., 1986). As this table shows, the minimum (fastest) travel time through the unsaturated zone of more than 9,000 yr greatly exceeds the 1,000 yr required to satisfy the issue.

In an attempt to assess the sensitivity of the travel time to variations in flux, Sinnock et al. (1986) used an initial flux of 1 mm/yr, double the bounding value estimated by Wilson (1985). This analysis resulted in a minimum travel time of about 3,700 yr, still greater than the amount of time required to satisfy the issue.

The analysis presented here and discussed in detail in Sinnock et al. (1986) and DOE (1986) is considered preliminary. However, the modeling effort has attempted to use the best available data, and it is believed that the results obtained are realistic. Plans to develop this model further and to extend the data base are presented in Section 8.3.1.2.2.

3.9.4.2 Ground-water travel time in the saturated zone

For the saturated zone, the assumed flow path extends from the eastern edge of the primary repository area southeastward for 5 km to the accessible environment (Figures 3-28 and 3-25) through the tuffaceous beds of Calico Hills and the welded Topopah Spring Member or the welded Crater Flat Tuff (Prow Pass or Bullfrog Member). Estimates for ground-water travel times along this travel path have been made using the following assumptions (DOE, 1986):

1. Darcian flow applies.
2. Flow paths are parallel to the hydraulic gradient and are nearly horizontal.
3. The water-level measurements shown in Figure 3-28 (Robison, 1986) provide a reasonable estimate for the hydraulic gradient along the flow path.
4. The system is isotropic within each unit, and hydraulic conductivity values obtained from hydraulic tests of wells in the southeastern part of the Yucca Mountain area are representative of the values along the flow path.
5. Calculated effective porosities from Sinnock et al. (1984) are of a magnitude that will tend to result in minimum travel times calculated for flow in the saturated tuffaceous beds of Calico Hills.

The hydraulic gradient has been estimated using water-level altitudes of 730.4 m at drillhole UE-25b#1 and 728.3 m at well J-13 (Figure 3-28). An

Table 3-31. Summary of ^aunsaturated zone travel time for vertical flux of 0.5 mm/yr^a

Travel path ^b	Travel time (yr)
Minimum	9,345
Mean	43,265
Maximum	80,095

^aSource: Sinnock et al. (1986).

^bTravel paths are from the repository horizon to the water table and are based on multiple modeling scenarios as discussed in Sinnock et al. (1986).

examination of the data for well UE-25P#1 (Craig and Robison, 1984) suggests that the water table is located approximately at the interface between the Calico Hills and the Topopah Spring units at this point (approximately 3,000 m from both the edge of the repository and well J-13). Using the principle of conservation of mass, hydraulic gradients have been estimated for each of the units as shown in Table 3-32. Uncertainties in these estimates include the following components: (1) with very low gradients, small errors in the measurements of water levels can have significant effects on the value of the gradient; (2) the measured water level at well J-13 could be expected to be lower than the static water level because of pumping, thereby resulting in a steeper estimated gradient; and (3) vertical components of flow may be present. Because vertical flow would have the effect of lengthening the flow path, the assumption of horizontal flow probably is conservative, although Waddell et al. (1984) indicate the controls on vertical and horizontal flow at Yucca Mountain are variable and generally unknown.

Table 3-32 shows these values and provides estimates for travel times through the Calico Hills and the Topopah Spring units. On the basis of these estimates, the cumulative travel time through the saturated zone is about 172 yr. This value is different from that presented in DOE (1986), due to the use of well UE-25b#1 instead of USW H-4; the former was chosen in this analysis as being more representative of regional trends in the hydraulic gradient (Figure 3-28).

This estimate of travel time is considered to be conservative, based on the assumptions mentioned above and on the fracture effective porosities reported by Sinnock et al. (1984). These effective porosities were calculated by multiplying the fracture density from Scott et al. (1983) by the effective aperture calculated from a relationship provided by Freeze and Cherry (1979). The resulting effective porosities are considered to be reasonable estimates for fracture flow in the saturated zone. However, empirical estimates of saturated effective porosity for both matrix and fracture flow range from 8 to 12 percent for the Topopah Spring welded unit (Sinnock et al., 1984); 2.7 to 8.7 percent for Topopah Spring Member

CONSULTATION DRAFT

Table 3-32. Estimates for ground-water travel times through the saturated zone

Parameter	Unit	
	Tuffaceous beds of Calico Hills	Topopah Spring Member
Length of path (m)	3,000	2,000
Hydraulic conductivity (m/yr) ^a	69	365
Hydraulic gradient ^b	5.9×10^{-4}	1.1×10^{-4}
Darcy velocity (m/yr) ^c	4.1×10^{-2}	4.0×10^{-2}
Calculated ^d bulk effective fracture porosity	4×10^{-4}	2.8×10^{-3}
Particle velocity (m/yr) ^e	103	14
Travel time (yr) ^f	29	143

^aHydraulic conductivity for Calico Hills from Lahoud et al. (1984) and Thordarson (1983); Topopah Spring from Thordarson (1983).

^bBased on water levels at drillhole UE-25 b#1 and at well J-13 (see Figure 3-28). The estimates for the hydraulic gradient for each of the units has been based on conservation of mass between drillhole UE-25b#1 and well J-13.

^cDarcy velocity = (hydraulic conductivity) x (hydraulic gradient).

^dData from Sinnock et al. (1984).

^eParticle velocity = (Darcy velocity)/(bulk effective porosity).

^fTravel time = (length of path)/(particle velocity).

(Thordarson, 1983); and 20 to 30 percent for the Calico Hills vitric unit (Sinnock et al., 1984). Use of more realistic (less conservative) values of effective porosity would be likely to lead to a saturated ground-water travel time at least ten times greater (i.e., approximately 1,700 yr) than that indicated in Table 3-32.

3.9.4.3 Retardation and thermal effects

The ground-water travel times estimated for the saturated and unsaturated zones at Yucca Mountain are presently based on Darcy's law and Richard's equation. In fact, ground-water flow through fractured unsaturated media tends to be dominated by the matrix hydraulic conductivity. Water moving through a fractured system may travel into and out of the matrix from

the fractures as it moves down gradient (DOE, 1986). This is due to the differences in matric, osmotic, thermal, and pressure potentials across the fracture-matrix interface. This leads to a lengthening of the ground-water travel path for individual ground-water molecules, thereby increasing ground-water travel times.

Other factors may affect ground-water travel time, including dispersion, the existence of faults or impermeable zones along the travel path, the vertical movement of water in the saturated zone, and the upward movement of moisture in the unsaturated zone. At this time it is uncertain whether some or all of these mechanisms exist along the travel path. Efforts will be made during site characterization to determine the existence of these and other features, and also to estimate the effect of these on pre-waste-emplacement ground-water travel times.

The possible effects of thermal gradients on the ground-water flow system at Yucca Mountain are unknown at this time. Thermal gradients may be causing upward flow in the unsaturated zone (Wilson, 1985), although this is only conjective at present. As mentioned in Section 3.7.3, little information exists on thermal gradients in the saturated zone. The information available is from isolated locations that give little indication of the thermal gradient in the vicinity of Yucca Mountain. Such information will be collected during site characterization, and its effect on the ground-water flow system will be evaluated at that time (Section 8.3.1.2.2).

3.9.5 HYDROCHEMICAL CONFIRMATION OF GROUND-WATER BEHAVIOR

This section presents information concerning hydrochemical evaluation of the ground-water system, and the question of whether the site will meet the performance objective for pre-waste emplacement ground-water travel time as required by 10 CFR 60.113(a)(2) (Issue 1.6). Section 3.9.1.3 provides data and descriptions of the hydrochemical evaluation of the ground-water system, including possible origin, travel paths and flow rates. Section 3.7.3 provides further hydrochemical data for the hydrogeologic study area. This section deals primarily with problems associated with interpreting ground-water velocities derived from radiocarbon data.

3.9.5.1 Unsaturated zone

Complete hydrochemical or isotopic analyses are not available for pore water taken from the site's unsaturated zone. Methods to extract this water in an uncontaminated state are still under development. Yang (1986) presents preliminary calcium and sodium concentration data; his studies indicate that calcium concentrations are elevated in the unsaturated zone and dominate over sodium. Section 8.3.1.2.2 presents plans for characterizing the chemical and isotopic composition of the pore water. This information will be used to understand potential interactions of pore water with the waste canister and to define flow paths and velocities in the unsaturated zone.

3.9.5.2 Saturated zone

Summaries of the hydrochemical characteristics of the water in the saturated zone in the Yucca Mountain area, of the hydrochemical mechanisms controlling these characteristics, and of the hydrochemical evidence for ground-water origin, aquifer mixing and ground-water residence time are presented in Sections 3.7.3 for the hydrogeologic study area and in Section 3.9.1.3 for the site. As discussed in Section 3.9.1.3, only limited data exist in the immediate vicinity of Yucca Mountain, and those data tend to be mainly radiocarbon data. This section elaborates on the difficulties in interpreting ground-water velocities from radiocarbon data which have not been adequately modeled for initial carbon-14 activity. Plans for obtaining additional hydrochemical data are presented in Sections 8.3.1.2.2 for the unsaturated zone and 8.3.1.2.3 for the saturated zone.

The reported radiocarbon dates of ground water at Yucca Mountain have been uncorrected, apparent ages (Section 3.7.3.2.1) (Benson and McKinley, 1985; Claassen, 1985). Given a series of assumptions examined in more detail below, such ages may represent the time since ground water was isolated from the CO_2 gas reservoir of the atmosphere or the unsaturated zone (i.e., the time since the water entered the water table beneath Yucca Mountain). Uncorrected ages are realistic when no carbon-12 dilution has occurred to the initial carbon-14 by the dissolution of carbon-14-free carbonate minerals. Claassen (1985) indicates that this is the case in the tuffaceous aquifer, since these formations are virtually carbonate-free. If such dilution occurs, the apparent radiocarbon ages may be an upper limit of the ground-water age (Section 3.7.3.2.2).

The other major assumption underlying the apparent ages relates to the initial carbon-14 composition of the infiltrating water. Uncorrected ages assume that this water was in isotopic equilibrium with a gaseous reservoir at 100 percent modern carbon (pmc). This is generally valid, since both the atmosphere (before nuclear weapons testing) and soil gas (gas in the unsaturated zone in a well mixed or low CO_2 gas-residence-time system) have this composition. In a poorly mixed system with long CO_2 gas residence times where the carbon-14 in the CO_2 gas has decayed, the radiocarbon reservoir is at less than 100 pmc. In such a case, the apparent age of the ground water is greater than the time it has been in the aquifer (by an amount proportional to the difference between 100 pmc and the carbon-14 concentration in the equilibrating CO_2 gas).

Yang et al. (1985) have shown that the carbon-14 concentration of the CO_2 gas in the soil zone can be as low as 34.2 (± 1.5) pmc in the unsaturated zone at 368 m below the surface. The apparent age of a ground-water sample equilibrated with such gas is really equal to the age of the gas in the unsaturated zone (without isotopic mixing) plus the residence time of the ground water in the aquifer since equilibration (without isotopic dilution). The apparent age is thus older than the true ground water residence time. Since isotopic re-equilibration with CO_2 gas can occur throughout the unsaturated zone, apparent ages of the unsaturated zone water (time since precipitation) are not reliable.

Well documented and tested models of initial ground-water carbon-14 activity exist. Some of these models (Mook, 1976, 1980; Fountes and Garnier,

1979) are particularly well suited for modeling waters which equilibrate with CO_2 gas in the unsaturated zone and then enter the water table. Such models can be used to examine the apparent difficulties in interpreting carbon-14 data from Yucca Mountain (Section 8.3.1.2 for plans for their use). Such models, which are said to give corrected ages, do not actually perform corrections of an error, but rather, they simply do not ignore the hydrochemical processes which control carbon-14 behavior.

Kerrisk (1987) has pointed out that at Yucca Mountain, gaseous CO_2 from the unsaturated zone may be added to or exchange with the carbonate of the saturated zone water. This process could lead to apparent ages that are younger than the true ages of the water in the saturated zone.

An accurate knowledge of initial carbon-14 activity is less important in calculating relative ground-water ages or velocities than it is in determining absolute chronologies. Ground-water velocities can be estimated by dividing the distance between two points along a hydrologic flow line by the carbon-14 age difference of the water at the two points. Consistent errors in estimating initial activity tend to cancel, resulting in relative ages which are more accurate than are the individual absolute ages. This is not the case when calculating the velocity of water to a point where the carbon-14 age is known from the assumed point of present day recharge (absolute age zero). Mixing (i.e., the two points are not actually on the same hydrodynamic flow line or the line has moved during the transit of the parcel of water) is also not considered in this calculation. These two difficulties apply to the results of Claassen (1985) who states, "For example, assume recharge near the head of Fortymile Canyon occurred 17,000 years ago and flowed to the lower end of the Canyon, where it is sampled. The velocity that is calculated, 7 m/yr, must be a maximum. Because recharge may occur anywhere but not necessarily everywhere along surface drainageways, no probable minimum velocity can be calculated. The absence of water older than about (apparent age) 10,000 yr. B.P. even near the head of Fortymile Canyon, would favor velocities slower than 4 m/yr." The planned use of initial carbon-14 modeling and relative dating along flow paths to resolve apparent difficulties of interpretation is described in Chapter 8.3.1.2.3.

3.9.6 MONITORING AND VERIFICATION

This section discusses the monitoring and verification program designed to assess hydrologic conditions at and near Yucca mountain. The program will also provide an historical background to permit detection of changes in baseline conditions during repository construction, operation, and following closure.

Data from the existing monitoring program are included in Section 3.9.1. The current program consists of 25 boreholes which penetrate the saturated zone and a single instrumented hole in the unsaturated zone. Details of the construction, instrumentation, and monitoring of these holes is discussed in Section 3.9.1.

CONSULTATION DRAFT

The primary functions of the program are to provide a data base against which future observations maybe compared, to provide information for verification of ground-water models, to evaluate parameters for use in travel-time calculations, and to collect data for the construction of a potentiometric-surface map and determination of ground-water flow directions. The current program provides this information and will be continued. However, additional data are required to provide the level of detail necessary for a thorough understanding of the hydrologic system at Yucca Mountain. Additional data on ground-water geochemistry, water potential temperature variations, etc., will be collected to fulfill the data requirements of the various conceptual models proposed for the site (Sections 8.3.1.2.2 and 8.3.1.2.3).

In order to provide an expanded data base, additional test holes will be added to the monitoring network during site characterization. The new holes will be used for various types of data aquisition, including determination of the hydraulic nature of structural features, measurement of the hydraulic gradient, estimation of ground-water flowpaths, and calculation of various hydraulic parameters in both the saturated and unsaturated zones. Detailed plans for these holes are included in Section 8.3.1.2 along with specific plans for monitoring and use of the resultant data.

3.9.7 LOCAL GROUND-WATER USERS

This section identifies all production wells that are located near the Yucca Mountain site. It is possible that withdrawals from these wells could affect the local flow system. Available pumping information that will aid in identifying these effects is presented. These data, in conjunction with results from planned modeling and field studies will be used to quantify and qualify any potential changes to the local flow system. Plans for further studies are presented in Section 8.3.1.2.

There are two wells located proximally to the Yucca Mountain site that withdraw water on a regular basis. Wells J-12 and J-13, located approximately 12 km and 8 km southeast of the Yucca Mountain site, respectively, are utilized and controlled by the DOE to support activities in the southwestern part of the Nevada Test Site (refer to Section 3.8.1 for a discussion of regional ground-water users).

Water from wells J-12 and J-13 is produced from the Topopah Spring Member of the Paintbrush Tuff, which constitutes the lower portion of the welded tuff aquifer (Section 3.6.1). The Topopah Spring Member occurs at a depth of approximately 180 to 347.2 m at well J-12, and 207.3 to 449.6 m at well J-13 (Thordarson et al., 1967; Claassen, 1973). Pumping-test and water-level data for both wells are shown in Table 3-33. Well-construction information can be found in Thordarson et al. (1967), Thordarson (1983), Young (1972), and Claassen (1973). Table 3-34 summarizes water production for wells J-12 and J-13 from 1983 through 1985. Water-quality data are presented in Section 4.1.2.2 of Chapter 4.

Wells J-12 and J-13 have been pumped intermittently since their completion. Before 1968, well J-12 was pumped at rates in excess of 2,000 m³/d. After the wellbore was cleaned and deepened in 1968, production capabilities

CONSULTATION DRAFT

Table 3-33. Pumping test and water-level data for wells J-12 and J-13 from 1960 through 1969^a

Date of pumping test	Water level (m) Datum: MSL ^b	Specific capacity (m^3/d)/m	Pumping rate (m^3/d)	Draw-down (m)	Duration of test (h)
WELL J-12					
01/27/60	727.6	-- ^c	--	--	--
11/01/60	--	1,700	2,080	1.2	2
08/22/62	726.7	1,000	2,110	2.10	1
07/27/65	--	2,800	2,020	0.73	83
07/25/68 ^d	727.0	220	4,250	19.54	1
08/24/68 ^d	727.3	850	4,880	5.78	5
04/21/69	727.3	1,600	4,850	3.02	4
WELL J-13					
02/18/64	728.5	420	3,800	9.11	3
02/18/64	727.9	310	2,800	12.31	96
03/31/64	729.4	--	--	--	--
04/21/69	728.2	540	3,640	6.77	4

^aModified from Claassen (1973).

^bMSL = mean sea level.

^c-- indicates no data.

^dWell bore cleaned and deepened to 347.2 m. Test data indicated downhole sloughing, resulting in decreased hydraulic conductivity. Original total depth: 270.4 m.

Table 3-34. Water production from wells J-12 and J-13 from 1983 through 1985^{a, b}

Well	1983		1984		1985	
	Gallons	Acre-feet ^c	Gallons	Acre-feet	Gallons	Acre-feet
J-12	25,498,500	78.2	26,058,400	80.0	25,049,800	76.9
J-13	42,148,700	129.3	40,349,300	123.8	37,811,000	116.0
Total	67,647,200	207.5	66,407,700	203.8	62,860,800	192.9

^aProduction figures are for water year, October 1 through September 30.

^bSource: Witherill (1986).

^cTo convert from acre-feet to cubic meters, multiply by 1233.489.

CONSULTATION DRAFT

exceeded 4,500 m³/d. Well J-13 has been shown to be capable of producing in excess of 3,500 m³/d of water (Claassen, 1973). The specific capacity of the well underwent a slight decline in 1964 shortly after completion. Claassen (1973) hypothesized that the well was not fully developed during the 1964 tests. Subsequent pumping tests performed in 1969 support this hypothesis, and the specific capacity of 540 m³/d/m is considered the most reliable measurement.

Pumping tests performed after 1968 on well J-13 produced drawdowns of less than seven meters with pumping rates in excess of 3,500 m³/d (Table 3-33). Figure 3-47 indicates that even with well J-13 in continuous service, declines in the water level have been minimal. This suggests that the short-term effects of pumping on the regional potentiometric surface are probably negligible. The effects of long-term pumping on the local potentiometric surface are unknown and remain to be investigated (Robison, 1984). Tests planned during site characterization for evaluating the potential effects of repository development on the local potentiometric surface are presented in Sections 8.3.1.2.3.1 and 8.3.1.16.2.1.

The water which will be needed for site characterization activities probably will be drawn from wells J-12 and J-13. It is possible that a substitute source may be located and developed (Morales, 1986). Preliminary estimates of the amount of water necessary for site characterization activities, and the construction and operation of a geologic repository at Yucca Mountain, are presented in Section 3.8.1.

Other drillholes exist near the site but are pumped only periodically to obtain hydrologic and hydrochemical information. These test holes have not been used to supply water. Section 1.6 provides a discussion of all existing exploratory drillholes in the Yucca Mountain area.

3.9.8 PALEOHYDROLOGY

Discussion of the paleohydrology of the south-central Great Basin, including Yucca Mountain, is presented in Section 3.7.4. In this section, the discussion is limited to those past and probable future hydrologic conditions that may directly affect a repository at Yucca Mountain.

Winograd and Szabo (1986) concluded that there was a progressive lowering of water table in the south-central Great Basin during the Quaternary, and that "a continued decline of the regional water table in the next 10⁵ to 10⁶ yr (and beyond?) in response to increasing aridity and to lowering of ground-water base level" is likely (Section 3.7.4). They noted, however, that this progressive water-table decline "does not preclude superimposed and relatively rapid cyclical fluctuations in water level in response to glacial (i.e., pluvial) and interglacial climates of the Pleistocene." Preliminary field evidence and the results of a computer-model analysis bearing on past and possible future water levels beneath Yucca Mountain are outlined below.

Bish and Vaniman (1985) noted that nonwelded glass occurs both above and below the welded zone in the Topopah Spring Member (Tptw) of the Paintbrush Tuff. Bish and Vaniman (1985) state "the lower nonwelded vitric zone thins

Figure 3-47. Water levels in well J-13 since completion. Modified from Thordarson (1983).

and disappears to the east where stratigraphic dip and structural displacements bring the basal Tptw glassy zone closer to the static-water level. The vitric nonwelded material may have important paleohydrologic significance because the preservation of open shards and pumice made of nonwelded glass is rare below past water levels (Hoover, 1968)." Where present, the base of the lower nonwelded vitric tuff occurs about 80 to 100 m above the present water table. On the basis of mineralogy, Bish and Vaniman (1985) concluded, "The only apparent change in phase assemblage near the water table in Yucca Mountain is the alteration of vitric tuff of Calico Hills and lower Topopah Spring Member." The observations by Bish and Vaniman (1985) suggest that past water levels beneath Yucca Mountain may never have been more than 100 m higher than the modern water levels.

To help assess the implication in Bish and Vaniman (1985) that the lowest position of vitric nonwelded tuff might be related to a stand of a paleo-water table, a contour map of the base of the lower vitric nonwelded tuff (in both the Topopah Spring Member and in the tuffaceous beds of the Calico Hills) will be prepared and contrasted with the modern water table. If such a contour map indicates (1) a relatively smooth surface at the base of the vitric nonwelded tuff, (2) a surface that locally cross-cuts stratigraphic horizons in the tuffs, and (3) a surface that climbs sharply to the north as does the modern water table beneath Yucca Mountain, then it is possible that the porous nonwelded vitric tuffs have indeed been altered below a paleo-water table, as suggested by Bish and Vaniman (1985). Such a contour map would then presumably mark the stand of water table at the time or times of alteration of the tuffs. The occurrence of such a stand probably could not be dated, but might have occurred several million years ago. Even in the absence of a date for the alteration, the base of the nonwelded vitric tuff shown on Bish and Vaniman's cross sections (1985) is tens of meters to more than 100 m below the base of the proposed repository horizon. Thus, if the previously discussed work demonstrates the likelihood of a relationship of the paleowater table to the lowest occurrence of vitric tuff, a possible maximum (though undated) water-table stand would have been identified that is significantly lower than the repository horizon under consideration. Section 8.3.1.3 discusses the investigations planned to characterize the mineralogy of Yucca Mountain.

Czarnecki (1985) addressed the issue of possible past, and by implication, future pluvial-related water-table rises beneath Yucca Mountain with a two-dimensional finite element ground-water flow model of the region, together with assumptions about pluvial recharge on, and underflow into and from the region. He calculated a maximum increase in water-table altitude of about 130 m beneath the site of the proposed repository. Inundation of any part of the planned repository would require a water-table rise of more than 200 m. His analysis also suggests a past shortening by two-thirds of ground-water flow paths from Yucca Mountain to discharge areas in the southern Amargosa Desert. These discharge areas are still beyond the boundary of the accessible environment.

The analysis by Czarnecki (1985) may be considered conservative for several reasons. A 100-percent increase in precipitation during pluvial periods was assumed, resulting in a 15-fold increase in recharge from the modern rate (0.5 mm/yr) to about 8 mm/yr at Yucca Mountain. Czarnecki (1985) acknowledges that this recharge value, which was derived from the empirical

Maxey-Eakin water-budget method, probably is high. Half of the calculated recharge flux in the model was applied directly east of the proposed repository site, along a segment of Fortymile Wash. This flux causes about three-quarters of the computed water-table rise of 130 m. Czarnecki (1985) notes, however, that under a 100-percent increase in precipitation, large quantities of runoff might flow away from the area down Fortymile Wash and other drainageways. This would have the effect of decreasing the effective ground-water recharge to much less than the calculated values (Czarnecki, 1985).

Other considerations were noted by Czarnecki (1985) that might have resulted in a greater simulated water-table rise. Recharge into Fortymile Wash was limited to the main stream channel near Yucca Mountain. If recharge had been included along the full length of Fortymile Wash and its distributaries, which extend beyond the town of Amargosa Valley, then a greater water-table rise would have been simulated. Simulated development of discharge areas southeast of the town of Amargosa Valley helped to limit the water-table rise beneath the primary repository area. If this discharge were decreased because of the possible existence of marsh deposits or eolian silts or because of a greatly decreased ratio of vertical hydraulic conductivity to horizontal hydraulic conductivity, then the simulated water-table rise might be greater. The model used by Czarnecki (1985) derived its parameters from a model presented in Czarnecki and Waddell (1984), which assumed that the ground-water system was in steady-state conditions. If the ground-water system was still equilibrating to recharge that may have occurred 10,000 to 20,000 years before present, then the values of transmissivity used in the model in Czarnecki (1985) might be too high. Large transmissivities used in recharge simulations would produce less water-table rise than if smaller transmissivities were used.

Extensive investigations are planned to better quantify the parameters of the hydrologic system in the Yucca Mountain region. These parameters will be used to model the hydrologic system more effectively and any variations induced by climatic or tectonic changes. Plans for these studies are discussed in Sections 8.3.1.2.1.4 and 8.3.1.2.2.9.

During pluvial periods of the Pleistocene, perched springs and seeps could have occurred along the flanks of Yucca Mountain and these might have resulted from recharge along the exposed up-dip portion of tuff units, with water moving down-dip to discharge sites along the flank of the mountain. No conclusive evidence for former springs has been observed, but additional investigations are planned, as outlined in Section 8.3.1.5.2.

Deposits along fault zones may provide additional evidence for changed hydrologic conditions. Along portions of the Bow Ridge fault, 1 km east of the eastern boundary of the proposed repository, Trench 14 has exposed deposits of carbonate, opal, and minor amounts of sepiolite (Vaniman et al., 1985; Taylor and Huckins, 1986; Voegelé, 1986a,b). These deposits are discussed more extensively in Section 1.2.2.2.10. The origin of these deposits is under study, and the following hypotheses are being considered:

1. The deposits are the result of ground-water discharge along the fault, from the deep regional flow system and reflect former higher water levels, tectonic uplift of the deposits, or a combination of both.

CONSULTATION DRAFT

2. The deposits are the result of discharge along the fault zone from a shallow ground-water flow system along the flank of Yucca Mountain.
3. The deposits are shallow soil features that resulted from rain water that moved downslope from the top of a nearby low hill, along the contact between the soil zone and bedrock, and accumulated at the fault zone. The water would have evaporated slowly, leaving precipitates of the minerals that were dissolved in the water. Present evidence from studies discussed in Section 1.2.2.2.10 supports this hypothesis (Vaniman et al., 1985; Taylor and Huckins, 1986; Voegele, 1986a).

Further investigations designed to determine the nature and origin of these deposits are discussed in Section 8.3.1.5.2.

3.10 SUMMARY

Chapter 3 discusses the data that are available and additional data that are needed for understanding the hydrologic aspects of the site. These data will be used to assist in resolving many of the postclosure and preclosure issues within the NNWSI Issues Hierarchy. This section summarizes the data that have been presented, and the plans identified in Section 8.3.1.2 for collecting the additional data needed to satisfy the performance and design issues.

3.10.1 SUMMARY OF SIGNIFICANT RESULTS

One of the primary issues to be satisfied during site characterization is: Do the data collected in order to describe the geohydrologic characteristics at the Yucca Mountain site provide the information required by the design and performance issues? Investigations of the geohydrologic system described in Section 8.3.1.2 include (1) understanding the regional hydrogeologic system; (2) describing the unsaturated-zone hydrologic system at the Yucca Mountain site; (3) describing the saturated zone hydrologic system at the site; and (4) presenting information required to support post-closure repository performance which requires that the geohydrologic setting be compatible with waste containment and isolation under both anticipated and unanticipated conditions and that ground-water travel time along the fastest pathway of likely and significant radionuclide travel be not less than 1,000 years.

The region surrounding Yucca Mountain is one of the most arid parts of the United States. The climate is typified by low rainfall, low humidity, high evaporation; high summer temperatures, and high wind velocities during the spring. Precipitation generally increases with elevation; average annual amounts range from 43 mm in Death Valley to 760 mm on the highest peaks of the Spring Mountains (Section 3.1).

Collection of precipitation data is planned throughout the region in order to characterize better the present-day precipitation as a function of

topographic setting; rainfall intensity and duration characteristics are needed for rainfall-runoff models and for infiltration studies. Modern meteorological conditions also form a basis for interpreting historical records and predicting future conditions. The plans for collecting these data are presented in Section 8.3.1.2.1, 8.3.1.2.2, and 8.3.1.12.1.

Because of generally arid climatic conditions, there are no perennial streams in the regional hydrologic area. The limited surface water that does exist is restricted to short spring-fed reaches of the Amargosa River, small pools at some large springs, and some large springs and marshes in Alkali Flat and Death Valley. However, arroyos occasionally flood during convective rainstorms. Long-term flood records as a basis for predictions are few, but occasional flash floods and debris flows are anticipated. Erosion and deposition from flash floods and associated debris flows are among the most active geomorphic processes in the region, and they have a major role in the development of alluvial fans, denudation of mountainous landscapes, and the evolution of drainage-channel morphology.

Gages and other instrumentation are planned to be installed and operated in order to correlate streamflow events with precipitation. Streamflow and its characteristics will be studied in order to determine inundation, sediment transport, and erosion in the vicinity of the proposed repository surface facilities (Section 8.3.1.2.1).

Two hydraulic-engineering studies of flood-prone areas near Yucca Mountain provide a basis for estimating magnitudes and recurrence intervals of future floods (Section 3.2). Further studies will determine the magnitudes of probable maximum floods near the surface sites of proposed repository facilities (Section 8.3.1.16.1).

The nature and quantities of present and future infiltration or recharge need to be known better, especially at the site, including Fortymile Wash. Studies that include soil-moisture evaluations and chemical and isotopic analyses of waters from both the unsaturated and saturated zones are identified in Section 8.3.1.2 to provide this information.

Hydrogeologic conditions at the Yucca Mountain site are affected by the regional ground-water flow systems. The regional ground-water flow system originates as recharge at topographic highs within the hydrogeologic study area, and generally discharges at topographic lows. Other discharge areas are located where structural barriers to flow cause ground water to discharge. These discharge points are near Beatty, in the southeastern part of the Amargosa Desert, and in adjoining parts of Death Valley. Flow to these points is generally southward near Yucca Mountain and southwestward in areas east of Yucca Mountain. The hydrogeologic study area, having an area of about 16,200 km², has 10 identified and described hydrogeologic units. These range from an uppermost, discontinuous valley fill aquifer, through volcanic-rock aquifers and aquitards to a sequence of carbonate-rock aquifers and clastic-rock aquitards.

The ultimate source of nearly all ground water in the study area is precipitation on the basin. Because high-altitude areas generally receive more average annual precipitation, they generate more ground-water recharge than the valley floors, although recharge may also occur by infiltration of

surface runoff in wash bottoms. Most of the higher mountains (recharge areas) are in the northern and eastern part of the study area. Flow from the higher areas to the north passes southward beneath Yucca Mountain to be discharged by springs and evapotranspiration.

The hydraulic conductivity of rock in the hydrogeologic study area is controlled by fractures, joints, faults, vapor-phase cavities, interstitial spaces, and bedding planes. The aquitards have low fracture hydraulic conductivity and negligible interstitial hydraulic conductivity. Where fractures and faults are present, open, and unsealed with rock material, hydraulic interconnection among the hydrogeologic units is effective in transmitting water if hydraulic gradients of sufficient magnitude are present. Ground water in transit within the system and beneath Yucca Mountain has been estimated to be 10,000 to 20,000 yr old.

A conceptual model for ground-water flow in the saturated zone in the vicinity of Yucca Mountain has been developed and is based on the following assumptions: (1) The rock containing the primary matrix porosity is homogeneous and isotropic, (2) Secondary porosity is controlled mostly by fractures; the volume of water stored in them is relatively small in comparison to that stored in the matrix, (3) Hydraulic conductivity of fractures is several orders of magnitude larger than the conductivity of the matrix, (4) Distances between fractures are small in comparison with the dimensions of the ground-water basin, (5) On a large scale, the orientation of fractures may be assumed random so that the system appears isotropic, i.e., a slightly different form of a porous medium with equivalent porous-media properties.

Estimated average annual ground-water recharge from precipitation for the hydrogeologic study area is $5.6 \times 10^7 \text{ m}^3$ (Rush, 1970). The largest natural discharge point in the study area is the southeastern Amargosa Desert, where estimated average annual ground-water discharge is $3.0 \times 10^7 \text{ m}^3$ (Walker and Ekin, 1963; Rush, 1970). The remaining discharge is distributed among several smaller discharge areas. At Yucca Mountain the specific fluxes based on a two-dimensional flow model range from $3.6 \text{ m}^2/\text{yr}$ to $1.1 \times 10^6 \text{ m}^2/\text{yr}$.

Major-ion chemistry of ground water in the region has been summarized but enough water chemistry data are not yet available to develop a hydrochemical model to support conceptual-flow models, or to verify suggested ground-water flow paths (Section 3.7.3).

There is evidence for former, higher levels of the water table in the study area during the Plio-Pleistocene probably because increasing aridity associated with uplift of the Sierra Nevada resulted in lower precipitation with the Great Basin, which in turn has lead to a lowering of the water table. Also, flow-path lengths of ground water may have been shorter during the Plio-Pleistocene. However, current data based on climatic trends or evidence for past hydrologic conditions are insufficient to make reliable projections of future hydrologic conditions (Section 3.7.4). Section 8.3.1.5.2 discusses plans to evaluate further the potential effect of future climatic conditions on the hydrologic system at Yucca Mountain.

At the Yucca Mountain site, the water table is about 250 m below the proposed repository horizon. The proposed repository horizon is in turn, is

about 300 m below land surface. It is not known at this time under what circumstances the water table might be expected to rise sufficiently to saturate an underground facility at the repository horizon. As discussed in Sections 3.7.4 and 3.9.8, there is no evidence that water levels were more than 100 m higher during the Quaternary Period, but evidence indicates that water levels may have declined during that time. The results of preliminary regional ground-water flow modeling studies indicate that the climatic regime is unlikely to produce water table rises over the next 10,000 years that could inundate the repository. This unlikely scenario is to be examined further, as described in Section 8.3.1.5.2.

The preliminary information presented in Section 3.8.1 and 3.8.2 suggests that siting a mined geologic disposal system at Yucca Mountain will not adversely affect local ground-water users or the ground-water system. As discussed in Section 3.8.1, the Amargosa Desert, located within the Alkali Flat-Furnace Creek Ranch subbasin, currently is experiencing overdraft problems. Repository-related ground-water withdrawals probably will not affect the overdrafted areas because of the distance between the overdraft area and well J-13, and the fact that withdrawals are made from different aquifers. Computer models of ground-water flow of Yucca Mountain and vicinity will be developed (Sections 8.3.1.2.2 and 8.3.1.2.3) that can be used to simulate the effects of repository-related withdrawals on the local ground-water system. These models will provide information on the potential for impacts to local ground-water users, the potential impacts from these users to repository water supply, and the possible changes in the ground-water flow direction that might occur as a result of the ground-water withdrawals.

As discussed in Section 3.8.1.4, wells J-12 and J-13, which produce from the welded tuff aquifer, are capable of supplying water for repository activities through decommissioning. If water supply problems in the area were to become more severe, or if pumping wells J-12 and J-13 were to prove to be too costly, an alternate source should be readily available. The information required to fully evaluate the potential for water supply problems, is not available at present. Population projections for the next 50 and 100 yr, and information regarding human settlement patterns will be needed. These data will be obtained and evaluated in the investigations described in Sections 8.3.1.9.2 and 8.3.1.16.2.

The discussions of recharge (Section 3.7.1) and of moisture fluxes in the unsaturated zone at Yucca Mountain (Sections 3.9.3 and 3.9.4) note that the present mean annual precipitation at Yucca Mountain is estimated to be about 150 mm/yr. A knowledge of the historical, present, and probable future precipitation and potential evapotranspiration rates at Yucca Mountain are important in characterizing fully the hydrology of the unsaturated zone. The state of the unsaturated-zone hydrologic system is determined by the rate at which water enters the system as net infiltration across its upper land-surface boundary. Over the Yucca Mountain block as a whole (Sections 3.9.3.4 and 3.9.4), neither the rate nor the areal distribution of net infiltration is presently known. Because the hydrologic system within the unsaturated zone is complexly three-dimensional (Section 3.9.3.4), it is important to establish limiting magnitudes in the rate of net infiltration and its likely spatial distribution. Consequently, as described in Chapter 8.3, an extensive program of in situ infiltration monitoring and evaluation is planned not

only to estimate better current rates and distribution, but also to improve understanding of the specific flow mechanisms that control surface infiltration at Yucca Mountain.

The unsaturated zone at Yucca Mountain consists of a stratified sequence of welded and nonwelded, fractured and unfractured tuffs that comprise a fault-bounded, eastward-tilted block extending from 300 to 750 m above the water table (Section 3.9.3.4). The rock-stratigraphic units have been grouped into hydrogeologic units on the basis of common, approximately uniform hydrologic and material properties. The hydrogeologic units comprising the unsaturated zone are divided into three broad categories: (1) densely to moderately welded tuffs, as exemplified by the TCw and TSw units, having low saturated matrix hydraulic conductivity (about 10^{-11} m/s) and high fracture densities (10 to 40 fractures/m³); (2) nonwelded tuffs, for example, the PTn, and CHn units, having relatively high saturated matrix conductivities (about 10^{-8} to 10^{-6} m/s); and (3) nonwelded, virtually unfractured zeolitized tuffs of low saturated matrix hydraulic conductivity (about 10^{-11} m/s) of which the CHnz unit is the prototype. The repository is proposed to be constructed within rocks in the TSw hydrogeologic unit, which is overlain by the PTn and TCw units and is underlain by the vitric (CHnv) or zeolitic (CHnz) facies of the CHn unit. The water table beneath Yucca Mountain transects the CHn or underlying CFu units beneath almost all of the Yucca Mountain block.

Because of its low matrix hydraulic conductivity, the TSw unit can transmit only small vertical fluxes (less than about 1 mm/yr) at low saturations. However, because of its high fracture density (10 to 40 fractures/m³), the unit is expected to become more transmissive at or near complete matrix saturation when liquid-water flow within the fractures becomes efficient. The hypothesis is borne out by the results of numerical simulations that were performed on the basis of the unsaturated zone conceptual model and hydrologic-property data. The numerical simulations presume steady-state flow and incorporate approximate, empirically and theoretically derived moisture-characteristics relations for the rock matrix and the fractures. The simulations test three fundamental flow-mechanism hypotheses: (1) that liquid-water flow is restricted predominantly to the rock-matrices in all unsaturated units, (2) that flow occurs predominantly within the fractures of welded TCw and TSw units, and (3) that combined matrix-fracture flow occurs in the TCw and TSw units. The results of these simulations are consistent with the hypothesis that matrix flow is presently dominant within the repository TSw unit for a vertically downward flux of less than 0.5 mm/yr at a matrix saturation of about 0.5, as indicated by presently available data pertaining to ambient conditions within the TSw unit. Flow in fractures and faults may also occur, however, and tests are planned to evaluate this phenomenon (Section 8.3.1.2.3).

The results of preliminary numerical simulations are highly approximate and uncertain because they are based on an overly simplified representation of the hydrologic system and on preliminary hydrologic-property data. Nevertheless, these numerical results are mathematically and physically internally consistent and yield predictions that are consistent quantitatively with the salient features of the conceptual model. Moreover, these results are compatible with the available data concerning ambient hydrologic conditions within the unsaturated zone at Yucca Mountain. There is a need

for additional, detailed hydrologic data by which to refine the mathematical model representations. A major concern is to develop an appropriate quantitative description of the flow of liquid water in variably saturated, fractured tuffs. Well established laboratory techniques are available for measuring most matrix hydrologic properties for small rock samples and cores with the possible exceptions of measuring saturated hydraulic conductivity for welded and zeolitized tuff samples. Similar techniques are lacking, however, for single fractures and fracture systems. In fractured media, the scale of measurement must be large compared to the mean fracture aperture and spacing. Hence, the direct empirical determination of the bulk hydrologic properties of the fractured units must rely strongly on hydrologic testing in surface-based boreholes and within the exploratory shaft, as described in Section 8.3.1.2.2. Sufficient data need to be collected to permit statistical analyses of the limits of uncertainty of these data. Uncertainties may be introduced by random errors of measurement, as well as by inherent heterogeneities within the hydrogeologic units. The resulting statistical distributions are needed to provide the requisite input data to stochastic models from which the overall uncertainty of quantitative predictions, such as ground-water travel times, can be assessed.

Ambient hydrologic conditions within the unsaturated zone at Yucca Mountain remain poorly known, although, as described in Chapter 8.3.1.2.2 considerable data on these conditions will be obtained during monitoring in surface-based boreholes and within the exploratory shaft. Preliminary data and inference from test wells USW UZ-1, UZ-6, and H-1 indicate ambient rock-matrix saturations in the range from 0.4 to 0.65 and ambient matrix potentials of about -0.3 MPa in the repository TSw unit. These data are consistent with the estimated value of 0.5 mm/yr or less for the present flux through this unit. It may be concluded from these conditions: (1) that the vertical flux of water within the repository TSw unit probably is low; (2) that the flow is restricted largely to the rock matrix with probably little significant flow within fractures; and (3) that the flux may be nearly constant, occurring under unit vertical hydraulic gradient. These conclusions are only indicative of existing conditions; considerably more data need to be acquired before they can be confirmed or refuted with sufficient certainty. Similar conclusions apply for the PTn and TCw units overlying the repository TSw unit; however, for these units, the paucity of reliable data precludes establishing little more than upper limits on possible fluxes based on estimates of the likely rate of net infiltration.

Because of the presence of the natural geothermal gradient, there is the potential for the upward movement of moisture as water vapor within the unsaturated zone. This potential probably is greatest within the fractured TSw unit in which vapor movement is envisioned to occur principally in the partially saturated fractures. Water-vapor movement may occur either as advection superimposed on an underlying bulk-gas flow or as diffusion along the water-vapor concentration gradient that would be coincident with the geothermal gradient if water vapor is in local thermodynamic phase equilibrium with liquid water. Bulk-gas flow may occur either as thermally driven convection or as forced convection impelled by land-surface barometric fluctuations. It is not yet known, however, to what depths the effects of land-surface barometric changes may penetrate and, thus, influence bulk-gas movement within the unsaturated zone. It should be emphasized that until compelling evidence to the contrary becomes available, the hydrologic system

within the unsaturated zone beneath Yucca Mountain must be conceived as a two-component (water, air) two-phase (liquid, gas), system involving liquid-water, water-vapor, and bulk-gas movement embedded within the fractured, solid-phase rock matrix. Although virtually no data are available pertaining to the gas-phase hydrologic system, these data are to be collected through gas-phase testing and sampling within surface-based boreholes and the exploratory shaft as described in Chapter 8.3.1.2.2.

Velocity of ground-water flow and resulting travel time in the unsaturated zone strongly depend on effective porosity and hydraulic conductivity of fractures and rock matrix. Based on an upper-bound flux of 0.5 mm/yr, ground-water travel time within the unsaturated zone from the proposed repository to the water table is estimated to range from about 9,000 to 80,000 yr.

The lateral direction of ground-water movement in the saturated zone near the Yucca Mountain site probably is to the southeast. Hydraulic gradient is variable, and southeast of the repository block it is nearly flat. Hydraulic conductivity and effective porosity, two of the parameters that determine ground-water velocities and travel times, are not adequately known yet. Calculations based on conservative values of the controlling parameters indicate that saturated zone ground-water travel time, from the repository block to the accessible environment, may be about 200 yr (Section 3.9.4).

Based on these estimates for travel times, the minimum ground-water travel time from the edge of the repository to the accessible environment under present conditions is approximately 9,200 yr, well in excess of the 1,000 yr limits required by 10 CFR Part 60.113(a)(2).

Uncorrected carbon-14 data indicate that ground water in the vicinity of Yucca Mountain ranges in age from about 10,000 to 20,000 yr, and that waters of either younger or older ages have not been identified. This may bear on ground-water velocities or location of recharge, but the possible implications have not been resolved yet.

Mineralogic evidence suggests that past water levels beneath Yucca Mountain may never have been more than 100 m higher than present levels. Conservative flow modeling of the regional ground-water system suggest that water-table altitudes (past or future) could be, at most, 130 m above present levels, whereas inundation of a repository would require a rise of 200 m or more (Section 3.9.8).

3.10.2 RELATION TO DESIGN

A number of aspects of the ground-water system at Yucca Mountain directly relate to the design of a repository and to waste container design, in particular:

1. Flooding.

2. Ground-water occurrence between the surface and the repository horizon.
3. Ground-water conditions in the waste package environment.

Sections 3.2.2 and 6.2.4 discuss the potential for flooding at the Yucca Mountain site surface facilities. Preliminary investigations discussed in these sections indicate that the surface facilities are located above the level of the probable maximum flood. Further studies are discussed in Section 8.3.1.16. In addition, flood protection measures are included in the design to prevent surficial sheet flow affecting the surface facilities.

The proposed repository at Yucca Mountain is situated in the unsaturated zone. Therefore, it is unlikely that any significant bodies of water would be encountered in the development of the repository. The possibility of the occurrence of perched-water bodies exists; this will be examined during site characterization, particularly through the development of the exploratory shaft. However, it is considered unlikely that even if perched-water bodies do exist between the surface and the repository horizon at Yucca Mountain, they will present a major concern in repository development.

The waste emplacement package environment is one of the more important aspects considered in the design of waste containers. The potential interaction between the container walls and moisture in the emplacement holes is a consideration in estimating the lifetime of the canisters. This aspect is discussed further in Chapter 7, and additional plans for further analyses are presented in Sections 8.3.4 and 8.3.5.

3.10.3 IDENTIFICATION OF INFORMATION NEEDS

As mentioned in the introduction to Chapter 3, the hydrologic system in the vicinity of Yucca Mountain is one of the more important aspects that affects the postclosure and preclosure performance of the proposed repository. In particular, hydrology directly relates to the nine performance issues and associated information needs of Key Issue 1, namely:

Short titles

- Issue 1.1: Total system performance (Section 8.3.5.13)
- 1.1.1 Site information needed for calculations (Section 8.3.5.13.1)
 - 1.1.3 Calculational models for release scenario classes
(Section 8.3.5.13.3)

Short titles

- 1.1.4 Radionuclide releases for scenario classes (Section 8.3.5.13.4)
- 1.1.5 Probabilistic release estimates (Section 8.3.5.13.5)

CONSULTATION DRAFT

- Issue 1.2: Individual protection (Section 8.3.5.14)
- Issue 1.3: Ground-water protection (Section 8.3.5.15)
- 1.3.1 Class I or special sources of ground water (Section 8.3.5.15.1)
- Issue 1.4: Containment by waste package (Section 8.3.5.9)
- 1.4.1 Waste package design features needed (Section 8.3.5.9.1)
- 1.4.3 Scenarios and models needed (Section 8.3.5.9.3)
- 1.4.4 Containment barrier degradation (Section 8.3.5.9.4)
- 1.4.5 Time to loss of containment (Section 8.3.5.9.5)
- Issue 1.5: Engineered barrier system release rates (Section 8.3.5.10)
- 1.5.1 Waste package design features (Section 8.3.5.10.1)
- 1.5.3 Scenarios and models (Section 8.3.5.10.3)
- 1.5.4 Release rates from waste package and engineered barrier system (Section 8.3.5.10.4)
- 1.5.5 Radionuclides leaving waste package near-field environment (Section 8.3.5.10.5)
- Issue 1.6: Ground-water travel time (Section 8.3.5.12)
- 1.6.1 Site and design information needed (Section 8.3.5.12.1)
- 1.6.2 Computational models to predict travel times (Section 8.3.5.12.2)
- 1.6.3 Identification of likely travel paths (Section 8.3.5.12.3)
- 1.6.4 Pre-waste emplacement ground-water travel time (Section 8.3.5.12.4)
- 1.6.5 Disturbed zone boundary (Section 8.3.5.12.5)
- Issue 1.7: Performance confirmation (Section 8.3.5.16)
- Issue 1.8: NRC siting criteria (Section 8.3.5.17)
- Short titles
- Issue 1.9a: Higher-level findings (postclosure) (Section 8.3.5.18)
- Issue 1.9b: 100,000-year release (Section 8.3.5.18)

In addition, hydrologic information is needed to satisfy the following postclosure design issues:

Issue 1.10: Waste package characteristics (postclosure) (Section 8.3.4.2)

1.10.4 Near-field environment (Section 8.3.4.2.4)

Issue 1.11: Configuration of underground facilities (postclosure)
(Section 8.3.2.2)

1.11.6 Thermal loading and thermomechanical rock response
(Section 8.3.2.2.6)

For the postclosure situation, most of the hydrologic information required to satisfy the performance and design issues identified above will be collected by the geohydrology program (Section 8.3.1.2) and its associated investigations:

SCP section

Short title

8.3.1.2.1	Regional hydrology system
8.3.1.2.2	Site unsaturated zone hydrologic system
8.3.1.2.3	Site saturated zone hydrologic system

In addition, Investigation 8.3.1.3.1 will provide information about water chemistry within the potential emplacement horizon and along potential flow paths, and Investigations 8.3.1.5, 8.3.1.6 and 8.3.1.8 will evaluate the potential effects of future climatic conditions, erosion, and igneous and tectonic activity on hydrologic characteristics, respectively.

Investigations 8.3.1.9 and 8.3.1.9.3 provide information on the present and future value of ground-water resources, and the potential effects of exploiting natural resources on the hydrologic characteristics in the vicinity of the Yucca Mountain site.

In the preclosure timeframe, Issues 4.1 and 4.4, along with the hydrology program (8.3.1.16), are associated with hydrologic characteristics, namely:

Issue 4.1: Higher level findings - ease and cost of construction
(Section 8.3.5.7)

4.1.1 Site not disqualified (Section 8.3.5.7.1)

4.1.2 Site meets qualifying conditions of technical guidelines
(Section 8.3.5.7.2)

4.1.3 Site meets qualifying conditions of system guidelines
(Section 8.3.5.7.3)

CONSULTATION DRAFT

Issue 4.4: Preclosure design and technical feasibility (Section 8.3.2.5)

4.4.7 Design analyses (Section 8.3.2.5.7)

<u>SCP section</u>	<u>Short title</u>
8.3.1.16.1	Flooding recurrence intervals
8.3.1.16.2	Water supplies
8.3.1.16.3	Ground-water conditions

The plans to collect this information and the methods that will be used to evaluate the effects of the hydrologic regime on performance and design are presented in Section 8.3. The details of each study and activity are presented under each investigation, together with listings of parameters to be collected or that are required, and associated technical procedures and schedules.

3.10.4 RELATION TO REGULATORY GUIDE 4.17

NRC Regulatory Guide 4.17 (NRC, 1987) was used as the basis for the Annotated Outline for Site Characterization Plans. Chapter 3 has been written to satisfy the Annotated Outline and has covered all of the aspects required by Regulatory Guide 4.17, (NRC, 1987) with the following additions:

1. Sections 3.0 and 3.10 have been added to provide an introduction to the material presented in Chapter 3, a summary of the material, and a tie to the material presented in Section 8.3, which discusses the investigations pertinent to hydrology.
2. Sections 3.2.1, 3.5, 3.9.3, 3.9.5 have had title changes from those presented in Regulatory Guide 4.17 to indicate more clearly the material covered in each section.
3. Some fourth order headings presented in Regulatory Guide 4.17 have been deleted because the material requested under each section seemed repetitive. Aside from these minor modifications, all of the material requested by Regulatory Guide 4.17 (NRC, 1987) presently appears in Chapter 3.

Nuclear Waste Policy Act
(Section 113)

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***Site Characterization
Plan***

***Yucca Mountain Site, Nevada Research
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Nuclear Waste Policy Act
(Section 113)

GEOCHEMISTRY

Consultation Draft

***Site Characterization
Plan***

***Yucca Mountain Site, Nevada Research
and Development Area, Nevada***

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CONSULTATION DRAFT

Chapter 4

GEOCHEMISTRY

INTRODUCTION

This chapter contains geochemical information about the Yucca Mountain site. The chapter references plans for continued collection of geochemical data as part of the site characterization program. Details of these plans are contained in Section 8.3.1.3.

This section provides a brief introduction to the geochemistry chapter. It contains discussions about the following: (1) the concerns that drive the collection of geochemical data, (2) the methods by which data presently available have been collected, (3) concepts of the site that influence geochemical data collection, and (4) the status of present data and models.

In addition to the discussions in this chapter, the following related concerns are addressed in Chapter 7:

1. Anticipated interactions among the waste form, engineered barriers, and environment.
2. The chemical composition and form of the waste.
3. The solubility of the waste form in ground water.
4. The radionuclide species released by leaching from the waste form.
5. Anticipated chemical and mineralogical compositions of natural and engineered barriers.
6. Anticipated interactions among the waste form, water, vapor, gas, and rock.

ISSUES

Geochemical information generated during the characterization of the Yucca Mountain site will play an important role in resolving questions about the ability of the site to isolate radioactive wastes from the accessible environment. This issue is addressed by Key Issue 1 (Will the mined geologic disposal system at Yucca Mountain isolate the radioactive waste from the accessible environment after closure in accordance with the requirements set forth in 40 CFR Part 191, 10 CFR Part 60, and 10 CFR Part 960?) (Section 8.2). Geochemical information is specifically addressed by Characterization Program 8.3.1.3 (geochemistry).

In addition, geochemical data will be used to evaluate the Yucca Mountain site in terms of the siting guidelines outlined in 10 CFR Part 960 and the siting criteria in 10 CFR Part 60. In particular, two issues will require site characterization data for their resolution: Issue 1.8 (Can the demonstrations for favorable and potentially adverse conditions be made as

CONSULTATION DRAFT

required by 10 CFR 60.122?) (Section 8.3.5.17) and Issue 1.9 (Can the higher level findings required by 10 CFR Part 960 be made for qualifying conditions on the postclosure system guideline and the disqualifying and qualifying conditions on the technical guidelines for geohydrology, geochemistry, rock characteristics, climate changes, erosion, dissolution, tectonics, and human interference; and can the comparative evaluations be made by 10 CFR 960.3-1-5?) (Section 8.3.5.18).

Specific to the geochemistry task are three favorable conditions and four potentially adverse conditions (10 CFR 60.122):

Favorable conditions:

1. The nature and rates of tectonic, hydrogeologic, geochemical, and geomorphic processes (or any of such processes) operating within the geologic setting during the Quaternary Period, when projected, would not affect or would favorably affect the ability of the geologic repository to isolate the waste.
2. Geochemical conditions that
 - a. Promote precipitation or sorption of radionuclides.
 - b. Inhibit the formation of particulates, colloids, and inorganic and organic complexes that increase the mobility of radionuclides.
 - c. Inhibit the transport of radionuclides by particulates, colloids, and complexes.
3. Mineral assemblages that, when subjected to anticipated thermal loading, will remain unaltered or alter to mineral assemblages having equal or increased capacity to inhibit radionuclide migration.

Potentially adverse conditions:

1. Ground-water conditions in the host rock, including chemical composition, high ionic strength or ranges of Eh-pH, that could increase the solubility or chemical reactivity of the engineered barrier system.
2. Geochemical processes that would reduce sorption of radionuclides, result in degradation of the rock strength, or adversely affect the performance of the engineered barrier system.
3. Ground-water conditions in the host rock that are not reducing.
4. Potential for the movement of radionuclides in a gaseous state through air-filled pore spaces of an unsaturated geologic medium to the accessible environment.

PAST DATA COLLECTION

The Yucca Mountain site was one of a number of sites on and near the Nevada Test Site (NTS) in southern Nevada that was considered as a potential location for a waste repository. Geochemical information had been collected about this area for some time as part of the nuclear testing program at the NTS. The information presently available is a combination of data obtained before Yucca Mountain was considered as a potential waste repository site (before 1977) and data that have been collected as part of the Nevada Nuclear Waste Storage Investigations (NNWSI) Project. Essentially all the data presented here have been collected as part of the NNWSI Project. Information obtained from other sources has primarily been used to aid interpretation or for confirmation of more recent data.

The geochemical data collected as part of the NNWSI Project have been obtained by an examination of samples taken from the surface or at depth from Yucca Mountain and vicinity and by laboratory experiments using these samples and similar materials. Mineralogy and petrology samples have come from drill cores, sidewall samples, drill cuttings, and surface outcrops. Ground-water samples have come from wells that tap the saturated zone at and near Yucca Mountain. Data collected for other aspects of site characterization are also used or referenced in this chapter. This is particularly true of Chapter 3, Hydrology, which is often referenced. Data describing the geochemistry of the Yucca Mountain site and the effects of geochemistry on radionuclide transport are contained in Section 4.1. In addition, assessments of the stability of geochemical conditions (Section 4.4) and of the effects of waste emplacement on geochemical conditions (Section 4.2) are being conducted to ensure that changes in geochemical conditions will not adversely affect the site. These assessments are primarily obtained from experiments that simulate expected conditions and from analyses of the effects of these conditions on the geochemistry of the site. Natural systems and large-scale field tests (Section 4.3) are also being studied to provide information over longer times and larger distances than can be obtained in the laboratory. Data obtained from these studies will be important for confirmation of conceptual models and for validation of mathematical models of important processes.

The geochemical characterization of the Yucca Mountain site is not complete. The discussions presented in this chapter raise a number of questions or identify areas where further information is needed. Specific plans for answering these questions and collecting the information needed are described in Section 8.3.1.3. Specific information needs relevant to geochemistry are also identified in Section 4.5.3 that summarizes the geochemical investigations to be performed during site characterization to address the information needs described in this chapter.

CONCEPTS OF THE SITE THAT INFLUENCE GEOCHEMICAL DATA COLLECTION

A definition of geochemical data needs depends on an understanding of the geochemical processes that affect radionuclide transport from the repository toward the accessible environment. A brief description of the present concepts of the site and of radionuclide transport processes that have influenced data collection is presented here to help orient the reader.

CONSULTATION DRAFT

The Yucca Mountain site is composed mostly of ash-flow and ash-fall tuffs in a number of distinct cooling units (Section 4.1.1). The resulting mineralogy primarily reflects the original bulk chemical composition, the extent of devitrification and welding of individual units, and the variable interaction with water during alteration. Where water has contacted vitric tuff, secondary minerals such as zeolites and clays have generally formed from the alteration of glass. Ground water at and near the site is mainly a dilute, near-neutral, sodium-bicarbonate water (Section 4.1.2).

The potential repository location is in the unsaturated zone, in the lower welded and devitrified part of the Topopah Spring Member of the Paintbrush Tuff (Chapter 1). The most likely transport path for radionuclides either in solution or as particulates is in water that might move through the repository and toward the accessible environment. In the presently preferred conceptual model (Montazer and Wilson, 1984), water in the unsaturated zone moves generally downward toward the water table; however, some lateral movement may occur where the lateral permeability greatly exceeds the vertical permeability. Although vertical components of flow occur locally beneath the water table, flow in the saturated zone is dominantly lateral in southerly to southeasterly directions from Yucca Mountain (Section 3.9).

The following geochemical processes that affect the transport or retardation of radionuclides in water have been studied at Yucca Mountain: sorption, precipitation or dissolution, and physical processes like diffusion, dispersion, or filtration of particulates (Section 4.1.3). Sorption of radionuclides provides a potential geochemical barrier at the Yucca Mountain site. Sorption is likely to occur on the highly sorptive minerals, such as the zeolites and clays found between the repository and the water table and below the water table. Understanding sorption processes requires information about the mineralogy (Section 4.1.1) and ground-water chemistry (Section 4.1.2), and knowledge of flow characteristics, particularly the relative importance of matrix (porous) and fracture flow. It is probable that flow in both the matrixes and fractures occurs in the unsaturated zone as well as the saturated zone. However, matrix flow is probably the dominant mechanism in the unsaturated zone and fracture flow is probably dominant in the saturated zone (Chapter 1, DOE, 1986). Therefore, sorption data applicable to both matrix flow and fracture flow are being obtained (Section 4.1.3). Sorption measurements have been determined using Yucca Mountain tuff and ground-water samples for essentially all experiments.

Because the potential repository location is in the unsaturated zone, radionuclides such as carbon-14 and iodine-129 may be transported away from the repository in both the gas phase and the aqueous phase. Exchanges between the gas phase and the aqueous phase present in the rock pores may provide an effective retardation mechanism, and is a topic of investigation (Section 4.1.3.6).

The solubility of waste elements can influence their transport by limiting the concentrations of the elements dissolved in water. Solubility depends primarily on the chemistry of the specific element and the local water chemistry. Solubility measurements under conditions characteristic of the Yucca Mountain site, in combination with chemical equilibrium modeling, are being used to characterize waste element solubility (Section 4.1.3.4).

In addition to being dissolved in water, waste elements can also exist as particulates (colloids) or be sorbed onto particulates that are present in the water. The formation of natural colloids of waste elements is discussed in Section 4.1.3.4; plans to study particulate material that is naturally present in the water from Yucca Mountain and sorption of radionuclides on particulate material are outlined in Section 8.3.1.3.4.

The relative importance of physical processes, such as diffusion, dispersion, or filtration of particulates in the water, that influence radionuclide transport by retarding the radionuclides is also strongly dependent on local flow conditions. As previously noted, the relative importance of matrix flow versus fracture flow is still somewhat uncertain (Section 3.9). Physical processes associated with both matrix and fracture flow are being studied (Section 4.1.3). Matrix diffusion, in which radionuclides being transported by water in fractures can diffuse into water in the surrounding porous rock matrix, is one of the physical processes being given special attention (Section 4.1.3.5). This retardation mechanism may be important if fracture flow is significant along portions of the flow paths.

The concepts presented above will be formalized into a number of conceptual models as data are collected that confirm the physical and chemical descriptions of the site and the operation of the various geochemical processes. The detailed plans for this work are described in Section 8.3.1.3. In particular, geochemical models of the unsaturated and saturated zones (Section 8.3.1.3.7), a conceptual model of mineral evolution at Yucca Mountain (Section 8.3.1.3.3), and a predictive model of sorption as a function of mineral content of the tuff (Section 8.3.1.3.4), are being developed.

PRESENT STATE OF DATA AND MODELS

The development of an understanding of the geochemistry of the Yucca Mountain site is an iterative process in which data collection and analysis are combined with conceptual models. A detailed discussion of questions that still exist and of investigations can be obtained from the remainder of this chapter and from Section 8.3.1.3. However, there are a number of general comments that can be made about the present state of geochemical data and models for the site:

1. The water chemistry in the saturated zone has been well characterized, but ground-water samples are needed from the unsaturated zone. They will be obtained during construction of the exploratory shaft. A qualitative understanding exists of the mineralogical controls on ground-water composition.
2. The mineralogy and petrology, based on data from the existing drill-holes, have been well characterized. Future work during construction of the exploratory shaft will provide additional data. However, characterization along flow paths away from the repository is not complete because flow-path definitions are uncertain and because there are few exploratory borings along potential pathways.
3. A conceptual model of mineral evolution is being developed. It must provide credible explanations of the present mineral distribution

CONSULTATION DRAFT

and must be capable of predicting future mineral evolution due to natural processes and the thermal loading of the repository.

4. Sorption data have been collected using representative ground water and tuffs. An understanding of the chemistry that controls sorption exists in some cases but is uncertain for many waste elements, such as the actinides, which have complex aqueous chemistry.
5. Solubility data appropriate to Yucca Mountain are being collected. Chemical equilibrium modeling is being done but it is uncertain whether equilibrium models will be adequate. If not, kinetic models will be required.
6. Physical processes (i.e., dispersion, diffusion, advective processes) that affect radionuclide transport are understood qualitatively, but quantitative data are lacking. Radionuclide transport depends strongly on local ground-water flow characteristics, which are presently not well understood.
7. Radionuclide retardation by all processes along the flow paths to the accessible environment is being assessed by performing integrated transport calculations. This includes an analysis of significant interactions among thermal, hydrologic, chemical, and mechanical processes that could have an impact on the geochemistry of the site and the transport of radionuclides as particulates, pseudocolloids, and colloids at Yucca Mountain.
8. The potential for gaseous transport needs to be assessed. No quantitative assessment has been made yet and the mechanism for isotopic exchange and gas and liquid-solid exchange is uncertain at this time.

Experimental methods and measurement techniques used to obtain the data in this chapter are discussed in detail in the references from which the data were taken. Data quality and related data uncertainty are also discussed in the references. The types of data collected range from mineral types and distribution to specific measurements (e.g., sorption ratios and solubility measurements). In the case of measured sorption ratios, the experimental data uncertainty is included in the data tables.

4.1 GEOCHEMISTRY OF THE HOST ROCK AND SURROUNDING UNITS

4.1.1 MINERALOGY AND PETROLOGY

The mineralogic and petrologic characteristics of Yucca Mountain are important for defining the site conditions existing before repository construction and operation, and for predicting the site response to disruption. Mineral distributions as a function of depth and distance from the potential repository are also important for calculating overall radionuclide retardation by sorption. Because sorptive behavior and ground-water chemistry depend on mineralogy, this chapter begins with a description of the site

mineralogy (Sections 4.1.1.1 to 4.1.1.3). Section 4.1.1.4 describes mineral stability at the site.

4.1.1.1 General description of the host rock and surrounding units

The most abundant rock type at Yucca Mountain is ash-flow tuff. The mineralogy of an ash-flow tuff is developed in several episodes of crystallization (Lipman et al., 1966; Byers, 1985). Crystallization from magma occurs primarily within the magma chamber before eruption. During this episode, phenocrysts (most commonly consisting of feldspars, quartz, and biotite) form along with less abundant minor and accessory minerals like hornblende, pyroxene, iron-titanium oxide minerals, sphene, allanite, and zircon. In the second episode of crystallization, glass shards and pumice commonly devitrify to form alkali feldspars and silica minerals. In portions of the ash flow where vapor concentrations are high, vapor-phase crystallization results in the growth of euhedral crystals of tridymite, cristobalite, and alkali feldspars in gas pockets. During vapor-phase crystallization, some rare silicate and oxide minerals are formed (e.g., andalusite and pseudobrookite) (Vaniman et al., 1984; Bish, 1987). Vapor-phase crystallization is common in the proposed host rock, the Topopah Spring Member of the Paintbrush Tuff (Lipman et al., 1966), where several zones develop spherulitic or lithophysal cavities in which pseudobrookite, rutile, manganese-rich augite, pargasite, and biotite have grown from vapor along with the more abundant alkali feldspars, tridymite, and cristobalite. In the third episode of crystallization, many of the initial tuff components may be altered late in the cooling sequence or after the loss of initial heat. It is in this late stage that the most sorptive minerals (i.e., zeolites and clays) are formed. The stratigraphic distribution of these crystallization episodes in the Topopah Spring Member is discussed in Chapter 1.

The petrologic features of ash-flow tuffs are largely determined by cooling history. Simple cooling units deposited at temperatures greater than 500°C (Section 1.2.2.2) typically have compacted or welded interiors between top and bottom zones of partially welded to nonwelded tuff. Compound cooling units occur where eruptive origins are more complex and the welding relations are not so simple; the proposed host rock for the repository in Yucca Mountain is within a compound cooling unit (Lipman et al., 1966). Even within this compound cooling unit, however, the general relationship is preserved in which the core of the cooling unit is more densely welded and the top, bottom, and distal edges of the ash flow are poorly welded or nonwelded. Immediately after the emplacement of an ash-flow tuff, primary glass in the more densely welded interiors of the cooling unit commonly devitrifies to intergrowths of silica minerals and alkali feldspars. The poorly welded or nonwelded margins of the cooling unit, however, often do not devitrify but remain glassy after the ash flow has lost its original heat. It is from the glass in these zones that the most significant intervals of secondary zeolites are formed. Thus, the zeolite intervals at Yucca Mountain (Section 4.1.1.3.2), are roughly correlated with the once vitric stratigraphic intervals that are at the margins of devitrified ash-flows (Vaniman et al., 1984).

Glass, feldspar, and mafic silicates may alter to smectites either within the feldspathic devitrified cores of cooling units or in their glassy margins. It might be expected, therefore, that smectites should occur throughout the tuff cooling units of Yucca Mountain, and in general this is true. In detail, however, the smectite abundances are lower in some zeolitized intervals than in adjacent nonzeolitized intervals; this is particularly true around the zeolitized tuff of Calico Hills (Caporuscio et al., 1982). This relationship is of interest in that the pH and ion concentrations during the hydrous alteration of glass are generally too low for zeolite formation unless the glass alters first to smectite (Hay and Sheppard, 1977). The negative correlation between clay and zeolite abundances at Yucca Mountain suggests that zeolite growth displaces smectite development soon after glass alteration begins along the cooling-unit margins.

Finally, it is important to note that although tuff mineralogy is closely related to cooling-unit petrology, the alteration of tuff to hydrous minerals is coupled with the availability of water. The minerals formed from altered tuff vary with depth. The transition from clinoptilolite to analcime to albite with depth is well documented at Yucca Mountain (Bish et al., 1981) and is discussed in Section 4.1.1.3.

4.1.1.2 Analytical techniques

This section discusses analytical techniques for mineralogic and petrologic analysis, as well as experimental techniques for studying mineral dehydration. Standard petrographic or mineral analysis procedures are used, although the quantitative x-ray technique of powder diffraction (XRD) involves some innovative data-reduction methods (Bish and Chipera, 1986). Special methods employed in the study of fracture minerals are discussed in Carlos (1985). Other techniques, some of which are unique and nonstandard, are described in Section 8.3.1.3.

The basic data on rock textures, mineralogy, and mineral compositions come from drill-core samples, supplemented by sidewall samples and drill cuttings where cores were not available, and samples from outcrops and trenches at Yucca Mountain. Standard optical petrography techniques were used in the study of thin sections made from these samples. Thin sections were used to determine textural relations, the degree of welding, and mineral growth distributions, with emphasis on the potential host rock in the Topopah Spring Member. Percentages of various textures and of phenocrysts were used to determine the stratigraphic position in the Topopah Spring Member (Byers, 1985).

Mineral occurrences and abundances are more generally determined by quantitative XRD. The methods used for quantitative XRD analysis are described by Bish and Vaniman (1985). When estimating the uncertainties of values determined by quantitative XRD, errors due to peak integration, crystalline solution, sample-standard variability, and peak overlap are considered. Relatively high precision is obtained for quartz, calcite, and analcime data (± 5 to 10 percent of amount present). Smectite, mica, alkali feldspar, clinoptilolite, and mordenite analyses have precisions that are lower (± 10 to 20 percent of the amount present). Representative XRD tables

can be found in Bish and Vaniman (1985). A new deconvolution technique is being developed to allow greater precision in the feldspar, cristobalite, and tridymite analyses, although feldspar data will still be limited by large amounts of crystalline solution and sample-to-standard variability. This new technique is outlined in Section 8.3.1.3.

For samples with more than 5 percent smectite, the smectites are dispersed and separated by sedimentation and centrifugation techniques. From these separations the type of smectite and the degree of interstratification with other clay minerals are determined by the XRD technique (Bish, 1981).

Mineral compositions for volcanic glasses, zeolites, and authigenic feldspars are determined on an automated Cameca electron microprobe. Instrument parameters for primary-phenocryst analyses are summarized in Caporuscio et al. (1982). Parameters for zeolite analysis are given by Levy (1984a) and Caporuscio et al. (1982). Parameters for the analysis of fine-grained devitrified ground mass are given by Byers (1985).

For zeolite-dehydration experiments, natural clinoptilolites were obtained. Cation-exchanged varieties of clinoptilolite were prepared by suspending 2.5-g portions in 1.0 M solutions of calcium, potassium, and sodium chloride. The treatment was repeated 3 times for 16 h each at about 50°C (Bish, 1984; 1985). The solids were centrifuged and rinsed after each ion exchange. Harsher treatments, such as autoclaving, were not used in the cation exchanges to avoid recrystallization or alteration of the clinoptilolites. Details of these methods are given by Bish (1984).

4.1.1.3 Mineralogical, petrologic, and chemical composition of the host rock and surrounding units

In this section, the topics of mineralogy, petrology, and chemical composition are discussed in two parts. First, the potential host rock within the Topopah Spring Member is described. The description begins with the primary phenocryst, devitrification, and vapor-phase constituents, and ends with a discussion of the secondary mineralogy within fractures. These data are focused on their usefulness in indicating stratigraphic position and in characterizing the mineralogy along potential radionuclide travel paths. The chemistry of the Topopah Spring Member is discussed in Chapter 1. In the second part, the surrounding units are described. This description begins with the primary chemistry and mineralogy and concludes with a discussion of the secondary minerals, emphasizing the sorptive zeolites and clays.

Data on mineral abundances are compiled in Figures 4-1 and 4-2 for drillholes USW G-2, USW GU-3, and USW G-3. The drillhole locations are shown in Section 1.2.1, Figure 1-18. Figures 4-1 and 4-2 represent mineral abundances versus depth in these drillholes from different parts of Yucca Mountain.

Figure 4-1. Abundance of minerals as a function of depth from surface; determined by x-ray diffraction for core samples from drillhole USW G-2. Modified from Caporuscio et al. (1982).

Figure 4-2. Abundance of minerals as a function of depth from surface, determined by x-ray diffraction for core samples from drillholes USW GU-3 and USW G-3. Modified from Vaniman et al. (1984).

CONSULTATION DRAFT

4.1.1.3.1 The potential host rock

The potential host rock is the lower, rhyolitic portion of the welded devitrified Topopah Spring Member. Present plans do not call for any waste emplacement in the upper, quartz latite portions of the welded devitrified Topopah Spring Member.

4.1.1.3.1.1 Mineralogy of the host rock matrix

The Topopah Spring Member of the Paintbrush Tuff is petrologically complex because it is a thick, compound cooling unit. The unit is compositionally zoned, from nearly aphyric (phenocryst-poor) high-silica (SiO_2) rhyolite (about 77 percent SiO_2) in the lower four-fifths grading upward through a zone of about 20 m into porphyritic (phenocryst-rich) quartz latite (about 69 percent SiO_2). The Topopah Spring Member includes a moderately to densely welded, devitrified interior that is bounded by a thin altered upper vitrophyre and a variably altered lower vitrophyre. The potential host rock for repository emplacement is within the devitrified zone of moderate to dense welding above the basal vitrophyre.

The devitrified high-silica rhyolite matrix above the vitrophyre consists of fine-grained intergrowths of 60 to 80 percent feldspar with 20 to 40 percent silica minerals (quartz and cristobalite with some tridymite). The cryptocrystalline groundmass of the overlying quartz latite is similar, although in most of the quartz latite, tridymite, and cristobalite are more abundant than quartz (Bish et al., 1984). The quartz latite also has higher phenocryst abundances than the high-silica rhyolite (6 to 16 percent versus about 1 percent; Byers, 1985). Phenocrysts throughout the Topopah Spring Member consist predominantly of sanidine and plagioclase; biotite, quartz, and iron-titanium oxide minerals are also common phenocrysts, although biotite is more abundant in the quartz latite and quartz is more abundant in the high-silica rhyolite.

A detailed description of the petrography of different stratigraphic intervals within the Topopah Spring Member, including major mineral variations and phenocryst compositions, and abundances is given in Chapter 1. Plans for more detailed description of petrologic variations are discussed in Section 8.3.1.3. The studies planned will also cover the limits of mineralogic variability in the host rock to help interpret the results of in situ experiments and to help determine the stratigraphic position during repository construction.

4.1.1.3.1.2 Mineralogy of fractures in the host rock

Detailed studies of fracture mineralogy have recently been completed on samples taken from above the static water level (SWL) in drillhole USW G-4 (Carlos, 1985). (The SWL approximates the position of the present-day water table, although it is a composite level affected by differences of hydraulic head in the open section of the hole.) These studies are being pursued because of the possible importance of fracture minerals in cation exchange

during fracture flow. Sequential generations of fracture-related mineral development in the unsaturated zone can be discerned (Figure 4-3), although the absolute ages of fracture formation and mineral deposition are not yet available.

Fractures are common in the moderately to densely welded host rock. Many of the fractures are lined or sealed by silica and feldspars. Fractures sealed by silica and feldspar probably formed by contraction during the cooling of ash flows, with the silica deposited by vapor-phase crystallization during initial cooling after the ash flow was emplaced. All other fracture coatings appear to be younger than the silica- and feldspar-filled fractures. Later fractures that reopened or cut across the early fractures are lined with zeolites (mordenite plus heulandite), sparse smectite, and rarely calcite (Carlos, 1985). Zeolite fracture linings are often discontinuous, and most fractures are only partially sealed. Smectite usually occurs as scattered late overgrowths on the other minerals and rarely fills fractures in the proposed repository horizon.

In Figure 4-3, the fracture mineralogy in tuffs between 770 ft (234.7 m) and the SWL correlate better with lithology than with stratigraphy. The numbers at the top of the figure indicate sequence of deposition, with type 1 (lithophysal type) representing earlier feldspar and silica deposits. The sequence varies between the welded-devitrified, glassy, and zeolitic lithologies. The same fracture minerals may have formed at different times in different units. It can be seen that zeolites occur in fractures throughout the depth interval sampled and are not restricted to the rocks in which the matrix is also zeolitized.

An understanding of fracture mineralogy will also provide information on the past movement of fluids in fractures, which will be important in evaluating the potential for radionuclide movement in the unsaturated zone. The minerals of fractures will also be an important contributor to sorption if fracture flow dominates in the transport of waste elements from the repository. Methods of dating fracture minerals are being investigated. Radiometric dating of smectites and electron-spin-resonance dating of quartz are both being pursued, as described in Section 8.3.1.3.

The information collected on fracture mineralogy and matrix alteration from several drillholes and the exploratory shaft will constrain our models of secondary-mineral formation at Yucca Mountain (Section 8.3.1.3) and models of the mineralogy along potential radionuclide transport paths.

4.1.1.3.2 Surrounding units

The Topopah Spring Member is overlain and underlain by other silicic volcanic units. The mineralogy of these bounding units is summarized briefly below. The stratigraphic framework is discussed in Section 1.2.2.

Above the Topopah Spring Member, the younger members of the Paintbrush Tuff include the Pah Canyon, the Yucca Mountain, and the Tiva Canyon members in order of decreasing age. The Pah Canyon and Yucca Mountain members form thin units (less than 71 m) at the site and are absent throughout much of the

CONSULTATION DRAFT

Figure 4-3. Sequence of fracture coatings in the unsaturated zone of drillhole USW G-4. Modified from Carlos (1985). Numbers across the top indicate inferred sequence of mineral growth, determined from overgrowth or cross-cutting relationships. Along the right side of the figure, major lithologic variations in the host rock (welded-devitrified, glassy, and zeolitic) are identified.

exploration block; the Tiva Canyon, however, is about 70 to 150 m thick and forms most of the outcrops on the eastern dip slope of Yucca Mountain (Section 1.2.2.2). These upper members are described in Chapter 1; more detailed descriptions are given by Orkild (1965), Lipman et al. (1966), Byers et al. (1976), and Scott et al. (1983). The mineral constituents and petrography of the upper members are summarized in Section 1.2.2.2.6. Electron-microprobe analyses of phenocrysts are summarized by Broxton et al. (1982), and the compositions of glasses and zeolites are described by Caporuscio et al. (1982). Chemical analyses are yet to be completed (Section 8.3.1.3.2).

The mineralogic compositions of the Topopah Spring Member and of the underlying units are summarized by Warren et al. (1984) and by Bish and Vaniman (1985). Detailed descriptions are published for samples from well J-13 (Heiken and Bevier, 1979; Byers and Warren, 1983; Bish and Chippera, 1986), drillhole UE-25a#1 (Sykes et al., 1979; Carroll et al., 1981; Bish and Chippera, 1986) drillhole USW G-1 (Bish et al., 1981; Bish and Chippera, 1986), drillholes USW G-2 and UE-25b-1H (Broxton et al., 1982; Caporuscio et al., 1982), and drillholes USW GU-3 and USW G-3 (Vaniman et al., 1984). Selected whole-rock compositions are summarized by Zielinski (1983) and cover the range of quartz latitic to rhyolitic tuffs that characterize the volcanic sequence at Yucca Mountain, plus the less common dacitic lavas or flow breccias. Additional chemical data will be obtained during site characterization (Section 8.3.1.3.2).

The petrography and petrology of the Topopah Spring Member were mentioned in Section 4.1.1.3.1 and described in Chapter 1. The symbols used for petrologic units are given in Table 4-1. The petrography of units underlying the Topopah Spring Member is summarized in Table 4-2 based on samples from drillhole USW G-1, and in Table 4-3, which is based on this drillhole and surrounding localities (Warren et al., 1984). Drillhole USW G-1 is centrally located within the site, as shown in Figure 4-4. The tables show that two feldspars, plagioclase and sanidine, plus quartz predominate in the phenocryst assemblages, with biotite forming the most common mafic phenocryst (but note the exceptionally high orthopyroxene content of the Prow Pass Member of the Crater Flat Tuff). Warren et al. (1984) conclude that these petrographic features vary slightly within a given unit, although at the southern end of the site, the tuffaceous beds of Calico Hills contain amphibole, orthopyroxene, and clinopyroxene (Vaniman et al., 1984).

Table 4-3 is a concise summary of the major phenocryst compositions (excluding quartz) in the Topopah Spring Member and in the underlying units. The anorthite content of plagioclase and the potassium and barium content of sanidine have been found to be distinctive for particular units (Warren et al., 1984).

Particular emphasis has been placed on studies of the alteration mineralogy in units surrounding the host rock. The alteration mineralogy is important because (1) it records the history of past chemical transport at Yucca Mountain, and (2) it provides a volumetrically abundant sorptive component to be considered in radionuclide retardation modeling.

CONSULTATION DRAFT

Table 4-1. Symbols used for petrologic units, rock types, alteration, and minerals listed in Table 4-2^a (page 1 of 3)

Unit, rock type, alteration, or mineral	Symbol
PETROLOGIC UNIT	
Paintbrush Tuff	TP
Topopah Spring Member	
Upper, mafic-rich portion	TPTu
Lower, mafic-rich portion	TPTl
Tuffaceous beds of Calico Hills	
Upper, mafic-poor portion	TH1
Lower, mafic-poor portion	TH2
Crater Flat Tuff	TC
Prow Pass Member	TCP
Bedded tuff	TCBa
Bullfrog Member	TCB
Upper portion, 2,152 to 2,317.4-ft depth, drillhole USW G-1	TCBu
Middle portion, 2,317.4 to 2,425-ft depth, drillhole USW G-1	TCBm
Lower portion, 2,425 to 2,601.6-ft depth, drillhole USW G-1	TCBl
Bedded tuff	TCTa
Tram Member	TCT
Upper portion, 2,601.6 to 3,083.0-ft depth, drillhole USW G-1	TCTu
Lower portion, 3,083.0 to 3,522.0-ft depth, drillhole USW G-1	TCTl
Flow breccia of drillhole USW G-1	TFB
Rhyodacite lava of drillhole USW G-2	TRD
Bedded tuff between the flow breccia of USW G-1 and the Lithic Ridge Tuff	TLRa
Lithic Ridge Tuff	TLR
Upper portion, 3,920.0 to $\approx$ 4,365-ft depth, drillhole USW G-1	TLRu
Lower portion, $\approx$ 4,365 to 4,940.2-ft depth, drillhole USW G-1	TLRl
Older tuffs of drillhole USW G-1	TT
Unit A	TTA
Unit B	TTB
Unit C	TTC

CONSULTATION DRAFT

Table 4-1. Symbols used for petrologic units, rock types, alteration, and minerals listed in Table 4-2^a (page 2 of 3)

Unit, rock type, alteration, or mineral	Symbol
ROCK TYPE	
Tuff	tf
Bedded tuff	b
Ash-flow tuff	t
Nonwelded	nwt
Partially welded	pwt
Moderately welded	mwt
Densely welded	dwt
Vitrophyric	vt
Lava	l
Intermediate-composition lava (groundmass plagioclase present)	intl
Lava-flow breccia	fb
Tuff breccia	tb
Argillite	ar
Granitoid	gr
ALTERATION ^b	
None	
Vitric	gl
Low temperature (secondary)	
Opaline	o
Argillite	ar
Zeolitic	z
Clinoptilolite	zc
Analcime	za
Calcite	cc
Pyritic	py
Silicic (Chalcedony)	q
High temperature (secondary)	
Albitic	ab
High temperature (primary)	
Vapor phase	vp
Granophyric	gr
Spherulite	sp
Axiolitic	ax

CONSULTATION DRAFT

Table 4-1. Symbols used for petrologic units, rock types, alteration, and minerals listed in Table 4-2^a (page 3 of 3)

Unit, rock type, alteration, or mineral	Symbol
MINERALS ^c	
Felsic phenocrysts	
Quartz	Q
Sanidine (+ anorthoclase, if present)	K
Plagioclase	P
Mafic phenocrysts	
Biotite	Biot
Hornblende	Hbld
Clinopyroxene	Cpx
Orthopyroxene	Opx
Accessory minerals	
Sphene	Sph
Allanite	All
Perrierite/cherkinite	Per
Apatite	Ap
Zircon	Zr

^aSource: Warren et al. (1984). To convert feet to meters, multiply by 0.328.

^bThe symbols m (minor) and μ (micro) are used as modifiers on the alteration symbols.

^cThe mineral volume abundances listed for most samples are for unaltered or partly altered phenocrysts. In some samples, phenocrysts are partly or completely altered. In these samples, the symbol "A" indicates that the accompanying concentration is for partly altered grains only, and the symbol "ps" indicates that the accompanying concentration is based entirely on completely altered (pseudomorphic) forms.

Table 4-2. Summary of petrography for core samples from drillhole USW G-1, determined by point count of glass-covered thin section^{a,b} (Page 1 of 2)

Sample depth (ft) ^c	Unit	Rock type	Alteration	Points counted	Major components (volume %)		Felsic phenocrysts (relative %)			Mafic phenocrysts (ppmV)				Fe-Ti oxide (ppmV)	Accessory minerals (grains identified)			
					Lithics	Felsics	Q	K	P	Biot	Hbl	Cpx	Opx		Sph	All	Ap	Zr
1,561.8	TH1	nwt	Zc	3,750	2.8	2.2	43	16	41	270	0	0	0	0	0	0	0	0
1,689.5	TH1	nwt	Zc	8,000	1.8	2.0	51	24	25	375	0	0	0	125	0	0	0	4
1,811.7	TCP	pwt	VP	3,600	0.14	6.2	15	53	32	<280	0(ps)	0	1,900(ps)	560	0	0	0	4
1,943.4	TCP	nwt		3,300	0.45	14.4	13	39	48	<280	0	0	910(ps)	610	0	0	0	8
2,009.8	TCP	pwt		3,750	2.6	7.9	15	47	38	<280	<280	0	2,100(ps)	530	0	1	Rare	5
2,124.7	TCP	nwt		3,600	0.6	8.8	6	50	44	280	0(ps)	0	0	1,100	0	2	Rare	7
2,231.0	TCBu	pwt	p-mwt	3,700	0.03	12.2	22	36	42	2,700	1,300(ps)	0	0	810	0	0	Rare	9
2,246.0	TCBu	p-mwt		3,000	0.00	11.8	19	45	36	4,300	1,000(ps?)	0	0	1,700	0	0	Gone	7
2,300.4	TCBu	pwt		3,750	0.00	13.6	24	31	45	1,600	2,400(ps)	0	0	1,600	0	0	Rare	13
2,354.6	TCBm	nwt	Gr	3,750	0.03	14.6	27	30	43	3,500	2,400(ps)	0	0	1,900	0	0	Rare	10
2,397	TCBm	pwt	VP	3,750	0.2	13.0	23	32	45	5,300	1,300(ps)	0	0	530	0	0	Rare	7
2,461.5	TCB1	nwt	VP	3,650	0.9	7.1	5	44	51	2,200	1,600(ps)	0	0	550	0	0	Rare	8
2,470.6	TCB1	pwt	VP	3,700	1.6	8.3	18	29	53	4,900	1,100(ps)	0	0	1,300	0	0	Rare	10
2,478.3	TCB1	pwt	VP	3,750	0.5	10.0	14	39	47	4,000	530(ps)	0	0	270	0	0	Rare	5
2,507	TCB1	nwt	Zc	3,700	1.2	9.6	13	36	51	4,100	3,200(ps)	0	0	2,400	0	0	Rare-sparse	12
2,555	TCB1	nwt		3,520	0.8	9.9	9	40	51	2,300	2,300	280(ps?)		280	0	0	Shot?	5
2,594.2	TCB1	pwt		3,750	2.2	7.1	17	32	51	4,300	270(ps)	0	0	800	0	4	Rare	10
2,678.0	TCTu	pwt		3,980	1.3	7.5	15	34	51	7,300	0	0	0	1,000	2	1	Sparse	6
2,772.6	TCTu	pwt	m-dwt	3,800	2.1	10.2	29	34	37	5,800	1,600(ps?)		0	530	0	0	Sparse	5
2,851.7	TCTu	nwt		3,900	4.5	13.6	41	37	22	4,400	0	0	0	1,800	0	0	Rare	11
2,869	TCTu	m-dwt		3,360	3.8	12.1	41	35	24	6,000	0	0	0	890	0	0		7
2,931.4	TCTu	m-dwt		4,000	3.3	12.8	37	32	31	7,000	0	0	0	500	0	0		9
3,013.9	TCTu	pwt		3,500	12.3	12.9	27	44	29	4,900	(ps)	0	0	860	0	10		>2
3,192.8	TCT1	pwt	Ar, cc	3,600	23.8	9.4	33	30	37	4,200	0	0	0	280	0	0	Many	3
3,197	TCT1	nwt		3,460	22.3	7.4	38	29	33	3,200	0	0	0	<280	0	0	?	3
3,284.5	TCT1	pwt		3,600	9.0	8.6	35	31	34	3,300	0(ps)	0	0	1,700(py)	0	4	Alt.	>8
3,515.1	TCT1	pwt		3,800	25.8	8.3	34	22	44	2,900	260(ps)	0	0	1,300(py)	0	1	Present	Present
3,724.0	TFB	fb	G1,0	3,700	0.00	8.8	0.0	0.0	100	0	11,100	15,100(ps)		8,100	0	0	Large	0
3,908.2	TFB	fb	G1	3,150	0.00	10.0	0.0	0.0	100	0	7,900	2,500	8,200(ps)	5,700	0	0	Large	0
3,956.9	TLR	pwt	Ar	0	0	0	0	0	0					2	2		Present	Present
3,969.9	TLR	nwt	Ar	3,300	9.0	17.5	2	34	64	5,500	0(ps?)	0	0	2,400	11	3	Present	Present
3,992	TLR	pwt		3,500	26.5	11.3	4	31	65	9,700	0	0	0	2,000	12	1	Present	Present
4,150.4	TLR	pwt		3,400	13.8	8.5	2	35	63	1,200	0	0	0	600	6	1	Rare	8
4,222.1	TLR	pwt		3,200	42.7	6.1	7	40	53	1,600	0	0	0	1,600	6	0	Sparse	5
4,408.4	TLR	pwt		1,800	11.8	9.2	1	37	62	5,000	0	0	0	1,100	1	1	Rare	7

Table 4-2. Summary of petrography for core samples from drillhole USW G-1, determined by point count of glass-covered thin section^{a, b} (Page 2 of 2)

Sample depth (ft) ^c	Rock Unit	Alter-ation	Points counted	Major components (volume %)		Felsic phenocrysts (relative %)			Mafic phenocrysts (ppmV)				Fe-Ti oxide (ppmV)	Accessory minerals (grains identified)			
				Lithics	Felsics	Q	K	P	Biot	Hbl	Cpx	Opx		Sph	All	Ap	Zr
4,471.0	TLR	pwt	3,600	26.4	5.1	7	36	57	3,600	0	0	0	560	8(ps)	1	Present	12
4,578.2	TLR	pwt	3,800	23.8	7.6	5	39	56	2,600	0	0	0	530	2(ps)	2	Sparse	9
4,758.4	TLR	pwt	3,800	19.0	9.2	9	38	53	2,600	0	0	0	1,500	6(5ps)	0	Sparse	10
4,849.0	TLR	pwt	3,900	13.0	8.9	10	35	55	3,300	0	0	0	1,000	7(ps)	0	Shot	10
4,917.0	TLR	nwt	3,800	5.9	5.6	14	51	35	3,200	0	0	0	1,800	11(ps)	3	Present	12
4,946.4	TTA	nwt	3,450	4.4	9.8	20	58	22	3,000	0	0	0	300	4(ps)	2	0	12
4,969.0	TTA	pwt	3,700	3.1	11.9	24	31	45	4,900	0	0	0	2,700	5(ps)	0	Partly shot	10
5,002.3	TTA	nwt	3,700	2.5	17.0	32	31	37	5,100	0	0	0	2,400	3(ps)	3	Rare	13
5,045.0	TTA	mwt	3,700	8.9	19.9	28	43	29	4,100	0	0	0	1,900	2(1ps)	0	1 ph	19
5,097.9	TTA	mwt	3,600	0.58	17.4	28	41	31	2,200	0	0	0	2,000	1(ps)	3	Rare	10
5,115.5	TTA	mwt	3,750	3.1	18.3	24	42	34	4,800	0	0	0	1,900	2	11	Sparse	20
5,141.5	TTA	mwt	3,700	9.2	14.1	27	33	40	2,100	0	0	0	1,200	4	4	Sparse	15
5,142.2	TTA	pwt	3,750	2.2	19.4	28	36	36	3,700	<270	0	0	1,900	7	5	Present	13
5,187.0	TTA	nwt	3,650	2.1	17.6	24	38	38	1,600	270	0	0	2,200	6	7	Sparse	17
5,265.6	TTA	pwt	3,400	5.4	17.4	35	31	34	2,400	590	0	0	1,800	6	8	Sparse	17
5,316.0	TTA	b	3,600	2.4	21.6	33	36	31	1,400	0	0	0	2,800	4	12	Rare	13
5,373.7	TTB	pwt	3,650	0.69	11.3	13	26	61	9,000	0	0	0	3,300	15	1	Sparse	20
5,400.0	TTB	pwt	3,800	2.3	13.6	12	30	58	4,500	0	0	0	3,200	8	1	Sparse	24
5,416.6	TTB	pwt	3,700	12.8	11.2	11	27	62	2,700	0(ps?)	0	0	3,200	5	1	Rare, shot	20
5,438.2	TTC	pwt	3,900	0.74	12.9	1	3	96	12,000	0	0	0	4,600	0	1	Present	50
5,454.1	TTC	b	3,800	1.9	14.9	16	18	66	7,100	0	0	0	5,000	9	9	Sparse	42
5,496.1	TTC	pwt	3,900	7.5	10.1	1	4	95	21,000	0	0	0	4,900	8	3	Sparse	21
5,517.3	TTC	pwt	3,600	10.3	11.8	2	9	89	11,400	0	0	0	4,500	9	2	Sparse	20
5,540.0	TTC	pwt	3,700	8.5	13.5	0.4	0.0	99.6	19,200	810(ps)	0	0	4,300	1	0	Sparse-common	29
5,558.7	TTC	pwt	3,600	0.64	19.1	4	4	92	16,400	1,100(ps)	0	0	5,300	12	3	Sparse (large)	20
5,600.0	TTC	mwt	3,750	8.4	14.8	5	7	88	10,100	800(ps)	0	0	6,400	16	4	Sparse	24
5,642.0	TTC	mwt	3,300	5.8	15.4	2	4	94	21,500	1,500(ps)	0	0	5,800	12	3	Sparse	22
5,728.0	TTC	mwt	1,650	21.1	21.6	0.0	1	99	18,200	0(ps)	0	0	7,900	4	3	Sparse (large)	24
5,841.0	TTC	mwt	1,650	10.8	18.0	0.0	4	96	12,100	600(ps?)	0	0	8,500	3	1	Sparse-common	26
5,894.3	TTC	mwt	1,650	7.2	15.5	0.0	4	96	6,100	6,100	0	0	3,600	6	1	Sparse	14
5,929.8	TTC	mwt	1,650	5.9	18.2	4	4	92	13,300	5,500(ps)	0	0	6,100	7	8	Sparse-common	24
5,944.9	TTC	m-dwt	1,600	6.3	21.6	1	6	93	6,900	7,500(ps)	0(ps)	0	5,000	8	5	Sparse-common	24
5,980.0	TTC	m-dwt	1,650	3.3	20.9	0.0	0.3	99.7	24,200	7,300(ps)	1,800(ps)	0	7,300	4	1	Sparse-common	24
5,984.7	TTC	m-dwt	1,650	2.3	25.3	0.0	0.0	100	27,200	8,500(ps)	1,200(ps)	0	9,700	1	3	Sparse-common	25

^aSource: Warren et al. (1984).

^bppmv = parts per million. Other symbols are defined in Table 4-1.

^cTo convert feet to meters, multiply by 0.328.

^dQuestion marks indicate uncertainty.

CONSULTATION DRAFT

CONSULTATION DRAFT

Table 4-3. Statistical modes^a for compositional parameters of phenocrysts for the lower Topopah Spring Member and for underlying units

Unit symbol ^b	Number of samples	Sanidine ^c		Plagioclase ^d		Mafic minerals ^e			
		mole%	wt%	mole%	An	Molecular Mg/(Mg + Fe)	Biot	Hbld	Opx
		Or + Cn	BaO	Rim	Core				Cpx
TPT1	5	59	0.00	15	17	0.43	-- ^f	--	--
TH1	3	68	0.18	19	21	0.37	--	--	--
TH2	1	73	1.0	--	37	0.45	--	--	--
TCP	14	53	0.14	11	11	0.42	--	0.29	--
TCB	22	61	0.56	15	16	0.40	0.44	--	--
TCT	18	67	0.55	21	21	0.42	--	--	--
TFB	4	--	--	61	61	--	0.66	0.70	0.73
TLR	15	65	0.67	18	19	0.59	--	--	--
TTA	8	64	0.55	17	19	0.55	--	--	--
TTB	2	66	0.96	23	29	0.59	--	--	--
TTC	6	72	3.4	29	32	0.62	--	--	--

^aThe median value very closely approximates the statistical mode for compositional parameters of sanidine and mafic minerals and is the value given for these minerals. Sample locations are shown in Warren et al. (1984).

^bUnit symbols are defined in Table 4-1.

^cOr = potassium feldspar; Cn = celsian; BaO = barium oxide.

^dAn = anorthite

^eBiot = biotite; Hbld = hornblende; Opx = orthopyroxene; Cpx = clinopyroxene.

^f-- = not present.

CONSULTATION DRAFT

Figure 4-4. Location map for the zeolite cross sections in Figures 4-5, 4-6, and 4-7. This figure and the stratigraphy shown in Figures 4-5 and 4-7 are based on Scott and Bonk (1984).

Distributions of abundant sorptive zeolites in the rock matrix are summarized in cross sections (Figures 4-5 to 4-7) from Bish and Vaniman (1985), which are located on Figure 4-4. The three cross sections transect the exploration block. Only zeolite abundances greater than 10 percent are shown in Figures 4-5 to 4-7, because lesser concentrations may not be laterally continuous and are therefore unreliable for predictions of abundance. From these figures four principal intervals of zeolitization can be identified (Bish et al., 1984). The intervals are as follows:

1. Interval I. This interval is the clinoptilolite-heulandite-smectite zone at the top of the lower Topopah Spring vitrophyre, shown near the base of Tptw. Although thin (generally less than 3 m) and irregularly developed (it is virtually absent in drillhole USW GU-3), this interval is important because it is at the immediate base of the host rock and is everywhere above the water table. This zeolitized interval is unique because it is within the densely welded region of a compound cooling unit, whereas zeolitization is generally restricted to the poorly welded or nonwelded margins of cooling units. Zeolitization in interval I may result from alteration concentrated along the devitrification front at the top of the vitrophyre. Other alteration episodes may be superimposed on this; (see Levy (1984b) for a detailed discussion).
2. Interval II. This interval is the relatively thick zone of clinoptilolite ($\pm$ mordenite) that may occur in the bedded, non-welded, and poorly welded tuffs that form the base of the Topopah Spring Member and the underlying tuffaceous beds of the Calico Hills; it often extends into the top of the Prow Pass Member of the Crater Flat Tuff. This zeolitized interval occurs within the "n" unit between Tptw and Tcpcw in Figures 4-5 and 4-7. The base of the Topopah Spring Member is the lower margin of a compound cooling unit, providing a predictable locale for zeolitization. The tuffaceous beds of Calico Hills are only partly welded to nonwelded throughout and are thoroughly zeolitized across the northern and eastern part of the exploration block. However, along the crest of Yucca Mountain this interval may be incompletely zeolitized (drillhole USW H-5) or nonzeolitized and vitric (drillhole USW GU-3) (Vaniman et al., 1984).
3. Interval III. This interval contains the partially welded and bedded tuffs at the base of the Prow Pass Member and at the top of the underlying Bullfrog Member compound cooling unit. This interval is consistently zeolitized throughout Yucca Mountain in the vicinity of the exploration block (Vaniman et al., 1984). In Figures 4-5, 4-6, and 4-7, it can generally be traced within the "n" unit between Tcpcw and Tcbw.
4. Interval IV. This interval contains the poorly welded and bedded tuffs at the base of the Bullfrog Member compound cooling unit and at the top of an underlying cooling unit within the uppermost Tram Member. This interval also consists of clinoptilolite plus mordenite. In Figures 4-5, 4-6, and 4-7, it can generally be traced within the "n" unit below the Tcbw.

Figure 4-5. Major zeolite distributions (Bish and Vaniman, 1985), north-south cross section, shown as a dark pattern overlying the stratigraphic cross section of Scott and Bonk (1984). Location of cross section is shown on Figure 4-4. Unit designations follow the usage of Scott and Bonk (1984). Tpcw, Tpyw, Tppw, and Tptw represent the welded zones of the Tiva Canyon, Yucca Mountain, Pah Canyon, and Topopah Spring members of the Paintbrush Tuff; Tcbw, Tctw, and Tfb represent the welded zones of the Prow Pass, Bullfrog, and Tram members of the Crater Flat Tuff; Tfb represents lavas and flow breccias; the Totw represents the welded zones of older tuffs. The symbol n is used for all nonwelded tuff zones. The subhorizontal barred line indicates the water table as inferred from static water levels in the drillholes.

Figure 4-6. Major clinoptilolite-mordenite distributions (Bish and Vaniman, 1985), northwest-southeast cross section. Location of cross section is shown on Figure 4-4. Information on drillhole USW H-5 below 760 m is from Bentley et al. (1983), and information on the central part of the drillhole USW H-6 is from Craig et al. (1983).

Figure 4-7. Major zeolite distributions (Bish and Vaniman, 1985), center to southeast cross section (Figure 4-5).

The weight percentages of zeolites in the rock in these intervals and the interval thicknesses are summarized in Table 4-4 for drillholes USW G-1, USW H-3, USW H-4, USW H-5, USW GU-3, USW G-3, USW G-4, UE-25a#1, and UE-25b#1 in and around the exploration block. Well J-13 is not included because the drillhole was not continuously cored and samples are lacking from certain critical intervals, particularly interval II within the tuffaceous beds of Calico Hills; well J-13 is also outside the Yucca Mountain exploration block. Drillhole USW G-2 is not included in Table 4-4 because it is 2 km north of the present exploration block and hydrothermal processes may have affected the deep alteration mineralogy at the northern end of Yucca Mountain (Caporuscio et al., 1982; Bish, 1987). The cumulative thicknesses of zeolitized tuff between the host rock and the static water levels in the drillholes are tabulated by Vaniman et al. (1984), and range from 44 to 142 m.

Much work remains to be done on the mineralogy along potential flow paths away from the host rock. Data from several drillholes drilled to the water table within and around the exploration block will be obtained during site characterization (Section 8.3.1.3). Other drillholes may be situated along possible transport pathways away from the exploration block (Section 8.3.1.3). Samples from the exploratory shaft facility will also become available. Present assessment of sorptive minerals along flow paths is still restricted to the unsaturated zone as long as flow paths within the saturated zone are unknown (Section 3.9.3.1). Mineral abundances along possible flow paths will be compiled as more data become available.

Yucca Mountain zeolites and smectites vary in composition, which may control many chemical and physical properties. The properties affected include cation exchange and selectivity (Sherry, 1969; Breck, 1974; Vaughan, 1978), thermal expansion (Bish, 1984), and dehydration behavior (Bish, 1985). Quantitative sorption ratios for different radionuclides as a function of individual clinoptilolite, heulandite, or mordenite composition are lacking, but data will be collected to define the dependence of sorption ratio on individual mineral composition (Section 8.3.1.3). Vaughan (1978) describes selectivity sequences for clinoptilolite in the order potassium>calcium>sodium. This selectivity sequence results in K-clinoptilolite being more difficult to cation exchange than Na- or Ca-clinoptilolite. The distributions and compositions of various zeolite types are being studied to determine their origins and the nature of chemical transportation during alteration, as well as to provide a basis for estimating overall retardation (Sections 4.1.3.3 and 8.3.1.3).

Present data on compositional variations in Yucca Mountain zeolites and clay minerals as a function of depth are summarized in the discussion that follows. The discussion includes consideration of the effects these variations in composition will have on rock properties. The zeolitized intervals I through IV, described earlier in this section and in Table 4-4, are used in the summary of sorptive zeolite compositions.

Sorptive zeolites (clinoptilolite, heulandite, mordenite). Representative compositions for the clinoptilolite-heulandite group minerals from Yucca Mountain are given in Table 4-5. Histograms showing the distribution of silicon-to-aluminum ratios in these minerals are presented in Figure 4-8; consistency in silicon-to-aluminum ratios may be one measure of consistency

Table 4-4. Zeolitized intervals for Yucca Mountain^{a, b}

Interval	USW G-1 SWL=577 m	UE-25a#1 UE-25b#1 SWL=471 m	USW G-4 SWL=541 m	USW H-4 SWL=519 m	USW H-5 SWL=704 m	USW H-3 SWL=754 m	USW GU-3 USW G-3 SWL=754 m
I: Above the lower Topopah Spring vitrophyre	Depth: 392-393 m (1 m thick) 15% cpt	Depth: 385-398 m (3 m thick) 7% cpt	Depth: 396-401 m (5 m thick) 10% (± 12) cpt	Depth: 357-361 m (5 m thick) No samples	Depth: 485 m 10% cpt	Depth: 367-371 m (4 m thick) No samples	Depth: 360-364 m (4 m thick) Trace cpt
II: Base of the Topopah Spring unit, tuff of Calico Hills	Depth: 425-565 m (140 m thick) 52% (± 17) cpt	Depth: 404-556 m (152 m thick) 67% (± 6) cpt 17% (± 20) mord	Depth: 420-545 m (125 m thick) 50% (± 19) cpt 5% (± 7) mord	Depth: 400-504 m (104 m thick) 63% (± 13) cpt 8% (± 12) mord	Depth: 584-594 m (10 m thick) 37%	Vitric (nonzeolitized)	Vitric (nonzeolitized)
III: Between the Prow Pass and the Bullfrog units	Depth: 622-706 m (84 m thick) 45% (± 12) cpt 18% (± 18) mord	Depth: 636-710 m (74 m thick) 60% cpt & mord	Depth: 600-682 m (82 m thick) 32% (± 12) cpt 19% (± 10) mord	Depth: 596-698 m (102 m thick) No samples	Depth: 665-689 m (34 m thick) 60% cpt	Depth: 549-610 m (61 m thick) 68% cpt	Depth: 557-613 m (56 m thick) 58% (± 18) cpt
IV: Between the Bullfrog and the Tram units	Depth: 779-823 m (44 m thick) 37% (± 6) cpt 15% (± 11) mord	Depth: 863-890 m (27 m thick) 4% (± 4) cpt 16% (± 10) mord	Depth: 828-860 m (32 m thick) 9% (± 9) cpt 35% (± 33) mord	Depth: 765-774 m (9 m thick) No samples	No samples	Depth: 732-760 m (28 m thick) 52% (± 17) cpt 18% (± 13) mord	Depth: 776-822 m (46 m thick) 36% (± 8) cpt

^aSource: Vaniman et al. (1984).^bcpt = clinoptilolite and heulandite; mord = mordenite; and SWL = static water level (given in depth below surface). Percentages represent weight.

Table 4-5. Representative clinoptilolite analyses for Yucca Mountain^a (page 1 of 2)

Mineral	Interval I		Intervals II, III, and IV and deeper occurring clinoptilolite								
	Pah Canyon Member (G2-584)	Topopah Spring Member (H5-1666)	Calic Suite, Eastern Yucca Mountain				Alkalic Suite, Western Yucca Mountain				
			Tuff of Calico Hills (25p1-1250)	Prow Pass Member (25p1-1700)	Bullfrog Member (25b1-2879)	Tuff of Lithic Ridge (25p1-3330)	Tuff of Calico Hills (G1-1774)	Prow Pass Member (G3-1874)	Bullfrog Member (G3-2615)	Tram Member (G3-3589)	Tuff of Lithic Ridge (G3-4423)
CONCENTRATION (WT%)											
SiO ₂	65.6	65.5	68.6	67.1	56.1	60.6	68.1	68.2	63.9	65.1	65.3
TiO ₂	0.00	0.02	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Al ₂ O ₃	13.8	13.3	12.4	12.9	16.9	16.0	12.2	11.9	11.7	11.4	12.6
Fe ₂ O ₃	0.00	0.00	0.00	0.00	0.05	0.00	0.00	0.00	0.00	0.51	0.12
MgO	1.47	0.86	0.07	0.32	0.22	0.08	0.09	0.00	0.00	0.12	0.22
CaO	4.23	5.19	3.59	4.37	6.77	7.22	1.11	1.95	1.57	0.79	2.35
BaO	0.02	0.00	0.00	0.05	0.21	0.19	0.03	0.00	0.00	0.19	0.27
Na ₂ O	0.18	0.17	1.13	1.55	0.89	0.85	2.84	1.69	2.88	2.23	3.21
K ₂ O	1.09	0.21	3.01	1.23	1.43	0.41	4.20	5.20	2.71	5.55	1.15
Total	86.3	85.2	88.8	87.5	82.6	85.4	88.6	88.9	82.8	85.9	85.2
CATION CONCENTRATION (%)											
Si	72.60	73.90	73.60	72.90	64.30	67.60	71.90	72.40	72.20	71.20	71.80
Ti	0.00	0.02	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Al	18.00	17.70	15.70	16.50	22.80	21.10	15.20	14.90	15.60	14.70	16.40
Fe ³⁺	0.00	0.00	0.00	0.00	0.04	0.00	0.00	0.00	0.00	0.42	0.10
Mg	2.43	1.45	0.11	0.52	0.38	0.13	0.14	0.00	0.00	0.20	0.36
Ca	5.02	6.27	4.13	5.08	8.31	8.64	1.26	2.22	1.90	0.93	2.77
Ba	0.01	0.00	0.00	0.02	0.09	0.08	0.01	0.00	0.00	0.08	0.12
Na	0.39	0.37	2.35	3.26	1.98	1.84	5.82	3.48	6.32	4.73	6.85
K	1.54	0.30	4.12	1.70	2.09	0.58	5.66	7.04	3.91	7.75	1.61

Table 4-5. Representative clinoptilolite analyses for Yucca Mountain^a (page 2 of 2)

Mineral	Intervals II, III, and IV and deeper occurring clinoptilolite										
	Interval I		Calcic Suite, Eastern Yucca Mountain					Alkalic Suite, Western Yucca Mountain			
	Pah Canyon	Topopah	Tuff of	Prow Pass	Bullfrog	Tuff of	Tuff of	Prow Pass	Bullfrog	Tram	Tuff of
	Member	Spring	Calico	Member	Member	Lithic	Calico	Member	Member	Member	Lithic
	(G2-584)	(H5-1686)	Hills	(25p1-1700)	(25b1-2879)	Ridge	Hills	(G3-1874)	(G3-2615)	(G3-3589)	Ridge
	(G2-584)	(H5-1686)	(25p1-1250)	(25p1-1700)	(25b1-2879)	(25p1-3330)	(G1-1774)	(G3-1874)	(G3-2615)	(G3-3589)	(G3-4423)
$(Al + Fe)/(2Mg + 2Ca + 2Ba + Na + K)$											
	1.07	1.10	1.05	1.02	1.05	1.05	1.06	1.00	1.11	1.01	1.10
$Si/(Al + Fe)$											
	4.03	4.18	4.69	4.42	2.82	3.20	4.73	4.86	4.63	4.71	4.35
RELATIVE ABUNDANCE OF EXCHANGEABLE CATIONS (MOLE %)											
K	16	4	38	16	16	5	44	55	32	57	14
Na	4	4	22	31	16	17	45	27	52	35	59
Ca+Mg	80	92	40	53	68	78	11	18	16	8	27

^aSample numbers are given in parenthesis under the unit name and consist of abbreviated drillhole designation (G2 = USW G-2, H5 = USW H5, 25p1 = UE-25p#1, 25a1 = UE-25a#1, 25b1 = UE-25b#1, G1 = USW G-1, G3 = USW G-3) followed by sample depth from surface in feet. Data are from Broxton et al. (1986).

^bTotal iron calculated as Fe_2O_3 .

INTERVAL I CALCIC CLINOPTILOLITES
AND HEULANDITES

INTERVALS II THROUGH IV
CLINOPTILOLITES ON EASTERN
SIDE OF YUCCA MOUNTAIN

INTERVALS II THROUGH IV
CLINOPTILOLITES ON WESTERN
SIDE OF YUCCA MOUNTAIN

Figure 4-8. Histograms comparing the silicon-to-aluminum ratios for heulandites and calcic clinoptilolites from Yucca Mountain, Nevada. Modified from Broxton et al. (1986).

in sorptive behavior (Rundberg et al., 1985). Triangular plots showing vertical and lateral variations in exchangeable-cation ratios are presented in Figures 4-9 and 4-10. In zeolitized interval I, clinoptilolite and heulandite are calcium rich and have silicon-to-aluminum ratios ranging from 4.0 to 5.0 (Figure 4-8 shows some examples). The magnesium content of these zeolites is generally high, ranging from 0.6 to 1.5 percent magnesium oxide by weight. Using the thermal stability criteria of Mumpton (1960), Levy (1984a) determined that at least some of the calcium-rich clinoptilolite group minerals that occur at the top of the Topopah Spring vitrophyre belong to the heulandite structural group. Clinoptilolite-group zeolites from zeolitized intervals II through IV show more chemical diversity than those in interval I. In these deeper intervals, clinoptilolite compositions are subdivided into an eastern calcic suite of samples and a western alkalic suite of samples based on their exchangeable-cation contents (Figure 4-10). A transitional zone, found in drillholes USW G-4 and USW H-4, has characteristics intermediate between the eastern and western suites. The eastern clinoptilolites show a strong trend toward calcium enrichment with depth. In contrast, the western clinoptilolites become more sodium rich with depth. Silicon-to-aluminum ratios for both suites are similar (Rundberg et al., 1985), falling mostly between 4.0 and 5.2 and clustering around 4.6 (Figure 4-8). However, the silicon-to-aluminum ratios in eastern clinoptilolites have a bimodal distribution with a smaller population of ratios ranging from 2.8 to 3.6. This smaller population of low silicon-to-aluminum ratios correlates with the most calcic clinoptilolite compositions (approximately 70 mole percent calcium plus magnesium) in interval IV.

The other major zeolite type at Yucca Mountain is mordenite. Its chemistry is poorly known because individual mordenite crystals are fine grained and intergrown with other authigenic phases so that analysis by electron-microprobe and mineral-separation techniques is not feasible. The determination of mordenite compositions by bulk-rock techniques is difficult because monomineralic beds of mordenite are extremely rare at Yucca Mountain. In drillhole USW G-4, a bulk-rock sample containing more than 80 percent mordenite contains relatively equal amounts of sodium, potassium, and calcium (Broxton et al., 1986). These mordenites appear to be somewhat more sodic and less calcic than coexisting clinoptilolites.

The data from drillholes (Broxton et al., 1986) show that the first clinoptilolite-heulandites encountered below the repository horizon contain major amounts of calcium and minor amounts of potassium and sodium. Clinoptilolites of this composition will undergo cation exchange more readily than the potassium-rich clinoptilolites below in zeolitized interval II. In addition, the calcic clinoptilolites close to the repository will contract less at elevated temperatures than the deeper sodium-rich zeolites (Section 4.2.2), and fewer rock mechanical problems and permeability changes will be encountered because of this. Below the lower Topopah Spring vitrophyre, in the nonwelded base of the Topopah Spring Member and in the tuffaceous beds of Calico Hills, clinoptilolites contain subequal amounts of potassium and calcium. Below the Prow Pass Member the clinoptilolites are generally sodium rich, and sodium is more easily exchanged than calcium or potassium. Sodium-rich clinoptilolites contract appreciably when temperatures are elevated (Bish, 1985), but these zeolites only occur beyond the anticipated thermally disturbed zone and should pose no mechanical problems. Plans to test this are discussed in Section 8.3.1.3.2.

- COMPOSITIONS FOR PAH CANYON MEMBER FOR DRILL HOLE USW G-2
- COMPOSITIONS OF THE ZEOLITES THAT OCCUR ON TOP OF TOPOPAH SPRING BASAL VITROPHYRE, FOR DRILL HOLES THROUGHOUT YUCCA MOUNTAIN

Figure 4-9. Triangular diagram showing the exchangeable-cation ratios in calcic-clinoptilolite and heulandite from zeolitized interval I at Yucca Mountain. Data plotted are in weight percent. Modified from Broxton et al. (1986).

- NONWELDED BASE OF TOPOPAH SPRING MBR., TUFF OF CALICO HILLS, AND TOP OF PROW PASS MBR.
- NONWELDED BASE OF PROW PASS MBR. AND TOP OF BULLFROG MBR.
- △ NONWELDED TOP OF BULLFROG MBR. AND TRAM MBR.
- PRE-TRAM VOLCANIC ROCKS

NOTE: SMALL ARROWS WITHIN PLOTS INDICATE CHEMICAL TRENDS WITH INCREASING DEPTH.

Figure 4-10. Triangular diagrams showing ratios of exchangeable cations in clinoptilolites at Yucca Mountain. Modified from Vaniman (1987).

Smectites. Smectites are widespread throughout the units underlying the Topopah Spring Member (Bish, 1987) and therefore occur along possible flow paths. Although smectite abundances are typically 1 to 10 percent, amounts as high as 50 percent occur in some units (Figures 4-1 through 4-3). The sorption properties of smectites are not as strong a function of composition as they are for zeolites, but smectite dehydration under slightly elevated temperatures (up to 100°C) is determined largely by composition (Bish, 1981). In addition, information on geologic processes in the past can be obtained from data on smectite compositions and structures (Bish, 1987). This information can aid in the prediction and understanding of the effects of future elevated temperatures.

Smectites in the host rock unit and along flow paths into the saturated zone have mixed potassic, sodic, and calcic interlayers. These cations are easily exchanged by many small and large cations and complexes. There is a minor increase in the potassium content of smectites with depth throughout most of the exploration block (Caporuscio et al., 1982). However, potassium contents of the smectites between the host rock and the Tram Member are still low enough so that the sorptive properties of the smectites should not significantly differ from the smectites at shallower depths. Clays below the Tram Member include a higher content of illite, which will be less sorptive. Only in the deepest rocks at the northern end of the repository block (drillhole USW G-2) are the smectites significantly transformed to illite (Caporuscio et al., 1982; Bish, 1987). In the bulk of the repository block, the smectites retain their highly sorptive nature. As a result of their minor compositional variations, the smectite sorptive properties should not vary significantly near a Yucca Mountain repository. Plans to test this are discussed in Section 8.3.1.3.

In addition to their importance for sorption, the smectites can provide information on the geologic history of Yucca Mountain. Within and north of Drill Hole Wash, the data show evidence of hydrothermal alteration at depth, with large potassium increases and expandability decreases in smectites more than 1,097 m below the surface (Bish, 1987). Smectites above the Bullfrog Member in drillhole USW G-2 are not altered, whereas deeper smectites and interstratified illite-smectites suggest that alteration occurred at temperatures as high as 230°C at 1,158-m depth. These data (Caporuscio et al., 1982) and preliminary potassium-argon radiometric dates for illites from drillholes USW G-1 and USW G-2 suggest that the hydrothermal alteration occurred 10.9 ± 0.2 million years ago in the northern end of the exploration block (Bish, 1987). Smectites at depth in the southern portion of the block (drillhole USW G-3) do not show evidence of this alteration and remain fully expandable to the bottom of the drillholes examined. Even in the northern end of Yucca Mountain, expandable sodium- and calcium-containing smectites are found at depth, along with potassium-rich, nonexpandable clay. This suggests that some smectite formation also occurred after hydrothermal alteration. The use of these deeper tuffs as natural analogs to repository-induced thermal alteration is discussed by Bish (1987).

4.1.1.4 Mineral stability

Mineral stability in Yucca Mountain must be considered in two ways. The first is to consider whether the present mineral assemblages in Yucca Mountain are stable under present conditions and, if they are not, what mineralogic changes are probable. The second consideration is what effect repository heating will have on mineral stability, due to both increased temperature and possible dehydration of the rock. This second consideration is covered in Section 4.2.

The mineralogy of the host rock is primarily quartz, cristobalite, and alkali feldspar with some clays present (Section 4.1.1.3). The stability of clays, cristobalite, glass and zeolites is of concern under present and expected conditions. The cristobalite in these rocks is metastable and will very slowly be converted to quartz. The slow rate of this conversion is evidenced by the persistence of cristobalite in these tuffs, which are about 12 million years old (Kistler, 1968). The disappearance of cristobalite with depth is probably a result of the higher temperatures experienced by tuffs deeper in Yucca Mountain. If the present geothermal gradient (Section 4.1.2.8) is responsible, then repository-induced temperature increases of a few tens of degrees could cause appreciable transformation of cristobalite to quartz. However, there is evidence of increased thermal gradients in at least some parts of Yucca Mountain in the past (Bish, 1987). Thus, the lack of cristobalite lower in the mountain may result in part from past hydrothermal alteration. Plans for obtaining the additional information on silica polymorph stability are outlined in Section 8.3.1.3.3.

There are two effects when cristobalite transforms to quartz. First, there is a 12-percent volume decrease (using molar volumes for quartz and cristobalite from LePage and Donnay (1976) and Peacor (1973), respectively) that might affect the mechanical and hydrologic properties of the tuff; and second, there is a lowering of the aqueous silica activity. No water compositions are available from the unsaturated zone in which the host rock is located, but water samples taken from below the water table where cristobalite is present show approximate saturation with respect to cristobalite despite the presence of a considerable amount of quartz. This observation suggests that the aqueous silica activity in a particular horizon will remain high until most, if not all, of the cristobalite has converted to quartz. The silica-activity evolution is of particular concern because silica activity appears to be important in controlling the stability of clinoptilolite (Keith et al., 1978).

Clays are present in small quantities in the host rock and occur throughout Yucca Mountain as late-stage alteration minerals (Caporuscio et al., 1982). Clays are also observed as alteration products of Topopah Spring tuff in hydrothermal experiments (Section 7.4.1.3). On the basis of these observations and on reaction-path studies of glass alteration (Kerrisk, 1983) it seems possible that clays may be continuing to crystallize in Yucca Mountain. The stability of clays is of concern because they are highly sorptive minerals. At present it seems likely that the amount of clays will remain constant or increase in Yucca Mountain. The possibility of clay dehydration or breakdown due to repository-induced heating is discussed in Section 4.2.2.

The smectites at or above the stratigraphic level of the Tram Member in Yucca Mountain have interlayer water that can be driven out with minor temperature increases or with minor changes in the partial pressure of water, causing a decrease in volume (Bish, 1981). This volume reduction is reversible at low temperatures. The implications of irreversible collapse are discussed in Section 4.2.2.

Below the host-rock horizon, successive zones of glass, heulandite and smectite, clinoptilolite, analcime, and authigenic albite are found in addition to devitrified zones dominated by alkali feldspar and silica phases. The stability of clinoptilolite is of particular concern because it has been identified as an important sorptive phase (Section 4.1.3.3).

Smyth (1982) has summarized the diagenetic zones beneath Yucca Mountain. This diagenesis begins with unaltered glass. Unaltered glass occurs in many rocks above the water table, including the vitrophyre of the host rock (where devitrification is not extensive). The next zone contains clinoptilolite or mordenite or both (Calico Hills and the upper zeolitic Crater Flat Tuff). Below these zeolites, analcime occurs, replacing clinoptilolite and mordenite. Finally, authigenic albite is typical of the deepest tuff units examined. This generalized zonation can be seen in Figures 4-1 and 4-2.

There are varying hypotheses to explain this diagenetic zonation. Smyth and Caporuscio (1981) give a plot of sodium-ion concentration versus temperature appropriate for the zeolite transition in this diagenetic sequence. Increases in sodium concentration appear to decrease the temperatures at which these reactions occur. Because the present ground water has a low sodium content (37 to 171 mg/L, Ogard and Kerrisk, 1984), the analysis of Smyth (1982) would predict that the clinoptilolite-to-analcime transformation, if it occurred, would be expected at temperatures of about 100 to 120°C.

There are, however, other factors that must be considered in this analysis. First, the stoichiometry of clinoptilolite and analcime does not suggest that the concentration of the sodium ion should affect the relative stability of the two phases. Second, the transition from clinoptilolite to analcime is also characterized by the disappearance of cristobalite, which suggests that silica activity with analcime is lower than that with clinoptilolite. Because silica evolves in the transformation of clinoptilolite to analcime, a lowering of the silica activity would move the system toward analcime stability relative to clinoptilolite. Reaction-path modeling for temperatures up to 175°C (Kerrisk 1983) indicates that increase in temperature alone does not account for the transition from clinoptilolite to analcime; rather, this transition can probably be accounted for by the change in silica activity that occurs with the change from cristobalite to quartz saturation. Field observations (Keith et al., 1978) also indicate that silica activity, rather than temperature, may be a factor controlling the transformation of clinoptilolite to analcime. Keith et al. (1978) observed clinoptilolite and analcime in drillhole Y-8 in Yellowstone National Park. Clinoptilolite was observed in a temperature interval from about 70°C to slightly above 150°C. However, throughout this interval, clinoptilolite zones were interlayered with analcime-bearing zones, indicating that temperature was not a controlling factor in determining whether clinoptilolite or

analclime was present. There was, however, an excellent correlation between the presence of clinoptilolite and that of cristobalite and between analclime and quartz (Figure 4-1), suggesting that silica activity was the controlling variable in determining the presence of clinoptilolite or analclime.

The observations on which the conclusions of Smyth (1982) were based probably reflect the effects of temperature on the kinetics of the transition from cristobalite to quartz more than an equilibrium temperature for the transformation of clinoptilolite to analclime. The apparent dependence of transition temperature on the sodium content of the solution may arise from two sources. The generally higher pH high-sodium solutions may increase the rate of transition from cristobalite to quartz. Also, as the sodium concentration increases, the activity of water is lowered by the increasing salinity of a solution, which would favor formation of the less hydrous phase, analclime.

Thus, it seems likely that the stability of clinoptilolite in Yucca Mountain is determined largely by the silica activity. If that activity in a particular horizon is at or above cristobalite saturation, the clinoptilolite is probably stable to at least 150°C. If clinoptilolite breaks down after the emplacement of waste in a repository at Yucca Mountain, it will probably be because repository heating will have increased the rate of cristobalite conversion to quartz. Plans to obtain the information to assess the likelihood of cristobalite transformation are discussed in Section 8.3.1.3.

The effects of composition on thermal expansion and thermal stability have been determined for several cation-exchanged clinoptilolites (Bish, 1984). It is known that sodium- and potassium-rich clinoptilolites are stable to higher temperatures than the calcium-rich clinoptilolites (Shepard and Starkey, 1966) that commonly occur above the basal vitrophyre in the Topopah Spring Member. However, the potassium-rich clinoptilolites in the base of and slightly below the Topopah Spring Member contract less with increased temperature than do the sodium-rich clinoptilolites farther along flow paths (Bish, 1984).

Glass is another phase in Yucca Mountain that is certainly metastable. The distribution and compositions of glass are discussed by Caporuscio et al. (1982), Vaniman et al. (1984), Bish and Vaniman (1985), and Broxton et al. (1986). Glass exists in the vitrophyre in the lower Topopah Spring Member and in parts of the tuffaceous beds of Calico Hills (Figures 4-1 and 4-2). Alteration of the glass originally present in the tuffaceous beds of Calico Hills in the northeastern part of Yucca Mountain has led to the extensive development of clinoptilolite (Section 4.1.1.3). The persistence of glass in the southwest area of Yucca Mountain may be due to the lack of water for hydration because the glasses are above the water table. However, since considerable pore water is present, the general lack of clinoptilolite formation may also be due to lack of reaction because of low temperature. There is a possibility that glass would react to form appreciable clinoptilolite in response to moderate heating from the repository, or to increased hydration that would result from a rise in the water table. Future glass stability studies are outlined in Section 8.3.1.3.3.

4.1.2 GROUND-WATER CHEMISTRY

The composition of water in the Yucca Mountain exploration block is an important parameter for evaluating the suitability of the site for the containment and isolation of radioactive waste. The water composition will affect the rate at which the waste containers corrode, the leach rate of the waste form, and the rate of radionuclide transport to the accessible environment. A reference ground water (i.e., water from well J-13) similar in composition to the water of Yucca Mountain is needed to conduct transport and retardation experiments.

Because the composition of ground water is controlled by a number of variables both internal and external to the repository horizon, it is necessary to define the range of ground-water compositions, in both the saturated and the unsaturated zones, from the exploratory block to the accessible environment. In addition, it is necessary to identify the main factors that would determine how changes in the natural (e.g., increased recharge) and man-induced conditions would affect the ground-water compositions.

At present, the only water available for chemical analysis has come from saturated-zone wells on and near Yucca Mountain. When the exploratory shafts are constructed, water from the pores in unsaturated tuff, any water flowing in fractures in unsaturated tuff, and water from any perched water zones in Yucca Mountain will be sampled, where possible, and analyzed. Plans for further analyses of the saturated-zone ground water can be found in Section 8.3.1.2. The remainder of this section discusses ground water from the saturated zone. Plans for studying the unsaturated-zone water chemistry can be found in Section 8.3.1.2.2.

4.1.2.1 General description of the hydrochemistry

Analyses of ground water from drillholes that penetrate the host rock and surrounding units in the area of the exploration block indicate that they are principally sodium bicarbonate waters (Benson et al., 1983; Ugard and Kerrisk, 1984) with low contents of total dissolved solids (200 to 400 mg/L). However, water from the carbonate aquifer (Drillhole UE-25p#1) contains over 1,000 mg/L total dissolved solids. Tables 4-6 and 4-7 summarize the geochemical data reported on ground-water samples from Yucca Mountain and its vicinity. The data are consistent with the data reported in Section 3.9.1.3, but include additional species (dissolved iron, manganese, aluminum, nitrite, nitrate, and oxygen) not reported there. The analytical techniques and sampling procedures by which these data were obtained are reported by Ugard and Kerrisk (1984). The locations of wells are shown in Figure 4-11.

4.1.2.2 Major inorganic content

The content of major and minor inorganic constituents of ground water from Yucca Mountain and vicinity are listed in Tables 4-6 and 4-7 and are discussed in the following paragraphs.

CONSULTATION DRAFT

Table 4-6. Element concentrations in ground water from the vicinity of Yucca Mountain

Well ^b	Field pH	concentrations (mg/L)								
		Ca	Mg	Na	K	Li	Fe	Mn	Al	Si
USW VH-1 ^c	7.5	10	1.5	80	1.9	0.090	-- ^d	--	--	23
USW H-6	7.4	5.5	0.22	74	2.1	0.10	0.12	0.04	0.12	20.0
USW H-3	9.4	0.8	0.01	124	1.5	0.22	0.13	0.01	0.51	16.9
USW H-5	7.1	1.1	0.03	54	2.3	0.04	0.01	--	0.17	17.4
USW G-4	7.1	9.2	0.15	56	2.5	0.08	0.04	0.02	0.02	19.6
USW H-1 ^c	7.5	6.2	<0.1	51	1.6	0.04	--	--	--	19
USW H-4	7.4	10.8	0.19	84	2.6	0.16	0.03	0.005	0.04	25.9
UE-25b#1	7.7	19.7	0.68	56	3.3	0.28	0.04	0.004	0.03	31.5
UE-25b#1 ^e	7.2	18.4	0.68	46	2.5	0.30	0.69	0.36	0.04	28.7
UE-25b#1 ^f	7.3	17.9	0.66	37	3.0	0.17	0.08	0.07	0.06	28.8
J-13	6.9	11.5	1.76	45	5.3	0.06	0.04	0.001	0.03	30.0
UE-29a#2	7.0	11.1	0.34	51	1.2	0.10	0.05	0.03	0.04	25.8
J-12 ^c	7.1	14	2.1	38	5.1	-	-	-	-	25
UE-25p#1 ^g	6.7	87.8	31.9	171	13.4	0.32	<0.1	<0.01	0.1	30

^aConcentrations from Ugard and Kerrisk (1984) unless otherwise noted.

All samples are integral water samples unless otherwise noted.

^bSee Figure 4-11 for locations.

^cData from Benson et al. (1983).

^d--indicates the element was not detected.

^eBullfrog zone, 4th day of pumping.

^fBullfrog zone, 28th day of pumping.

^gFrom dolomite aquifer.

CONSULTATION DRAFT

Table 4-7. Anion concentrations and other measurements for ground water from the vicinity of Yucca Mountain

Well ^b	Concentration ^a (mg/L)								Eh ^c
	F ⁻	Cl ⁻	SO ₄ ²⁻	HCO ₃ ⁻	NO ₂ ⁻	NO ₃ ⁻	O ₂	Detergent	
USW VH-1 ^d	2.7	11	44	167	--	--	--	--	--
USW H-6	4.1	7.7	27.5	--	-- ^e	5.3	5.6	--	395
USW H-3	5.4	8.3	31.2	245	<0.10	0.2	<0.1	<0.02	-123
USW H-5	1.3	5.7	14.6	--	--	8.6	6.3	<0.005	353
USW G-4	2.4	5.5	15.7	--	--	5.5	6.4	--	402
USW H-1 ^d	1.0	5.8	19	122	--	--	--	--	--
USW H-4	4.5	6.2	23.9	--	--	4.7	5.8	>2	216
UE-25b#1	1.2	7.1	20.6	--	--	0.6	1.8	--	220
UE-25b#1 ^f	1.5	9.8	21.0	--	0.5	2.2	<0.1	2.7	-18
UE-25b#1 ^g	1.2	6.6	20.3	--	--	4.5	1.8	0.02	160
J-13	2.1	6.4	18.1	143	--	10.1	5.7	--	--
UE-29a#2	0.56	8.3	22.7	--	--	18.7	5.7	--	305
J-12 ^d	2.1	7.3	22	119	--	--	--	--	--
UE-25p#1 ^h	3.5	37	129	698	--	<0.1	--	<0.2	360

^aConcentrations from Ogard and Kerrisk (1984) unless otherwise noted.

All samples are integral samples unless otherwise noted.

^bSee Figure 4-11 for location.

^cmV versus H₂ electrode.

^dData from Benson et al. (1983).

^e--indicates the element was not detected.

^fBullfrog zone, 4th day of pumping.

^gBullfrog zone, 28th day of pumping.

^hFrom dolomite aquifer.

Figure 4-11. Selected drillhole and well locations on and near Yucca Mountain exploration block (the names of the wells and drillholes have been shortened for clarity: for example, drillhole USW H-6 is listed as H-6 and drillhole UE-25p#1 as P-1, etc.). Modified from Ogard and Kerrisk (1984).

The data presented in Tables 4-6 and 4-7 are representative of the chemistry of the ground waters at Yucca Mountain. Kerrisk (1987) is a more recent review of the water chemistry at Yucca Mountain and in the surrounding area. This report references and compiles hundreds of analyses representing a large data base for the saturated zone water chemistry. More hydro-chemistry work is planned in Section 8.3.1.2 for the saturated zone and and more importantly the unsaturated zone. Section 8.3.1.3.1 will synthesize the present and future data into a ground-water chemistry model.

The dominant cations in the Yucca Mountain ground water are sodium, calcium, potassium, and magnesium. Sodium is the most abundant, accounting for 65 to 95 percent of the cations present. The relative concentrations of cations are shown in a ternary diagram in Figure 4-12. Data from three other wells are also shown on these figures; wells UE19e and U20a-2 are in tuffaceous rocks on Pahute Mesa (Claassen, 1973), and well 9 is in the Amargosa Desert (Claassen, 1985).

The additional wells were included because ground water at Yucca Mountain is probably derived by subsurface flow from recharge areas at higher altitudes to the north such as Pahute Mesa (Winograd and Thordarson, 1975), and because the southern Amargosa Desert is considered a discharge area for water moving beneath Yucca Mountain (Winograd and Thordarson, 1975). The shaded areas in Figure 4-12 represent the ranges of compositions of interstitial and fracture waters in ash-flow tuffs at Rainier Mesa north of Yucca Mountain (White et al., 1980). Wells on Yucca Mountain and just to the west (USW H-3, USW H-5; and USW H-6; Figure 4-11) plot nearest the sodium apex of Figure 4-12. Wells on the eastern slopes and washes (USW H-1, USW H-4, USW G-4, J-13, and UE-25b#1) show increasing levels of calcium. The carbonate aquifer well (UE-25p#1) and the well from the Amargosa Desert (well 9) show the highest relative calcium contents. The two wells on Pahute Mesa (wells UE19e and U20a-2) are similar to the wells on and just east of Yucca Mountain; this is consistent with a relation in actual recharge or recharge mechanisms between the two areas. The data for the wells plot in a band that is generally parallel to the sodium-calcium axis in Figure 4-12, which indicates that the relative potassium content of the waters is nearly constant.

The concentrations of iron and manganese are generally low (Table 4-6). The elevated concentrations in drillhole UE-25b#1 after the fourth day of pumping may reflect contaminants from the drilling operation because the concentrations dropped down to more typical values after 28 d. Both iron(III)/iron(II) and manganese(IV)/manganese(II) as well as nitrate/nitrite have been considered as oxidation-reduction (redox) couples that reflect the redox potential of the ground waters (Rundberg et al., 1985). Analytical data are available only for the nitrate/nitrite couple in drillhole UE-25b#1 water. The redox potential, Eh, calculated from these data (100 to 400 mV) is substantially different from the Eh measured in this water with a platinum electrode (80 to 170 mV) (Rundberg et al., 1985). Such differences are commonly observed in ground water (Lindberg and Runnells, 1984) and may indicate a lack of equilibrium due to slow reaction kinetics among the redox couples in the water. The positive Eh values and dissolved-oxygen content of waters listed in Table 4-7 indicate that most of the waters are oxidizing. The reducing character of the drillhole UE-25b#1 water collected on the fourth day after pumping was initiated probably reflects an initial disturbance of the producing horizon due to completion of the drillhole. The

CONSULTATION DRAFT

Figure 4-12. Relative sodium-potassium-calcium concentration in Yucca Mountain water (Ogard and Kerrisk, 1984). Shaded area represents range of compositions of interstitial and fracture waters at Rainier Mesa. Modified from White et al. (1980). Data shown are in mole percent.

reducing character of the drillhole USW H-3 water has not been fully explained but may reflect a reduction of recharge waters by organics during the last pluvial (Ogard and Kerrisk, 1984).

The anionic constituents of the ground water show a relatively uniform distribution in all the wells, with about 80 percent bicarbonate and the remainder as sulfate and chloride (usually present in nearly equal molar concentrations) and fluoride (in varying concentrations). The carbonate ion is a major complexing agent for the actinide elements, with the hydroxide ion (Section 4.1.3.4) and the fluoride ion being of less importance under oxidizing conditions (Ogard and Kerrisk, 1984).

Spatial variations in ground-water composition are implicit in Tables 4-6 and 4-7 because the ground waters listed are from wells surrounding the repository block. Overall, there is only minor variation in ground-water compositions in and adjacent to Yucca Mountain.

4.1.2.3 Trace elements

The trace elements detected in Yucca Mountain ground water are barium, bromine, lithium, phosphorus, strontium, titanium, and vanadium. The observed ranges in concentration (in mg/L) are less than 0.001 to 0.011 for barium, less than 0.2 to 14.5 for bromine, 0.04 to 0.95 for lithium, less than 1.2 for phosphorus, less than 0.001 to 0.07 for strontium, less than 0.012 to 0.03 for titanium, and less than 0.004 to 0.03 for vanadium (Daniels et al., 1983; Ogard et al., 1983b).

Because lithium was added to the drilling fluid as a tracer to monitor contamination of ground water by the fluid (Daniels et al., 1983), the high end of the range reported here may not represent natural concentrations.

4.1.2.4 Organic content

The in situ organic contents of the waters from drillhole UE-25b#1 and well J-13 were measured by Means et al. (1983). The total organic carbon contents were 0.14 ± 0.05 mg/L in well J-13 water and 0.58 ± 0.06 mg/L in drillhole UE-25b#1 water, respectively. The lower carbon content in well J-13 water is probably representative of the in situ content because the well is a producing well for the Nevada Test Site and all drilling fluids have been removed by extensive pumping. Because of the low concentrations, little can be said about the individual organic compounds other than that organics with high molecular weight (higher than 1,000) constitute 50 percent of the organics present in J-13 water and 33 percent in UE-25b#1 water. The low organic content in the saturated zone suggests that complexing of waste elements americium, neptunium, plutonium, and uranium with naturally occurring organic ligands will be insignificant.

4.1.2.5 Dissolved gas

Measurements of the concentrations of dissolved gases in Yucca Mountain ground waters are very limited. The presence of carbon dioxide in saturated-zone water is indicated by the change in pH (from 7 to 8.5) of ground water that is pumped from deep wells and allowed to stand at atmospheric pressure. A change in dissolved carbon dioxide from $10^{-2.5}$ to the normal $10^{-3.5}$ atmospheres would explain this shift in pH (Daniels et al., 1982). A direct measurement of the carbon dioxide content of the lowest zone of well UE-25p#1 produced a value of 242 mg/L (Bryant and Vaniman, 1984).

The dissolved-oxygen contents of several saturated-zone ground waters are shown in Table 4-7 (Ogard and Kerrisk, 1984). The presence of dissolved oxygen in the water indicates oxidizing conditions.

4.1.2.6 Background radioactivity

Well J-13 is the only well in the vicinity of Yucca Mountain for which data on radionuclide activity are available. The activities of radionuclides present in well J-13 water have been measured (EPA, 1976); results are shown in Table 4-8. Plans for obtaining additional data on the concentration of naturally occurring radionuclides in Yucca Mountain ground water are outlined in Section 8.3.1.2. These data will provide baseline values for future reference in monitoring the ground water at Yucca Mountain.

4.1.2.7 Particulates and colloids

It is necessary to measure the ambient concentration of particulates because particulate material in ground water at a repository site may be significant as a means of transport of radionuclides by sorption on the particulate material. At Yucca Mountain there are no springs or other accessible sources of flowing ground water. Therefore, well J-13, a water well adjacent to Yucca Mountain, was pumped and sampled for particulates by filtering 9,300 L of water through a 0.4-micrometer filter, and 5,300 L of water through 0.05-micrometer Nuclepore filters (Kerrisk, 1987). Material that was present on the filters ranged in diameter from tens of micrometers down to less than 0.01 micrometer, the resolving power of the microscope. A total quantity of 0.25 g of solid material was obtained, corresponding to a sediment concentration of about 2.7×10^{-5} g/L. The composition of individual particles varied considerably. The detectable cations in the >0.4 micrometer fraction were (in weight percent) silicon (60), iron (20), calcium (11), and aluminum (4), while analysis of the smaller size fraction gave somewhat different results: sodium (44), silicon (42), calcium (8), and iron (4) with no detectable aluminum. Because the amount of iron in the Yucca Mountain tuffs and ground waters is low, it is possible that the iron-rich particulates could have been contamination from the steel piping and pumping system. Filter samples also were obtained from two other wells at the NTS, both drawing from an aquifer in alluvial soil. The results were similar to those from well J-13. Because the samples were derived from water flowing in

Table 4-8. Background radioactivity

Number of samples analyzed	Radionuclide	Radioactivity concentration 10^{-9} $\mu\text{Ci/ml}$		
		Maximum	Minimum	Average
5	H-3	8	<7	<8
1	Sr-89	<2	<2	<2
1	Sr-90	<0.9	<0.9	<0.9
1	Ra-226	0.067	0.067	0.067
1	U-234	1.7	1.7	1.7
1	U-235	<0.02	<0.02	<0.02
1	U-238	0.22	0.22	0.22
1	Pu-238	<0.03	<0.03	<0.03
1	Pu-239	<0.04	<0.04	<0.04

^aData extracted from Table 8, EPA (1976) for five water samples collected from Nevada Test Site well J-13.

an artificially created gradient, it can be argued that the particulate content is not that of natural water. However, pumped well water may provide a worst-case situation. Plans for obtaining data on the sorption of radionuclides on colloids and particulates are outlined in Section 8.3.1.3.4.

4.1.2.8 Temperature and pressure

The temperatures in drillholes and wells on and near Yucca Mountain are shown in Figure 4-13. The temperature gradient varies between approximately 20° and 40°C/km at depths of less than or equal to 3 km. The variance suggests that the temperature in the repository horizon should be 25° to 35°C and it also provides a baseline range of temperatures for modeling the perturbation of the temperature gradient produced by the emplacement of the waste. This topic is discussed in Section 1.7 and the discussion here is taken from Sass and Lachenbruch (1982).

Figure 4-13. Plot of temperature versus depth in drillholes deeper than 600 m for Yucca Mountain and nearby areas. Modified from Sass and Lachenbruch (1982).

Because the repository horizon is in the unsaturated zone, the pressure on ground water in this horizon will be close to atmospheric at the appropriate elevation above sea level. Below the water table and along pathways to the accessible environment, the pressure of the ground water will correspond roughly to depth below the water table.

4.1.2.9 Mineralogical controls on water composition

The composition of ground water in the exploratory block and between the repository block and the accessible environment is determined by various factors, including the composition of water infiltrating from the surface, the composition of saturated zone water upgradient from the repository block, and interactions between ground water and the rocks through which it may flow. Ground-water interactions with rock and models describing these interactions are discussed in this section. These interactions between ground water and rock consist primarily of the dissolution of primary silicate phases contained in the volcanic rocks, the precipitation of secondary phases, and sorption and ion-exchange reactions with primary and secondary phases. Primary silicate phases include volcanic glass and various mineral types formed at high temperatures (higher than 500°C). Secondary phases include minerals such as clays, carbonates, and zeolites as well as amorphous solids such as amorphous silica and semiamorphous oxides (e.g., manganese-iron, oxyhydroxides). The models describing ground-water interactions with rock have not considered precipitation and ion-exchange reactions with these minerals.

Because reactions between ground water and silicate phases are generally slow at the temperatures and water-flow rates likely to prevail in rocks between the repository block and the accessible environment, models that assume chemical equilibrium between ground waters and host rocks may not be appropriate. Models of ground-water interactions with rock must incorporate reaction with rock kinetics. Because reaction kinetics cannot be reliably predicted on the basis of theoretical considerations alone, experimental data on the kinetics of appropriate interacting systems (ground water and rock) are required (Chapter 7 and Section 8.3.1.3.3).

White and Claassen (1979; 1980) have presented experimental data and discussed mechanisms for the dissolution of volcanic glass by water. Their glass samples were obtained from the Timber Mountain Tuff exposed on Rainier Mesa at the NTS. White et al. (1980) presented experimental data on the dissolution of glassy and nonglassy tuffs from Rainier Mesa. Using the data, they derived rate constants that were used to extrapolate the experimental results to longer time scales. White et al. (1980) concluded that the composition of interstitial and fracture waters obtained from the Rainier Mesa area could be explained on the basis of a glass-dissolution model. However, the model does not explain all the water compositions obtained by White et al. (1980), such as the highly sodic waters. In addition, the model does not attempt to explain the detailed variations in the pH and the Eh of the ground water. These two parameters are particularly important to radionuclide-transport calculations. Nonetheless, two important aspects of these studies are clear: (1) water composition varies with stratigraphic

CONSULTATION DRAFT

position in the tuffs of Rainier Mesa, suggesting local control, and (2) experimental data suggest that the variations are the result of dissolution of the tuffs through which the water flows.

The overall similarity of the rock types and alteration minerals in Yucca Mountain and Rainier Mesa (Byers et al., 1976) suggests that the controls on ground-water compositions in these two areas might be similar. This is supported by the overlap in the relative cation concentrations of Rainier Mesa and Yucca Mountain ground waters shown in Figure 4-12. Note that the overlap does not extend to ground water from the Paleozoic limestone aquifer (drillhole UE-25p#1) beneath Yucca Mountain (Figure 4-12). The concentrations of other cations, anions, and uncharged species in Yucca Mountain waters also overlap those in Rainier Mesa waters.

Although the work of White and Claassen (1979, 1980) has provided important data on the possible influence of glass dissolution on ground-water compositions in tuffaceous aquifers, further work is required to adequately understand the controls on the composition of Yucca Mountain saturated- and unsaturated-zone waters. This additional work is discussed in Section 8.3.1.3.

4.1.2.10 Reference ground-water composition

Many of the processes taking place in the transport and retardation of waste elements along pathways from the repository to the accessible environment will be investigated in a laboratory and then the results will be used in modeling the entire pathway. To conduct such experiments, a reference ground water or ground waters that will represent ground-water conditions that will be experienced along such a pathway is needed. The reference ground water for the NNWSI Project was chosen on the basis of meeting the following criteria:

1. It should be readily available because many laboratory experiments will use the reference ground water.
2. It should have a composition that is representative of compositions estimated for the interstitial and fracture waters in the Topopah Spring Member tuff in the unsaturated zone (the potential repository horizon).
3. It should have a composition that is representative of ground waters along the potential flow paths away from the repository.
4. It should be a natural ground water rather than synthetic water.

The ground water from well J-13 has been chosen as best meeting these requirements.

Well J-13 is a producing well located in Jackass Flats southeast of the Yucca Mountain site (Figure 4-11) which has been pumped since 1962. Because of the large amount of water produced to date, the well should be clear of

any drilling fluids that may have been used in the original drilling. The main producing zone in well J-13 is in the Topopah Spring Member of the Paintbrush Tuff, which corresponds to the potential repository horizon at Yucca Mountain. This aspect of well J-13 water makes it a useful reference water.

The compositions of interstitial and fracture waters in the unsaturated zone of Yucca Mountain are not yet available, plans for obtaining them are outlined in Section 8.3.1.3. However, White et al. (1980) have determined the composition of interstitial and fracture waters in the tuffaceous aquifers of Rainier Mesa. Because Rainier Mesa and Yucca Mountain are composed of similar rock types, the ground-water compositions in these two areas should reflect similar rock-water interactions. As shown in Figure 4-12, the composition of well J-13 water plots in the field of the Rainier Mesa interstitial and fracture waters reported by White et al. (1980), and may be representative of interstitial and fracture waters in the unsaturated zone of Yucca Mountain. Data for waters from the saturated zone beneath the Yucca Mountain area are also plotted in Figure 4-12. Because the data for well J-13 water plot near the middle of the range of compositions, this water should be adequate for use as a reference water. Other constituents in well J-13 waters (e.g, dissolved silica, bicarbonate, chloride, and sulfate) also show concentrations similar to those of Rainier Mesa interstitial and fracture waters.

4.1.3 GEOCHEMICAL RETARDATION PROCESSES

Under present baseline conditions, the expected flow mechanism in the unsaturated zone is matrix flow through the tuff. This does not exclude the possibility that fracture flow may be important under some potential conditions. In addition, it is possible that future hydrologic research may find that fracture flow under present conditions is also important. Fracture flow is important in the saturated zone where it is the significant flow mechanism. The flow of water through the unsaturated zone is driven by the hydrostatic head and by capillary suction. The expected flow path in the unsaturated zone is downward through the welded and highly fractured Topopah Spring Member tuff and then through the zeolitized bedded tuff of the Calico Hills. The median pore size for tuff reported by Walter (1982) is between 1.2 and 0.1 μm . The unsaturated flow will be drawn to smaller pores by capillary suction. Much of the smallest diameter porosity is intracrystalline pore space in the zeolites and clay minerals. This is particularly true in the nonwelded zeolitic tuff. The intracrystalline porosity of the zeolite clinoptilolite is about 36 percent based on the crystallographic structure. Zeolitic tuff, with 80 to 90 percent clinoptilolite, has a total porosity of 40 to 50 percent, and hence, in zeolitic tuff more than half the porosity is intracrystalline. Furthermore, the channels in the clinoptilolite framework are only 3.6 angstroms in diameter, and it is, therefore, expected that there would be good contact between water containing radionuclides and the strongly adsorbing zeolites and clays in the unsaturated zone.

The transport of contaminants by flow through either a porous matrix or a fracture system in a porous matrix will in each case be affected by various

geochemical and mechanical processes. Among the chemical processes considered in this section are adsorption on mineral surfaces (both internal and external to the crystal structure), including the kinetics of adsorption, and processes leading to precipitation. The mechanical processes considered are dispersive effects (hydrodynamic dispersion, channeling) and diffusion.

In the saturated zone, the flow is believed to be mainly through fractures in welded-tuff units, such as the upper part of the Prow Pass Member. The transport of radionuclides through such fractured units will be affected by sorption on fractured surfaces, but probably not to the same degree as sorption in matrix pore spaces. Since fluid velocities in fracture flow can be much higher than in matrix flow, local chemical equilibrium may not be attained and kinetics becomes an important consideration. Evidence from laboratory experiments shows that sorption reaction rates are fast and equilibrium could be achieved for quite high flow rates (Rundberg, 1985). An additional process that will affect transport through fracture systems is matrix diffusion into the adjacent matrix pore spaces. This process may be limited by channeling, which effectively reduces the surface area through which solutes diffuse from fracture flows to the matrix. Channeling also acts as a dispersion mechanism--that is, it broadens the pulse of solute concentrations.

4.1.3.1 General description of geochemical retardation

The most likely mechanism for the release of radionuclides from a repository is transport in the aqueous or vapor phase from the repository to the accessible environment. This transport can be retarded by three geochemical processes:

1. Sorption--the sorption of chemical species on mineral surfaces such as ion exchange, chemisorption, van der Waals attraction, etc., or ion exchange within the crystal structure.
2. Precipitation--the formation of less soluble compounds by chemical reaction with chemical entities present in the ground water.
3. Exchange between the vapor phase and the aqueous or solid phases.

Sorptive mineral phases, particularly zeolites and clays, along potential flow paths are expected to delay the rate at which radionuclides are released to the accessible environment. Flow paths from the repository to the saturated zone are through bedded tuff at the bottom of the Topopah Spring Member and through the tuffaceous beds of Calico Hills, both of which contain significant quantities of zeolites and clays (Section 4.1.1.3). Other formations that might be part of the flow path are the Prow Pass, the Bullfrog, and the Tram members of the Crater Flat Tuff; zones containing zeolites and clays in variable abundances are present in these units (Section 4.1.1.3).

Precipitation processes (i.e., solubility) will limit the concentration of radionuclides in water that moves from the repository toward the accessible environment. Solubility depends on the composition of the solution;

that is, formation of complexes can retain species in solution in greater amounts than would otherwise exist. Yucca Mountain ground water is primarily a sodium-bicarbonate water (Section 4.1.2.1); carbonate, bicarbonate, and hydroxyl are the most prevalent anions that form complexes with radionuclide cations. However, for some radionuclides, such as cesium or strontium, the solubility limits may not limit radionuclide transport. Quantitative data for the hydroxyl and carbonate complexes of most waste elements are insufficient and need to be determined. Plans for obtaining this information are outlined in Section 8.3.1.3:

Vapor-phase transport is generally in the opposite direction from aqueous transport and may proceed more quickly since the pores that are not filled with water are the larger pores and have a higher permeability than the water-filled pores. In the unsaturated zone, volatile elements, such as iodine, can be transported in the vapor phase. The vapor-transport process is impeded, however, by exchange between the slower moving aqueous phase and the solubility in the aqueous phase, which lowers the partial pressure of the volatile species.

There are physical processes such as matrix diffusion that are driven by advective forces in which radionuclides diffuse into water in the pores and may retard the transport of all species, including nonsorbing or low-sorbing species (Section 4.1.3.5). In general matrix diffusion can be regarded as the movement of dissolved species from water in connected pore spaces to water in "dead-end" pore spaces as a result of gradients in species concentrations. In particular, the connected pore spaces may be fracture networks in which flow is rapid, and the dead-end pore spaces may be the matrix pores in which flow is relatively slow. Dissolved species may also be transferred from matrix flow to fracture flow (and vice versa) by advective forces arising from pressure differences; however, advective transfer is not considered in this chapter.

The modeling of radionuclide transport requires that the reactions of radionuclides that occur between the solid phase (minerals, or precipitates) and aqueous phase (fracture or pore water) be coupled with the hydrologic flow and physical processes like diffusion and dispersion. Much of the basis for transport modeling has origins in chemical engineering and specifically the modeling of chromatographic columns. Chromatographic columns, however, are constructed so that dispersion is minimized. Moreover, the column is homogeneous and hence channeling is minimized. Homogeneity also prevents any effects that would be analogous to matrix diffusion and fracture flow. The columns are run at flow rates that do not exceed mass transfer kinetics and under chemical conditions (i.e., controlled pH, complexing agents, etc.) which ensure well-behaved adsorption. Thus, for column chromatography the coupling between chemistry and flow is simple and can be successfully modeled. The transport of tracers in geohydrologic systems, however, is not simple and simplifying assumptions must be tested by laboratory and field studies (Sections 8.3.1.3.6 and 8.3.1.3.4). In addition, transport by colloids must be examined because colloids may not interact with the stationary solid phase.

CONSULTATION DRAFT

4.1.3.1.1 Key radionuclides

The importance of the various radionuclides present in the waste can vary depending on time scales of interest, site conditions, transport mechanisms, health physics concerns, or waste types. A list of important waste elements has been developed for the Yucca Mountain site (Kerrisk, 1985) from the following criteria:

1. The quantity of various radionuclides present in the waste.
2. The release limits established by the EPA for the radionuclides (40 CFR Part 191).
3. Preliminary estimates of the solubility and sorption of waste elements.
4. Special problems that may arise for certain waste elements.

The following types of waste were considered: spent fuel and high-level waste (HLW) from pressurized water reactors (PWR) or boiling water reactors (BWR), and defense high-level waste (DHLW). Data for spent fuel and HLW were obtained from a compilation by Croff and Alexander (1980). The radionuclides include fission products, actinides and their decay products, and activation products present in the fuel, cladding, and subassembly hardware. Data for PWR waste (spent fuel and HLW) were used as characteristic of both PWR and BWR waste because the radionuclide inventories are similar (Kerrisk, 1985). Data for DHLW were obtained from calculations by Fowler and Boersma of the Waste Solidification Technology Division of the Savannah River Laboratory (Boersma, 1984). These calculations give radionuclide inventories that are characteristic of waste from the future Defense Waste Processing Facility and include radionuclides present in the waste glass only. The results compiled by Croff and Alexander (1980) are in curies per metric tons heavy metal (Ci/MTHM), where MTHM is metric ton of heavy metal (uranium and plutonium) originally charged to the reactor. The results can be directly compared with the EPA standards, where the release limits are given in terms of curies per 1,000 MTHM. The results for DHLW are in curies per container. For this assessment, the equivalent loading of DHLW containers was estimated as 0.5 MTHM per container based on a comparison of the total activity of PWR HLW and DHLW canisters.

A general criterion for the importance of the various radionuclides in waste in terms of the EPA standards can be obtained from the ratio of the quantity of the radionuclide in the waste to the EPA release limit for that radionuclide, where both are expressed in curies per 1,000 MTHM. Tables 4-9 through 4-11 list radionuclides ordered by this ratio along with the ratios for PWR spent fuel, PWR HLW, and DHLW, respectively (Kerrisk, 1985). Radionuclides near the top of the lists in the tables have larger values of this ratio. Thus, larger fractions of these radionuclides must be contained by the repository to meet the EPA standard. At 100 yr after discharge from the reactor, isotopes of cesium, strontium, americium, and plutonium head the lists. In the 1,000- to 10,000-yr range, isotopes of the actinides (americium, plutonium, neptunium, and uranium) head the lists. By 100,000 yr after discharge, the lists are headed by actinides as well as radionuclides in the actinide decay chains, but a few other radionuclides, such as isotopes of

Table 4-9. Radionuclides ordered by ratio of inventory to U.S. Environmental Protection Agency limit for PWR spent fuel for various decay times^a

1 x 10 ² yr		1 x 10 ³ yr		1 x 10 ⁴ yr		1 x 10 ⁵ yr	
Am-241	3.8 x 10 ⁴	Am-241	9.0 x 10 ³	Pu-239	2.4 x 10 ³	Pu-239	1.8 x 10 ²
Pu-238	1.1 x 10 ⁴	Pu-240	4.8 x 10 ³	Pu-240	1.8 x 10 ³	Th-230	1.0 x 10 ²
Cs-137	1.0 x 10 ⁴	Pu-239	3.1 x 10 ³	Am-243	6.7 x 10 ¹	U-234	1.6 x 10 ¹
^b Ba-137m	9.8 x 10 ³	Am-243	1.6 x 10 ²	U-234	1.9 x 10 ¹	Pu-242	1.5 x 10 ¹
Sr-90	6.8 x 10 ³	U-234	2.0 x 10 ¹	Pu-242	1.7 x 10 ¹	Np-237	1.1 x 10 ¹
^c Y-90	6.8 x 10 ³	Pu-242	1.8 x 10 ¹	Th-230	1.7 x 10 ¹	^d Ra-226	1.0 x 10 ¹
Pu-240	5.3 x 10 ³	^e Np-239	1.6 x 10 ¹	Np-237	1.2 x 10 ¹	U-233	4.1
Pu-239	3.1 x 10 ³	C-14	1.4 x 10 ¹	^e Np-239	6.7	U-236	4.0
Pu-241	1.0 x 10 ³	Np-237	1.0 x 10 ¹	Ni-59	4.8	Th-229	3.7
Cm-244	3.3 x 10 ²	Pu-238	9.7	C-14	4.6	U-238	3.2
Ni-63	3.1 x 10 ²	Ni-59	5.2	U-236	3.5	Ni-59	2.2
Am-243	1.7 x 10 ²	U-238	3.2	U-238	3.2	Zr-93	1.9

^aSource: Kerrisk (1985), PWR = pressurized water reactor.

^bShort-lived daughter of Cs-137.

^cShort-lived daughter of Sr-90.

^dDecay products of Ra-226 are in secular equilibrium. Each alpha emitter in the decay chain has the same ratio of inventory to EPA limit; others have a ratio that is 10 times lower. The Ra-226 decay products are generally short-lived.

^eShort-lived daughter of Am-243.

CONSULTATION DRAFT

Table 4-10. Radionuclides ordered by ratio of inventory to the U.S. Environmental Protection Agency limit for high-level waste from pressurized water reactors for various decay times

1 x 10 ² yr	1 x 10 ³ yr	1 x 10 ⁴ yr	1 x 10 ⁵ yr
Cs-137 1.0 x 10 ⁴	Am-241 1.8 x 10 ³	Am-243 1.6 x 10 ²	Pu-239 5.1
^b Ba-137m 9.8 x 10 ³	Am-243 1.7 x 10 ²	Pu-239 4.0 x 10 ¹	Np-237 3.4
Sr-90 6.8 x 10 ³	Pu-240 6.3 x 10 ¹	Pu-240 2.4 x 10 ¹	Th-230 2.4
^c Y-90 6.8 x 10 ³	Pu-239 2.1 x 10 ¹	^d Np-239 6.7	Ni-59 2.2
Am-241 2.0 x 10 ³	^d Np-239 1.6 x 10 ¹	Ni-59 4.8	Zr-93 1.7
Pu-238 5.2 x 10 ²	C-14 1.4 x 10 ¹	C-14 4.6	^e Nb-93m 1.6
Cm-244 3.3 x 10 ²	Ni-59 5.1	Np-237 3.5	U-233 1.2
Ni-63 3.1 x 10 ²	Np-237 3.5	Zr-93 1.8	^f Th-229 1.1
Am-243 1.7 x 10 ²	Zr-93 1.8	^e Nb-93m 1.7	Tc-99 9.4 x 10 ⁻¹
Sm-151 1.7 x 10 ²	^e Nb-93m 1.7	Tc-99 1.3	Sn-126 3.9 x 10 ⁻¹
Pu-240 6.8 x 10 ¹	Tc-99 1.3	Nb-94 9.1 x 10 ⁻¹	U-234 3.6 x 10 ⁻¹
Cm-242 2.0 x 10 ¹	Nb-94 1.2	Sn-126 7.3 x 10 ⁻¹	^g Pa-233 3.4 x 10 ⁻¹

^aSource: Kerrisk (1985).

^bShort-lived daughter of Cs-137.

^cShort-lived daughter of Sr-90.

^dShort-lived daughter of Am-243.

^eShort-lived daughter of Zr-93.

^fDecay products of Th-229 are in secular equilibrium. Each alpha emitter in the decay chain has the same ratio of inventory to EPA limit; others have a ratio that is 10 times lower. The Th-229 decay products are generally short lived.

^gShort-lived daughter of Np-237.

Table 4-11. Radionuclides ordered by ratio of inventory to the U.S. Environmental Protection Agency limit for defense high-level waste for various decay times

1×10^2 yr		1×10^3 yr		1×10^4 yr		1×10^5 yr	
Pu-238	8.7×10^3	Am-241	1.9×10^2	Pu-239	1.3×10^2	Th-230	7.5×10^1
Sr-90	7.8×10^3	Pu-239	1.7×10^2	Pu-240	3.8×10^1	U-234	1.1×10^1
^b Y-90	7.8×10^3	Pu-240	9.9×10^1	U-234	1.4×10^1	Pu-239	9.7
Cs-137	7.4×10^3	U-234	1.5×10^1	Th-230	1.3×10^1	^c Ra-226	7.4
^d Ba-137m	7.0×10^3	Pu-238	7.1	Ni-59	3.3	Zr-93	1.9
Am-241	7.8×10^2	Ni-59	3.5	Zr-93	2.0	^e Nb-93m	1.9
Ni-63	4.3×10^2	Zr-93	2.0	^e Nb-93m	2.0^c	Ni-59	1.4
Sm-151	2.1×10^2	^e Nb-93m	2.0	^c Ra-226	9.7×10^{-1}	U-236	6.3×10^{-1}
Pu-239	1.7×10^2	Th-230	1.2	U-236	6.2×10^{-1}	Np-237	2.6×10^{-1}
Pu-240	1.1×10^2	U-236	6.0×10^{-1}	Tc-99	3.3×10^{-1}	Tc-99	2.4×10^{-1}
Pu-241	2.4×10^1	Tc-99	3.4×10^{-1}	Np-237	2.7×10^{-1}	U-238	1.5×10^{-1}

^aSource: Kerrisk (1985).

^bShort-lived daughter of Sr-90.

^cDecay products of Ra-226 are in secular equilibrium. Each alpha emitter in the decay chain has the same ratio of inventory to EPA limit; others have a ratio that is 10 times lower. The Ra-226 decay products are generally short lived.

^dShort-lived daughter of Cs-137.

^eShort-lived daughter of Zr-93.

nickel (present in subassembly hardware) and zirconium (present as a fission product and cladding activation product), are near the top.

Two of the most important retardation mechanisms available in a geologic repository are precipitation and sorption. Radionuclides that are present in significant quantities in waste and have high solubility and low sorption must be considered important because they could pose problems in meeting release limits. A review of the waste inventories, the sorption data presented in Section 4.1.3.3 and the solubility data presented in Section 4.1.3.4 indicates that three waste elements (carbon, technetium, and iodine)

CONSULTATION DRAFT

may have high solubility and low sorption at the Yucca Mountain site. The radionuclides of these waste elements are not present in large quantities relative to the EPA limits, but they deserve consideration because they may not be strongly retarded by sorption or precipitation. Waste elements with moderately high solubilities and intermediate sorption coefficients are neptunium and uranium. The estimated solubility of nickel is relatively high; sorption data for nickel have not been measured as yet, but it probably should be considered in this category as well (Sections 4.1.3.4.2.3 and 8.3.1.3.4).

Radionuclides that can be released as gases must be given special consideration at Yucca Mountain. Most gaseous radionuclides are short-lived or are present in such small quantities that they do not represent problems. However, two radionuclides that may be transported in the gas phase and will be present for long periods of time are carbon-14 and iodine-129. The likely gaseous forms of both these radionuclides (carbon dioxide and iodine gas) are soluble in water. Carbon-14 as carbon dioxide (CO_2) may also undergo isotopic exchange with natural carbon in the surrounding rocks. Thus, for both waste elements, an exchange between the gas, aqueous, and solid phase may be possible.

On the basis of the quantity of radionuclides in the waste, the EPA release limits, solubility, sorption, the potential for vapor transport, and the potential for colloid transport, the following radionuclides should be considered important (Kerrisk, 1985):

1. Isotopes of the elements americium, plutonium, neptunium, and uranium are the most important because they are present in the largest quantities in waste relative to their EPA release limits and because of their potential for colloid transport. Solubility and sorption measurements for these elements should be of utmost importance.
2. Isotopes of the elements carbon, nickel, zirconium, technetium, thorium, radium (barium used as an analog), and tin are present in lesser amounts at most times, but are also important for solubility and for their transport potential.
3. Isotopes of the elements carbon, technetium, and iodine are important because these elements appear to have high solubility and low sorption at Yucca Mountain. This should be confirmed experimentally by determining their transport potential.
4. Isotopes of the elements carbon and iodine are important because they could be transported in the gas phase at Yucca Mountain. Studies of gas-phase transport should concentrate on these elements. Detailed discussions of the various retardation processes and the radionuclides considered for these processes are contained in Sections 4.1.3.3 through 4.1.3.8.

4.1.3.2 Analytical techniques

This section discusses the methods and procedures used to investigate geochemical retardation in the experiments reported here. (Results are reported in Sections 4.1.3.3. through 4.1.3.5.) The discussion emphasizes the retardation mechanism of sorption, but it also provides basic dynamic-transport information and includes the coupling of adsorption to dynamic processes. The procedures, and estimates of the attendant uncertainties, are discussed in detail by Daniels et al. (1982). The analytical techniques are discussed in Sections 4.1.3.2.1 through 4.1.3.2.5, but a general overview of the techniques and their interrelationships is given in this introduction.

The simplest mathematical approach to adsorption is to use the distribution coefficient, K_d , which is the approach used routinely by analytical chemists to model separations on chromatographic columns. The measurement of K_d allows the characterization of tuffs with different mineralogies. In a manner similar to chromatographic columns, the samples are crushed to a predetermined size fraction (typically between 75 and 500 micrometers) and carefully mixed to ensure homogeneity (on a small scale) of the column. The crushed sample will allow rapid equilibration. The heterogeneity of the tuff strata at Yucca Mountain requires a significant number of measurements to characterize the sorptive properties. The limitations of the distribution-coefficient approach to geologic investigations are as follows:

1. The assumption of a linear sorption isotherm. The terms "sorption isotherm," "Freundlich isotherm," or "Langmuir isotherm" are generally used to define the relationships between sorption and the concentration of the element being sorbed at a constant temperature.
2. The distribution between the solid phase and the aqueous phase may include precipitation or irreversible reaction or both.
3. The aqueous-phase species are not well known for many of the important radionuclides.
4. The assumption of equilibrium (i.e., rapid kinetics).

These limitations do not necessarily apply to all of the radionuclides of importance. For example, the solution chemistry of the alkali metals and the alkaline earths is well known, and the aqueous-phase species can be predicted with certainty. Average values are calculated for measured sorption ratios since measurements are made in duplicate. The uncertainties associated with averages of sorption data are the standard deviations of the means. When measured sorption values are very high or very low, the range of individual measurements may be quite large, resulting in a large standard deviation of the mean. This large range arises as a result of counting statistics. A very high or very low K_d means one phase, either the solution or the solid, has very little activity and, therefore, very few counts giving rise to counting errors larger than those obtained with sorption ratios close to unity.

Often, the surface area of the geologic media is thought to be important to total sorption. As a result, a surface-area sorption coefficient (K_a) is

used. However, for tuff, surface area does not appear to correlate with cation exchange capacity or particle size (Wolfsberg et al., 1979; Daniels et al., 1982). Furthermore, the data presented by Daniels et al. (1982) in their Appendix A show that particles in the size range of 75 to 500 micrometers have no effect on measured sorption ratios, yet the surface area is changing, therefore, K_d is not used for this size range. Other supporting evidence for the lack of correlation of surface area with cation exchange capacity can be found in Meyer et al. (1986). Analyses with neptunium (V) on crushed tuff to test effects of particle size resulted in only a slight increase in sorption. If the phenomena were controlled by surface area, a much larger effect of particle size would have been observed.

Three different sorption procedures were used: batch, crushed tuff column and circulating columns. These three sorption procedures are discussed in Sections 4.1.3.2.2, 4.1.3.2.3, and 4.1.3.2.4. Batch sorption methods approach equilibrium distribution coefficients for long-time experiments and yield information on kinetics for short-term experiments and on reversibility by desorption experiments. They are easy to conduct, but "equilibrium" sorption values may not be representative of a flowing system. The batch technique, in spite of its simplicity, is not without experimental pitfalls. To ensure that discrepancies between laboratory experiments are not due to experimental errors, the following must be carefully controlled:

1. Phase separation. When distribution coefficients (K_d) are high, the experiments are susceptible to contamination of the aqueous phase with colloidal material containing radionuclides leading to erroneously low K_d values.
2. Water composition. Comparison between experiments using radically different water compositions leads to large discrepancies.
3. Crushing and sieving. The rock must not be crushed below mineral crystal size. The samples should be sieved to ensure uniformity.
4. Mineral fractionation. This can occur either in the process of sieving or in the course of an experiment, such as a crushed tuff column, because the adsorption minerals (clays and zeolites) in tuff generally have small crystal sizes that will enrich in the smaller size fractions.
5. Speciation. The results of experiments where the complexing agents or oxidation-reduction conditions are not identical will show discrepancies for some radionuclides.

The above parameters must be controlled to study the distribution of geochemical properties in the geologic system. Where these parameters cannot be made to coincide exactly with the conditions in the field, compromises are made in a conservative direction for the purpose of applying the result to predictions in the field. For example, the smallest-size fractions are discarded, oxidizing conditions are used, etc. To systematically study which simplifying assumption may be valid and to experimentally examine the coupling of adsorption to dynamic processes the following types of experiments have been performed:

1. Crushed-tuff columns. The experiments allow the comparison of batch to transport through nearly identical material; they allow the study of kinetic effects, speciation, and other physical and chemical effects, such as anion exclusion. The advantage in using crushed tuff is that the results are free of ambiguities in the interpretation which would occur if the effect of heterogeneity and dispersion was not minimized.
2. Recirculating columns. The experiments were performed to test whether self-grinding, which might produce extremely small particles when stirred, had occurred in batch measurements. The technique is otherwise subject to nearly the same pitfalls as batch measurements.
3. Solid-tuff columns. The experiments have been and will continue to be performed to examine the validity of batch sorption measurements on tuff where homogeneity and dispersion are not controlled. The hydrologic flow is more constrained than that in the field. The filtration of colloids will also be studied to examine the potential for particulate migration.
4. Fractured-tuff columns. The experiments represent the most complex transport system that can be simulated in the laboratory.
5. Diffusion in solid tuff. The experiments are ancillary to the fractured-tuff columns but provide fundamental data for modeling diffusion in the tuff matrix. The experiments will also provide some kinetic data.
6. Solubility. The experiments are primarily directed toward understanding actinide solution chemistry in near-neutral solution. The understanding of actinide chemistry under near-neutral conditions is growing at a very slow pace because the low solubility of these elements does not allow the direct determination of oxidation state and molecular structure at present. Without a firm knowledge of the aqueous chemical species, the interpretation of the above transport experiments is hampered.

The major isotopes chosen for study come from the EPA critical element list. Additional radionuclides have been investigated when warranted (Section 4.1.3.3.1).

4.1.3.2.1 Preparation of ground-water samples for use in laboratory experiments

Before use, reference ground water from well J-13 is pretreated by putting it in contact with samples of tuff from individual strata in the vicinity of Yucca Mountain to obtain water chemistries representative of steady-state equilibrium for each tuff sample. After contact for at least 2 wk, the water is filtered through 0.05 micrometer Nuclepore membranes. Then, the water is analyzed and used in the sorption experiments. In general, after pretreatment sodium increases up to 50 percent and magnesium decreases by 50 percent probably due to ion exchange; chloride, fluoride, and

sulfate remain unchanged; and the pH (8.1 to 8.4) increases slightly. The changes, however, do not greatly affect sorption.

The spiked-feed solutions for each tuff sample were prepared by adding measured concentrations of radionuclide tracers to the pretreated water. Spiked-feed solutions are prepared as described by Daniels et al. (1982) for separate radionuclides. The final concentrations of the feed solutions generally range from 10^{-6} to 10^{-9} M; plutonium concentrations as low as 10^{-11} M have been studied.

4.1.3.2.2 Batch sorption methods

In batch measurements of sorptive properties, the distribution of a radionuclide is measured between ground water and crushed tuff as a function of contact time, the concentration of sorbing species, particle size, temperature, oxidation-reduction conditions, ground-water components, and lithology. In the batch procedure, a 1-g sample of crushed tuff is pretreated by contact with 20 ml of well J-13 ground water for at least 2 wk. The ground water used to prepare the solution containing radionuclides (spiked-feed solution) is also pretreated by contact for at least 2 wk with a crushed-tuff sample from the same section of core. The solids are separated from the liquid by centrifuging after the pretreatment period.

A sorption measurement starts when 20 ml of spiked-feed solution is added to the 1-g sample of tuff treated with ground water. The concentration of the spiked element ranges from 10^{-6} to 10^{-13} M, except in isotherm measurements where concentrations approaching the solubility limit are investigated. Care is taken to ensure that the spike concentration is well below predicted solubility limits in standard experiments. Except for the highest concentrations used in the isotherm studies (generally greater than 10^{-4} M), no significant change in sorption ratio has been observed with changing spike concentrations covering six to seven orders of magnitude (Wolfsberg et al., 1981; Daniels et al., 1982; Rundberg et al., 1985). Next, the two phases are thoroughly mixed and the sample is placed in a shaker to be agitated at about 200 rpm for selected times. At the end of the sorption period, the sample is removed from the shaker and the solids are separated by centrifuge. Tracer (spike) activity in the separated solid and liquid phases then is determined and the pH of the liquid is measured. The analytical techniques for each element are discussed by Daniels et al. (1982), who also discussed the technique used to completely separate the solids from the aqueous phase and to calculate the sorption ratios. The standard sorption and desorption times were 6 wk each. The data in Appendix A of Daniels et al. (1982) show that sorption either increases with time or remains unchanged. The choice of 6 weeks (or a total of 12 wk for both sorption and desorption measurements) was a compromise between reaching equilibrium, if that were even possible, and completing the measurements within a reasonable amount of time.

After the radioactivity of a solid phase has been measured, a desorption measurement is started. Tuff-pretreated ground water (20 ml) without tracer is added to the remaining solid phase. The sample is agitated again to ensure a thorough mixing of the two phases. The rock sample is separated

from the pretreated water and analyzed for radioactivity in the same manner as that used for the sorption sample.

The results are reported as sorption ratio, R_d , for each radionuclide as a function of the lithology of the crushed-tuff sample and various experimental conditions, where

$$R_d = \frac{\text{activity of radionuclide on solid phase per unit mass of solid}}{\text{activity of radionuclide in solution per unit volume of solutions}} \quad (4-1)$$

The sorption ratio, R_d , is equivalent to the distribution coefficient, K_d , only under equilibrium conditions. Sorption ratios determined by the batch method measure the distribution of an element between the solid and the liquid phases under specific conditions and do not necessarily represent equilibrium values. Uncertainties in the average sorption ratios reported for each set of test conditions are given as the standard deviation of the mean.

4.1.3.2.3 Crushed-tuff column method

The transport behavior of nuclides in columns of crushed tuff has been studied to characterize retardation in a flowing system. The column method is sensitive to speciation and also to reaction kinetics. The method will enable improvements to the transport model or empirical adjustments to the retardation data base so that the calculated bounds will encompass any reduced retardation from speciation or kinetics. Column studies are conducted to determine the retardation factor, R_f , rather than the sorption ratio, R_d . The velocity of a radionuclide is measured directly and then compared with the ground-water velocity as measured with nonreacting tracers like tritiated water or iodine-131. The retardation factor is related to the sorption ratio by the expression

$$R_f = 1 + \frac{R_d (\text{density}) (1 - \text{porosity})}{\text{porosity}} \quad (4-2)$$

where density equals the rock-column density (2.5 g/cm^3).

A thorough discussion of the crushed-tuff column methodology is given by Treher and Raybold (1982).

Crushed-tuff columns are limited by the time required to observe the elution of adsorbing species. For example, cesium and strontium elutions require approximately 1 yr at flow rates of 11 to 60 m/yr--a long time to keep a mechanical system operating without failure. Crushed tuff has the potential to create or expose surfaces that would not be exposed in situ. In addition, there can be a coupling between flow and mineralogy such that adsorbing minerals are either avoided or preferentially contacted in intact tuff.

4.1.3.2.4 Circulating system method

The circulating system for determining sorption ratios incorporates features of the batch and column methods. In the circulating system the solid phase remains stationary and thus is not subject to possible self-grinding during agitation. Also, the solution-to-rock ratio in the column itself is considerably lower than the 20 ml/g ratio used in batch experiments and is, therefore, closer to that in the true geologic setting; but the ratio in the entire system, including the reservoir, is the same as in the batch system. A description of the circulating-system technique is given by Wolfsberg et al. (1981) and Daniels et al. (1982).

4.1.3.2.5 Solubility

Solubility measurements that provide upper limits on the concentrations of waste elements were made by preparing oversaturated solutions of radionuclides and monitoring radionuclide concentrations as a function of time (Nitsche and Edelstein, 1985). During the equilibration time, the temperature and the pH of the solutions were controlled. For measurements in well J-13 water, the pressure of carbon dioxide over the water was also controlled. Solids that precipitated were identified by x-ray diffraction; some of the solids were amorphous and could not be identified in this manner (Section 4.1.3.4.2; Nitsche and Edelstein, 1985). At this time there is a lack of firm knowledge of aqueous chemical species (actinides) under near-neutral conditions. Further work is planned in Section 8.3.1.3.

4.1.3.3 Sorption

Ion exchange and sorption (chemical or physical) are surface processes by which ions or other solution species can be removed from ground water by rock. The measured sorption ratios (defined in Section 4.1.3.2.2) are primary data for the assessment of retardation along the flow path. An understanding of these processes may permit extrapolation of laboratory data to field experiments and will aid in estimates of uncertainty limits. Plans for obtaining the additional information required are outlined in Section 8.3.1.3.4. The sorption of radionuclides on colloidal material will be studied as discussed in Section 8.3.1.3.4.

Several parameters directly affect sorption measurements as they relate to Yucca Mountain. Among these are the composition of the ground water used in the sorption experiments, the cation-exchange capacity of the minerals, the pH, the temperature, the oxidation-reduction conditions, the waste element being sorbed, and the concentration of the waste element. Additionally, three primary experimental methods (batch, crushed-rock column, or circulating systems) are used for measuring sorption coefficients as discussed in Sections 4.1.3.2.2 through 4.1.3.2.4. Relevant data that characterize the ion-exchange and sorption capacities of the tuff at Yucca Mountain are taken primarily from Daniels et al. (1982).

4.1.3.3.1 Sorption data for tuff

Daniels et al. (1982, 1983); Ogard et al. (1983b), Wolfsberg et al. (1983), Wolfsberg and Vaniman (1984), Crowe and Vaniman (1985), Ogard and Vaniman (1985), Rundberg et al. (1985) present data concerning sorption on tuff and list the results of all previous experiments. These sorption data were obtained from samples representative of Yucca Mountain strata from well J-13 (JA samples) and drillholes UE-25a#1 (YM samples), USW G-1 (G1 samples), USW G-2, and USW GU-3. Variables include time, temperature, particle size, oxidation-reduction conditions, element concentration, ground-water composition, and test method (batch, column, or circulating system). The radionuclides investigated were americium, cesium, neptunium, plutonium, uranium, thorium, strontium, technetium, tin, barium, radium, cerium, europium, and selenium. Strontium and cesium, while being shorter lived than many potentially hazardous waste elements, were studied first because of their relatively simple sorption behavior. These elements continue to be used to help validate the experimental techniques because their behavior can be predicted fairly well. Barium is included as a chemical analog for radium to reduce the number of experiments with radium. A great deal of care must be taken in handling radium-226 samples to prevent the release of daughter element radon-222 which could contaminate low-level counting facilities. Uranium and thorium are long-lived parents of radium; cerium and europium have been studied as analogs of rare-earth radionuclides. Selenium was studied as a potentially important waste element.

4.1.3.3.2 Sorption data from batch experiments

Average sorption ratios (R_d) from batch sorption and desorption experiments on tuff samples for strontium, cesium, barium, radium, cerium, europium, americium, plutonium, uranium, selenium, technetium, and neptunium are given in Tables 4-12a, 4-12b, 4-13a, and 4-13b. Each sample is assigned a depth equivalent in drillhole USW G-1 on the basis of the relative position of the sample in the major stratigraphic units so that sorption results on samples from one drillhole can be compared with samples from another. Data are for contact times greater than one week and particle sizes greater than 75 micrometers in an air environment at 25°C unless otherwise noted. Mineral abundance data determined by x-ray diffraction and petrologic characterization of these samples (as well as for finer fractions) are given in Table 4-14.

The R_d values from sorption experiments are often significantly lower than those from desorption experiments. Although the differences in R_d are unexplained at present, they suggest that, at least for some elements, sorption is not readily reversible. The sorption of cesium, strontium, and barium, which are noncomplexed cations that probably sorb by ion exchange, appears reversible because the R_d values from sorption and desorption experiments generally agree within a factor of 2. The R_d values from desorption experiments for technetium, the lanthanides, and the actinides are as much as 10 times greater than R_d values from sorption experiments. The latter elements all have multivalent species, some of which are sparingly soluble, hydrated, or form colloids and complexes (Section 4.1.3.4.1), which can alter the sorption processes. Differences in sorptive behavior may be caused by

Table 4-12a. Average sorption ratios from batch sorption experiments on crushed tuff for strontium, cesium, barium, radium, cerium, and europium^{a, b} (page 1 of 2)
(See footnotes at the end of Table 4-12b)

Strati- graphic unit ^c	Sample	Depth (ft)	Sorption ratios (ml/g)					Europium
			Strontium	Cesium	Barium	Radium	Cerium	
Tpc	JA-8	606	270(5)	2,700(400) ^d	45(15)			2,100(300)
	YM-5	251	280(80)	5,800(800) ^d	1,100(200)		450,000(240,000) ^d	2,300,000(40,000) ^d
Tpp	G2-547	547	265(10) ^f	13,300(1,500) ^f	3,490(30) ^f			340(30) ^f
	G2-723	723	290(40) ^f	4,100(600) ^f	3,500(400) ^f			>10,000 ^f
	GU3-433	433	45(9) ^g	630(20) ^g	810(100) ^g			100(14) ^g
	GU3-855	855						
Tpt	YM-22	848	53(4)	290(30)	900(30)		1,270(40)	1,390(110)
	GU3-1203	1,203	42(1) ^f	350(30) ^f	640(40)			190(2)
	G1-1292	1,292	200(6) ^f	430(28) ^f	2,100(300)	1,500(100)	68(8) ^f	140(14) ^d
	GU3-1301	1,301	28(4) ^g	160(40) ^g	570(60) ^g			45(12) ^g
	YM-30	1,264	260(80)	855(5)	3,400(1,500)		230,000(100,000) ^e	160,000(50,000)
	JA-18	1,420	17,000(3,000)	16,000(1,000)	38,000(18,000)		2,800(1,400) ^e	1,400(200) ^e
Th	G1-1436	1,436	36,000(3,000)	7,800(500)	150,000(24,000) ^f		59,000(7,000)	30,000(2,000)
	G2-1952	1,952	2,200(400) ^f	63,300(1,100)	25,000(4,000) ^f			89(14) ^f
	GU3-1436	1,436						
Bt	GU3-1531	1,531						
	YM-38	1,504	17,000(2,000)	13,000(2,000)	100,000(10,000)		760(140)	1,600(200)
	YM-42	1,824	3,900(600)	17,000(1,000)	94,000(14,000)		49,000(7,000)	52,000(4,000) ^h
Tcp	G1-1854	1,854	60,000(14,000)	13,000(2,000)	45,000(7,000)			15,000
	YM-45	1,930	195(14)	520(90)	1,200(100)		730(100)	1,600(200)
	G1-1883	1,883	22(0.2)	187(3)	183(12)		1,420(20)	
	YM-46	2,002	190(60)	840(6)	14,000(6,000)		310,000(110,000) ^f	307,000(110,000)
	G1-1982	1,982	55(4)	1,120(110)	700(50)		560(40)	970(150)
	YM-48	2,114	2,100(400)	9,000(4,000)	18,000(6,000)		1,400(500)	2,200(500)
	YM-49	2,221	3,200(300)	36,000(3,000)	42,000(8,000)		550(100)	1,200(100)
	JA-26	1,995	95(35)	1,500(600)	800(300)			

Table 4-12a. Average sorption ratios from batch sorption experiments on crushed tuff for strontium, cesium, barium, radium, cerium, and europium^{a, b} (page 2 of 2)
(See footnotes at the end of Table 4-12b)

Strati- graphic unit ^c	Sample	Depth (ft)	Sorption ratios (ml/g)					Cerium	Europium
			Strontium	Cesium	Barium	Radium			
Tcb	JA-28	2,001	94(20)	1,640(210)	820(50)				2,100(1,000)
	G1-2233	2,233	48,000(3,000) ^f	13,500(800)	250,000(30,000)			1,400(300)	900(200)
	G1-2289	2,289	7,300(500)	37,000(13,000)	66,000(9,000)	46,000(20,000)			797(10)
	YM-54	2,491	62(12)	180(40)	400(150)			150(40)	470(40)
	G1-2333	2,333	180(20)	1,400(130)	1,500(200)				2,300(400)
	G1-2363	2,363	64(3)	470(40)	235(9)	540(60)			730(50)
	G1-2410	2,410	169(1)	1,250(50)	1,780				440(80)
	JA-32	2,533	57(3)	123(4)	380(30)			82(14)	90(20)
	G1-2476	2,476	41(1)	700(40)	385(11)				3,200(100)
Tct	G1-2698	2,698	42,000(3,000) ^f	7,700(400) ^f	63,000(5,000) ^f			240(30) ^f	200(30) ^f
	G1-2840	2,840	860(1)	2,200(200)	2,070(70)				4,900(400)
	G1-2854	2,854	94(1) ^f	1,080(120) ^f	1,000(50)				1,300(200)
	G1-2901	2,901	68(1) ^f	1,290(110) ^f	1,600(200) ^f			42,000(3,000) ^f	160,000(50,000) ^f
	G1-3116	3,116	2,400(17) ^f	6,600(500) ^f	12,000(4,000) ^f			100(10) ^f	760(60) ^f
	JA-37	3,497	287(14)	610(40)	760(150)				6,000(800)
T1	G1-3658	3,658	13,000(0)	4,950(50)	13,500(500)			1,000(200) ^f	530(40)
Tba	G2-3933	3,933	240(60) ^f	2,500(1,000) ^f	1,700(500) ^f				1,500(700) ^f

Table 4-12b. Average sorption ratios from batch sorption experiments on crushed tuff for americium, plutonium, uranium, selenium, technetium, and neptunium^{a, b} (page 1 of 3)

Strati- graphic unit ^c	Sample	Depth (ft)	Sorption ratios (ml/g)					
			Americium	Plutonium	Uranium	Selenium	Technetium	Neptunium
4-68	Tpc	JA-8 YM-5	606 251					
	Tpp	G2-547	547	13,000(110) ^f	1,200(120)	9.4(0.1)	2(2)	0 ^f
		G2-723	723	890,000(40,000)	>4,500	2.4(0.6)	19(2)	0 ^f
		GU3-433	433	3,400(200) ^d	330(60) ^g	0	15(3)	0
		GU3-855	855			10.0(0.7)	10(0.4)	7.9(0.1)
	Tpt	YM-22	848	1,200(130) ^{d, e}	64(20) ^d	1.8(0.2) ^d	0.3(0.14) ^d	7.0
		GU3-1203	1,203	1,100(120) ^g	360(40) ^g	0	0	2.7(0.1)
		G1-1292	1,292				(1)	
		GU3-1301	1,301	1,800(160) ^g	290(40) ^g	0	7(2)	0.03(0.001)
		YM-30	1,264					5.0(0.1)
		JA-18	1,420	180(30)	120(20)	2.5(0.4)		
	Th	G1-1436	1,436					
		G2-1952	1,952	1,700(70) ^f	66(6) ^f	0	2(1)	2.7(0.1)
		GU3-1436	1,436			20.0(2)	3(10)	
	Bt	GU3-1531	1,531			54.0(9)	5(1)	
		YM-38	1,504	14,600(1,000)	140(30)	5.3(0.2)		11.0
		YM-42	1,824					

Table 4-12b. Average sorption ratios from batch sorption experiments on crushed tuff for americium, plutonium, uranium, selenium, technetium, and neptunium^{a, b} (page 2 of 3)

Strati- graphic unit ^c	Sample	Depth (ft)	Sorption ratios (ml/g)					
			Americium	Plutonium	Uranium	Selenium	Technetium	Neptunium
Tcp	G1-1854	1,854						
	YM-45	1,930						
	G1-1883	1,883	470(300)					
	YM-46	2,002		77(11)				6.4
	G1-1982	1,982						
	YM-48	2,114						
	YM-49	2,221	4,300(1,400)				0.15(0.02)	
	JA-26	1,995		230(50) ^e			0.21(0.02)	9.0(3)
Tcb	JA-28	2,001						
	G1-2233	2,233						
	G1-2289	2,289						
	YM-54	2,491	153(6)	80(20)	1.3(0.3)	9(1)	4.2(0.5) ^h	
	G1-2333	2,333						
	G1-2363	2,363		110		25(5)		
	G1-2410	2,410			2.2(0.9)			
	JA-32	2,533	130(30)					
	G1-2476	2,476						
Tct	G1-2698	2,698						
	G1-2840	2,840						
	G1-2854	2,854						
	G1-2901	2,901		400(70) ^e	4.6(0.3)			
	G1-3116	3,116						
	JA-37	3,497	28,000(10,000) ^e					24

Table 4-12b. Average sorption ratios from batch sorption experiments on crushed tuff for americium, plutonium, uranium, selenium, technetium, and neptunium^{a, b} (page 3 of 3)

Strati- graphic unit ^c	Sample	Depth (ft)	Sorption ratios (ml/g)					
			Americium	Plutonium	Uranium	Selenium	Technetium	Neptunium
T1	G1-3658	3,658						
Tba	G2-3933	3,923	6,600(400)		1,600.0(30)	0	0.0(1)	0.1(0.006)

^aSources: Daniels et al. (1982); Ogard et al. (1983b); Wolfsberg et al. (1983). If no footnote is indicated, the sorption ratio in parentheses represents the standard deviation of the mean.

^bAmbient conditions, air, 20 ± 4°C; fractions do not contain particles smaller than 75 micrometers in diameter except in fractions designated by footnote (f). Sorption times greater than 1 week.

^cStratigraphic units are as follows: Tpc = Tiva Canyon Member; Tpp = Pah Canyon Member; Tpt = Topopah Spring Member; Th = tuffaceous beds of Calico Hills; Bt = bedded tuff; Tcp = Prow Pass Member; Tcb = Bullfrog Member;

Tct = Tram Member; Tl = older tuffs; Tba = bedded tuff.

^dNonweighted average; value in parentheses is the standard deviation of the mean.

^eSome data were rejected in averaging.

^fAverage of data for the fraction with particles smaller than 500 micrometers in diameter (contains some particles smaller than 75 micrometers).

^gNonweighted average of samples taken in two different positions.

^hPerformed under controlled atmospheric conditions of nitrogen with less than or equal to 0.2 parts per million O₂ and less than or equal to 20 parts per million CO₂.

Table 4-13a. Average sorption ratios from batch desorption experiments on crushed tuff for strontium, cesium, barium, cerium, and europium^{a, b} (page 1 of 2)
(See footnotes at the end of Table 4-13b)

Strati- graphic unit ^c	Sample	Depth (ft)	Sorption ratios (ml/g)				
			Strontium	Cesium	Barium	Cerium	Europium
Tpc	JA-8	606	311(3)	4,600(400)	480(50)	31,000(30,000) ^d	10,000(3,000)
	YM-5	251	320(30) ^d	8,900(600) ^d	1,200(120) ^d		36,000(14,000)
Tpp	G2-547	547	210(10) ^f	8,700(550) ^f	2,900(200) ^f		1,700(600) ^f
	GS-723	723	330(4) ^f	4,300(4) ^f	4,200(10) ^f		>10,000 ^f
	GU3-433	433	40(10) ^g	520(20) ^g	460(20) ^g		140(10) ^g
Tpt	YM-22	848	59(2)	365(7)	830(100)	6,500(800)	3,500(200)
	GU3-1203	1,203	47(1) ^f	340(10) ^f	720(30)	600(200) ^f	650(50)
	G1-1292	1,292	120(5) ^f	510(20) ^f	1,500(100)		600(70)
	GU3-1301	1,301	80(20) ^g	185(20) ^g	675(60) ^g		100(20) ^g
	YM-30	1,264	210(30)	1,500(100)	3,100(600)	170,000(15,000)	11,000(700)
	JA-18	1,420	15,000(2,000)	17,500(700)	280,000(50,000)	1,600(500) ^e	2,400(300) ^e
Th	G1-1436	1,436	87,000(12,000)	24,000(2,000) ^f	340,000(90,000) ^f	6,700(600)	5,300(600)
	G2-1952	1,952	4,200(200) ^f	46,000(1,400) ^f	40,000(1,000) ^f	2,600	1,600(200)
	YM-38	1,540	22,000	13,000	260,000		7,300
	YM-42	1,842	4,100(1,000)	21,000(2,000)	90,000(30,000)	44,000(5,000)	64,000(3,000)
Tcp	G1-1854	1,854	72,000(13,000) ^e	14,000(2,000)	150,000(40,000)	5,800(600) ^f	4,800(700)
	YM-45	1,930	210(20) ^f	620(110)	1,310(60) ^f		7,300(900) ^f
	G1-1883	1,883	59(1) ^f	430(4)	440(10) ^f	2,200(100) ^f	1,350(50) ^f
	YM-46	2,002	260(20) ^f	1,800(300) ^f	210,000(3,000) ^f	300,000(50,000) ^f	31,000(2,000) ^f
	G1-1982	1,982	322(8) ^f	2,300(200) ^f	2,780(120) ^f	7,000(800) ^f	6,370(130) ^f
	YM-48	2,114	2,700(200)	27,000(4,000)	34,000(7,000)	128,000(300)	8,100(1,200)
	YM-49	2,221	4,400(100)	39,000(1,000)	65,000(7,000)	1,040(40)	2,100(500)
	JA-26	1,995	39(3)	1,580(90)	450(13)		2,900(200)

Table 4-13a. Average sorption ratios from batch desorption experiments on crushed tuff for strontium, cesium, barium, cerium, and europium^{a, b} (page 2 of 2)
(See footnotes at the end of Table 4-13b)

Strati- graphic unit ^c	Sample	Depth (ft)	Sorption ratios (ml/g)				
			Strontium	Cesium	Barium	Cerium	Europium
Tcb	JA-28	2,001	114(3)	2,400(100)	1,160(20)		12,300(500)
	G1-2233	2,233	90,000(40,000) ^f	23,000(6,000) ^f	40,000(80,000) ^f	20,000(13,000) ^e	5,000(2,000) ^f
	G1-2289	2,289					
	YM-54	2,491	97(9)	310(20)	660(20)	1,000(200)	1,840(110)
	G1-2333	2,333	140(13)	1,230(100)	1,460(130)		9,900(1,200)
	G1-2363	2,363	150(6) ^f	1,200(30) ^f	820(20)	130,000(6,000) ^f	6,100(300)
	G1-2410	2,410	140(14)	1,120(100)	1,760(150)		6,000(3,000)
	JA-32	2,533	53(3)	175(11)	490(40)	530(120)	850(130)
	G1-2476	2,476	200(4)	1,520(0)			
Tct	G1-2698	2,698	210,000(50,000)	17,000(1,100)	190,000(80,000) ^f	2,000(400) ^f	
	G1-2840	2,840	1,540	2,300(130)	2,500(200)		9,000(1,100)
	G1-2854	2,854	96(1)	1,160(20)	1,300(0)		5,000(200)
	G1-2901	2,901	67(1) ^{e, f}	1,380(30) ^f	1,980(30)	39,000(1,000) ^f	210,000(50,000)
	G1-3116	3,116	24,000(13,000) ^f	11,000(3,000) ^f	160,000(80,000)	3,000(1,000) ^f	8,000(3,000) ^f
	JA-37	3,497	312(9)	850(50)	920(40)		11,000(2,000)
Tl	G1-3658	3,658	12,000(3,000) ^f	12,000(2,000) ^f	10,000(4,000)	9,000(4,000) ^f	9,000(3,000) ^f
Tba	G2-3933	3,933	140(20) ^f	1,400(350) ^f	1,100(200) ^f		3,000(1,100) ^f

Table 4-13b. Average sorption ratios from batch desorption experiments on crushed tuff for americium, plutonium, uranium, technetium, and neptunium^{a,b} (page 1 of 3)

Strati- graphic unit ^c	Sample	Depth (ft)	Sorption ratios (ml/g)				
			Americium	Plutonium	Uranium	Technetium	Neptunium
Tpc	JA-8	606					
	YM-5	521					
Tpp	G2-547	547	17,000(1,400)	1,200(170) ^f			
	G2-743	723	2.8×10^6 (2.6×10^4)	>47,000			
	GU3-433	433	9,300(1,780) ^g	920(40) ^g			
Tpt	YM-22	848	2,500(400) ^d	1,330(140) ^d	5(2) ^d	1.2(0.3) ^d	33(5) ^d
	GU3-1203	1,203	1,300(200) ^g	920(15) ^g			
	G1-1292	1,292					
	GU3-1301	1,301	2,500(600) ^g	1,300(460) ^g			
	YM-30	1,264					
	JA-18	1,420	1,100(300)	350(140)	9.4(1.4)		
Tht	G1-1436	1,436					
	G2-1952	1,952	5,800(1,100) ^g	350(45) ^g			
	YM-38	1,540	7,100(1,200)	1,600(300)	14.8(1.0)		24(2)
	YM-42	1,824					

Table 4-13b. Average sorption ratios from batch desorption experiments on crushed tuff for americium, plutonium, uranium, technetium, and neptunium^{a, b} (page 2 of 3)

Strati- graphic unit ^c	Sample	Depth (ft)	Sorption ratios (ml/g)				
			Americium	Plutonium	Uranium	Technetium	Neptunium
T _{cp}	G1-1854	1,854					
	YM-45	1,930					
	G1-1883	1,883	7,200(900)	890(60) = (61-1883)			36(10)
	YM-46	2,002					
	G1-1982	1,982					
	YM-48	2,114				1.6(0.2)	
	YM-49	2,221	3,400(400) ^e	720(90)		2.0(0.3)	12(4)
	JA-26	1,995					
T _{cb}	JA-28	2,001					
	G1-2233	2,233					
	G1-2289	2,289					
	YM-54	2,491	550(80)	720(40)	12(8)	38(-) ^h	
	G1-2333	2,333					
	G1-2363	2,363					
	G1-2410	2,410			8(2)		
	JA-32	2,533	2,200(600)				
	G1-2476	2,476					
T _{ct}	G1-2698	2,698					
	G1-2840	2,840					
	G1-2854	2,854					
	G1-2901	2,901					
	G1-3116	3,116					
	JA-37	3,497	32,000(10,000)	1,400(300)	9.9(0.4)		170(50)

Table 4-13b. Average sorption ratios from batch desorption experiments on crushed tuff for americium, plutonium, uranium, technetium, and neptunium^{a,b} (page 3 of 3)

Strati- graphic unit ^c	Sample	Depth (ft)	Sorption ratios (ml/g)				
			Americium	Plutonium	Uranium	Technetium	Neptunium
Tl	G1-3658	3,658					
Tba	G2-3933	3,933	1,200(410)	530(130)			

^aSources: Daniels et al. (1982); Ogard et al. (1983b); Wolfsberg et al. (1983). If no footnote is indicated, the sorption ratio in parentheses represents the standard deviation of the mean.

^bAmbient conditions, air, 20 ± 4°C; fractions do not contain particles smaller than 75 micrometers in diameter except in fractions designated by footnote f. Sorption times greater than 1 week.

^cStratigraphic units are as follows: Tpc = Tiva Canyon Member; Tpp = Pah Canyon Member; Tpt = Topopah Spring Member; Th = tuffaceous beds of Calico Hills; Tcp = Prow Pass Member; Tcb = Bullfrog Member; Tct = Tram Member; Tl = older tuffs; Tba = bedded tuff.

^dNonweighted average; value in parentheses is the standard deviation of the mean.

^eSome data were rejected in averaging.

^fAverage of data for the fraction with particles smaller than 500 micrometers in diameter (contains some particles smaller than 75 micrometers).

^gNonweighted average of samples taken in two different positions.

^hPerformed under controlled atmospheric conditions of nitrogen with less than or equal to 0.2 ppm O₂ and less than or equal to 20 ppm CO₂.

Table 4-14. Petrologic characterization of tuff samples from drillholes J-13, UE-25a#1, USW G-1, USW G-2, and USW GU-3^a (Page 1 of 4)

Unit ^b	Sample	Depth (ft)	USW G-1 depth (ft) ^c	Particle size (μm)	Abundance (weight %)								Degree of welding ^e	Oxidation state ⁱ	Crystals (%)	Lithics (%)
					Smectite	Illite Muscovite	Clinop- tilolite	Quartz	Cristob- alite	Alkali feldspar	Glass	Other ^d				
Tpt	JA-8	806	172	All ^g	25-50	-- ^h	--	--	10-20	tr ^h	25-50	--	M	(h)	8.9	6.7
Tpt	JA-8	806	172	75-500	30-60	--	--	<5	10-20	--	20-50	--				
Tpt	JA-8	806	172	75	30-60	--	--	--	10-20	--	30-60	--				
Tpc	YM-5	251	221	All	10	--	--	<5	<5	10-20	70	--	M		10.9	4.3
Tpp	G-2-547	547		<500	5-15	<5	--	--	5-15	10-20	40-60	--				
Tpp	G-2-547	547		75-500	14-84	--	--	3-7	2-8	43-63	--	6-16T				
Tpp	G-2-723	723		<500	5-15	<5	--	--	<5	5-10	15-30	30-70C				
Tpp	G-2-723	723		75-500	2-8	1-5	--	--	1-5	15-35	9-21	40-60C				
Tpt	GU-3-433	433	245	75-500	--	2-6	--	2-6	8-16	60-100	--	--				
Tpt	GU-3-855	855	758	<500	1-3	--	--	16-24	4-10	50-90	--	--				
Tpt	YM-22	480	888	All	5-10	--	--	40-60	--	40-60	--	--	0	G6(2-7)	1.0	0.4
Tpt	VII-22	848	888	108-500	<5	<2	--	30-50	--	30-50	--	--				
Tpt	VII-22	848	888	38-106	<2	tr	--	30-50	--	30-50	--	--				
Tpt	VII-22	848	888	<38	<5	<2	--	30-50	--	30-50	--	--				
Tpt	GG-3420 ⁱ	1,000	1,100	75-500	1	1	--	2-7	8-16	35-55	20-60	000				
Tpt	G1-1292	1,292	1,292	All	tr	--	--	--	5-10	10-20	80-90	--	V	G1		
Tpt	G1-1292	1,292	1,292	75-500	--	--	--	--	15-30	10-20	10-60	--				
Tpt	GU-3-1201	1,201	1,298	75-500	1-3	2-6	--	4-8	9-21	15-55	20-60	--				
Tpt	YM-30	1,264	1,318	All	5-10	5	5-10	40-60	5-15	30-50	--	--	0	G5(2-7)	2.1	21.6
Tpt	YM-30	1,264	1,318	75-100	--	--	15	30	20	35	--	--				
Tpt	YM-30	1,264	1,318	<75	--	--	15	30	20	35	--	--				
Tpt	JA-18	1,420	1,339	All	5	5	5-10	--	15-25	15-25	50	--	N	G3(2-5)	1.8	11.9
Tpt	JA-18	1,420	1,339	355-500	5	--	10-20	--	30-50	30-50	40	--				
Tpt	JA-18	1,420	1,339	108-150	5	5	10-20	--	30-50	30-50	40	--				
Tht	G1-1436	1,436	1,436	75-500	<5	<5	75-90	5-10	--	5	--	--		G6(5-7)	5.2	3.2
Tht	G2-1952	1,952	1,493	<500	--	1	25-45	3-7	3-7	20-40	--	15-35	M			
Tht	G2-1952	1,952	1,493	75-500 ^k	--	1	10-30	5-9	5-11	26-46	--	20-40				
Tht	G43-1436	1,436	1,504	<500 ^k	--	1	--	2-6	3-9	30-70	20-60	--				
Tht	YM-3d	1,504	1,538	108-500	5-10	<2	30-50	15-30	10-20	5-15	--	A, tr	N	G5(4-6)	4.0	7.7
Tht	YM-3d	1,504	1,538	38-106	5-10	<5	40-60	2-10	10-20	5-15	--	--				
Tht	YM-3d	1,504	1,538	<38	5-15	<5	40-60	2-10	10-20	10-10	--	A, tr				
Tht	VII-42	1,824	1,802	75-500	tr	<5	20	35-40	--	40	--	--			15.6	46.6
Tht	V14-42	1,824	1,802	<75	tr	tr	20	40	--	40	--	--				
Tba	GU3-1531	1,531		<500 ^k	--	3-7	--	15-25	3-9	20-60	10-50	--				

Table 4-14. Petrologic characterization of tuff samples from drillholes J-13, UE-25a#1, USW G-1, USW G-2, and USW GU-3^a (Page 2 of 4)

Unit ^b	Sample	Depth (ft)	USW G-1 depth (ft) ^c	Particle size (μm)	Abundance (weight %)								Degree of welding ^e	Oxidation state ^f	Crystals (%)	Lithics (%)
					Smectite	Illite	Clinop-tilolite	Quartz	Cristob- alite	Alkali feldspar	Glass	Other ^d				
Tcp	G1-1854	1,854	1,854	75-500	5-10	--	30-50	5-15	15-30	20-40	--	--				
Tcp	G1-1854	1,854	1,854	<75	5-10	--	40-60	20-40	15-30	10-30	--	--				
Tcp	YM-45	1,930	1,873	All	1-5	--	--	40-60	tr	30-50	--	--	M	C4(3-5)	13.5	0.6
Tcp	G1-1883	1,883	1,883	75-500	<2	<5	--	30-50	--	50-70	--	--	P	C4(3-5)	16.6	1.0
Tcp	G1-1883	1,883	1,883	106-500	<2	<2	--	20-40	0-10	40-60	--	--				
Tcp	G1-1883	1,883	1,883	38-106	<2	<5	--	30-50	0-10	30-50	--	--				
Tcp	G1-1883	1,883	1,883	<38	2-5	<2	--	20-40	0-10	40-60	--	--				
Tcp	VII-46	2,002	1,926	All	<5	<5	--	40-60	--	35-55	--	--	0		12.7	0.3
Tcp	G1-1982	1,982	1,982	All	tr	<5	--	5-15	40-60	30-50	--	--	N	C2	13.2	0.4
Tcp	G1-1982	1,982	1,982	75-500	5-10	<2	--	--	--	70-90	--	--				
Tcp	G1-1982	1,982	1,982	38-106	<2	--	--	--	40-60	40-60	--	--				
Tcp	G1-1982	1,982	1,982	<38	10-20	--	--	--	30-40	20-40	--	C, <2				
Tcp	YM-48	2,114	2,019	All	tr	--	10-20	--	--	20-30	40-60	--				
Tcp	YM-48	2,114	2,019	106-500	<2	--	20-40	5-10	5-15	20-40	10-30	--				
Tcp	YM-49	2,221	2,090	All	tr	tr	10-20	--	--	20-30	40-60	--	P	C6(5-7)	8.0	1.4
Tcp	JA-26	1,995	2,173	All	--	--	tr	30-50	tr	10-20	--	A, 30-50	N		18.3	1.0
Tcb	JA-28	2,001	2,178	All	tr	2.5	--	30-50	--	10-20	--	A, 30-50			20	5
Tcb	G1-2233	2,233	2,233	<500	<5	<5	20-40	11-20	10-20	10-20	--	H, 20-40	N	C6(6-7)	15.4	0.4
Tcb	G1-2233	2,233	2,233	75-500	<5	5-15	15-30	5-10	--	40-60	--	H, <5				
Tcb	G1-2233	2,233	2,233	38-106	<5	5	20-40	15-30	10-20	10-20	--	H, 20-40				
Tcb	G1-2289	2,289	2,289	75-500	--	5-10	30-50	<5	--	30-50	--	H, 10-20	N	C6(5-7)	15.8	1.0
Tcb	YM-54	2,491	2,330	All	tr	tr	--	50-70	--	20-40	--	--	W	C5(4-7)	17.8	0
Tcb	YM-54	2,491	2,330	106-500	5-10	2-5	--	30-50	--	25-45	--	--				
Tcb	YM-54	2,491	2,330	38-106	5-10	5-10	--	30-50	--	30-50	--	--	M	C6(5-7)	20.4	0.2
Tcb	YM-54	2,491	2,330	<38	5-10	2-10	--	15-30	--	40-60	--	--				
Tcb	G1-2333	2,333	2,333	75-500	2-5	2-5	--	15-30	10-30	50-70	--	--				
Tcb	G1-2333	2,333	2,333	<75	5-10	<5	--	15-30	20-40	20-40	--	--				
Tcb	G1-2363	2,363	2,363	All	10-20	5-10	--	30-50	--	30-50	--	--				
Tcb	G1-2363	2,363	2,363	106-500	5	<2	--	30-50	0-10	30-50	--	--				
Tcb	G1-2363	2,363	2,363	38-106	5	<2	--	30-50	0-10	30-50	--	--				
Tcb	G1-2363	2,363	2,363	<38	5-10	<2	--	20-40	0-10	40-60	--	--				

Table 4-14. Petrologic characterization of tuff samples from drillholes J-13, UE-25a#1, USW G-1, USW G-2, and USW GU-3^a (Page 3 of 4)

Unit ^b	Sample	Depth (ft)	USW G-1 depth (ft) ^c	Particle size (μm)	Abundance (weight %)								Degree of welding ^e	Oxidation state ^f	Crystals (%)	Lithics (%)
					Smectite	Illite Muscovite	Clinop-tilolite	Quartz	Cristobalite	Alkali feldspar	Glass	Other ^d				
Tcb	G1-2410	2,410	2,410	75-500	5-10	<2	--	20-40	0-10	30-50	--	--				
Tcb	G1-2410	2,410	2,410	<75	5-10	<5	--	20-40	5-15	30-50	--	--				
Tcb	JA-32	2,533	2,467	106-500	<5	5-15	--	30-50	--	30-50	--	--				
Tcb	JA-32	2,533	2,467	355-500	--	5-10	--	40-50	--	30-40	--	--				
Tcb	JA-32	2,533	2,467	150-180	--	10-15	--	35-50	--	40-65	--	--				
Tcb	JA-32	2,533	2,467	106-150	<5	5-15	--	30-50	--	30-50	--	--				
Tcb	JA-32	2,533	2,467	38-106	<2	tr	--	30-50	--	30-50	--	--				
Tcb	JA-32	2,533	2,467	<38	5-10	--	--	20-40	--	40-60	--	A, tr				
Tcb	G1-2476	2,476	2,476	75-500	<2	2	--	30-50	5-15	40-60	--	--				
Tcb	G1-2476	2,476	2,476	<75	2-5	2	--	30-50	5-15	40-60	--	--				
Tct	G1-2698	2,698	2,698	A11	<5	10-15	30-50	<5	--	30-50	--	H, 5	M	C6(5-7)	S13.2	1
Tct	G1-2840	2,840	2,840	75-500	2-5	2-5	--	40-60	0-10	30-50	--	--				
Tct	G1-2840	2,840	2,840	<75	2-5	2-5	--	40-60	0-10	30-50	--	--				
Tct	G1-2854	2,854	2,854	75-500	<2	5-10	--	30-50	0-10	30-50	--	--				
Tct	G1-2854	2,854	2,854	<75	<2	5-10	--	30-50	0-10	30-50	--	--				
Tct	G1-2901	2,901	2,901	A11	5-10	5-10	--	20-40	--	40-60	--	--	W	C6(6-7)	4.0	21.4
Tct	G1-3116	3,116	3,116	A11	5-10	5-10	5-15	20-40	--	20-40	--	A, 10-30				
Tct	JA-37	3,497	3,286	A11	20-40	5	5	30-60	--	15-30	--	--				
Tct	JA-37	3,497	3,286	355-500	10-15	--	tr	40-50	--	30-40	--	C, tr				
Tct	JA-37	3,497	3,286	106-150	5-10	--	--	40-50	--	30-40	--	C, tr				
TI	G1-3658	3,658	3,658	75-500	40-60	--	--	--	--	40-60				C3	23.4	0
TI	G1-3658	3,658	3,658	106-500	40-60	--	--	--	--	40-60						
TI	G1-3658	3,658	3,658	38-106	30-50	--	--	--	--	50-70						
TI	G1-3658	3,658	3,658	<38	50-70	--	--	--	--	30-50						

Table 4-14. Petrologic characterization of tuff samples from drillholes J-13, UE-25a#1, USW G-1, USW G-2, and USW GU-3^a (Page 4 of 4)

Unit ^b	Sample	Depth (ft)	USW G-1 depth (ft) ^c	Particle size (μm)	Abundance (weight %)							Degree of welding ^e	Oxidation state ^f	Crystals (%)	Lithics (%)
					Smectite	Illite	Clinop-tilolite	Quartz	Cristobalite	Alkali feldspar	Glass				
Tba	G2-3933	3,933		<500	10-20	--	--	20-40	--	20-40	--	10-40,A			
Tba	G2-3933	3,933		75-500	40-10	--	--	29-39	--	34-54	--	10-20,A			

^aAnalyses were performed by ESS division of the Los Alamos National Laboratory; methods are discussed in Caporuscio et al., 1982. Data from Daniels et al., (1982). To convert from feet to meters, multiply by 0.328.

^bTpc = Tiva Canyon Member of the Paintbrush Tuff; Tpt = Topopah Spring Member of the Paintbrush Tuff; Tht = Tuffaceous beds of Calico Hills; Tpc = Prow Pass Member of the Crater Flats Tuff; Tcb = Bullfrog Member of the Crater Flat Tuff; Tct = Tram unit of the Crater Flat Tuff; TI = lava flow and flow breccia; Tba = bedded tuff; and Tpp = Pah Canyon Member.

^cEquivalent depth in drillhole USW G-1 according to relative position in stratigraphic unit. The thickness of the Bullfrog and Tram units in drillholes UE-25a#1 and J-13, respectively, are assumed to be of the same thickness as the corresponding units in drillhole USW G-1.

^dA = analcime; C = calcite; M = mordenite; and T = tridymite.

^eN = nonwelded; P = partly welded; M = moderately welded; O = densely welded; V = very densely welded (vitrophyre); and W = intermediately welded.

^fThe empirical stage of oxidation of iron-titanium exsolution oxide phases; C1 denotes unoxidized and C7 denotes completely oxidized. See Caporuscio for a discussion of oxide mineral alteration trends.

^gStarting with whole rock.

^hA blank indicates that an analysis was not performed; dashes indicate that the mineral was not detected, and tr = trace (<1%).

ⁱTopopah Spring vitrophyre.

^jMineralogy in Tht in USW GU-3 is not similar to that in Tht in USW G-1. Stratigraphic equivalence is not valid.

^kSorption work was done with 75-500 μm fraction. Analysis still in progress.

differences in aqueous speciation or sorption reactions with the solid phases. Further studies are described in Section 8.3.1.3.4.

Only limited data are available on the effect of elevated temperature on sorption behavior in tuff. Data obtained at 70° and 85°C (Table XVII, Ogard et al., 1983b) suggest that higher temperatures will generally increase slightly the sorptive capacity of the tuff samples for all elements studied (Daniels et al., 1982; Ogard et al., 1983a). Samples studied for temperature effects were Topopah Spring Member tuffs containing equal amounts of cristobalite and alkali feldspar; Topopah Spring Member tuff largely composed of alkali feldspar; Topopah Spring Member tuff with equal amounts of glass and alkali feldspar; Bullfrog Member tuff containing equal amounts of quartz and alkali feldspar and a small amount of illite; and Tram Member tuff containing equal amounts of quartz, alkali feldspar, and a small amount of smectite.

The sorptive behavior of some multivalent radionuclides may depend on in situ oxidation-reduction conditions of the ground-water environment in the tuff. Data are available on sorptive behavior under conditions less oxidizing than atmospheric conditions (Daniels et al., 1982). These experiments were performed in a nitrogen atmosphere with less than 0.2 parts per million (ppm) oxygen and less than 20 ppm carbon dioxide. These conditions are somewhat different from those in deep geologic environments because of the low carbon dioxide content in the laboratory experiment and the possible influence on complex formation. The inert atmosphere environment should bracket the subsurface oxidation-reduction conditions. The results showed the absence of oxygen and carbon dioxide had little effect on sorption for most elements (Daniels et al., 1982). However, plutonium and technetium showed higher sorption ratios in these experiments (plutonium up to 800 ml/g and technetium up to 14 ml/g as compared with approximately 100 ml/g and less than 1 ml/g, respectively, under oxidizing conditions). Experiments performed under conditions that maintained the solution pH at 6 to 7 by varying the carbon dioxide content of the atmosphere also showed little changes in the sorption ratio in comparison with ambient experiments (Ogard and Vaniman, 1985; Rundberg et al., 1985).

It is important to determine the range of linearity of the sorption isotherms for several reasons. First, the concentration of waste elements will change considerably from a dilute solution at the repository to an even more dilute solution when the dissolved waste element becomes mixed with ground water. To model such a change in concentration in the transport process, the variation of sorption with concentration is needed. In the case of nonlinear sorption, the variation data may help to establish bounds on sorption. Second, mechanisms like those that lead to irreversible sorption are studied by measuring desorption ratios. Third, diffusion into the tuff matrix as would occur in fracture flow is interpreted and modeled. If diffusion is occurring, then the sorption mechanism is not a simple ion exchange.

In most transport codes, the sorption ratio does not depend on concentration. This assumption should be valid for ion-exchange equilibrium at tracer-level concentrations within zeolitized tuffs, but it may not be true for welded nonzeolitized tuffs. Therefore, the dependence on concentration must be determined. Sorption isotherms have been determined (Wolfsberg et al., 1981; Daniels et al., 1982) for three tuff samples from drillhole UE-25a#1 (YM-22, YM-38, and YM-49), and on a sample from drillhole USW G-1

(G1-2840) (Wolfsberg et al., 1982a) (Table 4-14 for sample descriptions). Sample YM-22 is a densely welded tuff from the Topopah Spring Member; sample YM-38 is a highly zeolitized tuff from the tuffaceous beds of Calico Hills; sample YM-49 is a partially zeolitized and partially welded vitric tuff from the lower Prow Pass Member; and sample G1-2840 is a devitrified tuff similar to YM-22. Sorption isotherms for strontium, cesium, barium, and europium were determined for samples YM-22 and YM-38 (Wolfsberg et al., 1981). Sorption isotherms were also determined for cesium for sample G1-2840 (Daniels et al., 1983). The experimental data were fitted to a Freundlich isotherm, an empirical formula of the form

$$q = kc^n \quad (4-3)$$

where q is the solute concentration on the solid phase, c is the solute concentration in the aqueous phase, and k and n are constants ($n = 1$ for a linear isotherm).

This isotherm can be used to summarize large amounts of data. The results by Wolfsberg et al. (1981, 1982a) and Daniels et al. (1983) for strontium, cesium, barium, and europium indicate a nonlinear isotherm ($n = 0.71$ to 0.92) for the coarse size fraction of sample YM-22 (welded tuff) and for sample G1-2840, both of which are nonzeolitized tuffs. With the possible exception of strontium ($n = 0.87$), a linear isotherm ($n = 0.98$ to 1.1) is indicated for the coarse-size fraction of sample YM-38 (zeolitized tuff). This latter result is expected because the sorption of these elements on zeolitized material should occur predominantly by ion exchange. The results of this study indicate that isotherms can be used to interpret the studies of sorption on both zeolitized and nonzeolitized tuff.

Sorption isotherms were determined for plutonium on tuff samples YM-22 and YM-49 for 10 concentrations from 2.9×10^{-8} to 5.9×10^{-13} M (Daniels et al., 1982). Except for the highest plutonium concentration, which generally exhibits the lowest R_d value, there was little correlation between sorption ratio and element concentration in the range studied. The Freundlich isotherm for tuff sample YM-22 (welded) indicates nonlinear sorption ($n = 0.84$ to 0.88), whereas sorption for tuff sample YM-49 (partially zeolitized and vitric) is approximately linear ($n = 0.96$ to 1.0). This is consistent with the results of measurements using strontium, cesium, barium, and europium. The sorption-isotherm studies are being extended to include the other waste radionuclides to fully model the retardation process. Plans for obtaining the additional information are outlined in Section 8.3.1.3.4.

The saturation of sorption sites can cause a sorption isotherm to be nonlinear, as observed in samples YM-22 and G1-2840. Some experimentally determined values for cation-exchange capacity (CEC), which is the measure of available sorption sites, are presented by Wolfsberg and Vaniman (1984). High concentrations (higher than 10^{-5} M) of sorbing species were used in these experiments to obtain the CEC. Sample YM-22 tuffs have a relatively low CEC while exhibiting nonlinear sorption isotherms; however, the zeolitized tuffs (YM-49, YM-38) have a relatively high CEC and exhibit linear sorption isotherms. However, it was determined that a simple retardation factor is valid for the alkali metals and alkaline earths at concentrations as high as 10^{-5} M for strontium and cesium and 10^{-8} M for barium and plutonium.

4.1.3.3.3 Sorption data from experiments with crushed-tuff columns

The sorptive behavior of radionuclides in elution experiments with columns of crushed tuff has been described by Treher and Raybold (1982). Several observations can be made based on data reported from approximately 40 column experiments. Sorption ratios (R_d) for strontium, cesium, and barium measured by the column method generally fall within the range of values measured with the batch technique when both column and batch measurements are made with crushed tuff washed free of fine particles (less than 35 μm). Wolfsberg et al. (1981) had indicated that the R_d values inferred from column studies were two to five times smaller than those from batch studies, but the comparisons were made with batch measurements on material containing fine particles, which are predominantly highly sorbing clays. Column-elution curves for uranium-237 result in an R_d value of 0.72 ml/g, in good agreement with the batch measurement of 1.5 ml/g. Experiments with actinide elements are planned (Section 8.3.1.3).

4.1.3.3.4 Sorption data from circulating-system experiments

Average sorption ratios determined by both the batch and the circulating-system methods are shown in Table 4-15 for strontium, cesium, barium, americium, and plutonium. Considering the spread of experimental values, the agreement between the two methods is acceptable. In most cases the results fall within the spread of individual experimental values reported by Daniels et al. (1982).

Devitrified tuffs give slightly higher sorption ratios in the batch method than in the circulating-system method. The R_d values for the smectite-bearing devitrified tuff and for the zeolite-bearing vitric tuff show no consistent pattern.

4.1.3.3.5 Comparison of sorption ratios from batch, circulating system, and column measurements

Batch and both types of column experiments on various tuffs with the same particle-size distribution yield sorption ratios for strontium, barium, cesium, americium, and plutonium that are within a factor of 2 (americium on sample JA-37, a devitrified tuff containing smectite, was an exception). This agreement means that self-grinding of the tuff in batch experiments is probably not a problem. The experiments with crushed-tuff columns show that batch sorption measurements are a valid representation of flowing experiments when the crushed tuffs are free of fine particles less than 35 micrometers in diameter.

4.1.3.3.6 Sorptive behavior as a function of stratigraphic position and mineralogy

This section discusses sorption ratios as a function of stratigraphic

Table 4-15. Average sorption ratios from batch, circulating-system, and crushed-tuff column sorption measurements at room temperature^a

Element	Sample	Petrologic character ^b	Batch ^c	Sorption ratio (ml/g)	
				Circulating system	Column
Sr	YM-22	D	53 (3) ^d	27 (2) ^c	22
	YM-54	D	62 (12)	45 (3)	43
	JA-37	D + S	287 (14)	390 (10)	123
Cs	YM-22	D	290 (30)	490 (50)	363
	YM-54	D	180 (40)	120 (10)	130
	JA-37	D + S	610 (40)	1,800 (10)	1390
Ba	YM-22	D	900 (30)	120 (10)	137
	YM-54	D	410 (150)	130 (10)	149
	JA-37	D + S	760 (150)	860 (40)	-- ^e
Am	YM-49	V + Z	4,300 (1,400)	2,200 (300)	--
	JA-37	D + S	28,000 (10,000)	3,400 (600)	--
	G1-1883	D	4,700 (300)	3,300 (100)	--
Pu	YM-49	V + Z	230 (50)	570 (170)	--
	JA-37	D + S	400 (70)	290 (170)	--
	G1-1883	D	77 (11)	56 (11)	--

^aSources: Daniels et al. (1982); Treher and Raybold (1982).

^bD = devitrified, V = vitric, S = contains smectite, Z = contains zeolite.

^cFrom Tables 4-12 and 4-13.

^dValue in parentheses is the absolute-value standard deviation of the means of duplicate measurements.

^e-- = no data.

position and mineralogy. These analyses are important for calculations of retardation along possible travel paths to the accessible environment.

Average sorption ratios determined from ambient-temperature batch sorption experiments on crushed-tuff samples from well J-13 and drillholes UE-25a#1, USW G-1, USW G-2, and USW GU-3 are presented in Table 4-12. In Figure 4-14, the data are plotted against equivalent sample depths in drill-hole USW G-1 for well J-13 and drillholes UE-25a#1 and USW G-1 by Daniels et al. (1982).

Figure 4-14. Variations in the sorptive behavior of strontium and cesium with the abundance of clinoptilolite and stratigraphic depth. Modified from Daniels et al. (1982). Dashed lines are data for a more complete set of samples from Bish et al. (1981).

Variations in the sorptive behavior of strontium and cesium with the abundance of the zeolite clinoptilolite are shown in Figure 4-14. The relative abundances of the minerals found at various depths given in Figures 4-1 and 4-2. The correlations are quite striking, showing an essentially direct correlation between an increase in strontium and cesium sorption ratios and an increase in clinoptilolite abundance. A similar correlation can be made for the sorption of barium and radium (Daniels et al., 1982; Ugard et al., 1983b). The correlations are even better when the abundances of clinoptilolite and mordenite are considered together. The increase in sorption for cesium, strontium, and barium on sample G1-3658 from the flow breccia below the Tram Member appears to be directly related to the high content of smectite clay in the sample (30 to 70 percent); the sample contains no detectable quantities of zeolite (Table 4-14) (Daniels et al., 1982; Ugard et al., 1983a). Section 8.3.1.3 describes plans for examining the sorption data using various statistical analyses, with emphasis on correlating sorption with mineralogy.

For cerium and europium, variations in the sorption ratios with stratigraphic position are not as regular as those for strontium, cesium, and barium. In broad outline, the stratigraphic positions of the maximum and minimum sorption ratios for cerium and europium coincide closely with the maxima and minima in R_d for cesium, strontium, and barium in units in the unsaturated zone and downward to the contact between the Bullfrog and Tram members. The Tram Member shows little correlation, with samples having the highest R_d for cerium and europium often giving the lowest R_d for cesium, strontium, and barium (and vice versa). The difference may be explained by mineralogic differences and will be investigated on that basis (Section 8.3.1.3). It may be difficult to correlate statistically this difference with mineralogic differences due to other processes that may be affecting the sorption; however, it is known that the analytical error inherent in these experiments is not the source of the difference.

Apparently there is a correlation between the sorption of cerium or europium and the presence of zeolites in the lower Topopah Spring Member, in the tuffaceous beds of Calico Hills, and in the upper Prow Pass Member (Table 4-14). The high sorption ratios in the upper Topopah Spring Member and the lower Tiva Canyon Member may be associated with the high content of smectite (10 to 60 percent) in the samples studied. The high sorption ratio measured for sample G1-2901 from the middle portion of the Tram Member unit does not appear to be correlated with mineralogic variations in drillhole USW G-1.

Sorption data for the actinides uranium, neptunium, plutonium, and americium are limited. The sorption ratios for uranium and neptunium are below 10 ml/g for most samples (Table 4-12b); sorption ratios for plutonium are near 100 ml/g (Table 4-12b) for all samples. However, there are no strong correlations with the mineralogy of the test samples. Sample JA-37 from the lower portion of the Tram Member and sample G2-723 from the lower portion of the Pah Canyon Member of the Paintbrush Tuffs generally give the highest R_d values for these actinides. The high R_d values may be due to the relatively high smectite content of sample JA-37 (10 to 40 percent), or the high calcite content (30 to 40 percent) of sample G2-723. The sorption ratios for americium may be mineralogically controlled, being generally higher than 1,000 ml/g in the zeolite tuffs of Calico Hills and the Prow Pass Member and decreasing to about 100 ml/g in the nonzeolitized lower Topopah

Spring and the middle Bullfrog members. The sorption ratio for sample JA-37 from the Tram Member is again the highest and may result from the presence of smectites.

The sorption ratios for plutonium are generally between 50 and 1,000 ml/g, independent of sample position and lithology (Daniels et al., 1982). In samples from nonwelded to partially welded zeolitized or vitric tuffs from the tuffaceous beds of Calico Hills, the Prow Pass, the Pah Canyon, the Topopah Spring, and the Tram members have R_d values for americium higher than 1,000 ml/g, whereas samples with lower R_d s are generally associated with welded and devitrified units. The sorption ratio for americium is highest in sample G2-723 from the lower Tram Member unit.

The sorption ratios for technetium are very limited. In most experiments under atmospheric conditions, the sorption ratios are less than unity and in some cases are zero. Because the sorption ratios for technetium are so low, long-term (12 mo) sorption experiments approaching equilibrium have been done and they show no change in R_d with time (Ogard and Vaniman, 1985). At these low sorption levels, correlation with stratigraphic position may not be valid or useful.

Very limited data for thorium and tin (Rundberg et al., 1985) do not show a correlation with mineralogy. The sorption ratios for thorium were higher than 140 ml/g, and those for tin were 150 ml/g or higher.

As described in Section 8.3.1.3, current studies will provide a more complete understanding of the sorption of the radionuclides listed in 40 CFR Part 191, Appendix A, including technetium, tin, and neptunium. Uranium and thorium are also being studied because they are the parents of radium. Regression analysis is being employed to determine the correlation of sorption with tuff mineralogy.

Because tuff samples may be composed of more than one sorbing mineral, a method has been developed to predict sorption ratios for samples of known mineralogy by combining the effects of several minerals. The combined effect is tentatively defined as a weighted sum called the SMC. Thus,

$$SMC = \sum W_i X_i, \quad (4-4)$$

where W_i is the weighting factor for each mineral phase relative to that of clinoptilolite and X_i is the abundance in percent of each phase. Further details are given in Daniels et al. (1982). Studies of sorption on pure minerals and selected core samples will be performed to quantify expected sorption along transport pathways using the SMC concept (Section 8.3.1.3.4).

The following correlations of mineralogy with sorptive behavior of radionuclides have been identified and can be summarized as follows:

1. The sorption of alkali metals (e.g., cesium) and alkaline earths (strontium, barium, and radium), which probably exist in ground water as uncomplexed ions (Section 4.1.3.4.2) and probably sorb by ion exchange, is directly related to the presence of minerals with exchangeable cations: the zeolite clinoptilolite and potentially the smectite clays. The sorption ratios on these minerals are from

1,000 to higher than 10,000 ml/g. Because of the large quantities of the zeolite clinoptilolite in the Calico Hills and Prow Pass units lying below the repository horizon, the waste elements cesium, strontium, and radium would be strongly sorbed and their movement greatly retarded if the flow path is through these units.

2. The correlation of sorption of cerium, europium, plutonium, and americium with mineralogy also exists, but the relation is not as clear as for the alkali metals. However, sorption ratios are high for the elements previously listed (from 100 to higher than 1,000 ml/g) and, therefore, the lack of correlation is not critical. The sorption of these elements, though, is undoubtedly influenced by the formation of hydroxyl and carbonate complexes.
3. It appears that sorption on tuff will not offer much retardation for such anionic species as oxidized technetium species (i.e., TcO_4^-).

4.1.3.3.7 Sorptive behavior as a function of ground-water composition

The composition of the ground water and matrix mineralogy controls the oxidation state and speciation of radionuclides as well as their solubility, while at the same time providing competing ion and species for sorption sites, all of which can affect sorption of the radionuclides by tuff. Because ground-water composition may vary between the repository and the accessible environment, studies have been performed to determine what effect changes in ground-water composition have on measured sorption ratios.

Three ground waters are being investigated: water from well J-13, located in Jackass Flats; water from drillhole UE-25p#1 and from drillhole USW H-3, both of which are located on Yucca Mountain. Water from well J-13 has been used as the reference ground-water composition partly because the approximate production region of the well is the Topopah Spring Member; which, at Yucca Mountain, lies in the unsaturated zone and is the proposed repository horizon, and because, at the time studies on ground water began, well J-13 was the nearest producing well to Yucca Mountain. Comparisons are now being made of sorption ratios obtained with J-13 water to those from drillholes UE-25p#1 and USW H-3. The composition of these waters is given in Tables 4-7 and 4-8. Water from drillhole UE-25p#1 has the highest ionic strength, whereas drillhole USW H-3 water has the highest pH measured thus far at Yucca Mountain and was measured as slightly reducing in the field. A thorough discussion of water chemistry at Yucca Mountain is contained in Ogard and Kerrisk (1984). Distilled water was also studied (in some instances) as a representative of an extreme in ground-water composition.

Elements studied thus far include strontium, cesium, barium, europium, and tin. Plans for studying the sorption of actinides as a function of ground-water chemistry is found in Section 8.3.1.3.4. The selection of tuff samples used was based on their predominant mineralogy. Sample GU3-1301 is a vitric tuff from the Topopah Spring Member; G1-2233 is a zeolitized tuff from the Bullfrog Member; G4-1504 is a zeolitized tuff from the Calico Hills; and G1-2901 is a devitrified tuff from the Tram Member. Sorption ratios were measured using standard batch technique (Section 4.1.3.2.2).

Tables 4-16a and 4-16b give the results obtained for well J-13, drillhole UE-25p#1, and distilled water (where available). Data from drillhole USW H-3 have not yet been published. For those elements presumed to sorb by an ion exchange mechanism (such as strontium, barium, and cesium), the lower ionic strength waters gave the highest sorption ratios. This is probably due to decreased competition for sorption sites. The exception appears to be barium in distilled water on sample G1-2233; however, the well J-13 results may be too high due to the inclusion of particles finer than 75 micrometers in that experiment. In contrast to these single oxidation state elements, europium and tin exhibit higher sorption ratios for the same initial concentrations in the higher ionic strength drillhole UE-25p#1 water than in the more dilute waters. It is postulated that europium and tin are precipitating out of solution in the drillhole UE-25p#1 water, perhaps as sulfates or fluorides. Plans for studying solubility and precipitation of the radionuclides are discussed in Section 8.3.1.3.5.

From the data presented, it is evident that ground-water composition can affect sorption ratios. The ground waters studied represent extremes in the water composition found so far at Yucca Mountain. When information on the vadose water composition (unsaturated-zone water) is available, studies will be made to characterize its effect on sorption ratios. Plans for obtaining the additional required information are outlined in Section 8.3.1.3.

4.1.3.4 Processes affecting radionuclide concentrations and speciation in solution

The solubility of the waste elements can influence their transport by limiting the maximum concentration of the elements dissolved in the aqueous phase. Speciation, defined as the formation of various complexes and oxidation states in the aqueous phase, in turn affects the solubility. Speciation and solubility of individual waste elements depend on the chemical properties of the waste element, on the state of the local water (composition, pH, oxidation state, and temperature), and, if nonequilibrium processes are important, on factors such as precipitation and dissolution kinetics, oxidation-reduction kinetics, the identity of the solids present, water-flow conditions, and colloid formation. The following discussion of solubility and speciation is in terms of waste elements rather than radionuclides because these processes are controlled by the total concentration of an element in solution--that is, the sum of the concentrations of all its isotopes. The concentrations of individual radionuclides depend on the isotopic composition of the element.

At this time it is possible to estimate the solubilities of many important waste elements under repository conditions. The estimates are primarily based on literature data interpreted in terms of a chemical-equilibrium model. However, solubility calculations alone will not be sufficient for determining the concentration limits of waste elements (NRC, 1984). For this reason, solubility data for important waste elements have been and will continue to be obtained by laboratory experiments under conditions that are characteristic of a Yucca Mountain repository. The experiments are just beginning, and only a few results are available. A combination of site-specific solubility and speciation data will be used

Table 4-16a. Sorption ratios for cesium and strontium in different waters^{a,b,c}

Core ^d		Cesium			Strontium		
		J-13	UE-25p#1	Distilled water	J-13	UE-25p#1	Distilled water
GU3-1301 (vitric)	Sorption	160(35)	46(5)	NA ^e	32(8)	10(2)	NA
	Desorption	180(40)	55(4)	NA	90(20)	32(1)	NA
G1-2233 (zeolitized)	Sorption	13,500(800)	7,500(1,100)	13,000(1,600)	48,000(3,000)	2,000(330)	56,000
	Desorption	23,000(6,000)	8,800(900)	8,700(3,400)	90,000(40,000)	3,000(500)	21,000(12,000)
G1-2901 (devitrified)	Sorption	1,290(110)	900(250)	3,350(170)	68(1)	44(13)	136(1)
	Desorption	1,380(30)	850(76)	3,800(670)	67(1)	35(4)	105(13)
G4-1502 (zeolitized)	Sorption	ND ^g	ND	ND	ND	ND	ND
	Desorption	ND	ND	ND	ND	ND	ND
G1-2840	Sorption	ND	ND	ND	ND	ND	ND
	Desorption	ND	ND	ND	ND	ND	ND

^aSorption ratios in ml/g.^bNumbers in parenthesis represent the standard deviation of the mean.^cCrowe and Vaniman (1985); Daniels et al. (1982); Ogard and Vaniman (1985); and Rundberg et al. (1985).^dThe cores are described in the text.^eNA = Not applicable.^fParticle size was <500 m rather than the standard 75 to 500 μ m.^gND = No data.^hOnly one measurement.

Table 4-16b. Sorption ratios for barium, europium and tin^{a,b,c}

Core ^d		Barium			Europium			Tin	
		J-13	UE-25p#1	Distilled water	J-13	UE-25p#1	Distilled water	J-13	UE-25p#1
GU3-1301 (vitric)	Sorption	570(60)	82(18)	NA	75(12)	>17,000	NA	160(8)	4,000(520)
	Desorption	660(100)	140(30)	NA	110(40)	>20,000	NA	1,280(180)	6,750 ^h
G1-2233 (zeolitized)	Sorption	250,000(30,000) ^f	41,000(6,300)	55,000(5,300)	900(200)	>5,600	516(100)	ND ^g	ND
	Desorption	240,000(80,000) ^f	59,000(6,000)	34,000(23,000)	5,000(2,000)	>7,900	1,400(600)	ND	ND
G1-2901 (devitrified)	Sorption	1,600(200)	1,400(520)	7,200(100)	160,000(50,000)	38,000(16,000)	37,000(400)	22,000(5,000)	35,800(900)
	Desorption	1,980(30)	1,100(90)	6,100(900)	210,000(50,000)	69,000(20,000)	27,500(6,200)	38,000(8,000)	52,500(1,900)
G4-1502 (zeolitized)	Sorption	ND	ND	ND	ND	ND	ND	215(56)	800(110)
	Desorption	ND	ND	ND	ND	ND	ND	500(8)	300(130)
G1-2840	Sorption	ND	ND	ND	ND	ND	ND	283(160)	20,000(2,700)
	Desorption	ND	ND	ND	ND	ND	ND	780(130)	18,400(6,900)

^aSorption ratios in ml/g.

^bNumbers in parenthesis represent the standard deviation of the mean.

^cCrowe and Vaniman (1985); Daniels et al. (1982); Ogard and Vaniman (1985); and Rundberg et al. (1985).

^dThe cores are described in the text.

^eNA = Not applicable.

^fParticle size was <500 μ m rather than the standard 75 to 500 μ m.

^gND = No data.

^hOnly one measurement.

together with literature data to develop solubility models. The modeling effort is basically an attempt to understand within a theoretical framework the parameters that control solubility and speciation. Thus, chemical-equilibrium (thermodynamic) models will be used wherever possible.

Section 4.1.3.4.1 gives brief discussions of speciation, solubility, natural colloid formation, and radiolysis, processes important in understanding the concentration limits of elements in solution. Section 4.1.3.4.2 discusses the results of solubility measurements conducted for this program as well as solubility and speciation data that are available in the literature. Section 4.1.3.4.3 briefly discusses the modeling program that is being conducted in conjunction with the solubility and speciation experiments. Section 4.1.3.4.4 lists estimates of the solubility of a number of important waste elements in well J-13 water. Plans for obtaining additional solubility data are outlined in Section 8.3.1.3.5.

4.1.3.4.1 Processes affecting waste element concentrations in solution

4.1.3.4.1.1 Speciation

Elements dissolved in water can exist as various chemical species such as different oxidation states or complexes with other ions in the water (Garrels and Christ, 1965; Baes and Mesmer, 1976; Stumm and Morgan, 1981). Solubility generally increases as the variety and concentration of complexes of an element increase; thus, the solubility is influenced by the tendency of a given element to form complexes and the concentrations of species with which it can complex. Sorptive behavior depends on the size and charge of the sorbing species; both of these quantities vary among the complexes of a given element. Thus, speciation can influence sorption. Aqueous species can be experimentally detected in solution by a number of techniques; spectroscopy is most commonly used (Rossotti and Rossotti, 1961). However, concentrations of aqueous species are normally calculated from a knowledge of the overall composition of the solution (total concentrations of the elements in solution) and the formation constants of possible aqueous species using equilibrium thermodynamic methods. Equilibrium thermodynamic methods work well (given the proper data) for the various complexes of a particular oxidation state, but may yield inaccurate results for the distribution of the element among possible oxidation states (Lindberg and Runnells, 1984).

4.1.3.4.1.2 Precipitation

Solubility, like speciation, depends on the composition, pH, and temperature of the water (Garrels and Christ, 1965; Sposito, 1981; Stumm and Morgan, 1981). If an element can also exist in various oxidation states, variables that control oxidation-reduction behavior also influence solubility. Unlike complex formation, which is essentially always an equilibrium process, precipitation processes are often not in equilibrium or in metastable equilibrium (Stumm and Morgan, 1981). If dissolution and precipitation kinetics are fast on the time scale of interest, equilibrium

behavior can be assumed. If kinetics are very slow, a metastable equilibrium may exist where the aqueous phase is in metastable equilibrium with some solid other than the most stable (least soluble) one. In some intermediate cases, the dissolution or precipitation rates may be comparable to the time scale of interest; for these cases, kinetic data are required to accurately describe aqueous concentrations.

Coprecipitation refers to a group of processes whereby more than one compound precipitates at one time (Sposito, 1981). Three examples are mixed-solid formation, adsorption during precipitation, and inclusion during precipitation. The importance of coprecipitation for waste-element solubility at Yucca Mountain is uncertain at this time.

4.1.3.4.1.3 Natural-colloid formation

A number of actinides, plutonium in particular, can form natural colloids under certain conditions (Avogadro and DeMarsily, 1984; Kim et al., 1984a; Olofsson et al., 1984). The questions of primary importance for a waste repository are whether natural colloids can form under the conditions expected at the repository or along flow paths to the environment--and whether the colloids would be transported by the aqueous phase. This section addresses only the question of natural-colloid formation; colloid transport is discussed in Section 4.1.3.6. Related, but different phenomena, the sorption of waste elements on particulate material normally present in the water and colloid formation during waste-form or waste-package dissolution, are discussed in Sections 4.1.2 and 4.1.3.3, and Section 7.4, respectively.

Plutonium(IV) readily forms colloids under conditions that are similar to Yucca Mountain water (Olofsson et al., 1984) near-neutral solutions of low ionic strength (about 10^{-3} mole/L) (Newton and Rundberg, 1984). These colloids are solution like sols that are optically clear, show a characteristic absorption spectrum, and do not settle out of solution. Colloidal plutonium shows x-ray diffraction patterns similar to crystalline plutonium dioxide; higher-order lines are missing, indicating small crystalline size (20 to 30 angstroms). There is also some indication that americium(III) may form colloids under similar conditions (Olofsson et al., 1984). Neither element has been studied to the extent that colloid properties (particle size and stability) can be predicted for known solution conditions. Colloid properties are, however, being examined. Plans for obtaining this information are outlined in Section 8.3.1.3.5.

4.1.3.4.1.4 Radiolysis

Radiation can affect the solubility of waste elements by altering the composition of the water or by influencing the crystallinity of the solids that form (Ausloos, 1969). The primary effect of gamma radiation will be a reduction of water pH and a trend toward more oxidizing conditions as long as air is present; secondary effects will be the production of nitrate (or nitrite) anions. Gamma radiation will be most important early in the life of the waste and will affect only the water near the waste. By 300 yr after

discharge from the reactor, the total gamma flux (in watts) of spent fuel is about 500 times less than the flux at 10 yr after discharge (Croff and Alexander, 1980). Because containment would not be significantly breached before 300 yr (10 CFR 60.113(a)(1)(ii)(A)), the effects of gamma radiation will be reduced considerably. The effects of gamma radiation on the far field is negligible; however, the effects on waste element dissolution are discussed in Section 7.4.3.

Alpha radiation can affect water compositions in ways that are similar to the effects of gamma radiation. The primary effects of alpha radiation are a decrease in pH and a trend toward more oxidizing conditions in the water. Solids composed of alpha emitters tend to show self-irradiation damage to their crystal structure; the solubilities of solids like the actinide oxides and hydroxides are affected in that amorphous solids, which are generally more soluble than crystalline solids, are more likely to be the natural precipitation products (Rai et al., 1983; Nitsche and Edelstein, 1985).

4.1.3.4.2 Solubility and speciation data

Solubility and speciation data are available for a number of important waste elements. Although most of the data have not been measured under conditions that are characteristic of a Yucca Mountain repository, thermodynamic models of solubility and speciation allow solubility estimates for these conditions. This section describes data that can be used to estimate waste element solubilities at Yucca Mountain.

4.1.3.4.2.1 Solubility of americium, plutonium, and neptunium in well J-13 water

Preliminary measurements of the solubility of americium, plutonium, and neptunium in a neutral electrolyte and in water from well J-13 have been completed (Nitsche and Edelstein, 1985). The measurements were made from oversaturation at 25°C and a pH of 7. For plutonium and neptunium, measurements were started with different oxidation states that might exist under natural conditions. Table 4-17 summarizes the results.

Under the conditions of the measurements, neptunium(V) was the only oxidation state finally observed in the aqueous phase, for neptunium(V) or neptunium(VI) as the initial aqueous oxidation state. Measurements were done starting with plutonium(IV), plutonium(V), and plutonium(VI) aqueous species; the final aqueous species were predominantly plutonium(V) plus plutonium(VI) (over 90 percent), with plutonium(V) ranging from about 40 to 80 percent. Some difficulties were encountered in identifying the solid phases that precipitated in the experiments. For americium solubility in well J-13 water, both americium(III) hydroxy carbonate (AmOHCO_3) and americium(III) carbonate-2-hydrate ($\text{Am}_2(\text{CO}_3)_3 \cdot 2\text{H}_2\text{O}$) were found. For neptunium in well J-13 water, sodium neptunyl carbonate-n-hydrate ($\text{Na}_3\text{NpO}_2(\text{CO}_3)_2 \cdot n\text{H}_2\text{O}$) was tentatively identified as the solid. For plutonium, the solids could not be identified.

Table 4-17. Analytical results of solutions in equilibrium with their solid phase (supersaturated starting conditions) in 0.1 M NaClO₄ and well J-13 ground water at pH = 7.0 ± 0.1, 25 ± 1°C

Test no.	Initial species	Solubility (M)		Final oxidation state ^a of soluble species ^a		Solid	
		NaClO ₄	Well J-13	NaClO ₄	Well J-13	NaClO ₄	Well J-13
1	NpO ₂ ⁺	(4.4 ± 0.1) × 10 ⁻⁴	(1.6 ± 0.6) × 10 ⁻³	+V = 100%	+V = 100%	Amorphous	Crystalline Na ₃ NpO ₂ (CO ₃) ₂ x nH ₂ O
2	NpO ₂ ²⁺	(3.5 ± 0.1) × 10 ⁻⁴	(7.2 ± 0.7) × 10 ⁻⁴	+V = 100%	+V = 100%	Crystalline unidentified	Crystalline unidentified
3	Am ³⁺	(3.0 ± 0.1) × 10 ⁻⁴	(1.1 ± 0.2) × 10 ⁻⁶	+III = 100%	+III = 100%	Mostly amorphous	Crystalline Am ₂ (CO ₃) ₃ x 2H ₂ O/AmOHCO ₃
4	Pu ⁴⁺	(3 ± 2) × 10 ⁻⁸	(1.6 ± 0.2) × 10 ⁻⁶	III + IV + p = 3 ± 2% V + VI = 95 ± 4% V = 68 ± 9%	III + IV + p = 2 ± 3% V + IV = 98 ± 3% V = 40 ± 5%	Amorphous	Amorphous
5	PuO ₂ ⁺	(2 ± 1) × 10 ⁻⁹	(8 ± 3) × 10 ⁻⁶	III + IV + p = 5 ± 4% + VI = 95 ± 4% V = 70 ± 17%	III + IV + p = 0 ± 2% V + VI = 100 ± 2% V = 64 ± 6%	Amorphous	Crystalline unidentified
6	PuO ₂ ²⁺	(1.0 ± 0.5) × 10 ⁻⁷	(3 ± 2) × 10 ⁻⁵	III + IV + p = 3 ± 2% V + VI = 97 ± 2% V = 82 ± 16%	III + IV + p = 2 ± 2% V + VI = 98 ± 2% V = 67 ± 6%	Amorphous	Crystalline unidentified

^ap = Pu(IV) polymer (Nitsche and Edelstein, 1985).

The results from these experiments are being reviewed in terms of a chemical equilibrium model of solubility. They will be incorporated into the solubility models for Yucca Mountain.

4.1.3.4.2.2 Solid waste dissolution experiments

In addition to direct solubility measurements, dissolution experiments on spent fuel and other high-level waste forms such as borosilicate glass are also being performed. The measurements are described in Section 7.4.3 (see also Bazan and Rego, 1985; Wilson and Oversby, 1985). The experiments expose bare spent fuel or glass to well J-13 water and other water compositions and essentially approach steady state from undersaturation. Many of the experiments contain other materials such as zircaloy cladding, stainless steel, or tuff. Thus, the experimental data may be more difficult to interpret in terms of fundamental mechanisms than simple solubility experiments.

Waste-dissolution experiments under conditions characteristic of Yucca Mountain will be reviewed and the data will be used, where appropriate, to help define solubility limits. Plans for this work are outlined in Section 8.3.1.3.5.

4.1.3.4.2.3 Literature data for waste elements

The measurement of solubilities and other thermodynamic properties of important waste elements is an active field of study. Thus, the literature contains data that are useful for understanding and modeling solubility and speciation. Although these data were not obtained under conditions that are characteristic of a Yucca Mountain repository, solubility models for repository conditions can be tested and validated using literature data.

Americium. The behavior of americium in natural waters is somewhat simpler than that of the other actinides (e.g., plutonium or uranium) because americium(III) is the only stable oxidation state observed under these conditions. There have been a number of recent measurements of the solubility of americium(III) in a neutral electrolyte and with carbonate present (Kim et al., 1984b; Rai et al., 1983; Silva and Nitsche, 1983; Bernkopf and Kim, 1984; Rai and Ryan, 1984; Nitsche and Edelstein, 1985). The studies have also been used to define or support formation constant data for a number of important americium(III) complexes. Experiments by Lundqvist (1982) and Bidoglio et al. (1983) were aimed at defining formation constants for carbonate complexes with americium(III). Experimental data indicate that in a neutral electrolyte at 25°C and pH 7, americium has a solubility of about 10^{-4} to 10^{-6} mole/L, and in well J-13 water, its solubility is about 10^{-6} mole/L.

Americium thermodynamic data have been reviewed by Kerrisk (1984a). A thermodynamic data base for the EQ3/6 chemical-equilibrium computer model (Wolery, 1983) was developed. However, the review did not include a number of the more recent publications cited above. Two problems identified in the review were a conflict in the identity and formation constants of americium

complexes with carbonate and a lack of data above 25°C. The americium data base is being updated and will be reissued in the near future. The review and comparison with experimental data as they become available will continue as described in Section 8.3.1.3.5. The solubility calculations reported in this section have used the data base prepared by Kerrisk (1984a).

Plutonium. The solubility of plutonium in a neutral electrolyte and in carbonate solutions has been studied by a number of investigators (Rai et al., 1980; Rai and Swanson, 1981; Rai and Ryan, 1982; Kim et al., 1984b, 1984; Ogard et al., 1983b; Rai, 1984; Crowe and Vaniman, 1985; Nitsche and Edelstein, 1985). In a neutral electrolyte, the solid that appears to control solubility is hydrous plutonium dioxide (sometimes called amorphous plutonium dioxide or plutonium(IV) hydroxide). The solubility is much higher than that calculated for crystalline plutonium dioxide and may be an example of metastable equilibrium because crystalline plutonium dioxide, the most stable solid under the experimental conditions, does not precipitate from solution. Under oxidizing conditions, plutonium(V) and plutonium(VI) are the primary aqueous species. Thus, oxidation-reduction reactions are important in determining plutonium solubility. There is evidence that with sufficient concentrations of plutonium in solution, the plutonium(V)/plutonium(VI) concentrations are in equilibrium with a stronger oxidizing agent than dissolved oxygen (Rai, 1984); this may result from alpha radiolysis of the water and formation of significant amounts of hydrogen peroxide or some other radiolysis product. In a neutral electrolyte in contact with air, measured total plutonium concentrations in solution at pH 7 are in the range of 10^{-8} to 10^{-6} M.

From solubility measurements with carbonate present, Kim et al. (1984b) have proposed the existence of solid plutonium(IV) dihydroxide carbonate ($\text{Pu}(\text{OH})_2\text{CO}_3$) and a very stable complex, plutonium(IV) carbonate ion (PuCO_3^{+}). Their value of the formation constant for the reaction $\text{Pu}^{4+} + \text{CO}_3^{2-} = \text{PuCO}_3^{+}$ is $\log(K) = 47.1$. This is similar to a value of $\log(K)$ of 41 or less at 25°C for the same reaction reported by Lemire and Tremaine (1980). If the formation constant of the plutonium(IV) carbonate ion (PuCO_3^{+}) is this large, the solubility of plutonium in water from Yucca Mountain could be quite high (Wolfsberg et al., 1982a). However, estimates of the formation constant based on analogies with other actinides give a $\log(K)$ value of about 10 (Wolfsberg et al., 1982a). A recent experiment by Silva and Nitsche (1985), designed to find evidence of the existence of plutonium(IV) carbonate ion (PuCO_3^{+}), placed an upper limit on $\log(K)$ of 13. Analysis of data taken during the preparation of solutions of plutonium in well J-13 water for sorption measurements indicates an upper limit of plutonium solubility in well J-13 water of 10^{-6} mole/L (Ogard et al., 1983b). This limit is consistent with the measurements of Nitsche and Edelstein (1985) noted above and with recent calculations of plutonium solubility in well J-13 water using $\log(K) = 10$ for the formation constant of plutonium(IV) carbonate ion (PuCO_3^{+}) and hydrous plutonium dioxide as the solid controlling solubility (Ogard and Kerrisk, 1984). Although some evidence has been published for the existence of a stable plutonium(IV)-carbonate complex, more recent data seem to be contradictory. The question will continue to be investigated as described in Section 8.3.1.3.5.

Another question that has been raised about plutonium is the stability of the hydrolysis product plutonium(IV) pentahydroxide ion ($\text{Pu}(\text{OH})_5$) (Allard,

1983). The complex has not been observed, but was assigned a formation constant by analogy with uranium (Baes and Mesmer, 1976). Plutonium(IV) pentahydroxide ion becomes important at high pH; if it is stable, plutonium solubility could increase with increasing pH beyond a pH of about 7 or 8. It is not clear at this time whether this problem is of concern at Yucca Mountain. It will continue to be reviewed, as described in Section 8.3.1.3.5.

Uranium. The solubility of uranium has been studied extensively both in neutral electrolytes and also with carbonate present (Gayer and Leider, 1955; Blake et al., 1956; Miller, 1958; Robins, 1966; Sergeyeva et al., 1972; Silva and Yee, 1981; Krupka et al., 1984). In a neutral electrolyte under oxidizing conditions, uranium(VI) is the predominant aqueous oxidation state and schoepite ($\text{UO}_2(\text{OH})_2 \cdot \text{H}_2\text{O}$) is the most stable solid phase at 25°C. At higher temperatures beta-uranium dioxide may be more stable (Robins, 1966). Some of the early solubility measurements were difficult because of a problem with precipitation of sodium uranates at high pH because sodium hydroxide was used to control solution pH. Measured uranium solubilities are in the range of 10^{-3} to 10^{-6} mole/L at 25°C and pH 7; however, the higher solubilities probably result from inadequate phase separation.

Uranium solubility under oxidizing conditions with sufficient carbonate present results in rutherfordine (UO_2CO_3) as the precipitate (Blake et al., 1956; Sergeyeva et al., 1972). The experiments were conducted under much higher carbonate concentrations than those existing in Yucca Mountain water and thus these data are not indicative of uranium solubility at Yucca Mountain. They are, however, useful for testing solubility models with carbonate present.

Experiments on spent fuel dissolution in well J-13 water showed uranium concentrations in solution that are less than about 2×10^{-5} mole/L (Wilson and Oversby, 1985). At present it is not clear what solid phase is controlling uranium solubility at this concentration. Although the test solutions are exposed to air, uranium in spent fuel is primarily as uranium(IV), that is as uranium dioxide; thus, it is possible that a uranium solid other than schoepite controls solubility. This measured concentration limit is about a factor of 100 below the calculated solubility of uranium in well J-13 water assuming schoepite as the controlling solid (Ogard and Kerrisk, 1984). The calculations indicate that one polynuclear complex, uranyl(VI) trihydroxide carbonate ion ($(\text{UO}_2)_2\text{CO}_3(\text{OH})_3^-$), incorporates almost all the aqueous uranium. An experimental determination of uranium solubility in well J-13 water may be needed to resolve this difference. Plans for obtaining the information are outlined in Section 8.3.1.3.5.

Thermodynamic data for uranium(VI) have recently been reviewed by Tripathi (1984). The thermodynamic data base for EQ3/6 is being updated using this review as a basis.

Neptunium. Solubility data for neptunium were discussed in Section 4.1.3.2.1 and in the work of Nitsche and Edelstein (1985). Other literature data on the solubility of neptunium under oxidizing conditions will be reviewed as part of the incorporation of neptunium data. Plans for the work are outlined in Section 8.3.1.3.5.

Other waste elements. A number of waste elements are likely to have high solubilities in water from Yucca Mountain. They include carbon, cesium, iodine, and technetium (Rai and Serne, 1978; Allard and Torstenfelt, 1983; Apps et al., 1983; Ogard and Kerrisk, 1984). The solubility of technetium, which can exist in a number of oxidation states, should be high because of the likelihood of oxidizing conditions in Yucca Mountain water (Section 4.1.2). Solubility measurements for the elements listed above are not planned under repository conditions, because there is little expectation that solubility will limit their concentrations.

Carbon (carbon-14) can be released as carbon dioxide gas or directly as a carbonate under certain circumstances. In either case, there is an equilibrium between carbonate in the aqueous phase and carbon dioxide gas. There is also a large reservoir of natural carbon as carbonate dissolved in the aqueous phase that must be considered. Solubility data for the most likely carbonate solids (calcite or dolomite) are well known (Stumm and Morgan, 1981). Based on these data it is unlikely that solubility will limit carbon-14 concentrations in water from Yucca Mountain. Thus, no solubility measurements are planned for carbon. Two alkali-metal waste elements, strontium and radium, have similar chemistries but radium is expected to have a much lower solubility than strontium in water from Yucca Mountain (Ogard and Kerrisk, 1984; Crowe and Vaniman, 1985). Solubility measurements are planned for radium, whose sulfate solid may provide a useful limit on the aqueous concentration. No solubility measurements are planned for strontium. The solubilities of tin and thorium are also expected to be low in Yucca Mountain water (Apps et al., 1983); solubility measurements are planned for these elements because they are in large quantities and because they are considered as important radionuclides (Section 4.1.3.1.1). Radioisotopes of nickel are present as waste elements in subassembly hardware from light water reactors and in defense high-level waste glass. No solubility data were found for nickel that could be applied to water from Yucca Mountain; solubility measurements are planned as outlined in Section 8.3.1.3.5.

4.1.3.4.3 Solubility modeling

A program to model solubility and speciation data is an important part of any effort to assess the concentration limits of elements in solution. It is necessary to model experimental data in terms of some theoretical framework: to test the data for inconsistencies; to assess the sensitivity of the data to variations in independent parameters; and to allow interpolation and some extrapolation of solubilities beyond the specific conditions used in the experiments. A chemical equilibrium (thermodynamic) model, EQ3/6, is being used to model solubility and speciation data (Wolery, 1983).

Past efforts in modeling have concentrated on assessments of thermodynamic data and calculations that highlight the importance of particular aqueous species or solids for solubility in water from Yucca Mountain (Wolfsberg et al., 1982b; Ogard et al., 1983b; Rard, 1983; Kerrisk, 1984a; Crowe and Vaniman, 1985; Isherwood, 1985). Calculated solubilities and important aqueous species have been reported (Ogard and Kerrisk, 1984) and used for further analyses (Kerrisk, 1984b). This effort is expected to continue as described in Section 8.3.1.3.5.

It is uncertain whether nonequilibrium or kinetic models will be required to model the solubility and speciation of waste elements at Yucca Mountain. Colloid formation and rapid fracture flow are areas where such modeling may be needed. Until the formation of waste element colloids is better understood, the requirement for the models will remain an open question. Plans to review this question are outlined in Section 8.3.1.3.5.

4.1.3.4.4 Solubilities of waste elements in Yucca Mountain water

The solubilities of a number of waste elements in well J-13 water have been estimated from some experimental measurements and some equilibrium-thermodynamics calculations. Table 4-18 lists the estimates along with their source. The water was assumed to be at pH 7, 25°C, and oxidizing.

Ogard and Kerrisk (1984) reported calculated solubilities for six waste elements (uranium, plutonium, americium, strontium, radium, and technetium) in well J-13 water. The solubilities of two of these elements (americium and plutonium) were subsequently measured in well J-13 water by Nitsche and Edelstein (1985). The measured solubility of americium was about 100 times higher than the calculated solubility. The EQ3/6 thermodynamic data base for americium is presently being revised. The measured solubility of plutonium varied with the starting material but the average value was about five times higher than the calculated solubility. For uranium, Table 4-18 shows the value calculated by Ogard and Kerrisk (1984) for uranium(VI). This solubility is about 100 times higher than measured uranium concentrations in the same water over spent fuel (Section 7.4.3 and Wilson and Oversby, 1985). The reason for this difference is uncertain. At this time there are no other direct measurements of waste-element solubilities in well J-13 water.

Ogard and Kerrisk (1984) also reported calculated solubilities of the same six elements in waters from two other drillholes in the vicinity of Yucca Mountain (UE-25p#1 and USW H-3). The water from these drillholes has somewhat different compositions than water from well J-13 (Tables 4-6 and 4-7); the calculated solubilities were significantly different in a number of cases. Particularly, the water from drillhole USW H-3 was assumed to be reducing. No waste element solubility measurements have yet been made in waters from drillholes UE-25p#1 and USW H-3. Future experimental solubility measurements are described in Section 8.3.1.3.5.

4.1.3.5 Matrix diffusion

Moving ground water would provide the principal means of transporting radionuclides from the repository. Consequently, an understanding of water movement through tuff formations is vital. The diffusion of radionuclides from fracture water into the pore water in the rock matrix is important for retarding the transport of dissolved radionuclides through fractures in tuff, particularly for nonsorbing or low-sorbing soluble species. The apparent diffusion coefficient for a given radionuclide in tuff matrix depends on properties that are intrinsic to the chemical species (e.g., ion mobility) and properties of the tuff (such as porosity, tortuosity, and sorption

CONSULTATION DRAFT

Table 4-18. Solubility of waste elements in well J-13 water^a

Waste element	Solubility (moles/L)	Source of data
Americium	1×10^{-6}	Nitsche and Edelstein, 1985
Plutonium	1×10^{-5}	Nitsche and Edelstein, 1985
Neptunium	1×10^{-3}	Nitsche and Edelstein, 1985
Uranium	4×10^{-3}	Ogard and Kerrisk, 1984
Curium	$^b_1 \times 10^{-6}$	Apps et al. 1983
Thorium	$^b_1 \times 10^{-9}$	Apps et al. 1983
Strontium	8×10^{-4}	Ogard and Kerrisk, 1984
Radium	3×10^{-7}	Ogard and Kerrisk, 1984
Carbon	High ^{b,c}	Kerrisk, 1984b
Cesium	High ^{b,c}	Rai and Serne, 1978
Technetium	High ^{b,c}	Ogard and Kerrisk, 1984
Iodine	High ^{b,c}	Apps et al. 1983
Tin	$^b_1 \times 10^{-9}$	Apps et al. 1983
Zirconium	$^b_1 \times 10^{-10}$	Rai and Serne, 1978
Samarium	$^b_2 \times 10^{-9}$	Benson and Teague, 1980
Nickel	$^b_1 \times 10^{-2}$	Benson and Teague, 1980

^aAt pH of 7, 25°C, and oxidizing conditions.

^bEstimated data.

^cSolubility of these waste elements is expected to be high enough so that solubility will not limit concentration at a Yucca Mountain site.

ratio). It is therefore necessary to measure the diffusion coefficient of waste-element species in various unsaturated and saturated fractured tuff units. Preliminary results from experiments on radionuclide diffusion in tuff matrix are described in detail by Daniels et al. (1982). The information presented here is based on that report. Further work on the dynamic transport of radionuclides is planned in Section 8.3.1.3.6. This work will encompass diffusion studies in saturated and unsaturated fractured tuff.

4.1.3.5.1 Diffusion into the rock matrix

In the current conceptual model of the hydrology of Yucca Mountain, water is believed to flow predominantly in the matrix of the tuffs in the unsaturated zone, and predominantly in the fracture networks of the tuffs in the saturated zone. The effective diffusion coefficient of a dissolved species in the tuff matrices is an important quantity in determining the rates at which that species is transported through the rock units of either zone. If the flux through unsaturated rocks is small (less than 0.05 mm/yr), ordinary diffusion will be the dominant transport mechanism, and measurements of the effective diffusion coefficient will be needed in order to model the transport of radionuclide-bearing compounds to the accessible environment. For the flows through the fracture networks of the saturated zones, the effective diffusion coefficient in the matrix is one of several quantities that determine the rate at which radionuclides dissolved in rapidly flowing fracture water can enter the pore water of the matrix, where chemical retardation is generally most pronounced; this is the "matrix diffusion" mechanism alluded to earlier in this section. Thus, the effectiveness of retardation in the transport of solutes through fracture networks also depends on the effective diffusion coefficient.

Diffusion through the liquid-filled pore spaces of the matrix may proceed so rapidly that the assumption of local chemical equilibrium is not valid. Thus, knowledge of the rates of the chemical reactions involved in sorption on the solid phases of the matrix is necessary to determine experimentally effective diffusion coefficients for dissolved species. The kinetics of the sorption reactions are also of intrinsic interest in determining the circumstances under which assumptions of chemical equilibrium (and therefore constant geochemical retardation) are valid.

Experiments were performed (Rundberg, 1985) to determine the diffusion coefficients and rate constants for the sorption of strontium, cesium, and barium in samples of Yucca Mountain tuff from drillhole USW G-1. The uptake of radionuclides with time was determined for tuff disks 2.54 cm in diameter and 2.0 mm thick. The data were then fitted to the one-dimensional diffusion equation for a plane sheet coupled to a linear reversible reaction with the solid phase, defined as follows:

The diffusion is governed by the equation

$$\frac{\partial C}{\partial t} = D \frac{\partial^2 C}{\partial x^2} - \frac{\partial S}{\partial t} \quad (4-5)$$

with the simultaneous reaction of the type

$$\frac{\partial S}{\partial t} = k_1 C - k_2 S \quad (4-6)$$

where

C = the concentration of the solute free to diffuse within the sheet

S = the concentration of the immobilized solute

D = the diffusion coefficient

x = the position in the disk

t = the time

k_1, k_2 = the rate constants of the forward and backward reactions.

The effect of slow reaction on the diffusion of a sorbing species into tuff is demonstrated in Figure 4-15. The uptake of cesium with a R_d of 230 ml/g and a diffusion coefficient of 8.33×10^{-6} cm²/s, assuming instantaneous or diffusion-limited sorption, is given by the broken line. The solid line is the best fit to the experimental data, using the solution to equations 4-5 and 4-6 given by Crank (1975). It is evident that the conventional treatment (instantaneous sorption) is inadequate. The shape of the curve does not match the experimental uptake well on a log-log plot, so varying the diffusion coefficient or the R_d value will not improve the fit. The solution with slow kinetics, however, appears to give an excellent fit to the experimental uptake. At early times, the discrepancy between the slow kinetics and the conventional treatment becomes quite large.

The rate constants for the sorption of cesium-137, strontium-85, and barium-133 are summarized in Table 4-19 with the diffusion coefficients determined by fitting the data to the Crank (1975) solution. The diffusion coefficients are all smaller than the ionic diffusion coefficient calculated from ionic mobilities. This is as expected because the pores in the tuff matrix represent a tortuous path that also may have constrictions to slow the diffusion of ions through the matrix. Details of the above analysis are given by Rundberg (1985). Plans for obtaining additional information are outlined in Section 8.3.1.3.6.

4.1.3.5.2 Retardation by matrix diffusion during fracture flow

Experiments comparing radionuclide transport by fracture flow in tuff (Daniels et al., 1982) with that in granite have provided a test of the validity of simple matrix-diffusion models. The principal purpose of the studies was to determine the degree of radionuclide retardation provided by geochemical processes (sorption) coupled to matrix diffusion along the flow path in a single fracture. The results provide information about the effectiveness of both diffusion and sorption processes in retarding radionuclides in rapidly flowing single-fracture systems. These results have been compared with transport-model predictions to partially validate the models.

The experiments employed artificial fractures induced in tuff and granite core samples. Artificial fractures are hydraulically smooth; thus, the

M/M_∞ = MASS OF SOLID OVER TIME (T) / MASS OF SOLID (∞)

DT/L^2 = DIMENSIONLESS TIME WHERE D = EFFECTIVE POROSITY,
T = TIME, AND L = THICKNESS OF TUFF WAFER

K_d = DISTRIBUTION COEFFICIENT (ASSUMED TO BE EQUAL TO
 R_d (SORPTION RATIO))

η = REACTION RATE OF DESORPTION/DIFFUSIVITY

D_e = DIFFUSION COEFFICIENT

Figure 4-15. Cesium diffusion with slow sorption (tuff sample number G1-1883). Modified from Rundberg (1985).

CONSULTATION DRAFT

Table 4-19. Sorption rate constants for tuff from drillhole USW G-1^{a,b}

Tuff sample	Radionuclide	K_d (ml/g)	k_1 (s ⁻¹)	k_2 (s ⁻¹)	D (cm ² /s)
G1-1436	Cesium	7,790	6.3×10^0	1.4×10^{-4}	2.78×10^{-6}
	Strontium	36,300	3.5×10^0	2.0×10^{-5}	2.11×10^{-6}
	Barium	148,000	1.2×10^1	1.7×10^{-5}	3.38×10^{-7}
G1-1883	Cesium	230	5.9×10^{-2}	4.2×10^{-5}	1.33×10^{-6}
	Strontium	27	1.1×10^{-3}	6.7×10^{-6}	1.33×10^{-5}
	Barium	210	5.7×10^{-2}	4.4×10^{-5}	8.89×10^{-6}
G1-1982	Cesium	1,000	4.0×10^{-1}	4.7×10^{-5}	9.44×10^{-6}
	Strontium	88	9.5×10^{-3}	1.4×10^{-5}	1.37×10^{-5}
	Barium	800	2.6×10^{-1}	4.4×10^{-5}	8.89×10^{-6}

^aSource: Rundberg (1985).

^b K_d = retardation coefficient; k_1 = rate constant for forward reaction; k_2 = rate constant for reverse reaction; D = diffusion coefficients.

pieces mate well and do not have fracture-fill material that would complicate the surface chemistry and diffusion process. Experiments were performed on partially welded tuff samples G1-2335 and G1-2840 from the Bullfrog and the Tram members, respectively. Water was pumped through the fractures at the relatively high velocity of 2.8×10^{-2} cm/s. The fracture apertures were estimated by applying Darcy's law to a series of permeability measurements (Freeze and Cherry, 1979). The effective apertures in both tuff samples were approximately 30 micrometers. The effective aperture in the granite sample was 16 micrometers. Water containing strontium-85 and cesium-137 tracers was monitored as it passed through the fractures.

The analytical solution proposed by Neretnieks (1979) for a one-dimensional fracture with matrix diffusion was applied to the data from the experiments. The granite data fit the solution with experimentally determined R_d values. However, the results of the fracture-flow experiments on tuff did not agree with those of flow calculations that used experimentally determined R_d values. The R_d values that gave the best fits to the breakthrough portion of the strontium elution in the fracture-flow experiments were 30 and 16 ml/g for samples G1-2335 and G1-2840, respectively. The R_d values from batch sorption measurements were 148 and 160 ml/g for the same samples. The shape of the elution curve calculated for the tuffs also does not agree with the observed elution data. The radionuclide activity desorbs more slowly than would be expected for reversible diffusion-controlled sorption. This observation is consistent with previous measurements of batch desorption on tuff.

The two important conclusions that can be drawn from these fracture-flow

studies are as follows:

1. Matrix diffusion is an important mechanism contributing to the retardation of radionuclides in fracture flow; however, simple analytic models may not be adequate to predict the transport of waste elements in tuff fractures.
2. Because matrix diffusion is directly related to porosity, it should more effectively retard the movement of soluble species in porous tuffs than in low-porosity crystalline rock like granite or densely welded, devitrified tuff.

Plans for further work on matrix diffusion coupled with fracture flow are given in Section 8.3.1.3.6.

4.1.3.6 Radionuclide transport

Ground water can carry material along in solution or as suspended solids. The rate at which the transported material moves is affected by a variety of factors, the most important being the velocity of the flowing water, and the partitioning of the material between liquid and solid phases. Other factors include:

1. Chemical species in the water flowing in fractures diffuse into the water contained in rock pores.
2. The negative charge that is present on many rock surfaces can restrict the approach of anions. The exclusion may limit the diffusion of anions into the matrix thereby allowing the anions to move at the higher velocity of the water flowing in the center of the fractures, or channels, away from the surface film. The same phenomenon can restrict the entry of anions into the smaller pores.
3. A somewhat similar effect, called hydrodynamic chromatography, can result in particles flowing along pores at velocities somewhat greater than the average water velocity. Filtration in rock pores, however, can remove or retard the flow of suspended particulates.
4. Dispersion can result from diffusion, channeling, and turbulent flow, but dispersion by itself does not affect the average rate at which the transported material moves.

Gaseous phase transport is another means of transporting radionuclides. Volatile elements such as iodine-129 and carbon-14 can be transported in the gaseous phase in the unsaturated zone.

Previous portions of Section 4.1.3 have discussed radionuclide transport in terms of the retardation processes of sorption, precipitation, and diffusion. The remainder of this section will discuss radionuclide transport by suspended solids and by gaseous transport.

4.1.3.6.1 Transport of suspended solids

A possible mode of transport involves the movement of radioactive particles suspended in the ground water. Colloidal particles (less than a few micrometers in diameter) remain suspended for long periods and hence may migrate with the ground water. As the solid waste form is leached, particles containing radionuclides may form by the sorption of dissolved radionuclides on nonradioactive particles. At this time it is believed that plutonium and americium are likely to contribute most to transport as colloids.

To estimate the amount of radionuclides that could be transported by colloidal suspension, it is necessary to determine whether colloidal-sized particles that could be sorption sites for waste elements exist in the ground water of Yucca Mountain. Then, the sorption ratios for waste elements on these particles must be measured or estimated from the composition of the particles. In addition, the conditions under which colloids could form from the waste elements or from the waste and their stability after formation must be determined (Section 4.1.3.4.1). Finally, the conditions necessary for filtration of the particles by the tuff itself must be defined.

Matter in the colloidal state has a large surface area ($300 \text{ m}^2/\text{g}$), thus it is not surprising that the most important properties of colloids are those that depend on surface interactions; such as adsorption. Drever (1982) discusses the nature and geochemistry of colloids, with emphasis on the charge surrounding colloids and its effect on suspension stability.

Olofsson et al. (1981, 1982a,b) classify radiocolloids (colloids containing radionuclides) as true colloids or pseudocolloids depending on their formation process. True colloids are formed by condensation of molecules or ions as a result of hydrolytic or precipitation processes. Colloids consist mostly of hydroxides or polymers formed by hydrolysis, and they have a very rapid formation rate. Pseudocolloids, on the other hand, are formed as a result of adsorption on impurities in the solution and tend to be much larger than true colloids (up to 5,000 angstroms). Pseudocolloids can be of two types, reversible and irreversible. The formation rate of pseudocolloids is basically determined by the sorption rate on colloidal impurities (Olofsson et al., 1982a,b).

In the event of a repository failure, radiocolloids are believed to be a significant factor for the transport of radionuclides, and might actually accelerate their transport away from the repository (Avogadro and Lanza, 1982; Avogadro and De Marsily, 1984). Radiocolloids may arise from a variety of sources. The corrosion of canister material can lead to the formation of absorbent colloids. Degradation of engineered backfills at the end of a repository's active life may also lead to colloid formation. If the waste form is leached by ground water, naturally occurring colloids derived from smectites, vermiculites, illites, kaolinite, and chlorite present in ground water may also adsorb radionuclides. Champ et al. (1982) have demonstrated experimentally the existence and rapid transport of plutonium colloids using core samples and ground water. Fried et al. (1975), in their 1975 laboratory studies, observed a fast migrating form of plutonium in their column experiments on Bandelier Tuff. Retardation of this polymeric form of plutonium was only one-tenth of the ionic form. A discussion of radiocolloid formation is found in Section 4.1.3.4.1.3.

Transport of particulates in geologic media will depend on aqueous flow rate, on pore and fracture size in the rock, on ions carried with the water, and on the nature of the particulate matter. Several mechanisms may remove colloidal particulates from ground water such as mechanical filtration by the rock matrix, sorption on the surface of rock pores (van der Waals), and neutralization of the repulsive charges on the colloids thus allowing them to coagulate.

Filtration will be an effective barrier to transport in the tuff matrix. Pore size is an important element in filtration. Pseudocolloids will not flow through pores that are smaller than the pseudocolloid carrier particles. In very small pores, true colloids will have a high probability of entering boundary layers around matrix particles, where they can be captured in the small pores. Pore-size distributions have been measured (Daniels et al., 1982) for several Yucca Mountain tuff samples. Mercury porosimetry data on a sample of Topopah Spring tuff show that most of the pores are less than 1.0 micrometer in diameter. There is a small percentage larger than 1.0 micrometer, but they should be dry because the amount of Topopah Spring saturation is less than 90 percent. More than 50 percent of the pore volume is made up of pores 0.1 micrometer in diameter or smaller. For samples from the Calico Hills, 50 percent of the pores are smaller than 0.05 micrometer, but there are somewhat more of the largest pores (4 micrometers). The data indicate that to pass through pores in the tuff matrix, colloids must be on the order of 0.1 micrometer or less.

Pore-water velocities, assuming matrix flow, are estimated to be less than 1.0 cm/yr (Travis and Nuttall, 1987). It is not clear whether such low velocities can provide sufficient momentum to allow bulk colloid movement. If fracture flow is significant, then colloid transport may be likely. If fracture flow is dominant in the Topopah Spring tuff, the average fracture-flow rate would be roughly 2 m/yr (Travis and Nuttall, 1987); this is still only on the order of transfer rates by molecular diffusion. If fracture flow is episodic, corresponding to a few intense rainfalls per year, flow rates in fractures might be much larger (tens of meters per year) but of shorter duration.

In addition, colloids will be subject to gravitational settling (Travis and Nuttall, 1987) for particles larger than about 0.1 micrometer. The transporting ability of very small colloids will be diminished by diffusion of radionuclides out of the carrier particles. Diffusion in solid phases is small, but not zero. Solid phase diffusion for some radionuclides on several Yucca Mountain minerals has been measured at about 10^{-16} to 5×10^{-13} cm²/s (Rundberg, 1985). A characteristic diffusion time (T) from 0.1 micrometer diameter particles is given by

$$D \approx \frac{R^2}{T} = \frac{L^2}{4T} \rightarrow T \approx \frac{L^2}{4D} \approx \frac{(10^{-5})^2}{4 \times 10^{-16}} = 2.5 \times 10^5 \text{ s} \approx 3 \text{ days} \quad (4-7)$$

where

- D = diffusion coefficient
- T = diffusion time
- L = particle diameter
- R = particle radius.

Small pseudocolloids, in fluids of low ionic concentrations, should not be able to retain absorbed radionuclides for more than a few days. Very small true colloids may be able to migrate rapidly if fracture flow is occurring; however, more experimental work is needed to determine the importance of true colloid transport. Plans for obtaining the additional information required are outlined in Section 8.3.1.3.

One additional mechanism that might allow transport of some radionuclides are microbes (West and McKinley, 1984). Studies measuring the uptake of radionuclides by bacteria and studies determining the mobility and radionuclide transport capability of bacteria in tuff are detailed in Section 8.3.1.3.

Because natural particulate material in ground water at a repository site may be significant as a means of transport of radionuclides by sorption on the particulate material, it is necessary to measure the ambient concentration of such particles. At Yucca Mountain there are no springs or other accessible sources of flowing ground water. Therefore, well J-13, a water well adjacent to Yucca Mountain, was pumped and the water was sampled for particulates by filtering several hundred liters of water through 0.05 micrometer Nuclepore filters, then examining the filters with a scanning electron microscope. An abundance of material was present on the filters, ranging in size from tens of micrometers in diameter down to less than 0.01 micrometer, the resolving power of the microscope. The composition of individual particles varied considerably, although most contained potassium, aluminum, and silicon as major constituents. Filter samples also were obtained from two other wells at the Nevada Test Site, both drawing from an aquifer in alluvial soil. The results of subsequent filter tests were similar to those from well J-13. Because the samples were derived from water flowing in an artificially created gradient, it can be argued that the particulate content is not that of natural water. However, pumped well water may provide a worst-case situation. Future experimental work is planned to determine whether particulates in well J-13 water can sorb a significant amount of dissolved radioactive material and whether these particulates are readily transported through crushed and fractured tuff. Plans for obtaining the additional information required are outlined in Sections 8.3.1.3.4 and 8.3.1.3.6.

4.1.3.6.2 Gaseous transport

A limited number of radionuclides can form volatile species that are capable of being transported in a moving vapor or gas. Among these are tritium, carbon-14, and iodine-129. In the far field, factors that affect transport in flowing ground water also affect transport in flowing gas, (i.e., the velocity of the gas determines the potential for advective transport). In the absence of flow, diffusion is the only mechanism for transport in the gaseous state. The processes of partitioning of the volatile species between the gaseous, liquid, and solid state and isotopic exchange must also be considered when assessing the impact of gaseous transport.

The probability of gaseous transport will be investigated to determine its applicability to Yucca Mountain. Presently, no work is actively being

done on gaseous transport; however, plans for obtaining information required to assess the probability of gaseous transport and the processes of exchange (i.e., gaseous phase and aqueous/solid) are outlined in Section 8.3.1.3.8.

4.1.3.7 Geochemical retardation in the host rock and surrounding units-- anticipated conditions

In the previous sections the principal processes that affect and retard radionuclide transport in ground water were discussed. Through the use and application of numerical models the data on sorption properties, matrix diffusion, flow rates (advection), radioactive decay, colloids speciation, and precipitation and dissolution can be synthesized with numerical models. Numerical models use the data concerning the processes as input and then calculate the total radionuclide transport and retardation through Yucca Mountain.

Currently available computer models can account for most of the known physical and geochemical transport processes (advection, diffusion, dispersion, matrix flow, fracture flow, and equilibrium sorption). More complex geochemical processes, such as speciation, dissolution and precipitation, and colloid formation have only very recently been incorporated into transport models. Plans for refining and improving already developed models, and for developing models that will incorporate all significant processes is outlined in Section 8.3.1.3.7. The development of such models is dependent upon results obtained from Investigations 8.3.1.3.1 through 8.3.1.3.6 and the geohydrology program (Section 8.3.1.2). In turn, the models can be used to interpret results arising from these information needs and to validate the assumption made in constructing the systems models for Information Needs 1.1.4 and 1.6.2; the models may also be useful in establishing the presence of favorable and potentially adverse conditions (Issue 1.8); and making the higher-level findings required by 10 CFR Part 960 (Issue 1.9).

One model that can be used to simulate the transport of radionuclides from the repository to the water table is TRACR3D (Travis, 1984; Perkins et al., 1985). As an example of model calculations under anticipated conditions, the transport of technetium-99 and uranium-238 was calculated (Vaniman, 1987). The processes assumed to affect transport in this example are advection, sorption, dispersion, matrix diffusion, and radioactive decay.

The relative importance of these processes is shown in Figures 4-16a, -16b and 4-17a, -17b. Figures 4-16a and 4-16b show the results of TRACR3D calculations after a 10,000-yr pulse of technetium-99 is introduced into a uniform vertical flow field of (1) 4.5 mm/yr and (2) 0.5 mm/yr. The normalized mass flux entering the water table is plotted against time. The expected flux has been shown to range from 0.1 to 4.5 mm/yr (Montazer and Wilson, 1984). If simple advection is the only transport process operating, the technetium-99 pulse will arrive at the water table without attenuation (the square pulses). Technetium-99 has a long half-life and is only slightly sorbed onto the Yucca Mountain tuffs. Dispersion will be an important attenuating mechanism (by spreading out the pulse and reducing amplitude) followed by matrix diffusion. In the calculations, a dispersivity of 10 cm was used. Comparison of Figures 4-16a and 4-16b show the magnitude of this effect. So

Figure 4-16a. The movement of a 10,000-yr pulse of technetium-99 for $V = 4.5 \text{ mm/yr}$. Modified from Vaniman (1987).

Figure 4-16b. The movement of a 10,000-yr pulse of technetium-99 for $V = 0.5 \text{ mm/yr}$. Modified from Vaniman (1987).

U-238 NORMALIZED MASS FLUX
ENTERING WATER TABLE

Figure 4-17a. Uranium-238 normalized mass flux entering water table for $V = 4.5 \text{ mm/yr}$. Modified from Vaniman (1987).

U-238 NORMALIZED MASS FLUX
ENTERING WATER TABLE

Figure 4-17b. Uranium-238 normalized mass flux entering water table for $V = 0.5 \text{ mm/yr}$. Modified from Vaniman (1987).

even where a small dispersivity is used, attenuation occurs. Also, in Figure 4-16b for the case of a slower flow rate, dispersion has a greater effect in attenuating the pulse because of the longer travel time. Therefore, for a conservative, nonsorbing radionuclide such as technetium-99, a slower flow rate will result in a longer travel time to the water table and a more attenuated, broadened pulse.

Figures 4-17a and 4-17b show the effects of sorption, dispersion, and diffusion on the transport of uranium-238. Again a 10,000-yr pulse is introduced in a uniform vertical flow field of (1) 4.5 mm/yr and (2) 0.5 mm/yr. Uranium-238 adsorbs readily to the Yucca Mountain tuffs. In Figure 4-17a, with the larger flow rate, sorption, dispersion, and diffusion act to delay the arrival of the peak by a factor of 10 and decrease the peak concentration by almost a factor of 10. When the flow rate is decreased as in Figure 4-17b, sorption, dispersion, and diffusion act to delay the arrival of the peak to the water table by a factor of 200 and decrease the peak concentration by more than a factor of 100.

It has been shown in this example calculation that under anticipated conditions the combined processes of sorption, dispersion, and diffusion have a significant effect on the transport of sorbing and nonsorbing radionuclides. Therefore, it is extremely important that these processes and their relative parameter values are well-defined. As the processes and their relative parameter values become updated more calculations similar to the examples given above will be done to investigate the sensitivity of radionuclide transport to sorption, dispersion, and diffusion as outlined in Section 8.3.1.3.7. The investigation into the sensitivity of radionuclide transport calculations to speciation, precipitation/dissolution, and colloid formation is planned and outlined in Section 8.3.1.3.7. Estimates of total radionuclide releases to the accessible environment based on transport calculations can only be completed after the sensitivity of radionuclide transport to geochemical processes is better known and more information concerning parameter values is obtained.

4.1.3.8 Geochemical retardation in the host rock and surrounding units--unanticipated conditions

To investigate the effects of unanticipated events on geochemical retardation into the host rock and surrounding units the following questions must be answered:

1. What is the potential for and likely magnitude of the unanticipated events?
2. What is the effect of these unanticipated events on such processes as sorption, diffusion, and dispersion?
3. What is the impact on the overall integrated transport due to the changes in the transport processes?

Examples of unanticipated events are an increase in water-transmission rates, changes in the water chemistry, changes in the water table, transformations of zeolites and clays to less sorptive minerals, and changes in permeability. Plans for obtaining the additional information required to determine the effects of these and other unanticipated conditions on transport processes are outlined in Section 8.3.1.3.

An example is given below to illustrate the interactions between the unanticipated condition, the effect on transport processes, and the effect on radionuclide migration. Preliminary calculations have been conducted to investigate the effect of an increase in water transmission rates on radionuclide transport. An increase in water transmission rate could occur, for example, from increased precipitation or by an elevation of the water table.

An increase in the water-transmission rate might result in fracture flow in the unsaturated zone, but in the absence of fractures, flow and transport will be diffusive. At the recharge rate in Yucca Mountain of a few millimeters per year (assuming current climatic conditions), radionuclides would not reach the water table via vertical porous flow for roughly 10,000 to 20,000 yr or more after the release (Travis et al., 1984). Computer simulations of transport in simple fracture systems using extreme values for flow rates (full saturated permeability applied constantly rather than intermittently) show that to get a travel time of less than 10,000 yr requires a water flux at least 100 times greater than the present water flux (Travis et al., 1984).

For radionuclide transport through fractures, more information is needed on fracture flow in Yucca Mountain tuffs. For example, additional information is needed to establish whether the recharge water moves vertically through unfractured layers and then enters fractures, or if water moves through fractures as a film on the fracture surface or as a pulse fully saturating the fracture. Film flow would be very much slower than slug flow (Travis et al., 1984). Plans for obtaining this information are outlined in Section 8.3.1.3.6.

Other factors that cannot be evaluated at present because of the lack of data are the connectivity of fractures and the dimensions of fracture apertures. If fractures are present but not connected or intersecting, their impact on flow and transport will be greatly diminished. If fractures are a connected network, the path that a water slug will travel will probably be tortuous and might involve branching into cracks in other directions. In regard to the dimensions of apertures, fracture flow depends on fracture permeability, which varies roughly as the cube of the aperture, a strong sensitivity. Plans for obtaining the required information about fracture aperture and fracture permeability are outlined in Section 8.3.1.3.6, and plans for obtaining such information for a hydrologic model are given in Section 8.3.1.4.

Given fracture flow, the contaminant will diffuse into the pore spaces of the rock. Because of this diffusion into the matrix, the effective flow rate for radionuclides must be modified (Sudicky and Frind, 1982). The modification introduces a retardation factor due to a matrix diffusion from fractures. One can then calculate an equivalent retardation value, applicable only when fracture flow is occurring, combining retardation through

sorption with retardation through diffusion. The calculated equivalent retardation factors are a factor of 10 to 100 larger than those retardation factors based on sorption alone (Travis et al., 1984).

From the preceding discussion it is evident that a change in one condition such as the water-transmission rate leads to a change in flow characteristics. The characteristics of the flow in turn will affect retardation processes such as sorption and finally the overall retardation. Plans for obtaining the impact on the overall, integrated transport due to the changes in the transport processes are outlined in Section 8.3.1.3.7.

Factors that may affect radionuclide transport include: (1) fracture fill, (2) nonequilibrium sorption, (3) radionuclide sorption in fractures, (4) colloid transport, and (5) geometric dispersion. Plans for obtaining this information are outlined in Section 8.3.1.3.

4.2 GEOCHEMICAL EFFECTS OF WASTE EMPLACEMENT

The thermal pulse from waste emplacement and its effects on water displacements and geochemistry must be characterized for intervals extending tens of meters away from the waste containers emplaced in a repository. Discussions of the interactions within the engineered barrier system are presented in Chapter 7.

This section describes the time-dependent temperature profiles for a reference repository in unsaturated devitrified tuff, with the attendant phenomena of drying near the waste containers and water condensation at boundaries that move inward as the thermal pulse from waste emplacement decays. Also discussed is the possible hydrothermal alteration due to the thermal pulse--a prospect that is not critical to the devitrified host rock but may be important to the hydrous alteration minerals and glasses beneath the host rock. Desiccation and rehydration of the hydrous alteration minerals, as well as other phase transitions that may affect permeability, rock strength, and radionuclide retardation, must be studied as well. Water chemistry as a function of temperature must also be known; therefore, an assessment of thermal effects on unsaturated-zone pore water and on recharge water is important. Finally, this section discusses the thermal dependence of factors important on constraining radionuclide migration--namely, sorption, precipitation, dispersion, diffusion, and advection on unsaturated tuff.

4.2.1 ANTICIPATED THERMAL CONDITIONS RESULTING FROM WASTE EMPLACEMENT

This section discusses the impact of expected thermal conditions resulting from waste emplacement. Waste emplacement leads to heat buildup due to radioactive decay. This heat influences the rate of the waste container degradation, and the compositions of the surrounding tuff due to hydrologic changes.

The magnitude of the effects resulting from the thermal pulse depends on (1) heat content of the waste, (2) rate of decay of the waste, (3) amount of pore water, (4) effective permeability of the tuff, (5) waste emplacement geometry, and (6) presence or absence of venting. These factors are discussed in detail in Section 7.4.1. However, time-temperature calculations described in the sources only extend out to 1 m from the borehole wall.

The following is an example of the type of calculation that can be done to investigate the qualitative behavior of the anticipated thermal conditions resulting from waste emplacement. Preliminary numerical calculations using a flow and transport code (WAFE) that incorporates a thermal component (Travis, 1985) have been made. Using a heat load of 50 kW/acre, Figure 4-18 shows a fluid temperature profile around a 2 by 6 m room after 50 yr. The absolute numbers shown should be considered preliminary because of the uncertainties in the rock property values. However, in these calculations, as heat diffuses away from the source, water vaporizes. The water vapor is then driven outward until it reaches cooler rock and then condenses. Temperatures are buffered by the boiling process because energy is going into vaporizing water rather than raising the rock temperature. Peak temperatures at the wet edge of the boiling region are only about 140°C, considerably lower than peak waste temperature. After a century or so, temperature and saturation profiles decay back to ambient. A temperature increase of 5°C or more extends to at least 30 m (Travis et al., 1984).

This same heat loading or thermal scenario can in turn influence the rate of breakdown of the waste container. For example, if heat loading were to cause vaporization of pore water near each container, the atmosphere surrounding a container could change from wet air to a dry-steam, oxygen-poor atmosphere, which would affect the degradation of the container. The problem is discussed in Section 7.4.

The presence of fractures in the rock can alter the thermally induced flow patterns and make the interpretation of the effects of anticipated thermal conditions on the natural environment more complicated. Many small cracks will increase the overall permeability, making a convective or circulating flow pattern more likely. However, few large cracks that promote conduction may prevent convective circulation, and more likely provide a path for a fairly rapid movement of water, vapor, and air. Other factors could also be important, such as the thermomechanical behavior of cracks. Heat may close cracks or open them depending on the mechanical properties of the host rock. Plans for obtaining the additional results on these scenarios are outlined in Section 8.3.1.3.7.

The sections that follow discuss the effects of the thermal pulse on rock alteration, water chemistry, and radionuclide retardation. As future experimental results are obtained, the data below will be updated. Plans for obtaining the information required for evaluating anticipated thermal conditions are outlined in Section 8.3.1.3.

Figure 4-18. Fluid temperature profile showing temperature contours 50 yr after emplacement of a 50 kw/acre heat load using the WAFE code. Dots generated by plotter indicate the direction of the gradient. Modified from Travis et al. (1984)

4.2.2 HYDROTHERMAL ALTERATION DUE TO THE THERMAL PULSE

Flow paths near the repository will pass through a zone in which temperatures are elevated above the ambient geothermal gradient. The elevated geothermal gradient may give rise to temperature-induced reactions that have the potential for altering various chemical and physical properties of the rocks in Yucca Mountain. In the repository environment, little alteration is anticipated among the abundant devitrified silica minerals and alkali feldspars because they are anhydrous and stable to high temperatures. Cristobalite may transform to quartz with about a 12-percent volume decrease. Opal may also dehydrate, with a comparable loss in volume. The abundances of cristobalite and opal range from 0 to 30 percent, indicating maximum whole-rock volume decreases can be expected to be less than 4 percent. The thermal influence of the repository, however, may extend to regions where zeolites, smectites, and glasses are abundant. The hydrous phases have much lower thermal stabilities and can be affected by a temperature rise of tens of degrees Celsius. Two major effects of temperature on the hydrous phase are contraction due to water loss and the alteration of sorptive minerals to minerals that are less sorptive.

When tuff that contains hydrous phases is heated, various processes may occur. A reversible dehydration of smectites and zeolites involving significant volume reductions (Bish, 1981; Bish, 1984) may possibly reduce rock strength and alter permeability. Irreversible dehydration of smectites, zeolites, and volcanic glass reduces both volume and cation exchange capacity. Glass devitrification is of particular interest because of the proximity of the host rock to the vitrophyre in the lower Topopah Spring Member and to the nonwelded, nonzeolitized tuffs at the base of the Topopah Spring Member and in the underlying tuffs of Calico Hills in the southern portion of the Yucca Mountain exploration block. Alteration of glass may have four important effects: (1) changing the volume, causing changes in rock property such as permeability; (2) maintaining a high silica activity; (3) raising the pH of acidic water; and (4) forming sorptive secondary phases like smectite and zeolites (Kerrisk, 1983). Finally, the alteration of minerals to less-sorptive phases (i.e., the alteration of clinoptilolite to analcime) can affect the zeolitic rocks influenced by the heat of the repository (Bish et al., 1982; Bish, 1987).

Since mineral reactions at repository temperatures may occur at rates comparable to the isolation period (thousands of years), an understanding of kinetics is also important. Also, because many of the effects of elevated temperatures on hydrous minerals result from mineral dehydration, the effect of the partial pressure of water on the dehydration reactions must be known in order to predict the differences between unsaturated-zone and saturated-zone mineral dehydration. The plans for obtaining information about these processes are outlined in Section 8.3.1.3.

4.2.2.1 Hydrothermal alteration of zeolites

Data on the kinetics of zeolite reactions at slightly elevated temperatures are very limited because of experimental difficulties; reactions occur very slowly at temperatures below 200°C. Hydrothermal experiments have been

conducted on zeolitic, vitric, and welded tuff at temperatures of 80 to 180°C (the temperatures necessary to make observable changes in short times) for several weeks to several months (Daniels et al., 1982; Knauss and Beiriger, 1984; Blacic et al., 1986). The results of the experiments pertinent to the near-field environment are discussed in Section 7.4.1.3; some zeolites have been formed during the experiments but the products require further characterization.

Assuming a water composition similar to well J-13 water, the geologic data from the deeper portions of drillholes in Yucca Mountain and the experimental data described by Smyth and Caporuscio (1981) indicate that reaction of clinoptilolite to analcime will occur above 85°C in the tuffaceous beds of Calico Hills, although silica activity may more directly control the reaction than temperature (Section 4.1.1.4).

Although the clinoptilolite-to-analcime reaction is not anticipated in most zeolitic tuffs, experiments in progress suggest that water loss with volume decrease (up to 10 percent) in clinoptilolite will begin below 50°C when the water-vapor pressure is low. The volume loss could occur in the first major zeolite horizon under the thermal influence of the waste emplaced in the repository. The opening of new flow paths in a zeolite horizon barrier is possible, but they will probably expand again and close during rehydration. The effects of temperature on number of natural clinoptilolites and sodium-, calcium-, and potassium-exchanged clinoptilolites have been examined by x-ray powder diffraction (Bish, 1984). The molar volumes of all clinoptilolites decreased, with sodium-clinoptilolite undergoing the greatest decrease and potassium-clinoptilolite undergoing the least. Calcium-clinoptilolite and the natural clinoptilolite samples underwent intermediate decreases in volume. Most of the volume contraction in most samples occurred on evacuation at room temperature, demonstrating the strong effect of water-vapor pressure upon dehydration. Samples heated under 1-atm water-vapor pressure underwent significantly less volume decrease at a given temperature than those heated in ambient air or in a vacuum. All samples showed the ability to rehydrate upon cooling.

The data of Bish (1984) show that heat effects on volume change of clinoptilolite are strongly related to the exchangeable-cation content. The volumes of natural, mixed sodium-potassium-calcium clinoptilolites in rocks could be reduced in a repository environment, particularly if the clinoptilolites occur in the rock being dehydrated by the thermal pulse. However, tests now in progress suggest that clinoptilolites occurring in partially saturated rocks at temperatures below 100°C should not decrease in volume. Thermogravimetric and differential scanning calorimetric experiments on natural and cation-exchanged clinoptilolites also demonstrate the strong dependence of dehydration on exchangeable-cation composition (Bish, 1985). The results show that clinoptilolites can yield water on heating and rehydrate on cooling if the dehydration process is reversible after long-term heating (more than 1 yr). The volume decreases observed for clinoptilolite were completely reversible over short heating periods (less than 1 week). However, long-term heating of clinoptilolite can cause modifications, including irreversible collapse and reduction in cation-exchange capacity. Sodium-rich clinoptilolites are most susceptible to large volume reductions and exchange-capacity modifications, and they can alter to a phase analogous to heulandite-B. Further experiments are examining both the dependence of

dehydration temperature on water vapor pressure and the possibility of irreversible dehydration at 100°C (Section 8.3.1.3).

4.2.2.2 Hydrothermal alteration of smectites

Although smectites are not as abundant as zeolites, they occur in small amounts in virtually all core samples from Yucca Mountain examined to date (Bish et al., 1982). The smectites are important to radionuclide retardation because they selectively sorb certain cations (cesium, strontium, barium, and some actinides) (Section 4.1.3.3). Bish (1981) reviewed the effects of changes in relative humidity and temperature on smectites and noted that decreases in ambient humidity alone can cause dehydration and large decreases in volume when the ambient humidity drops below 20 to 50 percent. Smectites typically undergo reversible crystal structure collapse with attendant volume decrease at about 100°C with humidities of less than 100 percent. However, Koster van Groos and Guggenheim (1984, 1986) showed for Na- and K-montmorillonite that an increase in water pressure significantly raises the temperature of dehydration and layer collapse. Thus, collapse as a result of dehydration will probably be unimportant except where temperatures are high and the partial pressure of water is low. Smectites will remain in an expanded state along the flow paths in the saturated zone where the clays will not be dehydrated. An expanded state may not, however, be the case for clays exposed to a steam environment; the ability of bentonite to swell osmotically has been shown to be greatly reduced after exposure to a steam environment (Couture, 1985).

With increasing temperature over extended times under unsaturated conditions such as those predicted for the near field, smectite dehydration and collapse become irreversible (MacEwan, 1972; Bish et al., 1982). Once the smectite layers have irreversibly collapsed, their sorption capacity is greatly decreased, although selectivity for cesium is dramatically increased (Bish et al., 1982). The dehydration reaction is very likely controlled by the diffusion of the interlayer cation into the silicate layers and can occur over short periods at temperatures as low as 150°C. However, the actual dehydration temperature for any given smectite is controlled by interlayer cation composition and kinetics and can be as high as 500°C under unsaturated conditions in the presence of large divalent cations (Bish et al., 1982). Experiments are under way to determine whether the kinetics of the irreversible-collapse reaction are fast enough for the reaction to be important in the thermally disturbed zone (Section 8.3.1.3.3). If smectites do collapse irreversibly in the thermally disturbed zone near the repository, the consequence could alter the physical properties of the rocks and decrease the cation exchange capacity in the near field. The effects should be on the very near field, and rocks farther along the flow paths will retain their expandability and sorptive capacity.

4.2.2.3 Hydrothermal alteration of rhyolitic glasses

Data in the literature indicate that volcanic glass naturally alters to zeolites, but the phases resulting from this alteration depend greatly on the

glass and the ground-water composition (Hay, 1978). Moreover, the alteration of glass to smectite is suggested by the smectites that occur in perlitic fractures in the basal vitrophyre of the Topopah Spring Member and supported by reaction-path modeling (Kerrisk, 1983). The alteration of glass to smectites or zeolites increases the sorptive capacity of the rock; this would be beneficial along flow paths. Alteration of glass to nonhydrous phases will probably lower the permeability along flow paths but the overall sorptive capacity along the flow path will not be significantly affected. The vitrophyre underlying the host rock contains approximately 3.5 percent water, and dehydration or recrystallization of the vitrophyre could produce significant amounts of water. The vitrophyre is also one of the few sources of reduced iron in Yucca Mountain, although it is doubtful whether the small amounts of ferrous iron available would result in less oxidizing conditions which could enhance actinide retardation (Caporuscio and Vaniman, 1985). Plans for further investigation of these topics are given in Sections 8.3.1.3.2 and 8.3.1.3.3.

4.2.3 CHANGES IN WATER CHEMISTRY DUE TO THE THERMAL PULSE

The heating of fluids in and adjacent to the repository will cause increased reaction with the host rock, changing the compositions of both. Experimental results on the interaction of well J-13 water with crushed and solid samples of devitrified Topopah Spring tuff at 90, 150, and 250°C are discussed in Section 7.4.1.3. Briefly, tests at 150°C for 66 days showed changes from 0.01 to 0.5 parts per million (ppm) for aluminum, from 38 to about 120 ppm for silicon, from 12 to 6 ppm for calcium, from 1.9 to near 0 ppm for magnesium, from 5.1 to 4.4 ppm for potassium, and a pH change from 7.1 to 6.8. The concentrations of sodium ion, fluoride, chloride, nitrate, and sulfate showed no significant changes. The dissolved-carbonate content decreased slightly. These changes were associated with the appearance of secondary phases, including a potassium-rich clay or zeolite; clays rich in magnesium, calcium, and/or iron gibbsite; calcite; and a pure silicon dioxide phase (possibly cristobalite). The total amount of secondary phases was quite small, with clay being dominant. Tests run at 90°C showed only minor changes in water composition for durations of 66- and 300-days.

The modeling of hydrothermal reactions with the EQ3/6 reaction-path code is also reported in Section 7.4.1.3. This code, unlike the experiments, allows the modeling of open system behavior. Open system behavior is expected at least for gases such as carbon dioxide in the unsaturated zone in Yucca Mountain. The results of modeling the interaction of Topopah Spring tuff with well J-13 water at 150°C over long reaction times are shown in Table 7-10. The differences between the long-term open-system model and the short-term experiments are as follows: (1) the degassing of carbon dioxide with a concomitant change in pH, (2) very high aluminum concentrations, (3) higher sodium concentrations, and (4) very low calcium and dissolved carbon concentrations in the model. The modeling results could be improved with the following: (1) better data on dissolution kinetics, (2) more data on precipitation kinetics, (3) better high-temperature data on aluminum hydroxide complexes, and (4) experimental data on the reaction of tuff with water in the presence of gamma radiation. Plans for obtaining the needed

data on hydrothermal interactions are outlined in Section 8.3.4.2 and Section 8.3.1.3.

4.2.4 EFFECTS OF THE THERMAL PULSE ON RADIONUCLIDE MIGRATION

The preceeding sections discussed the anticipated thermal conditions, the hydrothermal alteration due to a thermal pulse, and the effects of the thermal pulse on ground-water chemistry. To date little work has been done on investigating the effects of the thermal pulse on overall radionuclide migration. To affect radionuclide migration the thermal pulse must alter the characteristics of the host rock which in turn would alter the characteristics of the transport processes, such as matrix diffusion, colloid formation, and advection (e.g., matrix flow, fracture flow).

One preliminary investigation into the effects of the thermal pulse on radionuclide migration has shown that silica mobility and redistribution is expected. A heat alteration study described by Daniels et al. (1982) reports that, particularly for the vitrophyre (G1-1292), a great amount of silica was dissolved at elevated temperatures (152°C) and then was precipitated upon cooling globules analyzed as pure SiO_2 . In addition, greatly increased amounts of clays or other fine grained sheet silicates were deposited on glass edges (examined by scanning electron microscope). The placement of waste containers will lead to the development of a temperature gradient in the surrounding host rock which will induce a solubility gradient. If water also flows, silica will be redistributed and depending on the magnitude and duration of the heat-induced flow, silica may accumulate in cooler regions. This deposition will result in reduced pore size which subsequently will lower permeability. A region of reduced permeability around a waste container will tend to isolate the container from infiltrating water. Thus the infiltrating water will be deflected away from the container, thereby reducing the transport of radionuclides from the waste container.

The scenario described above was qualitatively investigated by a numerical study (Travis and Nuttall, 1987) with the WAFE computer code. The WAFE code (Travis, 1985) can calculate heat and mass transport in porous media. Silica and chloride were included in the study. The inclusion of chloride is conservative because it does not adsorb and is not present, to any significant extent, in tuff minerals; thus, its potential source is from the surface. With an initial buildup of the pressure and temperature, the results of the study predicted an outward migration of silica. A narrow elevated saturation region developed with silica dissolving in a region near the container and migrating outward where it precipitated upon reaching a cooler environment. Some silica returned to the initially hot region, but a net outward flux of dissolved silica occurred. Figures 4-19 and 4-20 show the calculated transient dissolved silica concentration distribution and the total (dissolved and precipitated) silica concentration distribution at 10 and 100 yr. Note that in Figure 4-19 there is a zone of no dissolved silica where $x = 4$ m. This corresponds to a region where the silica has precipitated. The sum of the dissolved and the precipitated silica is responsible for the peaks shown in Figure 4-20.

Figure 4-19. Calculated dissolved silicon dioxide distribution at selected times. Modified from Travis and Nuttall (1987).

Figure 4-20. Calculated total silicon dioxide (dissolved and precipitated) at selected times. Modified from Travis and Nuttall (1987).

Figure 4-21 shows the calculated distribution for the transient chloride concentration. Initially, the chloride concentration increases in the hot-test region because of a decrease in water saturation from a loss of water through vaporization. The concentration of chloride in the higher saturation region decreases because of the addition of liquid water as the vapor condenses. The chloride concentration is a minimum where $x = 4$ m. This region corresponds to a region of elevated saturation. As the saturation returns to ambient levels, so, too, does the chloride concentration profile.

Significant changes in the amount of dissolved or precipitated silica can lead to large permeability changes. Assuming that permeability is related to the pore diameters and that a silica precipitate will coat pore walls uniformly, then permeability change as a function of porosity change can be calculated (Travis and Nuttall, 1987). Figure 4-22 shows the permeability change predicted for the end of the heat decay period by the WAFE calculation. The silica-precipitate distribution leads to an elevated-permeability region close to the container and a reduced-permeability region immediately outside the elevated-permeability region. Both of these regions are only a few centimeters thick. The outer reduced permeability shell will tend to deflect the downward-flowing water away from the container region rather than through the region. In addition, the inner elevated permeability ring will tend to channel water that had penetrated the low permeability shell, also around the container rather than through it. By reducing the amount of available water, the amount of radionuclides dissolved and transported will be decreased. The exact magnitude of this effect depends on the details of the hydrologic behavior of the host rock, and, in particular, on the presence or absence of fractures.

The numerical calculation of Travis and Nuttall (1987), summarized above, did not treat the full complexity of the water composition and reactions. Instead, the calculations were simplified by considering only silica dissolution, transport, and precipitation. As the potential hydrologic behavior of the Topopah Spring Member is better estimated (Section 8.3.1.2), the information concerning the presence or absence of fractures is obtained (Section 8.3.1.2), and as the ground-water transport models are further developed (Section 8.3.1.3.7), the effects of thermal pulses on radionuclide migration due to all significant changes in transport processes will be further investigated (Section 8.3.1.3.7).

4.3 NATURAL ANALOGS AND RELATED FIELD TESTS

The long period of time over which a repository must provide waste isolation and the large volume of rock present between a repository and the accessible environment pose particularly difficult problems in the prediction and assessment of repository performance. Although laboratory studies provide data on fundamental chemical and physical processes, modeling of the total repository system will require the extrapolation of these data in space and time. However, the study of natural systems possessing one or more features that mimic critical parts of a repository environment may provide

Figure 4-21. Calculated chloride distribution at selected times. Modified from Travis and Nuttall (1987).

Figure 4-22. Calculated permeability assuming relationships between pore volume and permeability. Modified from Travis and Nuttall (1987)

additional confidence in the required extrapolations. This section describes natural analogs and related field tests that possess these attributes.

4.3.1 NATURAL ANALOGS

Although there are no natural systems available for study that are totally analogous to a repository environment, the following systems possess features that are pertinent to a repository in tuff: (1) warm and hot springs in tuffaceous rocks, and (2) uranium and thorium ore bodies and their host rocks. Only a small group of potentially analogous sites are discussed in this section.

4.3.1.1 Warm and hot springs

The study of warm and hot springs in tuffaceous rocks provides information about several important aspects of a repository environment in tuffaceous rock including (1) the effect of the thermal pulse on the chemistry of ground water; (2) the effect of heated ground water on the host rock including dissolution and precipitation reactions; (3) the transport of certain elements (e.g., strontium, cesium, uranium, thorium, etc.) found in radioactive waste in a hydrothermal environment; and (4) hydrothermal fluid flow in fractured tuff. In addition, the studies will provide case studies for testing computer codes like EQ3/6.

Warm and hot springs are present in tuffaceous rocks at many locations in the western United States. Several of the locations have been studied in detail, including (1) the geyser basins in Yellowstone National Park (Honda and Muffler, 1970; Thompson et al., 1975; Keith and Muffler, 1978; Keith et al., 1978; Thompson and Yadav, 1979; Bargar and Beeson, 1981; Bargar and Muffler, 1982; Keith et al., 1983; Bargar and Beeson 1984a,b); (2) the Valles caldera in northern New Mexico (Goff and Sayer, 1980; Goff et al., 1981; Goff and Grigsby, 1982; Goff et al., 1982; Phillips et al., 1984; Hulen and Nielson, 1985); and (3) the Newberry Volcano in central Oregon (Bargar and Keith, 1984; Keith et al., 1984a,b). Although the Newberry Volcano is not composed of typical silicic ash-flow tuff, it contains intermediate to silicic lavas and tuffs of the type that also occur in and around Yucca Mountain. The results obtained from the studies have been largely empirical and have included (1) descriptions of the petrography and the primary and secondary mineralogy of core and surface samples, (2) chemical analyses of rocks and thermal waters, and (3) temperature profiles.

The warm and hot springs of the geyser basins in the Yellowstone National Park are discussed here because they have been studied in the most detail. The springs emerge from silicic volcanic rocks and volcanogenic sediments that are less than 1 million years old. In fact, the springs are likely surface manifestations of recent igneous activity. The USGS has drilled 13 drillholes (75 to 320 m) to investigate the hydrothermal systems in the Yellowstone National Park. These holes have penetrated volcanogenic sediments (partly glacial), rhyolitic lava flows, volcanic breccias, and various types of silicic tuffs. Because the drillholes are located in parts

of the Yellowstone caldera that have been filled in with postcaldera rhyolite flows, ash-flow tuffs are not the dominant rock type encountered (Keith and Muffler, 1978). However, drillholes Y-5 (Keith and Muffler, 1978) and Y-9, Y-12, and Y-11 (Bargar and Muffler, 1982) do contain ash-flow tuffs (part of the Lava Creek Tuff of the Yellowstone Group; Christiansen and Blank, 1972) similar in composition and texture to the tuffs exposed at Yucca Mountain.

The warm and hot springs have altered the Lava Creek Tuff in drillholes Y-5 and Y-11 to a suite of secondary minerals including alpha-cristobalite, beta-cristobalite, quartz, chalcedony, opal, hematite, goethite, pyrite, manganese oxides, clays, zeolites, calcite, gypsum, and other minor phases. Because data are available on the alteration paragenesis, compositions of the secondary minerals, present day temperature profiles, and the compositions of water in the drillholes, the relative stabilities of the secondary mineral assemblages in the Yellowstone drillholes can be derived. In this context, the capability of the EQ3/6 code will be used and tested. Kinetic effects should be apparent in comparisons of the code predictions with the observed mineral assemblages.

Once the dominant chemical controls on the secondary mineral assemblages in the Yellowstone drillholes have been identified, this information can be used to characterize the stability ranges of the secondary minerals (e.g., silica, polymorphs, and zeolites) in Yucca Mountain under ambient conditions as well as in the thermal regime resulting from waste emplacement. In addition, the modeling will result in a better understanding of the effects of the thermal pulse on the chemistry of ground water in the near field. Plans for additional studies are described in Sections 8.3.1.3.1, and 8.3.1.3.3.

4.3.1.2 Uranium and thorium ore deposits and their host rocks

Uranium and thorium ore deposits, are a source of data on the following: (1) the long-term stability of radioactive solids (Naudet, 1978); (2) the long-term release of radionuclides from these solids (Curtis, 1985); (3) the transport of radionuclides under various pH, Eh, temperature, and pressure conditions, ground-water and host rock compositions, and hydrologic regimes (Ruffenach et al., 1980); and (4) the long-term effects of radiolysis (Curtis and Gancarz, 1983). Items 1, 2, and 4 relate to near-field conditions in a repository environment while Item 3 relates to near- and far-field conditions under which radionuclides may be transported from a repository to the accessible environment.

Known uranium and thorium ore deposits are separated into two types for the discussion that follows: those that reached nuclear criticality (i.e., experienced a sustained nuclear fission reaction), and those that did not reach criticality. The first type yields information on many of the radionuclides present in modern radioactive waste, while the second type yields information on a more limited number of radionuclides. Only one of the first type of deposit has been identified to date, the Oklo uranium deposit in the Republic of Gabon, Africa. Although there are many deposits of the second type, only a few have been studied from the viewpoint of the four items

listed above (e.g., Osmond and Cowart, 1981; Eisenbud et al., 1984; Airey et al., 1984, 1985).

4.3.1.2.1 The Oklo uranium deposit

Studies of the Oklo deposit indicate that, under the conditions in which the deposit was formed and subsequently evolved, uraninite remained generally stable over a period of 2.0 billion yr (Gancarz, 1978). The critical stage for the deposit lasted only 100 to 800 thousand yr (Ruffenach et al., 1980). Tellurium, uranium, thorium, niobium, zirconium, yttrium, the rare-earth elements, tin, plutonium, and the noble metals ruthenium, rhodium, and palladium, either remained in the uraninite lattice or were partly redistributed within the critical reaction zones over time (Table 4-20). The low solubility of most of these elements in aqueous fluids under the ambient chemical conditions was largely the reason for their retention in the zones. Chalcophilic elements such as cadmium, molybdenum, silver, and lead were transported out of the critical reaction zones by hydrothermal fluids and deposited in the surrounding host rocks at distances dictated by the solubilities of their respective sulfide or oxide phases under the temperature, pressure, and chemical conditions in the surrounding rocks. The alkali and alkaline earth elements and the noble gases were largely transported out of the Oklo deposit. Annual release rates cannot be derived for these elements because the timing of the losses cannot be constrained.

The effects of radiolysis in the Oklo ore deposit are discussed by Curtis and Gancarz (1983). The authors calculated the alpha- and beta-particle doses in the critical reaction zones during criticality and the energy imparted to the fluid phase by these particles. The energy caused radiolysis of water and the production of reductants and oxidants. The effect of these reductants and oxidants on the transport of radionuclides within and outside the reactors has been difficult to quantify. Analysis of iron(III)/iron(II) ratios in bulk rock samples from the reactor zones (Branche et al., 1975) indicate that the iron in these zones is generally more reduced than the iron in the host rocks well away from the reactor zones. In fact, iron is most reduced in the samples that show the greatest uranium-235 depletion. This suggests that the reduction of iron in the reactor zones was contemporaneous with the nuclear reactions and not a later supergene phenomenon of secondary enrichment. Curtis and Gancarz (1983) suggest that radiolysis of water resulted in the reduction of iron(III) in the reactor zones and the oxidation of uranium(IV) in uraninite. Furthermore, Curtis and Gancarz suggest that the oxidized uranium was transported out of the critical reaction zones and precipitated through reduction processes in the host rocks immediately outside the zones. The reduction processes likely involved organics or sulfides present in the host rocks. However, if the host rocks around the natural reactor cores had not contained species capable of reducing the oxidized uranium transported out of the cores, the uranium could have been transported much farther from the critical reaction zones. The important point is that even with intensive radiolysis very little of the uranium in the natural reactors was mobilized (several percent--Naudet, 1978).

CONSULTATION DRAFT

Table 4-20. Behavior of fission products in industrial reactors and in the Oklo natural reactors^a

Chemical state	Elements	Ionic radius (A)	<u>Industrial reactors</u>		<u>Oklo natural reactors</u>	
			Retention in the fuel	Localization in the free volume of the cladding	Retention	Migration
Gaseous compounds	Kr	1.8 ^b		x		x
	Xe	2.0 ^b		x		x
Volatile compounds	I ₂	2.5 ^b		x		x
	Te	0.60		x	x	
	Cs	1.8		x		x
Oxides insoluble in the matrix	Sr	1.21		x		x
	Ba	1.44		x		x
	Mo	0.73		x		x
Metallic inclusions	Tc	0.72		x	x	
	Ru	0.70		x	x	
	Rh	0.71	x		x	
	Pd	0.70	x		x	
Oxides soluble in the matrix	Y	0.98	x		x	
	Nb	0.72	x		x	
	Zr	0.92	x		x	
	Rare-	1.02-				
	earths	1.13	x		x	
	Pb	1.02				
	Th	1.12				
	U	1.08				
	Pu	1.04				

^aSource: Ruffenach et al. (1980).

^bMolecular radius (Johnston, 1966).

4.3.1.2.2 Other uranium and thorium deposits

Studies of uranium and thorium deposits that have not undergone a critical reaction phase have provided data on the mobility of various shortlived radionuclides in the uranium and thorium decay series. The deposits may also yield data on the mobility of technetium, neptunium, and plutonium, although

no such data have been reported to date. The ore deposits of Alligator Rivers are discussed here because they have been studied in considerable detail. Other deposits that have been investigated are the Colorado Plateau uranium deposits (e.g., Osmond and Cowart, 1981) and the Morro de Ferro thorium deposit in Brazil (Eisenbud et al., 1984).

The Alligator Rivers region of northern Australia is underlain by lower Proterozoic sediments and tuffs that overlie late-Archean granitic complexes. Approximately 1,800 million yr ago, these rocks were subjected to regional metamorphism and deformation. After subsequent uplift and some retrograde metamorphism, the rocks were covered by middle-Proterozoic and younger units (Needham and Stuart-Smith, 1980).

The uranium deposits of the Alligator Rivers area are found mainly in the lower Proterozoic Cahill Formation. The formation has a lower member of carbonate rock and carbonaceous schists that is overlain by an upper member composed of mica-quartz schist and gneiss. It was metamorphosed to amphibolite facies and subsequently retrograded to greenschist facies assemblages. Uranium mineralization is found in zones of brecciation, faulting, and ferromagnesian alteration within the formation (Needham and Stuart-Smith, 1980). The primary uranium mineralization appears to have occurred approximately 1,400 million yr ago. It consists mainly of uraninite that is locally replaced by coffinite and brannerite.

Because the Alligator Rivers region has a monsoonal climate, deep lateritic weathering profiles with anomalous uranium concentrations overlie the primary ore bodies. The profiles are generally composed of four zones (Airey et al., 1985): (1) a zone of oxidized ferricrete or a ferruginous zone grading into a mottled zone (these zones are unsaturated in the dry season and generally have an oscillating water table at other times - uranium is leached in these zones); (2) a pallid zone from which iron has been leached (some uranium is deposited in this zone - this and the following zones are all permanently saturated); (3) a transition zone between weathered and unweathered rock (conditions are generally oxidizing in the zone with some uranium leaching); and (4) unweathered rock of the Cahill formation.

The Australian Atomic Energy Commission under a contract with the NRC has conducted studies (Airey et al., 1984, 1985) of radionuclide transport in the rocks, sediments, and waters in and around the Alligator Rivers uranium deposits. The results of the studies include the following: (1) in the zone of weathering, uranium and thorium are associated principally with iron minerals; (2) ground waters in the region are generally oxidizing (+60 to +400 mV) and contain dissolved oxygen; (3) although uranium and thorium are mobilized in solution, a more significant mode of transport is by erosion; (4) the concentration of uranium in ground water ranges from approximately 0.25 to 190 ppb; (5) the rate at which uranium is leached or deposited in and around the ore bodies in percent per 1,000 yr ranges from a deposition rate of 0.4 to a leach rate of 2.5; and (6) colloidal transport does not contribute significantly to uranium transport but is important for the transport of thorium.

4.3.2 RELATED FIELD TESTS

Field tests at other sites have yielded data relevant to migration of radionuclides and to evaluation of models to be used for performance assessment. The following sections discuss the results of studies of radionuclide migration at (1) the site of the underground Cambrian nuclear explosion on the Nevada Test Site, and (2) the DP site at Los Alamos National Laboratory where liquid radioactive waste has infiltrated downward into Bandelier Tuff.

4.3.2.1 Cambrian site

Data from a field experiment at the site of the Cambrian underground nuclear test have provided direct information on the transport rates of radionuclides through alluvium. The data can help confirm that the transport model in the performance assessment code TRACR3D can accurately model radionuclide transport in tuffaceous media. The data also provide some indication of the limitations of laboratory sorption measurements; however, no further work is planned.

A field study of the distribution of radionuclides around the Cambrian underground cavity was initiated in 1974, about 9 yr after detonation of the Cambrian nuclear test. The field study provides an actual in situ field test in rocks that have some of the same sorption characteristics as those at or along flow paths from the potential repository at Yucca Mountain. Additionally, although the medium is dominantly tuffaceous alluvium, the reactive surface area is higher and effective porosity lower than is expected for the rocks in and around the exploratory block at Yucca Mountain.

The following information was taken from Hoffman and Daniels (1981). The explosion at the Cambrian site was a 0.75-kiloton test fired in May 1965, at a depth of 294 m beneath the water table in tuffaceous alluvium (Figure 4-23). Tests began after the cavity and chimney were predicted to be filled with ground water to the preshot static water level, 73 m above the detonation point. It was assumed that tritium, plutonium, and uranium fission products would be present in the cavity and in the ground water within the cavity and could be used to study possible migration from the cavity.

The field study began with the completion of a satellite well (RNM-2S) 91 m from the Cambrian cavity followed by drilling of a reentry well (RNM-1) into the cavity itself. Both solid and liquid samples were taken from well RNM-1 to determine the distribution of radionuclides between the solid material and the ground water. Water was then pumped from the satellite well to induce an artificial gradient sufficient to draw water from the Cambrian cavity through the surrounding rocks. Pumping from well RNM-2S began in October 1975. The original pump rate was $1 \text{ m}^3/\text{min}$ until October 1977 when the pump rate was increased to about $2.3 \text{ m}^3/\text{min}$.

The various radionuclides present in water drawn from well RNM-2S reflect the extent of radionuclide migration. Approximately 2 yr after pumping started, significant amounts of tritiated water were found in water from the satellite well, signaling the arrival of water from the Cambrian cavity region. By that time, approximately 1.44 million m^3 of water had been

Figure 4-23. Schematic of Cambrian explosion cavity and drillholes RNM-1 and RNM-2S. Modified from Hoffman and Daniels (1981).

pumped from the satellite well. After almost 6 yr of pumping, the tritium concentration in the pumped water reached a maximum. The original amount of tritium produced during the Cambric event was about 6×10^4 curies (Hoffman et al., 1977). By the end of September 1984, slightly more than 3.7×10^4 curies of tritium had been pumped out through well RNM-2S (about 60 percent of the initial inventory). The result of a transport calculation based on Sauty's model for hydrodispersive transfer in aquifers (Sauty, 1980) was presented by Thompson (1985). Travis (1984) has modeled the tritium elution at Cambric using the TRACR3D code. According to Travis (1984), the difference between the calculated and observed breakthrough curves is approximately equal to the margin of error in the input parameters used in the code.

Other radionuclides have also been measured in water from well RNM-2S including chlorine-36, krypton-85, ruthenium-106, and iodine-129. Radionuclides present in the Cambric cavity in cationic form (e.g., strontium-90 and cesium-137) have not been detected to date in well RNM-2S water (Thompson, 1985). The chlorine-36 results presented in Figure 4-24 indicate that the chlorine-36 pulse preceded the tritium pulse in well RNM-2S. This suggests that tritiated water is exchanging with water in the pores in the tuffaceous alluvium which, to some degree, exclude chlorine-36 due to the repulsion of the negative ion by negative surface charges. This suggests that matrix diffusion of water does occur in tuffaceous alluvium.

Krypton is a noble gas and might be expected to show very little retention in the alluvial aquifer. Although the krypton data show more scatter than the tritium data, the behavior of the elements is correlated. The observations are compatible with the hypothesis that some krypton is sorbed on the alluvium and is therefore retarded relative to water (Thompson, 1985). The fact that the krypton-tritium ratios in well RNM-2S water are lower than the calculated source term ratio of 1.22×10^{-4} while they are higher than the calculated krypton-tritium ratio in well RNM-1, also suggests krypton is sorbed onto the alluvium relative to tritium (Thompson, 1985).

Ruthenium-106 has been detected in well RNM-2S water (Daniels, 1981; Coles and Ramspott, 1982). Although the data are limited, it is clear that some ruthenium from the Cambric cavity has been transported to well RNM-2S. Sorption measurements for ruthenium on tuffaceous alluvium indicate ruthenium has a high sorption coefficient ($K_d = 10$ to 30 ; Wolfsberg, 1978) under ambient conditions, which suggests ruthenium should not have migrated more than a fraction of a meter from the Cambric cavity since 1965. However, batch sorption measurements may not distinguish which species of a given element sorb or to what extent the species sorbs. Therefore, there may be species present in the batch sorption experiments which sorb very little, if at all. Daniels (1981) suggests that the ruthenium-106 present in well RNM-2S water exists as an anionic species which does not sorb on the alluvium.

Analyses of iodine-129 in water from well RNM-2S show that iodine, like chlorine, appears to have arrived at well RNM-2S sooner than tritium. However, the calculated fraction of iodine-129 that has been eluted is only 39 ± 11 percent in comparison with approximately 60 percent for tritium. The cause for this difference has not been determined (Thompson, 1985).

Figure 4-24. Chlorine-36 and tritium concentrations in well RNM-2S water. The error bars are the standard deviations for single measurements; they were obtained from the errors associated with measurements and estimated uncertainties for the various parameters entering into the calculation. The estimated uncertainties were propagated in quadrature. Modified from Daniels and Thompson (1984).

4.3.2.2 Los Alamos DP site

The distribution of radionuclides and water beneath a liquid waste disposal site at the Los Alamos National Laboratories was investigated by Nyhan et al. (1985). The site, known as the DP West site, was used from 1943 to 1951 to dispose of untreated radioactive liquid wastes resulting from plutonium-purification operations. A liquid-waste-treatment plant was installed in 1952 to handle the increased load but discharges continued until 1967, when a new treatment plant was built. The liquid wastes were discharged into absorption beds 36.6 by 6.1 m and 1.2 m deep. Approximately 69,260 m³ of liquid effluent was discharged into the beds, with 89 percent added between 1945 and 1960. An estimated 10 curies of plutonium was discharged, with about 98 percent of the figure having been discharged, as untreated waste between 1945 and 1952. In August 1961, 10.8 m³ of plutonium-bearing effluent followed by 9.7 m³ of tap water were added to absorption bed 1. The fluid pulse resulted in formation of a saturation front that subsequently traveled downward through the unsaturated zone beneath the site.

The absorption beds consist of 60 cm of stone cobbles overlain by 15 cm of gravel, 15 cm of sand, and 30 cm of soil that were excavated out of the Tshirege Member of the Bandelier Tuff. The Bandelier Tuff at this location consists of an upper 3.6 m of moderately welded, light-brownish-gray tuff underlain by a 3-m-thick layer of reworked tuff and pumice which fills a channel (or cut) in the underlying tuff unit and is separated by a sharp contact zone. The underlying unit consists of 33 m of moderately welded, light-gray tuff. It is in turn underlain by a moderately to densely welded zone that forms the lower portion of the 250 m of the Bandelier Tuff at this location. The zone of water saturation is 350 m below the surface of the site. Therefore, at this location the Bandelier Tuff is entirely in the unsaturated zone.

Nyhan et al. (1985) drilled two 30.5 m drillholes through absorption beds 1 and 2 in area T of the DP West site. Continuous core samples were obtained and water samples were collected. A total of 800 samples were analyzed for americium-241 and plutonium. The results are shown in Figures 4-25 and 4-26. In drillhole 1 of bed 1, plutonium and americium-241 were detected to a sampling depth of 30.0 m and 30.5 m, respectively, while in drillhole 2 of the bed plutonium and americium-241 were detected to a depth of 14.5 m and 30.8 m, respectively. In bed 2, plutonium and americium-241 did not penetrate as far into the tuff because the bed did not receive the water pulse added to bed 1 in 1961. Plutonium penetrated 6.55 m in drillhole 1 and americium-241 penetrated 13.41 m. In drillhole 2 of bed 2, plutonium was detected only to a depth of 5.2 m and americium-241 to a depth of 12.8 m. The observations indicate that americium-241 has a higher mobility in this setting than does plutonium.

The spikes in the radionuclide-concentration profiles (Figures 4-25 and 4-26) are to some degree associated with geologic conditions. For instance, at 4 to 5.5 m in bed 1, a highly weathered, light orange-gray tuff layer with a high clay content was found, while at 8 to 9 m a contact zone was found separating a brownish-gray, moderately welded tuff from an underlying light gray, moderately welded tuff. Both the clay-rich layer and the contact zone have decreased hydraulic conductivities. Fractures, though present, did

Figure 4-25. Concentration of plutonium as a function of sampling depth for absorption beds 1 and 2 in 1978. Modified from Nyhan et al. (1985).

Figure 4-26. Concentration of americium-241 as a function of sampling depth for absorption beds 1 and 2 in 1978. Modified from Nyhan et al. (1985).

not appear to be associated with enhanced radionuclide transport. In fact, the fractures appeared to be barriers to unsaturated flow. Overall, Nyhan et al. (1985) concluded that "from 0.3 to 5.1 percent of the plutonium and from 3.0 to 49.0 percent of the americium-241 initially absorbed can be mobilized with the addition of less than 1 column volume of water."

Travis (1985) has attempted to model the migration of radionuclides at the DP West site with the TRACR3D code. He compared concentration profiles calculated on the basis of a constant adsorption coefficient and a colloid-population-balance approach. His results are shown in Figure 4-27. The colloid-population-balance approach clearly produces a better fit to the observed profile. This is consistent with the laboratory results of Fried et al. (1975) in which a rapidly migrating form of plutonium was found in experiments on Bandelier Tuff. Unfortunately, insufficient data exist on the forms of plutonium and americium likely to be released from waste packages at a repository at Yucca Mountain for extrapolating the results of Travis (1985) and Fried et al. (1975).

4.4 GEOCHEMICAL STABILITY

Man-induced factors that could potentially affect the geochemical stability of the site include mining, exploring for minerals, drilling unrelated to NNWSI Project operations, and withdrawing ground water. Natural changes that could affect the geochemical stability of the site include changes in the water regime as a result of tectonic or climatic variations and the introduction of magmatic fluids into the area. However, natural chemical buffers present in the water and rock help to resist changes in the geochemistry. The geochemical stability of the Topopah Spring tuff can be understood from the data presented in Section 4.1.1; chemical interaction between water and rock are discussed in detail in Section 4.1.1.3.

4.4.1 MAN-INDUCED EFFECTS

The potential for mining at the Yucca Mountain site is discussed in Section 1.7. As there is no compelling evidence that major, significant, or unique deposits of mineral or energy resources exist in the vicinity of the repository, the potential for mining is not considered important for geochemical stability.

The record of subsurface drilling is discussed in Section 1.6. Injection disposal wells are not present in the vicinity of the Yucca Mountain repository site. There is no reason to expect that the site has particularly favorable characteristics so that future injection disposal wells would be located in the area. Therefore, this factor is not a consideration for geochemical stability. Further discussion of this topic can be found in the Yucca Mountain environmental assessment (EA) (DOE, 1986).

Pumping of ground water may cause changes, usually minor, in ground-water chemistry by modifying flow paths or by inducing drainage of poorly permeable beds in which stored water has higher concentrations of solutes

Figure 4-27. The measured and simulated plutonium concentration profiles for the region below bed 1 at the DP West sites. K_d = distribution coefficient. Modified from Nyhan et al. (1985).

CONSULTATION DRAFT

than actively flowing water. In the region around Yucca Mountain, ground water is used primarily for irrigation. Lesser uses include municipal, domestic, livestock, mining, and support of operations at the Nevada Test Site.

Projections of future ground-water withdrawal in the region are discussed in Section 3.8. Although these projections and simulations of their effects at Yucca Mountain have not been completed, the effects of long-term pumping close to Yucca Mountain have been insignificant. Extensive pumping from well J-13, which produces from tuffs correlative with those at Yucca Mountain, has supported operations at Jackass Flat for more than two decades without significant changes in water levels or water chemistry (Thordarson, 1983). Although small, the hydraulic stresses around well J-13 are very likely greater than those resulting at Yucca Mountain from reasonably expectable future ground-water development. Therefore, disruption of geochemical stability by hydraulic stresses seems highly unlikely.

The construction and operation of the repository could affect radionuclide transport in two ways: first, the introduction of drilling fluids may provide a nutrient source for microbes; therefore, the effect of microbial activity on radionuclide migration is being assessed (Section 8.3.1.3.4); and second, the construction of the repository will produce particulates. The sorption and transport of radionuclides by particulates and colloids will also be studied (Section 8.3.1.3.4). Plans for studying the filtration of particulates by Yucca Mountain tuff are discussed in Section 8.3.1.3.6.

4.4.2 POTENTIAL EFFECTS OF NATURAL CHANGES

Natural processes of potential importance for geochemical stability include tectonism and volcanism (Section 1.3), flooding (Section 3.2), glass or mineral alteration that have effects on water composition (Sections 4.1 and 4.2), and climatic changes (Section 5.2). Where appropriate, specific processes that require further study are indicated.

Tectonism, volcanic activity, and structural features have already been treated in terms of possible physical disruption of the repository (Section 1.3.2); the displacement of flow paths by such processes could alter the effectiveness of natural geochemical barriers at the site. Among structural features, faults are important in bounding the exploration block at Yucca Mountain. Because variably mineralized faults may act either to enhance or retard ground-water flow, it is important to know not only what types of faults occur, but also what changes in flow and geochemical retardation might occur as a result of seismic activity and fault movement (Section 1.4). The problem will be addressed among the site characteristics contributing to performance assessment (Section 8.3.5.2.3), based on data obtained from the site geohydrology program (Section 8.3.1.2). The potential effects of igneous intrusion have been evaluated by Link et al. (1982), and the probabilities of volcanic activity have been estimated by Crowe et al. (1982). The risk from future igneous activity is low, but risk assessment is continuing (Section 8.3.1.8).

CONSULTATION DRAFT

Because the potential repository horizon at Yucca Mountain is in the unsaturated zone, there are some conceivable changes in saturation that would affect geochemical processes. First, any possible flooding of the repository would affect geochemical stability. A future rise in the water table is one possibility that has been evaluated, and it is considered that inundation of the repository is improbable (Chapter 3, DOE, 1986). Zones below the repository that were previously unsaturated may become saturated, however, and such a change will be considered in assessing the balance between saturated- and unsaturated-zone processes of retardation. Another possible mechanism of repository saturation would be through hydrothermal activity. There is some record of past hydrothermal activity at depth, well below the present water table. Current studies indicate that hydrothermal activity is old (Bish, 1987) and is related to the Timber Mountain igneous activity about 11 million yr ago (Kistler, 1968). Minerals with moderately high temperatures of formation (between ambient temperature and 200°C) (Levy, 1987) do occur in some voids and fractures above the water table; the timing of formation for such minerals will be evaluated to determine whether they formed during the waning stages of tuff cooling or whether they may have formed from later hydrothermal fluids (Section 8.3.1.3.2). The test will determine whether post-Miocene hydrothermal activity will have to be considered to be a potential risk at the site. Finally, the possibility of deep-seated spring activity transgressing the repository horizon at some time in the past is being assessed by the study of fault-related mineral deposits (Section 8.3.1.3.2).

The geochemistry (nonhydrothermal) of rock and water interactions at Yucca Mountain involves a number of metastable mineral phases. As such, the chemical rock and water system will change toward more stable assemblages. The changes, however, have been occurring for several million years and can be expected to continue so slowly that there will be no significant effect on repository performance. In particular, it is important to note that the potential repository horizon is located in devitrified tuff with a relatively stable assemblage of feldspar minerals and quartz and some metastable cristobalite and opal along with very small amounts of other metastable minerals. The impact of the natural evolution of silica activity on the metastable assemblages is the major factor of concern. This natural change has been investigated (Section 4.1.1.4), and will be a part of future studies (Investigation 8.3.1.3.3).

Long-term climatic changes are discussed in Section 5.2. In terms of geochemical effects that influence repository performance, a renewed pluvial episode like those of the Pleistocene Epoch could result in increased recharge through Yucca Mountain, a higher water table, or the expansion or appearance of zones of perched water (Winograd and Doty, 1980). Any of these changes could affect flow rates or flow pathways that are important for integrating the site geochemistry into performance assessment. For example, a recharge-related transition from matrix flow to fracture flow may result in more rapid transport, along with changes in colloid retardation properties and dispersion/diffusion properties. The bounds on effects from such potential changes are a part of the performance-assessment program plan (Section 8.3.5).

4.5 SUMMARY

4.5.1 SUMMARY OF SIGNIFICANT RESULTS

The results summarized here are those related to the geochemical retardation processes that are considered the most significant in terms of the performance objectives and conceptual models. Thus a comprehensive summary of Chapter 4 is not presented. Sections 4.1.1 (Mineralogy and petrology), 4.1.2 (Ground-water chemistry), and 4.1.3 (Geochemical retardation processes) are briefly summarized with emphasis on performance objectives.

The mineral and petrologic results, as related to performance objectives and conceptual models, are best summarized by describing the distribution of sorptive zeolites at Yucca Mountain. The sorptive zeolites in the units surrounding the host rock provide part of the geochemical barrier for the retardation of radionuclides. Four principal intervals of zeolitization have been identified: (1) the clinoptilolite-heulandite-smectite zone at the top of the lower Topopah Spring vitrophyre; (2) the relatively thick zone of clinoptilolite that occurs in the bedded, nonwelded, and poorly welded tuffs that form the base of the Topopah Spring Member and the underlying tuffs of the Calico Hills, and extends into the top of the Prow Pass Member of the Crater Flat Tuff; (3) the partly welded and bedded tuffs at the base of the Prow Pass Member and at the top of the underlying Bullfrog Member; and (4) the poorly welded and bedded tuffs at the base of the Bullfrog Member compound cooling unit and at the top of the underlying cooling unit within the uppermost Tram Member.

The ground-water chemistry of the site is summarized from the results obtained from the saturated-zone wells on and near Yucca Mountain. Analyses of ground water from the drillholes that penetrate the host rock and surrounding units in the area of the exploration block indicate that they are principally sodium bicarbonate waters with low total dissolved solid content (200 to 400 mg/L). However, water from the carbonate aquifer contains over 1,000 mg/L total dissolved solids. Yucca Mountain ground waters are oxidizing. The dominant cations in the Yucca Mountain ground water are sodium, calcium, potassium, and magnesium with sodium the most abundant, ranging from 65 to 95 percent of the cations present. The concentrations of iron and manganese are generally low. The anionic constituents of the ground water show a relatively uniform distribution in the wells with about 80 percent bicarbonate and the remainder as sulfate and chloride (usually present in nearly equal molar concentrations) and fluoride (in varying concentrations). Overall, only minor variations in ground-water compositions occur in and adjacent to Yucca Mountain. Temporal variations in ground-water composition are also minor. Finally, the water from well J-13 has been chosen as the reference ground water that will represent the ground-water conditions along the transport pathways from the repository to the accessible environment.

The geochemical retardation processes described in Section 4.1.3 include sorption, precipitation, and matrix diffusion. Sorption data have been obtained with samples representative of Yucca Mountain strata. The variables include time, temperature, particle size, oxidation-reduction conditions, element concentration, ground-water composition, and test method (batch,

column, or circulating system). The radionuclides investigated were americium, cesium, neptunium, plutonium, uranium, thorium, strontium, technetium, tin, barium, radium, cerium, europium, and selenium. A correlation of sorptive behavior with mineralogy has been identified. Sorption of alkali metals (e.g., cesium) and alkaline earths (strontium, barium, radium) is directly correlated with the presence of minerals with exchangeable cations such as the zeolite clinoptilolite and potentially the smectite clays. Sorption ratios on these minerals vary from 1,000 to more than 10,000 ml/g. The correlation of sorption of cesium, europium, plutonium, and americium with mineralogy exists, but the relation is not as clear as for the alkali metals. However, sorption ratios are high for the elements listed above (100 to more than 1,000 ml/g); therefore, the lack of correlation is not critical. The sorption of these elements though, is undoubtedly influenced by the formation of hydroxy and carbonate complexes. It appears that sorption on tuff will not offer much retardation for such anionic species as an oxidized technetium species (i.e., TcO_4^-). Precipitation processes (i.e., solubility) will limit the concentration of radionuclides in water that moves from the repository toward the accessible environment. The solubility of americium, plutonium, neptunium, uranium, curium, thorium, strontium, radium, carbon, cesium, technetium, iodine, tin, zirconium, samarium, and nickel in well J-13 water has been compiled from available measurements and calculations. Diffusion of radionuclides from fracture water into the pore water in the rock matrix is important for retarding the transport of dissolved waste elements through fractures in tuff.

4.5.2 RELATION TO DESIGN

The geochemical characteristics of the backfill materials and the anticipated chemical interactions among the waste container, backfill, ground water, and the host rock under assumed waste emplacement conditions are described in Sections 6.3.3 and 7.4. The geochemical characteristics of the seal material as well as the anticipated chemical interactions among the seal materials, ground water, host rock, and backfill under assumed emplacement conditions are described in Section 6.3.5. Waste container design and geochemical interactions are described in Section 7.4. The mineral stability studies may have an impact on design considering that there is a reported volume change as minerals undergo transitions. This volume change should be assessed in order to determine whether the change is significant to design.

4.5.3 IDENTIFICATION OF INVESTIGATIONS

The investigations under the geochemistry program are listed below and will be discussed in detail in Section 8.3.1.3. Investigations 8.3.1.3.1 (ground water chemistry), 8.3.1.3.2 (mineralogy, petrology and rock chemistry) and 8.3.1.3.3 (stability of minerals and glasses) will provide the base site characterization data on the natural geochemical environment at Yucca Mountain. Investigations 8.3.1.3.4 (radionuclide sorption), 8.3.1.3.5 (radionuclide solubility), 8.3.1.3.6 (radionuclide dispersion, diffusion, and advection), and 8.3.1.3.8 (gaseous radionuclide transport) will investigate the geochemical processes that are acting and may potentially act at Yucca

Mountain. Finally, Investigation 8.3.1.3.7 will support the listed investigation and assess the physical/chemical processes through integrated transport calculations. This investigation will synthesize the geochemistry test program data and provide a geochemical/physical model of Yucca Mountain. The investigations are the following:

Investigation 8.3.1.3.1 Water chemistry within the potential emplacement horizon and along potential flow paths.

The water chemistry within the saturated zone has been well characterized, but more data are needed for the unsaturated zone (Section 8.3.1.2.2). A model of the ground-water chemistry for Yucca Mountain relating composition to the geologic location along potential pathways is also needed.

The information needed is as follows:

1. Water chemistry of the unsaturated zone.
2. Additional baseline data on naturally occurring radionuclides in Yucca Mountain ground water.
3. Particulate content of Yucca Mountain ground water and collection.
4. More data on the mineralogic controls on Yucca Mountain ground water.
5. Model of the ground-water chemistry at Yucca Mountain.

Investigation 8.3.1.3.2 Mineralogy, petrology, and rock chemistry within the potential emplacement horizon and along potential flow paths.

The host rock in the Topopah Spring Member has been described, including the primary phenocryst and vapor-phase constituents. The discussion is found mostly in Chapter 1. Also discussed are the secondary minerals within fractures. The data were assessed for usefulness in indicating stratigraphic position and in characterizing the mineralogy along potential radionuclide travel paths. The chemistry and mineralogy of the surrounding units were detailed including a discussion of the secondary minerals with emphasis on the sorptive zeolites and clays.

More data are still needed to assess (1) how the mineral distributions at Yucca Mountain are going to affect retardation by sorption, (2) what processes account for the minerals found at Yucca Mountain, (3) whether those processes have been completed or are still operating, and (4) how the projected processes would affect a repository at Yucca Mountain. The following list of parameters address these needs:

1. Mineral distributions in bulk rock.
2. Mineral distributions in fractures.
3. Bulk rock chemistry.
4. Fracture chemistry.

5. Mineral origins and alteration history.
6. Mineralogy of transport pathways.

Investigation 8.3.1.3.3 Stability of minerals and glasses.

Mineral stability in Yucca Mountain was considered in two ways. The first consideration was whether the present mineral assemblages in Yucca Mountain are stable under present conditions and, if they are not, what mineralogic changes are probable. The presence or absence of clays, cristobalite, glass, and zeolites was studied to assess past thermal gradients. The transformation of cristobalite to quartz was studied in order to understand the evolution of silica activity, which controls the stability of clinoptilolite, a sorbing mineral. The second consideration was what effect repository heating will have on mineral stability, due to both increased temperature and possible dehydration of the rock. The hydrothermal alterations of zeolites, smectites, and rhyolitic glasses were studied to determine alteration products, volume changes, and to understand the kinetics of these mineral reactions. More work is needed to further define the stability of minerals and glasses in every area mentioned previously.

The information needed is as follows:

1. Data on mineral assemblages and temperatures in active hydrothermal systems in vitric tuff.
2. Data on dehydration in smectites, zeolites, and glasses. (From 8.3.1.3.2)
3. Data on the kinetics of silica phase transformations and correlations to aqueous silica activity.
4. Data on the kinetics of transformations in smectites, zeolites, and related tectosilicates.
5. Data for constructing thermodynamic models of clinoptilolite, analcime, and albite.
6. Development of a conceptual model of mineral evolution.

Investigation 8.3.1.3.4 Radionuclide retardation by sorption processes along flow paths to the accessible environment.

The retardation process of sorption has been thoroughly considered in this chapter. Sorption as a function of ground-water composition, mineralogy, pH, temperature, oxidation-reduction conditions, sorbing species, and waste element concentrations have been considered. Detail was provided on the analytical techniques used to determine sorptive coefficients and a discussion of the techniques for elucidating dynamic processes and how they are coupled to sorption.

Work must continue on all the previously discussed process parameters in order to assess the interrelationships of water chemistry, mineralogy,

solubility, and speciation, and the effect of these processes on the retardation process of sorption in expected and unexpected conditions.

The information needed is as follows:

1. Data on sorption as a function of ground-water composition, mineralogy, and sorbing element concentration.
2. Extension of the sorption data base, specifically sorption coefficients correlated with stratigraphy.
3. Data on sorption on particulates and colloids and its relation to transport.
4. Data on the affects of microbial activity on sorption and the subsequent relation to transport.
5. Work on the statistical analysis of sorption data.
6. Model for sorptive behavior of important waste elements.

Investigation 8.3.1.3.5 Radionuclide retardation by precipitation along flow paths to the accessible environment.

Precipitation processes will limit radionuclide transport when the solubility limits of waste elements are exceeded in water as it moves toward the accessible environment. Speciation, defined as the formation of various complexes and oxidation states in the aqueous phase, in turn affects the solubility. This chapter briefly discussed speciation, solubility, radio-colloid formation, and radiolysis, all processes important in understanding concentration limits of elements in solution. The few solubility measurements that have been conducted were discussed along with the modeling work needed to further calculate waste-element solubilities.

To continue the experimental work it is necessary to define the conditions that control solubility (water chemistry, temperature, radiation field) and to define the waste element behavior in order to best understand precipitation processes.

The information needed is as follows:

1. Experimental data on solubility or concentration limits of important waste elements.
2. Experimental data on the ability of waste elements to form natural colloids.
3. Data on, or theoretical understanding of, the sensitivity of solubility limits in controlling parameters. Theory will be developed from speciation measurements and solubility modeling.
4. Solubility calculations.

Investigation 8.3.1.3.6 Radionuclide retardation by dispersive, diffusive, and advective transport processes along flow paths to the accessible environment.

The retardation of radionuclides by dispersive, diffusive, and advective processes was discussed in this chapter in relation to the anticipated and unanticipated conditions at Yucca Mountain. The chapter discussed the processes of radionuclide transport (diffusion, fracture flow, anion exclusion, hydrodynamic dispersion, and particulate (colloids or solids) transport. Some of the transport processes retard the rate of radionuclide transport. Matrix diffusion was discussed along with matrix diffusion coupled with fracture flow. Although dispersion is not a transport mechanism, the process also retards radionuclides by channeling (Non-Fickian dispersion) and the broadening effect of heterogeneously distributed material properties. The transport processes are also intimately associated with sorption and with geochemical processes that result in precipitation.

The transport of tracers in geohydrologic systems is not simple and the simplifying assumptions must be tested by laboratory and field experiments to understand fully the relationship of these processes to the retardation of radionuclides. Extensive laboratory work is needed to successfully model the processes.

The information needed is as follows:

1. Data on diffusion processes (advection and no advection).
2. Data on matrix diffusion and its relationship to fracture flow.
3. Data on the relationship of adsorption to these transport processes.
4. Data on sorption kinetics.
5. Data on anion exclusion.
6. Data on colloidal movement and particulate transport.
7. Data on dispersion (hydrodynamic, channeling).

Investigation 8.3.1.3.7 Radionuclide retardation by all processes along flow paths to the accessible environment.

The modeling of radionuclide transport requires that chemical reactions of radionuclides occurring between the solid phase (minerals or precipitates) and aqueous phase (fracture or pore water) be coupled with the hydrologic flow and physical processes like diffusion and dispersion. The combination of all processes can be used to assess the retardation of radionuclides. This chapter discussed geochemical retardation in the host rock for anticipated and unanticipated conditions. The computer models discussed have the capability to include most of the known physical and geochemical transport processes. The preliminary work on radionuclide retardation was reported. Also discussed was the effect of the thermal load on radionuclide migration.

A conceptual geochemical-geophysical model must be developed to provide the baseline data needed for integrated radionuclide transport calculations. This baseline data set is incomplete; however, the calculation can be an iterative process as better and more complete data from geochemical, mineralogical-petrological, hydrological, and other pertinent NNWSI Project tasks are obtained. The significance and relative importance of each physical and geochemical process affecting transport also needs to be assessed.

The information needed is as follows:

1. Data from previous investigations listed and other NNWSI Project tasks (geochemical characteristics, source terms, rock characteristics, and geohydrologic characteristics).
2. Geochemical-geophysical model of Yucca Mountain.
3. Integrated transport calculations.
4. Significance and relative importance of processes affecting transport (sensitivity analysis).
5. Continued development of transport models.
6. Integration of field test results (geochemical).

Investigation 8.3.1.3.8 Retardation of gaseous radionuclides along flow paths to the accessible environment.

Because the repository may be in the unsaturated zone, this chapter discussed gaseous transport as a potentially significant mechanism for the transport of volatile radionuclide. Two mechanisms are likely to be effective: (1) isotopic exchange between the gas phase and aqueous phase, and (2) the solubility of the gaseous species in the aqueous phase. No experimental work has been done to date on the geochemical aspects of gaseous transport.

The information needed is as follows:

1. Gaseous species.
2. Gas-phase composition.
3. Physical transport mechanisms and rates.
4. Retardation mechanisms and transport with retardation.
5. Gaseous radionuclide transport calculations.
6. Experimental verification of gaseous transport calculations (gas-transport measurements).

4.5.4 RELATION TO REGULATORY GUIDE 4.17

Regulatory Guide 4.17, Part B, Section 6 (NRC, 1987) provides generic guidance for geochemical topics to be addressed in a site characterization report to the NRC. All of the geochemical topics called for in Regulatory Guide 4.17 are relevant and applicable to the Yucca Mountain site. Most of these are addressed primarily in this chapter, although some are discussed only briefly and with reference to more complete treatments in Chapters 3, 7, and 8 of this document.

Nuclear Waste Policy Act
(Section 113)

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***Site Characterization
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***Yucca Mountain Site, Nevada Research
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**CLIMATOLOGY AND
METEOROLOGY**

Consultation Draft

***Site Characterization
Plan***

***Yucca Mountain Site, Nevada Research
and Development Area, Nevada***

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Chapter 5

CLIMATOLOGY AND METEOROLOGY

INTRODUCTION

Past, present, and future climatic conditions at Yucca Mountain need to be characterized in order to design and predict the performance of a geologic repository. Meteorological conditions, or weather forecasts, must be considered in engineering design, surface facilities placement, and radiological safety assessment, and will serve as input to investigations, including rainfall-runoff assessments, to be performed during site characterization. Climatology involves the study of long-term manifestations of weather. Further evaluation of the site must address the potential for climatic changes that could alter the long-term waste isolation capability of the site. This chapter describes and evaluates data on the existing climate and site meteorology, and outlines the suggested procedures to be used in developing and validating methods to predict future climatic variation.

This chapter addresses the following performance and design issues:

<u>Issue</u>	<u>Short title</u>
1.1	Total system performance (Section 8.3.5.13)
1.6	Ground-water travel time (Section 8.3.5.12)
1.8	NRC siting criteria (Section 8.3.5.17)
1.9	Higher-level findings (postclosure) (Section 8.3.5.18)
1.10	Waste package characteristics (postclosure) (Section 8.3.4.2)
1.11	Configuration of underground facilities (postclosure) (Section 8.3.2.2)
1.12	Seal characteristics (Section 8.3.3.2)
2.1	Public radiological exposures--normal conditions (Section 8.3.5.3)
2.2	Worker radiological safety (Section 8.3.5.4)
2.3	Accidental radiological releases (Section 8.3.5.5)
2.5	Higher-level findings--preclosure radiological safety (Section 8.3.5.6)
2.7	Repository design criteria for radiological safety (Section 8.3.2.3)

The information in this chapter will also be used by the geohydrology, geochemistry, climate, erosion, meteorology, population density, offsite installations, surface characteristics, and preclosure hydrology testing programs. Connections between climatology and meteorology and these issues and testing programs will be identified, and plans for the collection of additional data will be referenced to sections of Chapter 8.

Meteorological data have been collected from monitoring stations operated by the National Weather Service, and at stations on the Nevada Test Site (NTS) and Yucca Mountain. Key meteorological data that indicate local climate are temperature, precipitation, and atmospheric moisture, and to a lesser extent, wind speed and wind direction. Data from a range of elevations are needed to describe the site climate (Figure 5-1). Table 5-1 provides specific information (elevation, period of record, etc.) on each of these stations. Some of the parameters listed in Table 5-1 for the site monitoring program are not specifically related to climate determinations or design considerations but have been included because they will be used in subsequent permitting and licensing activities that are not directly related to site characterization. The environmental monitoring and mitigation plan (EMMP) is being developed to ensure that site characterization activities will not result in significant adverse environmental impacts. The EMMP will describe how the site-specific meteorological data will be used in determining impacts associated with site characterization. The meteorological data will also be needed in obtaining permits, as outlined in the environmental regulatory compliance plan (ERCP).

Section 5.1 describes the recent local climate based on temperature, precipitation, upper air, surface winds, atmospheric moisture, and severe weather data collected at the monitoring stations. However, climatic classifications and climatological summaries only provide a general, time-averaged indication of the meteorological conditions at any given point. Because local meteorological data are needed for several of the investigations (e.g., hydrologic studies described in Chapter 3) and because of the uncertainty associated with applying non-site-specific data to Yucca Mountain, climate information must be supplemented with site-specific meteorological data. The meteorological monitoring program at Yucca Mountain is designed to provide such data (refer to Section 8.3.1.12 or the Meteorological Monitoring Plan (SAIC, 1985) for more information on the program).

Fully assessing the Yucca Mountain site as a potential repository must, however, go beyond the essentially short-term operational-phase concerns related to site meteorological conditions. The performance of the repository system over the next 10,000 yr, including climatic variations, will be included in the assessment as required by 10 CFR Part 60. In addition, 10 CFR Part 960 requires a determination of climatic changes over the next 100,000 yr. To address the performance and design issues listed at the beginning of this chapter, the objective of the climate assessment described in Section 5.2 is to provide climatic data that will be used to estimate infiltration parameters. In turn, these hydrologic infiltration parameters will be used to estimate the resulting effects on the nature and rates of erosion and on the hydrologic and geochemical characteristics at Yucca Mountain.

Figure 5-1. Meteorological monitoring stations in the vicinity of Yucca Mountain. Stations NTS-10 Yucca Mountain, NTS-10 Coyote Wash, NTS-10 Alice Hill, NTS-10 Fortymile Wash and NTS-60 Repository are operated as part of the NNWSI Project. Stations at Yucca Flat, Beatty, Desert Rock, and Amargosa Valley are or were operated by the National Weather Service. Stations T-6, Area 12 Mesa, BJJ, 4, 4JAn, 4JAo, and 5A were or are operated in conjunction with various NTS activities. Stations YA (too close to NTS-60 Repository to be shown on this map) and YR (too close to NTS-10 Yucca Mountain to be shown on this map) were also operated as part of the NNWSI Project.

CONSULTATION DRAFT

Table 5-1. Information on meteorological monitoring stations in the vicinity of Yucca Mountain

Station and location ^a	Elevation in meters, above MSL ^b	Meteorological parameters	Period of record
Yucca Flat (UCC) ^c 680,875 ft E 803,600 ft N	1,196	Temperature, relative humidity, precipitation at surface Wind speed and wind direction at surface Wind speed, wind direction, temperature, relative humidity at upper levels	1962-1971 1961-1978 1957-1964
Beatty ^c 481,250 ft E 795,830 ft N	1,006	Temperature, precipitation	1922-1960 1931-1960
BJY ^d 679,100 ft E 842,300 ft N	1,241	Precipitation, wind speed, wind direction	1960-1981 1957-1964
Desert Rock (DRA) ^c 686,719 ft E 682,790 ft N	1,005	Precipitation	1963-1981
4JAn ^d 610,605 ft E 740,840 ft N	1,043	Precipitation	1967-1981
4JAo ^d 617,000 ft E 748,000 ft N	1,100	Precipitation	1957-1967
T6 ^d 458,789 ft E 745,662 ft N	992	Precipitation	1958-1964
Area 12 Mesa ^d 631,450 ft E 889,090 ft N	2,280	Wind speed, wind direction	1957-1964
4 ^d 620,000 ft E 752,000 ft N	1,138	Wind speed, wind direction	1956-1962

Table 5-1. Information on meteorological monitoring stations in the vicinity of Yucca Mountain (continued)

Station and location ^a	Elevation in meters above MSL ^b	Meteorological parameters	Period of record
5A ^d 599,150 ft E 742,050 ft N	1,111	Wind speed, wind direction	1958-1966
YA (Yucca Alluvial) ^e 569,722 ft E 761,794 ft N	1,128	Precipitation	1983-1984
YR (Yucca Ridge) ^e 559,238 ft E 763,555 ft N	1,469	Precipitation	1983-1984
Amargosa Valley ^c 578,819 ft E 689,580 ft N	817	Temperature	1949-1976
NTS-60 Repository ^f 569,127 ft E 761,795 ft N	1,143	Wind speed, wind direction, standard deviation of wind direction, tempera- ture, temperature difference, net radiation, standard deviation of vertical wind speed, precip- itation, dew point	December 1985- present
NTS-10 Yucca Mountain ^f 558,862 ft E 766,434 ft N	1,463	Wind speed, wind direction, standard deviation of wind direction, tempera- ture, relative humidity, precipita- tion	December 1985- present
NTS-10 Coyote Wash ^f 562,876 ft E 766,195 ft N	1,274	Wind speed, wind direction, standard deviation of wind direction, tempera- ture, relative humidity, precipita- tion	December 1985- present

CONSULTATION DRAFT

Table 5-1. Information on meteorological monitoring stations in the vicinity of Yucca Mountain (continued)

Station and location ^a	Elevation in meters, ^b above MSL	Meteorological parameters	Period of record
NTS-10 Alice Hill ^f 576,810 ft E 769,661 ft N	1,234	Wind speed, wind direction, standard deviation of wind direction, tempera- ture, relative humidity, precipita- tion	December 1985- present
NTS-10 Fortymile Wash ^f 580,882 ft E 733,230 ft N	953	Wind speed, wind direction, standard deviation of wind direction, tempera- ture, relative humidity, precipita- tion	December 1985- present

^aAll coordinates are based on the Nevada Central grid.

^bMSL = mean sea level.

^cSite operated by the National Weather Service.

^dSite operated in conjunction with various NTS activities.

^eSite previously operated as part of the NNWSI Project.

^fSite operated as part of the Yucca Mountain meteorological monitoring program.

The data set will consist of seasonal averages of air temperature, relative humidity, cloud cover, surface wind speed, and the type, amount, duration, and intensity of precipitation. Plans for collection of such data are given in Sections 8.3.1.5 and 8.3.1.12. Using these and other data sets as input, potential changes in the rate of infiltration (flux) will be estimated.

To estimate the range and recurrence intervals of future climatic variations and the impact that the variations would have in the vicinity of Yucca Mountain, the nature and potential effects of paleoclimatic variation over the Quaternary Period must be evaluated as required by 10 CFR 960.4-2-4 and 10 CFR 60.122. Because records of Quaternary climate do not exist, climatological proxy data, derived primarily from geologic investigations of past biota and lakes, must be relied upon.

The use of paleobotanic proxy data relies on the fact that climate influences the type and amount of vegetation in an area and also influences the altitudinal range of various types of vegetation. The use of

paleohydrologic proxy data is based on the fact that the climate has a significant influence on surface-water systems. In principle, knowledge of past changes in plant distributions and of changes in lake levels and water chemistry will provide the basis for reconstructing the underlying casual climatic variations.

Section 5.2 discusses the availability of published proxy data and describes a general strategy for using these data in the reconstruction of past climatic variations and in the characterization of future climates. The climatic interpretation of the proxy data involves the following steps:

1. Using transfer functions or response surfaces to correlate statistically present-day meteorological data and plant distributions.
2. Obtaining dated records of paleobotanic variations through the analysis of plant macrofossils from pack rat middens and of fossil pollen from cores of lacustrine sediment.
3. Applying the transfer functions, response surfaces, or both to the paleobotanic time series to obtain estimates of past climatic variations.
4. Constructing climatic descriptions (synoptic snapshots) for critical time periods for the region over which the paleobotanic data are available.
5. Validating these climatic reconstructions with information for the paleolimnological data.
6. If possible, subjecting paleoclimatic time series to appropriate forms of spectral or statistical analyses to determine the frequencies of climatic variations.

The paleobotanical and paleolimnological data, and the associated paleoclimatic reconstructions, have three main applications:

1. Documentation. The reconstructions of past climatic variations based on the proxy data will serve to document the extent of past climatic variations in the vicinity of Yucca Mountain. These past climatic variations will illustrate the probable future climatic conditions that may occur under boundary conditions of the climate system similar to those that have occurred in the past.
2. Input to hydrologic studies. The climatic interpretations derived from proxy data will provide information such as estimates of precipitation, temperature, and seasonality for studies on the paleohydrology of the Yucca Mountain region. The relationship(s) established among past climatic variations and the resultant hydrological changes will be used in the prediction of future hydrologic variations.
3. Model validation. Mathematical models must be validated before they can be used to predict future climatic variations or scenarios. In

CONSULTATION DRAFT

other words, their performance in simulating the climate must be measured under conditions other than those used to formulate or calibrate the models. The paleobotanical and paleolimnological records will provide tests of simulations of past climatic variations.

The future climate model and its assumptions are discussed in Sections 5.2 and 8.3.1.5.1.

The uncertainty associated with data presented in Chapter 5 is primarily related to inherent inaccuracies in measurements and an insufficient number of samples and gaps in the paleoclimatic record. The testing, sampling, and modeling plans discussed in Sections 8.3.1.2, 8.3.1.5, and 8.3.1.12 are structured (1) to reduce these data uncertainties associated with the paleoclimatic and modern meteorological data and (2) to improve modeling of future climate. Uncertainties associated with data presented in this chapter include:

1. Modern meteorological data--the historical records cover only limited time periods and limited areal and elevational ranges.
2. Modern ecological data--the data do not sufficiently cover the necessary elevational and geographic ranges required for developing climate-proxy data calibration equations.
3. Paleoclimatic proxy data--these data are found in very restricted depositional environments and cover limited temporal and spatial ranges. In addition, limited abundance of these proxy data in the Great Basin and especially in the Yucca Mountain area results in gaps in the record, increasing the uncertainty associated with paleoclimatic reconstructions.
4. Age assignments for pack rat middens, lake cores, and pollen samples--radiometric dating techniques introduce uncertainty with increasing age, decreasing sample size, and contamination.
5. Modeling--both global climate models and regional models rely on input in the form of boundary conditions that are either estimated or else provided by other modeling activities that have their own levels of uncertainty. The models contain many assumptions and simplifications that limit their resolution.

These, and any other areas of uncertainty will be reduced through further data collection, additional dating techniques, and model sensitivity studies described in Sections 8.3.1.2, 8.3.1.5, and 8.3.1.12.

5.1 RECENT CLIMATE AND METEOROLOGY

Although a major emphasis throughout the site characterization process will be on assessing the ability of the selected host rock to contain stored wastes, the climate and site-specific meteorology of the Yucca Mountain area

can influence some important aspects of repository development. The design and operation of the repository must consider climatic influences to ensure that surface facilities are capable of withstanding expected meteorological conditions (e.g., the design of the ventilation system). The potential for flooding must also be considered in the siting of surface facilities. In addition to these short-term design considerations, defining the existing climatic conditions establishes the basis for comparing the future and past climates with the present climate. Further, establishing the existing climatic conditions and the infiltration rates associated with these conditions is important in evaluating whether climatic variations will affect infiltration rates and subsequent rises or declines in the water table in the Yucca Mountain area. The sections that follow provide both a general description of the climate of the Yucca Mountain area (Section 5.1.1) and discussions of specific atmospheric parameters that are important in establishing the climatic conditions at Yucca Mountain (Sections 5.1.1.1 through 5.1.1.6).

5.1.1 CLIMATE

Long-term site-specific climatological data for Yucca Mountain are not presently available. Five monitoring stations have been established but the period of record is less than 2 yr. Therefore, data from two weather stations near Yucca Mountain that were operated by the National Weather Service (NWS) have been used to provide a general description of the climate in the area (Figure 5-1). One of these stations is located approximately 32 km northeast of the Yucca Mountain site in Yucca Flat, a broad alluvial basin. The other station is near Beatty, Nevada, approximately 24 km west of the Yucca Mountain site. Supplemental information collected at various other locations on the Nevada Test Site covering varying time periods is also available and has been used to describe the climate of the Yucca Mountain site for comparative purposes. A meteorological monitoring program, described in Sections 5.1.3 and 8.3.1.12, will provide site-specific data on the meteorological conditions that are likely to influence site characterization activities or repository development.

The Yucca Mountain site is situated in an area bordering two NWS climatological zones of Nevada: south central and extreme southern (Bowen and Egami, 1983b). The distinction between these two classifications is governed mostly by elevation. Lower elevations in the vicinity of Yucca Mountain experience conditions typical of southwestern desert zones within the United States that are characterized by hot summers, mild winters, and limited amounts of precipitation. Higher elevations have less severe summer temperatures and greater but still limited amounts of precipitation. The general climatological classification of midlatitude desert, in a modified Koeppen system presented in Critchfield (1983), also can be used to describe conditions in and around the Yucca Mountain site. Midlatitude desert areas are far removed from windward coasts and are dominated by tropical and polar air masses. For areas classified as midlatitude desert, summers are dominated by continental tropical air masses and winters are dominated by continental polar air masses. Large annual and diurnal fluctuations in temperature are characteristic of midlatitude desert areas, as is significant variability in precipitation from year to year.

The major air masses affecting the weather of the Yucca Mountain area during winter months originate either over the Pacific Ocean or over polar continental regions. Most of the moisture carried by the Pacific air masses does not reach the Yucca Mountain area because the physical (orographic) lifting effect of the Sierra Nevada forces the air masses to higher elevations as they move eastward, thus cooling the air masses and lowering their ability to hold moisture. The moisture that cannot remain in the vapor phase, due to this cooling, falls as precipitation on the western slopes of the Sierra Nevada. When an air mass has passed the ridge of the mountains, it descends along the eastern slopes and warms again, creating what is called a rain shadow in the lee of the Sierra Nevada (Wallace and Hobbs, 1977). Yucca Mountain and the surrounding areas lie within this shadow. Polar continental air masses bring cold, dry air into the area but are not as common a winter phenomenon as are the Pacific air masses.

A thermally induced area of low pressure is created during the summer months over most desert regions (Wallace and Hobbs, 1977) and prevails over the southwestern United States during summer. Although this thermal low is generally associated with weak cyclonic motion (Huschke, 1959), it brings south to southwesterly winds to the Yucca Mountain area. However, this circulation pattern is essentially nonfrontal. Although summer is generally the driest time of the year, this circulation pattern can bring tropical moisture originating over the Pacific Ocean off the lower coast of California to the area, which in combination with the strong solar insolation during the summer can create thunderstorm activity. Another less frequent summer circulation pattern that brings moisture to the area is a semipermanent subtropical high-pressure system called a Bermuda High (Huschke, 1959). If this system becomes well developed, it can bring moisture from the Gulf of Mexico to the Yucca Mountain area with southeasterly winds, again resulting in thunderstorm activity.

In addition to these synoptic-scale climatic influences, the rugged terrain of the Yucca Mountain area can create micrometeorological variations of a given parameter within relatively short distances. Drainage winds are an example of this phenomenon. These terrain-dependent winds can locally affect wind speed, wind direction, and temperature (Eglington and Dreicer, 1984), and are most pronounced under calm (synoptic-scale) conditions during cloudless nights (Huschke, 1959). The ground surface quickly cools under these conditions by radiating its heat into the atmosphere. Air very near the surface is subsequently cooled. This cooler, denser air then drains down the terrain. In closed topographic basins such as Yucca Flat, the cold air essentially fills the basin and can significantly lower the temperature in the basin with respect to temperatures at the rims of the basin. The temperature inversion (temperature increasing with height) thereby created can limit effective atmospheric dispersion within the basin. However, these conditions generally dissipate quickly after sunrise as the ground surface is heated by the sun.

Another example of micrometeorological variations induced by the terrain is the variability in precipitation amounts between stations at different elevations and between those with differing exposure to prevailing storm tracks or occurrences. This issue is dealt with in detail in Sections 8.3.1.2 and 8.3.1.12, which outline plans for a precipitation monitoring

network designed to characterize the influence that storm location and precipitation amounts would have on runoff and infiltration.

Table 5-2 provides a general outline of the climatic conditions experienced at Yucca Flat during the 10-yr period from 1962 to 1971 (Bowen and Egami, 1983a). Three important parameters at the Yucca Mountain site that will probably differ from the Yucca Flat summary are temperature minimums, wind speeds, and direction. Also, precipitation amounts are expected to be greater at Yucca Mountain because of its higher elevation (1,463 m at the NTS-10 Yucca Mountain station versus 1,196 m at the Yucca Flat station). These parameters are expected to exhibit terrain influences that will be characterized through analysis of the data collected in the site meteorological monitoring program.

The link between synoptic-scale processes and their effect on site-specific meteorology is not known at this time because the data are not yet available. The multitower monitoring program implemented at Yucca Mountain (Section 8.3.1.12) is designed specifically to collect data that can be used to characterize the relationships between site conditions and regional weather systems. Plans for collecting regional data and establishing this relationship are presented in Section 8.3.1.12. In addition, the monitoring program will provide general data on terrain influences and specific data on drainage winds.

Terrain influences are an important factor to consider in site meteorology (Quiring, 1968). Such influences can result in wide variations of any particular parameter within relatively small differences in location or elevation. The temperature at the site is expected to fluctuate between wide limits under predominantly clear skies and low relative humidities. Summer temperatures in excess of 100°F (38°C) are expected to be common, as are winter temperatures below 32°F (0°C). Although extreme lows may reach below 0°F (-18°C), such occurrences are not common (Quiring, 1968).

Precipitation is expected to be minimal but flash flooding, due to thunderstorm activity, may occur irregularly. Normal precipitation patterns for the area indicate a relative maximum in January or February dropping to a low in June. A secondary peak occurs in July and August (or both) with a secondary low in October. However, significant storm events can disrupt this pattern. Snow does occur in the area, particularly at elevations higher than those at Yucca Mountain.

Predominantly northerly winds prevail because of synoptic influences in the fall and winter months, but south to southwesterly winds become dominant during the spring and summer. If the mountain-valley winds experienced elsewhere in the vicinity of Yucca Mountain (Quiring, 1968) occur at the site as expected, the winds during daylight hours will almost always be upslope during the warmest part of the day and will vary in direction, depending on the orientation of the terrain.

Records of meteorological parameters that together describe the climate of the region include temperature, precipitation, atmospheric moisture, surface wind speed and direction, upper air wind speed and direction, and severe weather phenomena. Each of these variables is discussed in detail in the sections that follow.

Table 5-2. Climatological summary for Yucca Flat, Nevada, 1962 to 1971^a (page 1 of 2)

TEMPERATURE ^b (°F)										DEGREE DAYS (Base 65°)		PRECIPITATION ^{b,c} (INCHES)									
MONTH	AVERAGES			EXTREMES				AVERAGE	GREATEST MONTHLY	YEAR	LEAST MONTHLY	YEAR	GREATEST DAILY	YEAR	AVERAGE	GREATEST MONTHLY	YEAR	GREATEST DAILY	YEAR		
	DAILY MAXIMUM	DAILY MINIMUM	MONTHLY	HIGHEST	YEAR	LOWEST	YEAR													HEATING	COOLING
JAN	52.1	20.8	36.5	73	1971	-2	1970	877	0	.53	4.02	1969	T	1971#	1.25	1969	0.9	4.3	1962	4.3	1962
FEB	56.7	25.8	41.3	77	1963	5	1971#	662	0	.84	3.55	1969	T	1967#	1.16	1969	1.9	17.4	1969	6.2	1969
MAR	60.9	27.7	44.3	87	1966	9	1969	634	0	.29	.60	1969	.02	1966	.38	1969	2.0	7.5	1969	4.5	1969
APR	67.8	34.4	51.1	89	1962	13	1966	411	1	.45	2.57	1965	T	1962	1.08	1965	0.7	3.0	1964	3.0	1964
MAY	78.9	43.5	61.2	97	1967	25	1967	147	38	.24	1.62	1971	T	1970#	.86	1971	0	T	1964	T	1964
JUN	87.6	49.9	68.8	107	1970	29	1971#	35	154	.21	1.13	1969	T	1971	.45	1969	0	0		0	
JUL	96.1	57.0	76.6	107	1967	40	1964#	0	366	.52	1.34	1966	0	1963	.77	1969	0	0		0	
AUG	95.0	58.1	76.6	107	1970	39	1968	1	368	.34	1.04	1965	0	1962	.35	1971#	0	0		0	
SEP	86.4	46.7	66.5	105	1971	25	1971	51	103	.68	2.38	1969	0	1968#	2.13	1969	0	0		0	
OCT	76.1	36.9	56.5	94	1964+	12	1971	266	9	.13	.45	1969	0	1967#	.42	1969	0	T	1971	T	1971
NOV	61.8	27.6	44.7	82	1962	13	1966	602	0	.71	3.02	1965	0	1962	1.10	1970	0.5	4.8	1964	2.3	1964
DEC	50.7	19.9	35.3	70	1964	-14	1967	914	0	.79	2.66	1965	T	1969#	1.31	1965	2.3	9.9	1971	7.4	1971
ANN	72.5	37.4	54.9	107	AUG 1970#	-14	DEC 1967	4600	1039	5.73	4.02	JAN 1969	0	SEP 1968#	2.13	SEP 1969	8.3	17.4	FEB 1969	7.4	DEC 1971

Table 5-2. Climatological summary for Yucca Flat, Nevada, 1962 to 1971^a (page 2 of 2)

MONTH	RELATIVE HUMIDITY (%)				WIND b,d (SPEEDS-IN MPH)				STATION PRESSURE (INCHES)			(e) AVERAGE SKY COVER SUNRISE TO SUNSET	AVERAGE NUMBER OF DAYS ¹														
	HOUR (PACIFIC STANDARD TIME)				AVERAGE SPEED	PEAK SPEED	YEAR	RESULTANT (DIR/SP)		AVERAGES	HIGHEST		LOWEST	SUNRISE TO SUNSET			PRECIPITATION					1.0 INCH OR MORE OF SNOW	THUNDERSTORMS	TEMPERATURE			
								23-02 PST	11-14 PST					CLEAR	PARTLY CLOUDY	CLOUDY	.01 INCH OR MORE	.10 INCH OR MORE	.50 INCH OR MORE	1.00 INCH OR MORE	90° F OR MORE			32° F OR LESS	32° F OR LESS	0° F	
JAN	67	49	35	60	6.6	58	1965	233/0.7	135/2.6	26.10	26.54	25.42	4.9	13	8	10	2	1	*	*	*	*	0	1	29	*	
FEB	67	45	32	56	6.9	52	1967	275/1.1	118/2.7	26.05	26.42	25.56	5.0	11	8	9	3	2	*	*	1	0	0	*	23	0	
MAR	58	31	23	44	8.4	55	1971	240/1.8	186/4.5	25.99	26.43	25.48	4.8	12	9	10	3	1	0	0	1	1	0	0	24	0	
APR	52	27	21	38	9.1	60+	1970*	250/2.2	198/5.1	25.96	26.39	25.50	4.5	13	9	8	3	1	*	*	*	1	0	0	12	0	
MAY	46	22	17	31	8.3	60+	1967	260/1.5	179/7.2	25.94	26.39	25.47	4.3	14	11	6	2	1	*	0	0	1	4	0	2	0	
JUN	39	19	14	26	7.9	60+	1967	272/1.9	185/8.2	25.92	26.20	25.56	3.0	19	7	4	2	1	0	0	0	2	14	0	*	0	
JUL	40	20	15	28	7.5	55	1971	278/0.9	185/12.0	26.00	26.19	25.68	3.0	19	9	3	3	2	*	0	0	4	29	0	0	0	
AUG	44	23	16	30	6.7	60+	1968	222/1.5	182/12.0	26.00	26.22	25.71	3.0	20	8	3	3	1	0	0	0	4	27	0	0	0	
SEP	43	21	17	32	7.0	52	1970	281/1.3	163/6.4	26.00	26.36	25.56	2.1	22	6	2	2	1	1	*	0	2	11	0	1	0	
OCT	46	24	19	36	6.8	60	1971	286/1.3	138/3.7	26.06	26.40	25.52	2.9	20	7	4	1	1	0	0	0	*	2	0	9	0	
NOV	61	39	31	52	6.1	51	1970	234/1.2	152/4.1	26.08	26.58	25.64	4.8	13	7	10	3	2	*	*	*	*	0	0	23	0	
DEC	68	50	41	64	6.6	53	1970	288/1.9	109/1.0	26.07	26.59	25.49	4.6	14	8	9	3	1	1	*	1	*	0	1	29	1	
ANN	53	31	23	41	7.4	60+	APR 1970*	—	—	26.01	26.59	25.42	3.9	190	97	78	30	14	3	1	3	14	87	2	152	1	

^aSource: Bowen and Egami (1983a). Blanks indicate not applicable.

^b# = most recent of multiple occurrences.

^cT - trace (amount too small to measure).

^dAverage and peak speeds are for the period December 1964 through May 1969. The directions of the resultant wind are from a summary covering the period December 1964 through May 1969.

^eSky cover is expressed in the range from 0 for no clouds to 10 when the sky is completely covered with clouds. Clear, partly cloudy, and cloudy, are defined as average daytime cloudiness of 0-3, 4-7, and 8-10, respectively.

^f* = one or more occurrences during the period of record but average less than one-half day.

5.1.1.1 Temperature

Temperature in the site vicinity varies widely on both a diurnal and annual basis (Quiring, 1968; Eglinton and Dreicer, 1984). Temperature data from the 10-yr climatological summary for Yucca Flat and the 39-yr (1922 to 1960) period of record at Beatty are presented in Table 5-3. These data suggest that Beatty is generally warmer and experiences higher maximum temperatures and less severe minimum temperatures than Yucca Flat. Temperature at the Yucca Mountain site is expected to closely resemble the Yucca Flat data. General temperature cycles and ranges of expected temperature values for the Yucca Mountain site are discussed below.

The lowest temperatures generally occur during the months of November through March. During this period minimum temperatures of $<32^{\circ}\text{F}$ (0°C) occur on 85 percent of the days at Yucca Flat and 57 percent of the days at Beatty. On an annual basis, Yucca Flat experiences 152 days (42 percent) with minimum temperatures $<32^{\circ}\text{F}$ (0°C) and Beatty experiences 88 days (24 percent) in the same range. The lowest average monthly temperature of 35.3°F (1.8°C) at Yucca Flat occurs in December, which is also the month that has the lowest average daily minimum temperature of 19.9°F (-6.7°C) and the lowest recorded temperature (for the 10-yr summary period) of -14°F (-26°C). Although temperatures below 0°F (-18°C) do occur, they are infrequent and commonly occur on less than 1 percent of the days from November through March. For Beatty, the data suggest that January is the coldest month, averaging 40.6°F (4.8°C) with an average daily minimum of 26.7°F (-2.9°C). However, the lowest temperature recorded at Beatty of 1°F (-17°C) occurred during February. Differences in both magnitude and duration of low temperatures between these recording stations are probably due to the drainage flow conditions present at the Yucca Flat site (as discussed previously), which are also expected to occur at the Yucca Mountain site. Despite the apparent frequency of temperatures below 32°F (0°C), daily maxima during the winter months remain quite mild, averaging at least 50°F (10°C).

Summer temperatures at these two recording stations reach relatively extreme levels but are comparable to temperatures observed elsewhere on the NTS (Quiring, 1968). The hottest months for both Beatty and Yucca Flat are June, July, and August, during which daily maximum temperatures $\geq 90^{\circ}\text{F}$ (32°C) occur on 76 percent of the days at Yucca Flat and on 87 percent of the days at Beatty. The highest temperature recorded at Beatty was 114°F (46°C) and occurred in July, which has an average daily maximum of 99.5°F (37.5°C). The highest temperature recorded at Yucca Flat is 107°F (42°C). Although this maximum temperature has occurred in June, July, and August, the warmest month overall is July, with an average daily maximum of 96.1°F (35.6°C) and a monthly average of 76.6°F (24.8°C).

Data from Amargosa Valley, Nevada (formerly Lathrop Wells), for the period 1949 to 1976 are presented in Nichols (1986) and are quite similar to data from Beatty and Yucca Flat. The Amargosa Valley monthly minimums occur in January (as at Beatty) with an average daily minimum temperature of 27°F (-3°C). As at Yucca Flat and Beatty, summer temperatures at Amargosa Valley peak in July with an average daily maximum of 99°F (37°C).

Substantial temperature differences between average daily maximum and average daily minimum are characteristic of the area due to high insolation

CONSULTATION DRAFT

Table 5-3. Yucca Flat and Beatty temperature data^a

Month and Station ^b	Temperature °F				
	Average daily maximum	Highest daily	Average daily minimum	Lowest daily	Monthly average
January					
Yucca Flat	52.1	73	20.8	-2	36.6
Beatty	57.5	78	26.7	7	40.6
February					
Yucca Flat	57.7	77	25.8	5	40.3
Beatty	58.2	80	24.6	1	43.9
March					
Yucca Flat	60.9	87	27.7	9	44.3
Beatty	65.6	85	33.9	16	49.7
April					
Yucca Flat	67.8	89	34.4	13	51.1
Beatty	74.0	98	41.0	17	57.5
May					
Yucca Flat	78.9	97	43.5	25	61.2
Beatty	82.0	103	47.9	26	65.2
June					
Yucca Flat	87.6	107	49.9	29	68.8
Beatty	92.3	112	55.5	33	74.0
July					
Yucca Flat	96.1	107	47.0	40	76.6
Beatty	99.5	114	62.1	36	80.8
August					
Yucca Flat	95.0	107	58.1	39	76.6
Beatty	97.2	113	59.9	41	78.6
September					
Yucca Flat	86.4	105	46.7	25	66.5
Beatty	90.5	109	53.4	34	72.0
October					
Yucca Flat	76.1	94	36.9	12	56.5
Beatty	77.4	98	43.6	22	60.5
November					
Yucca Flat	61.8	84	27.6	13	44.7
Beatty	65.3	86	33.5	10	49.4
December					
Yucca Flat	58.0	70	19.9	-14	35.3
Beatty	56.3	80	28.0	4	42.2
Annual average					
Yucca Flat	72.5	107	37.4	-14	54.9
Beatty	76.4	114	42.9	1	59.5

^aSource: Eglinton and Dreicer (1984).^bYucca Flat period of record: 1962 to 1971; Beatty period of record: 1922 to 1960.

CONSULTATION DRAFT

rates and generally low relative humidities. For the data presented in Table 5-3, the Yucca Flat temperature difference between maximum and minimum values of 39.7 Fahrenheit degrees (22.1 Celsius degrees) in September is the most pronounced, but it differs only slightly from the smallest difference of 30.8 Fahrenheit degrees (17.1 Celsius degrees) in December. In comparison, the Beatty data indicate only a slightly more moderate maximum difference of 37.4 Fahrenheit degrees (20.8 Celsius degrees) in July to a minimum of 28.3 Fahrenheit degrees (15.7 Celsius degrees) in December. As with previously discussed phenomena, the wide temperature range at Yucca Flat is probably due to its location in a basin, where the daily minimum temperatures are lower than at Beatty.

Although the exact relationship between the recording stations and Yucca Mountain is not known, general trends can be inferred from the known data. Summer temperatures in excess of 100°F (38°C) should be expected. Winter minimum daily temperatures below 32°F (0°C) will be relatively common, but temperatures below 0°F (-18°C) will be infrequent. There will probably be substantial temperature ranges, with cool summer nights and mild winter daily maximums during most years.

5.1.1.2 Precipitation

Precipitation in the Yucca Mountain area is associated with two distinct atmospheric circulation patterns. The first of these patterns creates winter frontal passages that are associated with Pacific air masses moving toward the area from the west. Approximately 50 percent of the precipitation in the vicinity of Yucca Mountain occurs as a result of these systems during the months of November through April, even though the entire area lies in the rain shadow of the Sierra Nevada. The second type of circulation pattern that occurs in the area creates a secondary peak in precipitation in the late summer (July and August) and is a result of thunderstorm activities. These storms have a much greater flood potential than the frontal precipitation that occurs during the winter months because the storms can release significant amounts of moisture in relatively short periods of time.

The low pressure area that typically dominates the southwestern United States during the summer results in south to southwesterly winds in the Yucca Mountain area that can transport moisture-laden air into the area. The convective activity associated with this moist air flow can result in strong (intense), isolated downpours throughout the area. If the ground surface cannot absorb this moisture quickly, runoff and flooding can occur. Flooding potential is high; a large number of variables influence thunderstorm activities and govern precipitation types, amounts, and intensity, and individual storms generally complete a thunderstorm cycle in about 2 h (Huschke, 1959).

Specific precipitation amounts for the seven stations located in the vicinity of Yucca Mountain that are considered most representative of conditions at the NTS are presented in Table 5-4 (Eglington and Dreicer, 1984). Monthly and annual average and maximum precipitation amounts at these stations are shown in the table. The data cover periods of record ranging from 5 yr at tower T6 to 29 yr at Beatty (see Figure 5-1 for locations).

Table 5-4. Monthly and annual average and maximum precipitation for sites in the vicinity of Yucca Mountain^a

Month	Precipitation (in.) ^{b,c}													
	BJY 1,241 mMSL (1960-1981)		Yucca Flat 1,196 mMSL (1962-1971)		Desert Rock 1,005 mMSL (1963-1981)		4JAn 1,043 mMSL (1967-1981)		4JAo 1,100 mMSL (1957-1967)		T6 992 mMSL (1958-1964)		Beatty 1,006 mMSL (1931-1960)	
	Avg.	Max.	Avg.	Max. ^d	Avg.	Max.	Avg.	Max.	Avg.	Max.	Avg.	Max.	Avg.	Max.
January	0.76	3.41	0.53	4.02	0.64	2.15	0.63	2.29	0.27	0.62	0.37	0.88	0.60	NA ^e
February	0.87	3.42	0.84	3.60	0.78	2.57	1.08	3.45	0.39	1.01	0.59	1.20	0.70	NA
March	0.73	3.58	0.29	3.50	0.70	3.08	0.83	3.00	0.16	0.35	0.16	0.30	0.48	NA
April	0.34	2.40	0.45	2.70	0.33	1.45	0.18	0.63	0.34	1.91	0.10	0.45	0.47	NA
May	0.33	2.02	0.24	1.62	0.35	1.57	0.31	1.41	0.11	0.28	0.15	0.48	0.23	NA
June	0.21	1.22	0.21	2.66	0.14	0.56	0.13	0.67	0.07	0.26	0.14	0.55	0.09	NA
July	0.48	1.54	0.52	1.87	0.34	1.46	0.35	1.50	0.19	0.48	0.50	2.29	0.20	NA
August	0.45	2.38	0.34	2.52	0.52	1.57	0.31	1.97	0.25	0.71	0.22	0.54	0.20	NA
September	0.53	1.89	0.68	2.38	0.38	2.28	0.28	2.13	0.47	1.68	0.50	1.62	0.19	NA
October	0.36	1.49	0.13	1.69	0.25	1.05	0.32	1.42	0.21	0.63	0.22	0.76	0.30	NA
November	0.50	2.37	0.71	3.02	0.50	2.07	0.33	1.22	0.54	1.67	0.61	1.49	0.43	NA
December	0.57	2.61	0.79	2.66	0.46	2.45	0.33	1.78	0.63	3.03	0.44	1.14	0.58	NA
Annual	6.03	12.13	5.73	14.05	5.39	10.08	5.08	11.62	3.63	8.06	4.00	4.61	4.47	NA

^aSource: Eglinton and Dreicer (1984). The locations of the monitoring stations are shown on Figure 5-1.

^bAll values are monthly or annual averages. To convert in. to mm, multiply by 25.4.

^cmMSL = meters above mean sea level; Avg.=average; Max.=maximum.

^dPeriod of record is 1958-1981.

^eNA = not available.

The maximum 5-yr average precipitation amount of 6.03 in. (153 mm) occurred at tower BJY, and the minimum of these stations was 3.63 in. (92 mm) at tower 4JAo. All the stations follow the characteristic annual precipitation cycle with a winter peak, an early summer minimum, a secondary peak in late summer followed by a secondary minimum in October. This cycle is clearly illustrated in Figure 5-2, which shows precipitation amounts for each month, based on the average data presented in Table 5-4. The maximum values for each month do not exhibit the same annual cycle evident in the monthly averages. Instead, these data indicate that maximum precipitation generally occurred during winter months with maximums for the various towers having occurred in January, February, March, July, and December (Table 5-4). This variability is indicative of individual frontal systems or storms having occurred during these months.

While the data presented thus far are a reasonable indication of overall precipitation cycles, estimating precipitation at the Yucca Mountain site requires further analysis. The major variable that must be considered is elevation as noted by several investigators (Quiring, 1983; Nichols, 1986). A common means of estimating the precipitation at an elevation where no data are available is to perform linear, loglinear, or exponential regressions using precipitation and elevation data from locations in the vicinity of the site in question. Eglinton and Dreicer (1984) have performed such an analysis using data from 14 stations near Yucca Mountain. At the elevation of the surface facilities, 1,143 m above mean sea level, the regression analysis predicts an annual precipitation of 160 mm. Quiring (1983) performed a similar regression analysis for 11 stations located in various parts of the NTS. His information suggests that at the surface facility elevation, approximately 150 mm of precipitation can be expected annually. While these two analyses are in close agreement with regard to the expected amount of precipitation, site-specific data for various locations and elevations are needed to fully evaluate precipitation in the vicinity of the site.

For the operation phase of the repository and during site characterization, the precipitation resulting from short-term thunderstorms of high intensity is a consideration in design and placement of surface facilities, including those for the exploratory shaft, because these storms can lead to flash flooding (sheet flow, stream flow, and debris flow). Data from two monitoring stations that were operated at the Yucca Mountain site from late 1982 until October of 1984 provide some indication of the amount and intensity of precipitation that has occurred at Yucca Mountain (Chu, 1986). The data, however, represent conditions for only approximately 2 yr. Thus, it is impossible to establish trends or develop corollaries between the short-term data and the long-term conditions at Yucca Mountain. The data are included only as an indicator of the potential thunderstorm-related precipitation. One of the monitoring sites, called Yucca Alluvial (YA), was near the location proposed for the surface facilities, and the other site, called Yucca Ridge (YR), was slightly east of the ridge of Yucca Mountain.

The most significant storm event recorded during the 2-yr period occurred on July 21, 1984. On that date, the YA site recorded a 24-h total precipitation amount of 2.54 in. (64.5 mm), 1.75 in. (44.5 mm) of which fell in 1 h. The YR site registered a similarly significant amount of precipitation of 2.73 in. (69.3 mm) for the 24-h period and an hourly maximum of 1.37 in. (34.8 mm). With a predicted annual average of between 6.30 and

Figure 5-2. Monthly average precipitation (in inches) for stations in the vicinity of Yucca Mountain (modified from Eglinton and Dreicer, 1984).

5.70 in. (160 and 145 mm), the individual storm represents a significant and high-intensity event. Several less significant events were recorded during the 2-yr period, in most instances affecting both of the monitoring sites. However, there were occurrences in which one of the sites received precipitation and the other received less, or none at all.

The Yucca Mountain area does receive precipitation in the form of snow, but such occurrences are uncommon (Nichols, 1986). Yucca Flat averages only about 2 in. (50 mm) of snow per month during the winter months. The snow that does fall persists for only a few hours. The greatest daily amount of snowfall recorded at Yucca Flat is 7.4 in. (188 mm), which occurred in December 1971. Snow is not important in terms of overall precipitation amounts but should be considered in the design of the surface facilities.

Because the repository would be located in the unsaturated zone beneath Yucca Mountain, evaluation of the long-term ability of the site to contain stored waste must include a determination of how much of the precipitation falling at the surface infiltrates as potential recharge to the ground water. While thunderstorms are significant events and potentially damaging, they occur in the summer months when soil moisture is low and potential evapotranspiration is high and thus are not likely to result in significant ground-water recharge (Nichols, 1986). The most likely events leading to infiltration that exceed soil moisture deficit and evapotranspiration and could thus lead to percolation through the repository horizon would occur during the winter months. A series of precipitation events with no intermediate drying-out period would represent high-recharge potential (Nichols, 1986). A discussion of hydrologic infiltration and flux through the unsaturated zone is found in Chapter 3.

A more comprehensive precipitation monitoring network is needed both in the immediate vicinity of Yucca Mountain and in sections of the Fortymile Wash drainage to fully evaluate the recharge potential. Plans for such a network are given in Sections 8.3.1.2 and 8.3.1.12, and geohydrologic investigations needed in conjunction with this precipitation data are discussed more completely in Section 3.7.

5.1.1.3 Atmospheric moisture

The same processes that restrict precipitation in the vicinity of Yucca Mountain provide for generally low relative humidity throughout the year. The general diurnal and annual cycle of relative humidity for the area can be derived from the data contained in the climatological summary for Yucca Flat shown in Table 5-2. On the basis of the Yucca Flat data, relative humidity is expected to reach a minimum monthly average of approximately 25 percent in June or July and reach a maximum of around 55 percent in December. Diurnal trends (also shown in Table 5-2) are for the highest relative humidity to occur during the early morning and late evening hours, with the lowest relative humidity occurring in the afternoon hours.

Although relative humidity, a temperature-dependent variable, is commonly used as an accepted measure of atmospheric moisture, it can be a somewhat misleading indicator. A less temperature-dependent indicator of

atmospheric moisture is the wet-bulb temperature or depression, which can either be measured directly or derived from relative humidity and dry-bulb temperature data using a psychrometric chart for the elevation. Because the wet-bulb temperature is always less than or equal to the ambient (dry-bulb) temperature (Huschke, 1959), the difference between the dry-bulb temperature and the wet-bulb temperature is referred to as the wet-bulb depression.

The monthly average relative humidity at hour 1600 and the maximum average temperature data given in the climatological summary for Yucca Flat (Table 5-2) have been used to calculate approximate wet-bulb depression values expected to occur in the vicinity of the Yucca Mountain site. These wet-bulb depression data were calculated using a psychrometric chart given in Weast (1972). The data are shown in relation to the ambient temperature and humidity values for Yucca Flat in Table 5-5. As shown in the table, the wet-bulb depression varies inversely with the relative humidity.

5.1.1.4 Wind speed and direction

Yucca Mountain lies in a geographical region of generally linear mountain ranges that dissect alluvial piedmont valleys with rugged, complex terrain features, as described in Chapter 1. Wind speeds and, more importantly, wind direction will be heavily influenced by the site-specific terrain features. Therefore, extrapolation of wind data from other sites to Yucca Mountain may not be as accurate as extrapolation of other meteorological parameters.

An analysis of winds at four locations on the NTS by Quiring (1968) provides data on conditions that might be experienced at Yucca Mountain. Two of the monitoring towers (4 and 5A) used for this study were located in Jackass Flats, 18 km to the east of the Yucca Mountain site. The third tower (BJY) was located in the middle of Yucca Flat, 12 km north of the Yucca Flat weather station. The fourth tower used in the study was placed high on a mesa overlooking Yucca Flat and is referred to as the Area 12 Mesa tower. The locations of the towers are shown in Figure 5-1.

Data from the towers indicate that wind direction is influenced primarily by two general types of atmospheric activity. First, large-scale pressure systems govern seasonal variations in wind direction and produce predominantly northerly winter winds and predominantly southerly summer winds. Secondary to the overall patterns are terrain-induced wind flow patterns and the effects of ground surface heating and cooling. The influence of these effects is evident in the diurnal wind flow reversal from up the terrain during the day to drainage flow at night (Quiring, 1968). The directions associated with these flows are an artifact of the terrain in the vicinity of the tower. The tower overlooking Yucca Flat was high enough above the valleys and basins to be less influenced by the daily flow-reversal patterns.

Wind speeds associated with the various flow patterns in the Yucca Mountain area are highest during the midafternoon hours and reach minimum speeds both shortly after sunrise and shortly after sunset. This pattern is also due to the terrain influences in which minimum wind speeds represent

CONSULTATION DRAFT

Table 5-5. Wet-bulb depression values calculated for Yucca Flat

Month	Relative humidity at hour 1600 (%)	Average daily maximum temperature ^a °F (°C)		Wet-bulb depression F° (°C)		Wet-bulb temperature °F (°C)	
January	35	52.1	(11.16)	10.7	(5.94)	41.4	(5.22)
February	32	57.7	(14.28)	14.0	(7.78)	43.7	(6.50)
March	23	60.9	(16.06)	16.1	(8.95)	44.8	(7.11)
April	21	67.8	(19.89)	18.9	(10.50)	48.9	(9.39)
May	17	78.9	(26.06)	24.2	(13.45)	54.7	(12.61)
June	14	87.6	(30.89)	28.8	(16.00)	58.8	(14.89)
July	15	96.1	(36.11)	32.4	(18.50)	63.7	(17.61)
August	16	95.0	(35.00)	30.6	(17.00)	64.4	(18.00)
September	17	86.4	(30.22)	27.0	(15.00)	59.4	(15.22)
October	19	76.1	(24.50)	22.5	(12.50)	53.6	(12.00)
November	31	61.8	(16.56)	14.3	(7.95)	47.5	(8.61)
December	41	58.0	(14.44)	16.8	(9.33)	41.2	(5.11)

^aSource: Eglinton and Dreicer (1984).

flow reversals or directional changes, and maximum speeds correspond to periods when intense surface heating initiates upgradient air movement. This pattern is common to the data collected from three of the four towers covered in Quiring's (1968) study during all months of the year. The exception to this pattern is the data collected from the tower located on the basin rim. At this tower location, wind speed minimums occur during afternoon hours in November through February.

Surface wind data from the Yucca Flat station have been summarized by frequency of occurrence for the period 1961 to 1978 (DOC, 1986) and show patterns quite similar to the towers evaluated in Quiring's (1968) study. Although not site specific, these data should be indicative of conditions occurring at Yucca Mountain because of the physical similarity between the two locations. The U.S. Department of Commerce (DOC, 1986) data, summarized by season and year, are shown in Figure 5-3 as wind rose plots.

CONSULTATION DRAFT

SITE:YUCCA FLAT - SPRING

TOTAL OBS - 12345

% CALMS - 8.8

SITE:YUCCA FLAT - SUMMER

TOTAL OBS - 11659

% CALMS - 11.0

SITE:YUCCA FLAT - ANNUAL

TOTAL OBS - 47620

% CALMS - 13.1

WIND SPEED CLASS
(METERS/SEC)

>11.0
8.5-11.0
5.4-8.4
3.4-5.3
1.8-3.3
0.0-1.7

OBS - OBSERVATIONS

SITE:YUCCA FLAT - FALL

TOTAL OBS - 11629

% CALMS - 13.1

SITE:YUCCA FLAT - WINTER

TOTAL OBS - 11987

% CALMS - 14.1

Figure 5-3. Seasonal and annual surface wind distributions for Yucca Flat (1961-1978). Note: scale is not the same for all distributions. Based on data from DDC (1986).

CONSULTATION DRAFT

The spring (March through May) distribution shows that winds from the south occur most frequently and account for 15.7 percent of the observations during the season. Winds from the north also occur quite regularly, accounting for 10.8 percent of the spring observations. Overall, winds from the south through the southwest and northwest through north are dominant during the season, with nearly 60 percent of the recorded observations having occurred in one of these six directions.

The summer season (June through August) distribution is again dominated by winds from the south, accounting for 16.3 percent of the distribution. Winds from the south through southwest are clearly the most common during the summer months and represent 38.3 percent of the total observations. The northerly component of the summer wind flow pattern is not nearly as pronounced as in the spring distribution, with winds from the north accounting for 8.0 percent of the observations. This northerly component in the summer months is most likely due to the nighttime drainage winds at this site, which were discussed previously.

Winds from the north during the fall (September through November) were most common and were observed 13.0 percent of the time, with winds from the northwest through the north accounting for 28.9 percent of the observations. There is also a strong southerly element of the fall wind rose, representing 11.0 percent of the recorded observations. The fall data show a general variability in wind direction that is not as evident in the other seasonal distributions.

The winter (December through February) distribution shows a predominance of northerly winds that alone account for 14.5 percent of the total observations. Winds from the northwest through north are the most common observations and account for 36.7 percent of the total distribution. The southerly component evident in the fall wind rose is also significant during winter months, occurring 11.6 percent of the time.

On an annual basis, winds from the south occur somewhat more frequently than winds from the north (14.0 percent from the south versus 11.6 percent from the north). Excluding these two dominant wind flow quadrants, the balance of the annual wind rose exhibits a relatively uniform distribution, somewhat skewed to the northwest and southwest due to the terrain in the vicinity of the Yucca Flat tower.

Wind speeds occurring at Yucca Flat are generally less than 12 mph (5.4 m/s) during all seasons, with generally higher wind speeds occurring in the spring and somewhat lower wind speeds (overall) occurring in the fall and winter seasons. Although there will be differences between the data collected at Yucca Flat and conditions at Yucca Mountain, the general seasonal trends and variability are assumed to be similar.

The wind speeds and directional variability at the Yucca Mountain site will be thoroughly evaluated through the monitoring program that has been implemented at Yucca Mountain. The program is described in the Meteorological Monitoring Plan (SAIC, 1985) in detail and discussed in Section 8.3.1.12. The general information evaluated thus far, however, indicates that although extreme wind phenomena (discussed in Section 5.1.1.6) are

potentially damaging, they do not present obstacles to design, construction, or operation of the proposed repository.

5.1.1.5 Upper air data

Data on meteorological conditions, specifically wind speed and direction, at levels above those measured by the various towers on the NTS are useful in assessing the possibility of long-range transport of potential repository emissions. Data collected at the Yucca Flat weather station from 1957 to 1964 and summarized in Quiring (1968) provide upper air wind data. The upper air data from the Yucca Flat weather station are for midseason months only (January, April, July, and October). Times of collection begin at 0400 Pacific standard time (PST) and end at 1600 PST at 3-h intervals, and it is assumed that the midseason data represent an entire season.

Data at 5,000 ft (1,524 m) above mean sea level (328 m above ground level) are similar to surface observations and are shown in Table 5-6, and are plotted as wind roses in Figure 5-4. Winds from the northwest through northeast during winter (January) are shown to occur 60.3 percent of the time, with average speeds of approximately 12 mph (5.4 m/s). Summer patterns at the 5,000-ft (1,524 m) level are virtually opposite the winter data, with southeasterly to southwesterly winds occurring 74.9 percent of the time at average speeds of about 13 mph (5.8 m/s). Spring southerly (southeast through southwest) winds occur somewhat more frequently than do northerly (northwest through northeast) winds, 46.2 percent versus 36.1 percent. Fall patterns balance just slightly opposite the spring data, with 44.5 percent northerly winds compared with 39.4 percent southerly winds. On an annual basis, the seasonal north-to-south change in wind direction is about equal: 43.5 percent of the time from the north (northwest through northeast) and 41.2 percent of the time from the south. Data at 6,000 ft (1,829 m) above mean sea level (633 m above ground level) presented in Table 5-7 and shown in Figure 5-5 show patterns similar to the data at the 5,000-ft (1,524-m) levels, but they are considered to be more synoptically influenced than at lower levels.

5.1.1.6 Severe weather and obstructions to visibility

Occurrences of severe weather are superimposed on the average or normal climatic conditions in the Yucca Mountain area. These events generally include thunderstorms and associated lightning and flash flooding, hail storms, tornadoes, straight-line extreme winds, sandstorms, temperature extremes, freezing rain, and fog. The most important of these phenomena, with respect to development of Yucca Mountain as a repository, are the thunderstorm-derived conditions such as high winds, hail, and flash flooding.

Tornadoes, a possible source of high winds, are considered rare in Nevada but have been observed within a radius of 250 km of Yucca Mountain (Eglinton and Dreicer, 1984). The most severe of these tornadoes was classified as F-0 on the Fujita tornado intensity scale. This scale was developed to classify tornado intensity and maximum wind speed based upon the extent of

CONSULTATION DRAFT

Table 5-6. Yucca Flat upper air data for 5,000 ft (1,524 m) above mean sea level (328 m above ground level)^a

Direction ^b	Winter		Spring		Summer		Fall		Annual	
	(%)	Avg. speed (m/s)	(%)	Avg. speed (m/s)	(%)	Avg. speed (m/s)	(%)	Avg. speed (m/s)	(%)	Avg. speed (m/s)
N	21.8	5.6	9.8	5.7	3.2	5.9	13.5	6.3	14.2	5.8
NNE	14.9	5.5	8.7	5.1	1.6	4.1	13.1	5.5	11.2	5.4
NE	7.9	5.0	5.6	4.6	1.5	1.5	8.9	4.8	6.7	4.8
ENE	2.9	3.2	2.9	3.5	1.7	0.6	4.9	3.6	3.2	3.4
E	1.5	2.6	1.5	2.0	1.2	1.7	1.3	2.8	1.4	2.4
ESE	1.7	2.0	2.3	2.4	2.5	2.5	2.7	2.8	2.2	2.5
SE	1.9	4.7	3.5	3.2	4.6	3.6	3.6	3.2	3.1	3.1
SSE	3.0	3.6	6.1	5.0	9.0	4.9	5.9	4.6	5.4	4.6
S	6.4	5.0	12.0	7.5	18.3	6.5	12.6	6.4	11.1	6.4
SSW	9.0	5.8	15.1	8.4	26.5	6.9	10.9	7.2	13.2	7.2
SW	6.4	5.4	9.5	7.8	16.5	6.7	6.4	6.7	8.4	6.7
WSW	3.5	4.0	3.8	4.7	5.8	5.0	2.7	4.3	3.6	4.5
W	1.7	3.1	2.0	3.5	2.6	2.8	1.5	3.3	1.8	3.2
WNW	2.4	2.8	4.5	5.9	1.2	3.0	2.3	3.3	2.7	4.1
NW	5.1	4.5	5.3	6.2	1.7	3.3	3.1	3.6	4.1	4.7
NNW	10.6	5.2	6.7	6.2	2.8	3.8	5.9	4.9	7.3	5.3
Calm	1.4	NA ^c	1.4	NA	1.2	NA	1.9	NA	1.5	NA

^aCalculated from Quiring (1968).

^bWinds blow from indicated direction.

^cNA = not applicable.

resultant damage. An F-0 tornado on this scale is classified as a very weak tornado; it has winds of between 40 and 72 mph (18 and 32 m/s), a path length of less than 1 mi (1.6 km), and a path width of less than 17 yd (16 m) (Ludlum, 1982). Dust devils, which are small whirlwinds containing sand or dust, occur in and around the Yucca Mountain site during the summer months. Dust devils occasionally develop wind velocities in excess of that associated with an F-0 tornado, but they dissipate rapidly (Eglinton and Dreicer, 1984).

Lightning is frequently associated with thunderstorm activity but, because cloud-to-cloud lightning occurs nearly 10 times as frequently as cloud-to-ground lightning, strikes of consequence (i.e., resulting in measurable damage) in Nevada only average 18 per yr (Eglinton and Dreicer, 1984). However, the sparse observational network may reflect a somewhat lower frequency of occurrence than actually might be experienced at the Yucca Mountain site.

CONSULTATION DRAFT

SITE:YUCCA FLAT-SPRING
TOTAL OBS - 433
% CALMS - 1.4

SITE:YUCCA FLAT-SUMMER
TOTAL OBS - 249
% CALMS - 1.2

SITE:YUCCA FLAT-ANNUAL
TOTAL OBS - 1922
% CALMS - 1.5

WIND SPEED CLASS
(METERS/SEC)

>11.0
8.5-11.0
5.4-8.4
3.4-5.3
1.8-3.3
0.0-1.7

OBS = OBSERVATIONS

SITE:YUCCA FLAT-FALL
TOTAL OBS - 577
% CALMS - 1.9

SITE:YUCCA FLAT-WINTER
TOTAL OBS - 663
% CALMS - 1.4

Figure 5-4. Seasonal and annual wind distributions at 5,000 ft (1,524 m) above mean sea level (328 m above ground level) for Yucca Flat (1957 to 1964). Note: Scale is not the same for all distributions. Based on data from Quiring (1968).

CONSULTATION DRAFT

Table 5-7: Yucca Flat upper air data for 6,000 ft (1,829 m) above mean sea level (633 m above ground level)^a

Direction ^b	Winter		Spring		Summer		Fall		Annual	
	(%)	Avg. speed (m/s)	(%)	Avg. speed (m/s)	(%)	Avg. speed (m/s)	(%)	Avg. speed (m/s)	(%)	Avg. speed (m/s)
N	15.3	7.6	7.4	7.8	1.8	6.7	10.8	5.9	10.4	7.1
NNE	17.0	6.8	8.6	5.7	1.6	5.9	14.0	6.6	12.2	6.3
NE	10.7	6.2	6.5	4.7	1.3	4.0	10.2	6.2	8.4	5.7
ENE	4.4	4.6	4.2	3.5	1.1	1.7	5.7	4.8	4.3	4.3
E	1.4	2.2	1.4	1.7	0.6	1.0	1.5	3.5	1.3	2.5
ESE	1.3	2.8	2.1	2.7	1.7	3.4	2.3	3.1	1.8	3.0
SE	1.4	3.0	1.9	3.1	2.9	4.8	3.2	3.6	2.3	3.6
SSE	2.4	4.8	3.6	6.3	7.4	6.0	5.4	5.0	4.2	5.5
S	5.9	7.1	11.9	8.2	21.1	6.8	10.4	7.1	10.6	7.3
SSW	9.8	7.3	19.7	8.3	31.2	7.4	13.7	7.4	16.0	7.6
SW	6.8	6.6	12.4	7.7	18.8	7.1	8.4	6.9	10.1	7.1
WSW	3.6	4.6	4.2	5.1	5.5	5.0	3.0	4.7	3.8	4.8
W	2.9	4.2	1.5	5.8	2.6	4.3	2.2	3.5	2.4	4.2
WNW	3.7	4.2	4.5	6.4	1.0	3.6	2.2	5.2	3.1	5.1
NW	5.4	5.2	4.9	6.5	0.6	3.4	2.4	4.8	3.7	5.5
NNW	8.9	6.3	5.6	6.8	0.9	4.9	4.2	5.0	5.7	6.1
Calm	0.6	NA ^c	0.0	NA	0.0	NA	0.9	NA	0.5	NA

^aCalculated from Quiring (1968).

^bWinds blow from indicated direction.

^cNA = not applicable.

Hail is a variety of thunderstorm activity that can have quite damaging effects, but only one occurrence is expected annually at Yucca Mountain (Eglinton and Dreicer, 1984).

Obstructions to visibility in the vicinity of Yucca Mountain could temporarily disrupt activities at the proposed repository. The most likely conditions that could obstruct visibility appreciably are sandstorms or fog. The conditions conducive to fog formation occur only about twice a year in this area of Nevada and sandstorms of sufficient magnitude to reduce visibility occur only a small fraction of the time (Eglinton and Dreicer, 1984). More detailed discussions of obstructions to visibility as they relate to safe operation of the repository will be included in the environmental impact statement.

Predictions of severe weather needed in design considerations are generally derived by extrapolating past recorded data. The associated probability of occurrence of an event is then calculated based on the frequency of occurrence of the event during the period of record. Of most importance in

CONSULTATION DRAFT

SITE:YUCCA FLAT-SPRING
TOTAL OBS - 433
% CALMS - 0.0

SITE:YUCCA FLAT-SUMMER
TOTAL OBS - 248
% CALMS - 0.0

SITE:YUCCA FLAT-ANNUAL
TOTAL OBS - 1918
% CALMS - 0.5

WIND SPEED CLASS
(METERS/SEC)

>11.0
8.5-11.0
5.4-8.4
3.4-5.3
1.8-3.3
0.0-1.7

OBS - OBSERVATIONS

SITE:YUCCA FLAT-FALL
TOTAL OBS - 577
% CALMS - 0.9

SITE:YUCCA FLAT-WINTER
TOTAL OBS - 660
% CALMS - 0.6

Figure 5-5. Seasonal and annual wind distributions at 6,000 ft (1,829 m) above mean sea level (633 m above ground level) for Yucca Flat (1957 to 1964). Note: Scale is not the same for all distributions. Based on data from Quiring (1968).

designing the surface facilities of the proposed repository at Yucca Mountain are estimates of extreme winds, temperature maximums and minimums, and extreme precipitation events.

Extreme wind speeds and associated probabilities of occurrence have been calculated for the NTS and are presented in Table 5-8 (Quiring, 1968). These data are for a fastest mile of wind, which represent an average highest wind velocity as 1 mile of air passes the measurement point. Eglinton and Dreicer (1984) discuss the potential for both straight-line winds and tornadic (cyclic) winds expected to occur at Yucca Mountain. The probability of a tornado strike at Yucca Mountain, given in Eglinton and Dreicer (1984), is approximately 7.5×10^{-4} in any given year. The maximum design wind speeds cited in Eglinton and Dreicer (1984) for the NTS are a straight-line wind speed of 94 m/s (210 mph) and a tornadic (cyclic) wind speed of 28 m/s (63 mph). Both phenomena are given as having a probability of 1×10^{-6} of occurring in 1 yr. These extrapolations, however, do not take into account the possibility of climatic change in the future.

The probability of occurrence of extreme temperatures, also given in Eglinton and Dreicer (1984), is shown in Table 5-9. These data are estimated on the basis of measured extreme temperatures but do not account for the influence of climatic change.

Another design consideration is extreme precipitation and the potential for flooding that could occur as a result. Because the flooding potential from short-duration, high-intensity storms is high, 24-h average precipitation amounts (Section 5.1.1.2) are not a realistic indicator of potentially damaging extreme precipitation and subsequent flood events at Yucca Mountain. Both 1- and 24-h maximum precipitation and associated probabilities of occurrence, again based on measured extremes, are given in Table 5-10 (Hershfield, 1961). The flooding potential is discussed in more detail in Section 3.2. Although this sort of extrapolation of extreme events is useful, more detailed and site-specific data on precipitation are needed. The plan for collecting such data is given in Section 8.3.1.12.

5.1.2 LOCAL AND REGIONAL METEOROLOGY

The meteorological monitoring program (Section 5.1.3) will provide data to be used in characterizing atmospheric dispersion processes. Aside from establishing the link between site meteorology and general (long-term) climatic conditions at the site, the data will be used in satisfying permit requirements and as input to the environmental impact statement. In general, meteorological conditions experienced at the site are expected to be quite similar to those of the stations used in describing the climate. Winds will be governed to a significant extent by the terrain with regard to direction and speed. The highest winds will be associated with winter frontal passages and thunderstorms. One meteorological parameter not previously discussed, because it is not a characteristic of climate, is atmospheric dispersion or stability. A discussion of this parameter follows.

Atmospheric stability is an important parameter with respect to dispersion of emissions (e.g., particulates and exhaust gases) from the proposed

Table 5-8. Annual extreme wind speed at 30 ft (9.1 m) above ground level and probability of occurrence for Yucca Flat, Nevada^a

Probability of occurrence in 1 yr	Fastest mile ^b	
	mph	m/s
0.5	48	21
0.2	55	25
0.1	61	27
0.02	75	33
0.01	82	37

^aSource: Quiring (1968).

^bFastest mile is defined as an average highest wind velocity as 1 mile of air passes the measurement point.

Table 5-9. Extreme maximum and minimum temperatures and probability of occurrence for Beatty, Nevada^a

Probability of occurrence ^b in 1 yr	Temperature (°C)	
	Maximum	Minimum
1.0	40.2	-6.6
0.5	42.3	-10.2
0.2	43.6	-12.3
0.1	44.4	-13.7
0.05	45.2	-15.1
0.04	45.5	-15.4
0.02	46.2	-16.8
0.01	47.1	-18.1
0.005	47.8	-19.4
0.002	48.8	-21.2
0.001	49.6	-22.4
0.0001	52.2	-26.8

^aSource: Eglinton and Dreicer (1984).

^bThese probabilities of occurrence do not reflect potential climatic changes.

Table 5-10. Maximum 1- and 24-h precipitation and probability of occurrence for Yucca Flat^a

Probability of occurrence in 1 yr	Maximum precipitation			
	1 h		24 h	
	in.	mm	in.	mm
1.0	0.30	7.6	0.75	19.1
0.5	0.40	10.2	1.00	25.4
0.2	0.60	15.2	1.25	31.8
0.1	0.70	17.8	1.50	38.1
0.04	0.80	20.3	1.75	44.5
0.02	0.90	22.9	2.00	50.8
0.01	1.00	25.4	2.25	57.2

^aSource: Hershfield (1961).

repository and has been analyzed using estimates of cloud cover, ceiling height, and net solar radiation for the 18-yr period of record at Yucca Flat (DOC, 1986). Annual average stability distributions similar to the data presented in Figure 5-3 can be constructed for only those occurrences of wind speed and direction falling into each of the six Pasquill stability classes (A through F). Stability class A defines an extremely unstable (highly convective) atmosphere, class B is for unstable conditions, class C is slightly unstable, class D is neutral, class E is slightly stable, and class F is stable. Wind roses for each of these stability classes are shown in Figures 5-6 (A through C) and 5-7 (D through F). The most significant features of the stability distributions are summarized in Table 5-11 and discussed in the following paragraphs.

Class A stability was most frequently associated with winds from the east, while winds from the east through southeast were most frequently associated with class B stability. For class C stability, winds from the south were the most common occurrence, but winds from the southeast through the southwest were also quite frequently associated with class C stability. Occurrences of class D stability were the second most commonly observed stability classification and were commonly associated with winds from the north and the south through southwest. The distribution for class E stability indicates that there is a distinct shift from the generally southerly winds associated with the unstable and neutral stability classifications (classes A through D) to predominantly northwesterly winds for class E and class F. Stable atmospheric conditions (class F) were the most commonly observed stability class and the distribution clearly shows the predominance of winds from the west through the north. Because stability classes E and F are both associated with relatively light winds, the predominance of winds from the west through the north for these classes is most likely due to drainage winds at this site that develop under synoptically calm conditions.

CONSULTATION DRAFT

SITE:YUCCA FLAT-A STABILITY
TOTAL OBS - 710
% CALMS - 20.7

SITE:YUCCA FLAT-B STABILITY
TOTAL OBS - 5436
% CALMS - 15.6

SITE:YUCCA FLAT-C STABILITY
TOTAL OBS - 5983
% CALMS - 5.5

WIND SPEED CLASS
(METERS/SEC)

OBS - OBSERVATIONS

Figure 5-6 Distribution for Pasquill stability classes A, B, and C for Yucca Flat (1961 to 1978). A = extremely unstable, B = unstable, C = slightly unstable. Note: Scale is not the same for all distribution. Based on data from DOC (1986).

CONSULTATION DRAFT

SITE:YUCCA FLAT-D STABILITY
TOTAL OBS - 14643
% CALMS - 0.8

SITE:YUCCA FLAT-E STABILITY
TOTAL OBS - 5621
% CALMS - 0.0

SITE:YUCCA FLAT-F STABILITY
TOTAL OBS - 15227
% CALMS - 27.0

WIND SPEED CLASS
(METERS/SEC)

>11.0
8.5-11.0
5.4-8.4
3.4-5.3
1.8-3.3
0.0-1.7

OBS - OBSERVATIONS

Figure 5-7. Distribution for Pasquill stability classes D, E, and F for Yucca Flat (1961 to 1978). D = neutral, E = slightly stable, F = stable.

Table 5-11. Yucca Flat Pasquill stability class distributions for the period 1961 to 1978^a

Stability class ^b	Percentage of total observations	Predominant direction ^c	Percentage of stability class observations	Predominant quadrant ^c	Percentage of stability class observations
A	1.5	E	22.9	E-S	64.5
B	11.4	E	22.9	E-SE	51.0
C	12.6	S	19.0	SE-SSW	47.8
D	30.7	S	21.2	S-SW, N	42.0, 14.2
E	11.8	NW	16.6	WNW-N, S-SW	51.4, 25.8
F	32.0	NW	14.2	W-N	51.7

^aSource: DOC (1986).^bA = extremely unstable, B = unstable, C = slightly unstable, D = neutral, E = slightly stable, F = stable.^cWind blows from indicated direction.

In summary, neutral and stable conditions (classes D, E, and F) were by far the most commonly experienced at Yucca Flat and account for 74.5 percent of the total observations. Stable conditions tend to be dominated by winds generally from the northwest, while neutral conditions had a significant south to southwesterly component in addition to a strong northerly component. Unstable conditions (classes A, B, and C) occurred only 25.5 percent of the time, had virtually no northerly component, and were dominated by winds from the east through the south, with the most unstable classes having a stronger easterly element than the slightly unstable category.

Determining the stability distributions is important from the standpoint of evaluating the potential impacts of particulate and gaseous emissions from the repository. The data are required as input to the dispersion models that will be used in acquiring permits for both the site characterization activities will be discussed in the environmental monitoring and mitigation plan and the repository (through the environmental impact statement (EIS) process). Although these data from Yucca Flat provide a preliminary indication of stability and dispersion characteristics of the area, site-specific data are needed. Determination of atmospheric stability was a significant consideration in development of the meteorological monitoring program being operated at Yucca Mountain (Section 5.1.3) and will be fully evaluated throughout site characterization activities.

5.1.3 SITE METEOROLOGICAL MEASUREMENT PROGRAM

A monitoring program was operated at Yucca Mountain for approximately 2 yr. It consisted of two 10-m towers instrumented to collect data on temperature, wind speed and direction (3-m and 10-m levels), relative humidity, insolation, ground surface infrared radiation, soil temperature, precipitation, and barometric pressure. The towers were installed to collect preliminary meteorological data and were decommissioned at the end of October 1984. However, most of the data from this program are still being reduced and are not available.

A new, extended monitoring program has been designed to collect data on both synoptic-scale meteorological influences and specific terrain-induced perturbations. The program includes four 10-m towers designated NTS-10 plus an area designation (Yucca Mountain, Coyote Wash, Alice Hill, and Fortymile Wash) and one 60-m tower designated NTS-60 Repository. The locations of the towers are described in Table 5-1 and shown in Figure 5-1. The 60-m tower is placed near the proposed surface facility location, and the other 4 towers are placed at various locations in the vicinity of Yucca Mountain. Wind speed, wind direction, and standard deviation of wind direction (sigma-theta) data will be collected at each of the 4 remote sites at the 10-m level and at both the 10- and 60-m level at the main site. All the sites are instrumented to collect data on precipitation, relative humidity (dew point at the main site), and ambient temperature. In addition, the 60-m tower instrumentation includes a net radiation sensor (for solar and terrestrial radiation), a vertical wind speed sensor, and circuitry for determining the temperature difference between the 10- and 60-m levels. Most of these parameters are recommended or required to be monitored for regulatory compliance (EPA, 1980). In addition, the hourly average wind speed, wind direction, and temperature data are required as input to dispersion models that will be used in assessing the ambient air quality impacts of the proposed activities. The sigma-theta, vertical wind speed, temperature difference, and net radiation data can all be used to determine atmospheric stability, which is a very important factor in determining ambient impacts through dispersion modeling. The relative humidity and dew point will be used for climatological comparisons, as will the precipitation data. The precipitation data will also be used as input to other studies that will be conducted during site characterization. The other programs are the infiltration studies and the surface water hydrology investigations.

The instruments used to collect data on the various meteorological parameters will meet the following specifications:

1. Wind direction: $\pm 3^\circ$ of true azimuth (including sensor orientation error) with a starting threshold of less than 0.45 m/s.
2. Wind speed: ± 0.22 m/s for speeds above the starting threshold of 0.45 m/s but less than 11.1 m/s, and ± 5 percent of true speed, not to exceed 2.5 m/s, at speeds greater than 11.1 m/s.
3. Sigma-theta: wind-vane damping ratio of between 0.4 and 0.6 (inclusive) with a 15° deflection and delay distance not to exceed 2 m.
4. Dry-bulb temperature: $\pm 0.5^\circ\text{C}$.

5. Temperature difference (between levels): $\pm 0.003^{\circ}\text{C}/\text{m}$.
6. Radiation (solar and terrestrial): ± 5 percent.
7. Precipitation: resolution of 0.25 mm with a recorded accuracy of ± 10 percent of total accumulated catch.
8. Relative humidity: ± 6 percent.
9. Dew point temperature: $\pm 1.5^{\circ}\text{C}$.
10. Time: within 5 min of actual time for all recording devices.

These specifications apply to digital systems; analog backup systems can deviate by up to 1.5 times these values.

The monitoring program will be run in accordance with quality assurance regulations, rules, and guidelines developed to ensure the validity and traceability of all collected data. This and other information regarding the monitoring program is contained in the Meteorological Monitoring Plan (SAIC, 1985), which is discussed in Section 8.3.1.12.

Air quality monitoring, although not directly related to site characterization, may be needed to fulfill permitting requirements for both the site characterization activities and repository development. The environmental monitoring and mitigation plan will contain information on those aspects of site characterization that may require air quality data, and the environmental impact statement (EIS) process will define air monitoring needs with respect to repository development.

5.2 LONG-TERM CLIMATIC ASSESSMENT

An assessment of the long-term climate in the Yucca Mountain area is necessary to resolve several issues. The nature and rates of change in past climates must be understood to allow the prediction of future climate conditions. An understanding of future climate conditions is needed to evaluate the potential effects of climatic change on the location and rates of erosion and on the hydrologic and geochemical characteristics in the vicinity of the Yucca Mountain site. The hydrologic system may be especially susceptible to changes in climate. This section discusses the present understanding of the nature and rates of past climate change and discusses the strategy for developing scenarios for future climatic variations. These scenarios can be used to evaluate anticipated and unanticipated future climate conditions that can be used to assess the potential changes in the hydrologic characteristics in the vicinity of Yucca Mountain (Section 8.3.1.5.2). These estimates of future climate conditions will also be used to determine potential changes in the location and rates of erosion (Section 8.3.1.6.2) at Yucca Mountain.

Paleoclimatic reconstructions discussed in this section focus on the latter part of the Quaternary Period. Figure 5-8 shows the relationship of the geologic time periods discussed in this chapter. The Quaternary Period includes the last 1.6 million years and is subdivided into the Pleistocene

CONSULTATION DRAFT

Figure 5-8. Time lines indicating geologic time periods discussed in text. Based on data from GSA (1983).

Epoch (from 1.6 million to 10,000 yr ago) and the Holocene Epoch (from 10,000 yr ago to the present). Further subdivisions of the Pleistocene Epoch are based on periodic glacial advances and recessions. The most recent of these glacial advances, known in North America as the Wisconsin Stage, is important in the reconstruction of the paleoclimatic history of the Yucca Mountain site because the conditions that accompanied this glacial regime may represent the factors that will affect future hydrologic conditions at the NTS, should the Holocene close with a return to a global glacial period. Paleoclimatic investigations must extend back at least to the previous interglacial period (known in North America as the Sangamon) in order to provide a climatic analog for possible future climate states that are warmer than those of today, which could occur in a carbon-dioxide-enhanced world.

Because it is the most recent glacial and pluvial event in the earth's history, the Wisconsin has been intensively studied and is the best understood of the glacial and interglacial stages that make up the Pleistocene. The beginning of the Wisconsin is correlated by Ruddiman and McIntyre (1981) with the 5d-5e boundary of the deep-sea oxygen isotope record, which occurred about 115,000 yr ago. During the late Wisconsin (23,000 to 10,000 yr ago) more than 100 closed basins in the northern and western Great Basin contained lakes (Smith and Street-Perrott, 1983), while in the southern Great Basin, at Yucca Mountain, a marked increase in the area of woodland species of vegetation suggests more available moisture and a general decline in temperature (Spaulding and Graumlich, 1986).

In the following sections, climate variation during the Quaternary Period will be discussed. Because past climatic variability is probably the best indicator of future climatic conditions, understanding this climatic variability is important in assessing future climatic variation. Ensuing sections are structured in terms of three principal topics: paleoclimatology (Section 5.2.1), future climate variation (Section 5.2.2), and paleoclimatic variations and their relation to the Yucca Mountain site (Section 5.2.3).

5.2.1 PALEOCLIMATOLOGY

5.2.1.1 Quaternary global paleoclimate

Climate varies on all temporal and spatial scales, ranging from inter-annual variations at a particular location related to atmospheric circulation anomalies, to the very long period variations at the global scale related to the evolution of the atmosphere and lithosphere (Webb et al., 1985). An understanding of climate variation on the global scale is necessary for the prediction of local climate. This is because climate at any point on the earth's surface is the result of processes that occur over the entire global area.

In assessing the environmental stability of a particular location, the key climatic variations are those that occur on the time scale of centuries to 100,000 yr (Crowley, 1983). During the Quaternary, the principal climatic variations at such time scales have been those associated with the repeated fluctuations between glacial and interglacial conditions.

The glacial cycles of the Quaternary are part of a long period of cooling, commencing in Late Cretaceous time (Lloyd, 1984). During the past 80 million yr, global average temperature declined, and first Antarctica and later Greenland, North America, and Europe became glaciated. Glaciation of the northern hemisphere probably began about 3.2 million years ago (Crowley, 1983), with clear evidence about 2.5 million years ago for glaciation on land areas contributing sediments to the North Atlantic (Shackleton et al., 1984).

The global volume of glacial ice has varied continuously throughout the Quaternary, but with characteristic quasi-periodicity correlated with variations of the earth's orbital elements, i.e., eccentricity (about 100,000 yr), obliquity (about 40,000 yr), and precession (about 20,000 yr) (Hays et al., 1976; Imbrie et al., 1984). During the past 120,000 yr, global ice volume increased from the low values of the last interglacial (125,000 to 118,000 yr before present) to the high values of the last glacial maximum about 18,000 yr ago, then decreased to approximately its present level by 6,000 yr ago. The paleoclimatic record of the past 18,000 yr provides an illustration of the extremes of climate to be expected in a single glacial to interglacial transition but does not necessarily indicate the full range of conditions that might occur in an interglacial to glacial transition (Spaulding, 1983).

At 18,000 yr before present, global average temperatures were lower than at present, and large ice sheets covered parts of North America and Europe (Denton and Hughes, 1981). Consequently, sea levels were lower, exposing much of the continental shelves (Bloom, 1983). Sea surface temperatures as a whole were lower, and extensive sea ice formed in the northern oceans (CLIMAP Project Members, 1981). Eighteen thousand years ago vegetation and the hydrologic cycle differed markedly from those of today (Street-Perrott and Harrison, 1984). The Laurentide ice sheet in eastern North America reached its maximum extent about this time, while the Cordilleran ice sheet in the Northwest approached its maximum by 15,000 yr B.P. (Mayewski et al., 1981; Waitt and Thorson, 1983).

In the North Pacific Ocean adjacent to western North America about 18,000 yr ago, sea-surface temperatures were more than 4 Celsius degrees lower than today over large areas, and oceanic circulation was altered as well (Imbrie et al., 1983). Land temperature declines ranged from 0 to 10 Celsius degrees during the last glacial maximum (Von Neumann, 1960). In southern Nevada, average annual temperatures were 6 to 7 Celsius degrees lower than the current average annual temperature (Spaulding, 1985). Experiments with climate simulation models (Section 5.2.2.2) suggest that 18,000 yr ago the expanded ice sheets, extensive sea ice, and colder oceans greatly influenced atmospheric circulation around North America (Gates, 1976b; Manabe and Broccoli, 1985; Kutzbach, 1985; Kutzbach and Wright, 1985). For example, to simulate the climate of 18,000 yr ago, Manabe and Broccoli (1985) used a Geophysical Fluid Dynamics Laboratory general circulation model coupled with a static mixed-layer ocean model (in which the sea surface temperatures were predicted by the model), and Kutzbach and Wright (1985) used the National Center for Atmospheric Research Community Climate Model (NCAR CCM) (in which sea surface temperatures were prescribed using the reconstructions by CLIMAP Project Members, 1981). Both models simulated similar changes in the atmospheric circulation across North America (relative to today) resulting from the imposition in the models of the Laurentide ice sheet, including (1) a split in the jet stream in the upper atmospheric flow, with one weaker branch

crossing the continent to the north of the ice sheet, and a second stronger branch crossing the continent from southwest to northeast to the south of the ice (Kutzbach and Wright, 1985); (2) the development of a strong ridge over western North America and a deep trough over eastern North America in the upper atmospheric circulation; and (3) the development of strong anticyclonic circulation at the surface over northern North America. Kutzbach and Wright (1985) describe the general compatibility between the climate simulated by the NCAR CCM and the geologic evidence for 18,000 yr ago in North America. Vegetation patterns in North America at 18,000 yr B.P. differed considerably from those at present, reflecting the great differences between modern and full-glacial regional climates (Spaulding et al., 1983; Kutzbach and Wright, 1985; VanDevender et al., 1987; SCP Section 5.2.1.2.3). Large-scale changes in regional hydrology also occurred, as is illustrated by the differences in lake levels between 18,000 yr B.P. and today (Smith and Street-Perrott, 1983; Street-Perrott and Harrison, 1984; Benson and Thompson, 1987; Forester, 1987; SCP Section 5.2.1.2.2).

Between 18,000 and 6,000 yr ago, the global ice volume decreased to approximately its present amount, the sea level rose, and the oceans warmed (Kutzbach and Guetter, 1986). Over the same interval the seasonal cycle of solar radiation was amplified (relative to both 18,000 yr ago and today) by the shift in perihelion from January to July, and by changes in the tilt of the axis (Kutzbach, 1981). At 10,000 yr ago, for example, global solar radiation at the top of the atmosphere was about 8 percent greater in summer than today and about 8 percent less in winter (Kutzbach, 1985). Accompanying these changes was an increase in the concentration of carbon dioxide in the atmosphere (Shackleton et al., 1983) in a manner consistent with the changes in global climate during this interval. The amplification of the seasonal cycle of solar radiation has been linked with the intensification of the summer monsoonal circulation of Africa and southern Asia about 9000 yr B.P. Greater monsoonal precipitation at that time is evidenced by the geologic record of lake-level variations in the tropics (Kutzbach and Street-Perrott, 1985). Similar increases in monsoonal rainfall have been postulated for the Yucca Mountain region (Spaulding and Graumilch, 1986), and this possibility will be examined during site characterization.

Astronomical forcing of climate provides a quasi-periodic physical mechanism for the simulation of past and future climate change. The potential causes of the climatic variations of the Quaternary will be discussed in Section 5.2.2.1 as they pertain to the prediction of future climatic variations. In the following sections, the regional expression of the large-scale climatic variations described above will be reviewed.

5.2.1.2 Quaternary regional paleoclimate

Global paleoclimatology contributes to an understanding of regional and site-scale climatic conditions principally by defining the broad chronology and large-scale mechanisms that control or place bounds on local climates. Although the global factors are necessary for long-term climate prediction, they are insufficient because they do not provide the information required from site characterization as discussed in the introduction to this chapter and in Section 5.2.

There are two primary types of geologic data that provide information on regional and site-specific paleoclimatic fluctuations. First, reconstructions of the past extents and hydrological environments of paleolakes and marshes provide an integrated history of hydrologic response to past climates. Second, paleontological evidence of the local terrestrial biological response to paleoclimates not only provides indications of general climatic conditions, but also suggests ranges for specific variables. By applying quantitative relationships derived from comparisons between modern plant distributions and modern climates, paleobotanical data may be evaluated in terms such as temperature extremes, available moisture, and relative moisture contributions by rain and snow.

Whereas paleobotanical data in the form of pack rat middens are available near the proposed repository (Spaulding, 1985), the adequacy of near-site paleolimnological data has not yet been fully assessed. Paleomarch deposits are available near the southern and western peripheries of the NTS (Haynes, 1967; Quade, 1986), and potentially promising playa sites occur to the north and east. Paleolacustrine data have been obtained from areas farther north, east, and west in the Great Basin, where pluvial lake systems (Figure 5-9) provide an abundance of datable deposits and micropaleontological data. Paleoclimatic investigations associated with the NTS will include efforts to obtain as much data as possible from the immediate proximity of Yucca Mountain. These data will include plant remains, pack rat middens, and microfossils from playa and paleomarch sediments. To reduce the uncertainty associated with site-specific studies, these local records will be supplemented with, and tested against, regional paleobotanical and paleolimnological records.

The remainder of this section is organized in the following format: Section 5.2.1.2.1 discusses the relationships among lake size (area and volume), climatic parameters, and hydrologic parameters, including historical data on the variability of these parameters; this review is necessary to establish a framework within which the significance of data from paleolakes can be understood. The presently available paleolacustrine data and their application in defining paleoclimates are presented and discussed in Section 5.2.1.2.2. In Section 5.2.1.2.3 the available results of paleobotanical studies in the Great Basin are summarized and compared with modern relationships of vegetation and climate, followed by a discussion of alternative applications to paleoclimate reconstructions. Other lines of paleoclimatic evidence, such as regional glacial features, are presented in Section 5.2.1.2.4. Finally, in Section 5.2.1.2.5, the present understanding of the paleoclimatology of the Great Basin is synthesized.

5.2.1.2.1 Historical lake, climatic, and hydrologic data

Within the Great Basin the most widespread and continuous data pertaining to paleoclimates are derived from studies of Quaternary lakes. However, even if it were possible to reconstruct complete histories of the shoreline altitudes, volumes, sedimentation, and biota of the lakes, the link to paleoclimates would still be indirect. An understanding of the relationship of lake size to the most important climatic parameters, precipitation and evaporation, and to the principal hydrologic parameters, streamflow and

Figure 5-9. Great Basin lakes of presumed Pleistocene age (modified from Spaulding et al., 1983).

ground-water flow, provides a framework for evaluating the significance of paleolacustrine data.

Historical records of precipitation, streamflow, and lake-level fluctuation reflect the hydrologic effects of climate change on large areas. These data, if corrected for the impact of man's activities, can be used to

1. Estimate the short frequency and amplitude of small-scale (<100 yr) changes in climate.
2. Estimate the spatial (area and altitude) variability of climate.
3. Provide data for the creation of synoptic snapshots of climate extremes.

These synoptic snapshots of climate and surface hydrology will serve as test scenarios for modeling future climate (Section 8.3.1.5.1).

Hydrologic balance of lake systems

Lake size is a function of a dynamic equilibrium resulting from changes in the magnitudes of water sources and sinks. The principal water sources for a lake are (1) streamflow into the lake, whether as direct runoff from the watershed or as emergence of ground water to stream channels; (2) precipitation on the lake surface; and (3) ground-water discharge directly to the lake. The principal processes by which water leaves a lake are (1) evaporation from the lake surface, (2) streamflow from the lake, and (3) ground-water outflow from the lake.

The paleolakes that persisted as relatively permanent bodies of water, and therefore provide the best preserved and most continuous record, are those that were contained in large drainage basins. Under these conditions ground-water inflow to the basin and ground-water outflow from the lake can be ignored without a significant loss of precision in water-budget calculations. Further, most but not all of the paleolakes of interest in the region occupied closed topographic basins, eliminating streamflow from the lake as a factor and leaving evaporation from the lake surface as the only process by which water leaves the lake.

For a lake and its surrounding large closed drainage area, the mean-annual hydrologic balance for the steady state can be written in terms of total discharge to the lake and evaporation from the lake surface:

$$(E_L) (A_L) = (P_L) (A_L) + D_s + D_g \quad (5-1)$$

where

E_L = evaporation rate from lake

A_L = area of lake surface

P_L = precipitation on lake

D_s = stream discharge into the lake

D_g = direct discharge of ground water to lake.

A knowledge of man's impact on the hydrologic balance is necessary for the correction of historical records of streamflow and lake levels (Whitaker, 1971; Stauffer, 1985). Figure 5-10 shows the historical and pristine (corrected) records of fluctuations in the water level of the Great Salt Lake for the period from 1851 to 1984. These corrected records can be subjected to spectral analysis to determine the characteristic temporal scales of variation in the hydroclimatological data during the monitored period.

Precipitation and streamflow data

Precipitation data exist for weather stations scattered throughout the western United States (Hershfield, 1961). The data have record lengths of up to 100 yr. Some weather stations located in California and Nevada will be used to provide vegetation-climate calibrations (response surfaces) for future climate prediction (Section 8.3.1.5.1.). Figure 5-11 shows the areal distribution of presently active precipitation and streamflow gaging stations in the basin once occupied by the Pleistocene Lake Lahontan.

The quantity of precipitation that falls directly on the lake basin compared with the inflow from streams varies from basin to basin and is important climatologically. For Lake Lahontan, precipitation that falls directly on the lake surface today is only about 15 percent of the total water input to the basin (Benson, 1986); for Lake Bonneville (Great Salt Lake) on-lake precipitation is about 27 percent of today's total water input (Stauffer, 1985); while for Lake Russell (Mono Lake), on-lake precipitation is about 19 percent of the total water input (Mason, 1967).

The relationship between on-lake precipitation and stream discharge can be examined by comparing precipitation data from a weather station located in the Sierra Nevada (Tahoe City, California), data from a station located on the floor of the Lahontan Basin (Fallon, Nevada), and discharge data for the Carson River that drains a watershed adjoining the Tahoe City site. Discharge data are mainly from gaging stations located upstream of agricultural regions. Figure 5-12 shows the good correlation between the precipitation at Tahoe City and the discharge of the Carson River. The climate (precipitation and temperature) of mountain ranges bordering a lake basin controls, to an extent, the amount of water input to a lake. This is especially true for lake basins such as Lahontan and Mono that lie in rain shadows. This relationship may be relevant to the paleoclimate at Yucca Mountain.

In evaluating the potential effect of climatic change on discharge to closed-basin lakes, it is instructive to examine records of gaged streamflow.

Figure 5-10. Historical and pristine (corrected) records of fluctuations in the level of Great Salt Lake, Utah. Source: Stauffer (1985).

Figure 5-11. Locations of active river streamflow-gaging and precipitation-collection stations. Source: Benson (1986).

Figure 5-12. Precipitation records for Fallon, Nevada, and Tahoe City, California, compared with Carson River discharge, 1942 to 1983. Streamflow data from Benson (1986).

Gaged streamflow, corrected for consumptive use, is a useful indicator of climate change (Meko and Stockton, 1984). Bartlein (1982) analyzed monthly streamflow over Canada and the United States for the period 1951 to 1970 by means of principal components analysis. He found that 75 percent of the total variance was explained by only 23 large scale anomaly patterns (of 102 potential components) that are, in turn, related to climatic anomalies in atmospheric circulation.

Meko and Stockton (1984) show that carefully selected records of streamflow in the western United States can provide useful information on climatic variation, especially for mountainous areas where high-elevation weather stations are sparse. They found that major low-frequency variations in streamflow have tracked variations in precipitation and temperature, and that secular trends in streamflow in various parts of the West have sometimes paralleled one another and at other times have diverged. This behavior appears to be related to major shifts in regimes of the general atmospheric circulation.

Many semiarid areas now produce up to 10 percent runoff and the most arid areas produce less than 5 percent. When viewing either the past or future, however, it is difficult to estimate how much additional precipitation, lowering of temperature, etc., could increase runoff in terms of either percentage or absolute volume. Langbein et al. (1949) constructed an empirical diagram that allows one to estimate runoff variation in nonarid regions caused solely by temperature or precipitation change. The arid end of this diagram suggests that in an area now averaging 5 percent runoff, an 18 Fahrenheit degree (10 Celsius degrees) drop in temperature (weighted according to precipitation) would increase annual runoff by factors ranging from about 3 to 6; in the same area, a 50 percent increase in precipitation could increase runoff by a factor of about 5. The diagram also implicitly suggests that regions having 15 in. precipitation or less and mean annual temperatures of 60°F (15.6° C) or more produce no runoff until those threshold values are exceeded. Langbein et al. (1949) also notes that drainage areas having similar altitudes and precipitation, but different rock types, exposures, and vegetation, can have runoff percentages that differ by 15 percent or more. Schumm (1965) extends these relations to include areas characterized by 400 mm (15.74 in) or less precipitation, showing that increasing precipitation by 50 percent or more and decreasing temperatures by about 5 Celsius degrees could increase runoff by factors ranging from about 5 to 20. The total sediment outflow from a watershed is a function of soil resistance, basin topography, and plant cover. The outflow reaches a peak at an effective precipitation of about 300 mm (11.8 in.), trailing off at lower values due to increased vegetation cover that protects against erosion (Knighton, 1984). Further discussion of erosion at Yucca Mountain can be found in Chapter 1 and Section 8.3.1.6.

The study of Meko and Stockton (1984) demonstrates that long-term streamflow records corrected for diversion and consumption can be used to reconstruct synoptic climates for the western United States. The results of this study will be combined with further studies presented in Section 8.3.1.5.1 to assist with the reconstruction of past synoptic climates.

Evaporation from lake surfaces

As was discussed in Section 5.2.1.2.1, evaporation is the principal

CONSULTATION DRAFT

process for water loss from closed lake basins. Therefore, it is essential in paleolacustrine studies to determine the surface area of paleolakes in order to use the equivalence of evaporation to estimate total runoff (surface water and ground water) in the basin.

The most direct means of measuring paleolake area is to trace shorelines by detailed geologic mapping within the context of a well-dated stratigraphy. Studies of cores obtained from lake beds provide information throughout the depositional life of a lake, but single-point depth sampling does little to establish lake area and provides incomplete data on other lake characteristics. Subsurface sampling with good areal distribution and detailed dating and correlation of deposits allow the reconstruction of sedimentary environments from which shoreline positions and paleohydrologic conditions can be estimated. However, studies of this type are very expensive and should be justified only by clear expectations of significant increases to our knowledge of paleoclimates in the southern Great Basin.

Finally, variations in the size of a lake lead to changes in the chemical conditions and thermal structure of the lake. These changes in turn lead to changes in the structure of plant and animal communities confined to the lake environment. The chemistry of minerals precipitated from a lake is also a function of the temperature and the chemical composition of lake water. In some instances, therefore, lake size can be inferred within broad limits from the fossil and mineralogic record preserved within lake sediment.

To estimate the evaporative discharge from paleolake surfaces, factors other than lake size that influence the volume of evaporation must be considered. Many empirical methods have been developed for estimating evaporation from water surfaces using commonly measured meteorological parameters, such as humidity or air temperature. Studies by Blaney (1957) concerning evaporation measurements at Lake Mead and Silver Lake are especially relevant to the review of the Yucca Mountain site because these lakes lie less than 200 km to the south. In both areas, annual lake surface evaporation exceeds 2 m. In reconstructing paleoclimates, several authors (Leopold, 1951; Broecker and Orr, 1958; Galloway, 1970; Mifflin and Wheat, 1979) have attempted to estimate the amounts of evaporation from various paleolake systems of probable late Wisconsin age. Approaches to the estimation of evaporation have been based primarily on estimates of maximum atmospheric temperature lowering during the last glacial or pluvial episode. Some authors used the lowering of past treelines as the evidence, while others used the past lowering of snowlines. These data were then used to estimate present and past temperature contrasts by assuming the constancy of lapse rates and the equivalence of free-air and ground-level rates, the mean and monthly distribution of the lowered temperatures, and the total dominance of air temperature over the other factors influencing evaporation rates. Later investigations have shown that these assumptions may lead to significant errors.

For example, Dohrenwend (1984) demonstrated that free-air lapse rates when measured at ground level are not necessarily representative of temperature variation with altitude. He proposed the following generalized model of temperature variation with altitude based on empirical data from eight mountain areas in the western United States: up to 300 m above basin floors, mean annual lapse rates are approximately zero; mean annual lapse rates are -0.057 K/100 m for 300 m to 2,000 m of additional altitude, and -0.76 K/100 m

for altitudes greater than 2,000 m above basin floors. However, even if this lapse-rate model is correct, it may not be applicable to times of high lake level and mountain glaciation. During such times, temperature at high altitudes would be influenced by the nearby mass of glacial ice, and temperature above the lake would be influenced by the heat content of water stored in the lake so that both the highstand lake and the mountain glacier would buffer the lapse rate in their immediate vicinity. Thus, extrapolating snowline air temperature estimates to surface air masses located immediately above lakes situated on distant basin floors seems questionable.

Galloway (1970) collected a large base of data relating mean monthly evaporation to mean monthly temperature. As shown on Figure 5-13, for any mean monthly air temperature, the range of mean monthly evaporation is approximately 0.12 m. Although air temperature is an important factor affecting evaporation rates, several studies have shown that other factors are not insignificant. Recent investigations that will continue during the site-characterization phase (Section 8.3.1.2) include studies of the influence of water temperature, humidity, amount and nature of cloud cover, and amount of heat stored in lakes on evaporation rates. Investigations are also evaluating various other methods (covariance, aerodynamic, Dalton, energy-balance, and combination approaches) for calculating evaporation rates. For calculating the sensitivity of evaporation rate to variation in one or more commonly measured climate parameters, the energy-balance method appears the most promising. Each of the heat terms contained within the energy-balance equation has a theoretical or empirical relation to one or more commonly measured climate parameters. The method is theoretically sound, and when applied to computational periods greater than 1 week, satisfactorily estimates evaporation rates that are nearly the same as those obtained from water budget calculations (USGS, 1954). Historically, the energy-balance method has been the standard to which other evaporation computation methods have been compared. Wind speed does not appear explicitly in equations used in the energy-balance method.

Morton (1986) uses a method of computing evaporation that adjusts the humidity of air passing over a lake for the change that occurs as a result of evaporation from the lake surface. Lake size thus becomes a factor. Using the parameters noted above, plus altitude, depth, salinity, and available evaporation data, Morton (1986) calculates evaporation according to three models and compares them with water budget estimates. He lists monthly evaporation estimates for Winnemucca, Pyramid, and Walker lakes in the Lahontan Basin and for Utah and Great Salt lakes in the Bonneville Basin. Averages of the calculated annual rates using the two methods that include data on lake size differ at most by 7 percent from water-budget estimates; calculations using pan evaporation data without lake-size corrections produce estimates that are 51 to 67 percent greater than water-budget estimates. Therefore, pan evaporation data should be reduced by about 30 percent in estimating the evaporation from natural lake surfaces.

Table 5-12 is a compilation of equivalent annual lake-surface evaporation expressed as 70 percent of the pan evaporation values from three locations in Nevada: Boulder City, at the southern extreme; Fallon Experimental Station, near midlatitude Nevada; and Rye Patch Dam, in the northern part of the state. From the range of the mean annual rates (193.44 to 86.93 cm/yr),

CONSULTATION DRAFT

Table 5-12. Annual equivalent lake surface evaporation data (calculated from pan evaporation data) in centimeters for three Nevada locations^a

Year	Boulder City	Fallon Experiment Station	Rye Patch Dam
1984	209.23	87.24	64.15
1983	181.10	94.12	53.91
1981	194.66	103.98	70.62
1978	175.47	98.34	71.60
1977	177.02 ^b	107.70	71.51
1976	NA ^b	110.36	65.32
1975	189.78	111.24	75.98
1974	174.82	115.61	81.82
1973	183.13	166.13	81.91
1972	184.70	104.82	84.78
1971	186.30	102.93	72.02
1970	202.80	111.18	60.44
1969	195.29	115.13	65.47
1968	200.72	110.02	64.69
1967	194.06	108.82	67.01
1966	199.19	125.38	102.87
1965	186.38	108.83	61.68
1964	210.26	107.92	69.78
1963	178.70	104.26	74.34
1962	210.68	104.45	80.04
1961	200.91	111.85	71.48
1960	204.16	114.18	93.86
1959	203.22	115.02	73.65
1958	193.86	105.84	81.15
1957	192.72	101.74	87.84
1956	201.43	73.02	87.35
1955	199.12	84.46	97.48
1954	199.40	82.08	100.58
1953	203.75	94.98	94.59
1952	186.05	78.87	93.31
1951	193.77	87.32	115.74
1950	209.32	84.42	112.73
1949	194.68	87.28	109.97
1948	210.31	87.82	72.91
1947	208.25	95.44	112.88
1946	200.53	87.93	113.99
1945	193.33	76.45	111.03
1944	214.44	85.81	117.90
1943	217.29	91.91	129.49
1942	217.96	97.50	128.32
1941	183.41	89.11	122.37
1940	220.47	86.36	78.53
1939	206.79	75.94	NA
Mean annual	193.44	98.46	86.93

^aBased on data from U.S. Weather Bureau, 1938-1965, and U.S. Department of Commerce, 1966-1984.

^bNA = not applicable.

Figure 5-13. Relation of mean monthly evaporation and temperature in the western United States. Source: Galloway (1970).

CONSULTATION DRAFT

it is evident that the climatic difference between southern and northern Nevada affects lake evaporation rates.

Using the Pyramid Lake, Nevada, area as a reference lake-climate system, the sensitivity of evaporation rate to change in each of six climate parameters will be investigated: (1) amount of sky cover, (2) type of sky cover, (3) air temperature, (4) water temperature, (5) dew point temperature (humidity), and (6) insolation at the top of the atmosphere.

Using established, empirical relations for components of the energy balance, the response of the calculated evaporation rate to change in the measurable climate parameters allows the following preliminary conclusions (Benson, 1986):

1. Evaporation rate strongly depends on the difference between air temperature and water temperature.
2. The use of insolation values for 12,000 and 18,000 yr ago results in small increases in the calculated rate of evaporation.
3. Relatively large absolute changes in relative humidity result in rather small changes in the calculated rate of evaporation.
4. Change in the fractional distribution and absolute amount of sky cover can bring about a significant reduction in the calculated evaporation rate.

These studies also tentatively demonstrate the fundamental role that reduced evaporation may have played in the creation and maintenance of high-stand closed-basin lakes such as Lake Lahontan. Using precipitation and runoff data for 1969, a particularly wet year, calculations for the period 14,000 to 12,500 yr ago indicate that evaporation must have been reduced to about 40 percent of its present-day value to support the large surface area of Lake Lahontan. Since this result rests on the assumption of a particular discharge rate, more work needs to be done to reconstruct actual discharge rates and thus test the importance of evaporation rate reduction in the creation of large lake systems.

Lake-level data

Historical records of lake-level change can be used to estimate times of climatic extremes. The surface altitudes of most of the lakes in the Great Basin before 1925 are only roughly known. Harding (1965) compiled existing lake-level data for Pyramid, Winnemucca, Walker, Mono, and Great Salt lakes. The data (Figure 5-14) indicate that before 1860 all lakes were at low levels. At that time, lake levels rose rapidly in all basins indicating a marked change in climate. From about 1870 until about 1915 most of the lakes were at high levels. After about 1915, lake levels began to decline, much of which probably can be ascribed to diversion and consumptive use associated with agriculture. The Great Salt Lake record shows a significant recent rise in lake level (Figure 5-14) due in part to an increase in precipitation (Arnow, 1984).

CONSULTATION DRAFT

Figure 5-14. Historical fluctuations in the levels of Great Salt, Walker, Pyramid, Mono, and Winnemucca Dry lakes.

5.2.1.2.2 Prehistoric lake-level fluctuations

As discussed in Section 5.2.1.2.1, changes in the sizes of lakes occur as a result of changes in the regional hydrologic balance. The hydrologic balance of the basin in which the lake is located is, in turn, a function of climate. While several criteria can be used to reconstruct past changes in hydrologic balances, paleolake reconstructions may be the most relevant criterion because long-term changes in lake areas, to a large extent, reflect changes in the total amount of runoff (both by surface water and ground water) from the surrounding basins.

Records of prehistoric fluctuations in lake characteristics can be used to

1. Determine the nature of climate variability expressed as the hydrologic balance of a closed-basin surface-water system on the scale of 100 to 100,000 yr.
2. Provide information on the relationships between climatic variations reconstructed from other proxy data (such as paleobotanical records) and the hydrologic system.
3. Provide data necessary to validate climate models.

Dating of paleolake deposits

A chronology of variation in lake size can be obtained through the use of various methods of dating, including radiocarbon, chlorine-36, tephrachronology, uranium series, and magnetostratigraphy. Low lake levels in some closed basins can be identified by changes in the carbonate mineralogy or by the presence of salt layers, which indicate either desiccation or a shallow, saline lake.

Changes in the level and thus the size of a lake for the past 30,000 yr can often be determined by radiocarbon dating of materials, such as tufa or gastropods, deposited near the air and water interface near the paleoshoreline. Sediment cores recovered from the deepest portion of lake basins contain the remains of microscopic lacustrine plants and animals. Variations in the relative abundances and types of these organisms are a function of lake chemistry and physics and, to some extent, size. By examining the biotic content of cores taken from the deepest part of lake basins, it is possible to reconstruct continuous records of lake size over time periods possibly extending back to the beginning of the Quaternary.

Broecker and Orr (1958), Benson (1978), and Stuiver and Smith (1979) have outlined why the radiocarbon activities of lacustrine samples may not always be accurate measures of sample age. The radiocarbon activity can be affected by the following processes:

1. The introduction of carbon-bearing older detritus into a sediment sample, or the solution of that older detrital carbonates after deposition in the lake, altering the apparent age of its water.
2. Precipitation of younger secondary carbonate in the sample.

3. Recrystallization of metastable carbonate phases, exchanging carbonate in the process.
4. Mobilization and reprecipitation of inorganic carbon below the sediment and water interface.
5. Introduction into the lake of waters containing carbon-14, free bicarbonate, and carbonate leached from old carbonate rocks within the watershed (Broecker and Walton, 1959).
6. Nonequilibration between contemporary radiocarbon of the atmosphere and the carbon dissolved in the lake water. For example, Broecker and Kaufman (1965) report that modern Mono and Pyramid lake waters have apparent ages of 1,800 and 800 yr, respectively.

Distribution and age of paleolakes in the Great Basin

The pluvial lake systems in the Great Basin thought to be of Pleistocene age are shown in Figure 5-9. The names and summary hydrographic characteristics of these paleolakes have been given by Mifflin and Wheat (1979).

Only a few of the Great Basin paleolake systems have been sufficiently dated. Nonetheless, most of the radiocarbon dates point toward the existence of lakes in the basins within the last 30,000 yr.

Of the many basins, four of the paleolake systems (Bonneville, Lahontan, Russell, and Searles) have been intensively studied using radiometric dating techniques. Their chronologies for the past 20,000 yr are known well enough for the purpose of model validation. Of these systems, Russell and Searles are close to the Yucca Mountain region and near the same latitude.

Before discussing individual records of lake-level change, it is helpful to consider certain general relations between basinal topography and hydrology that influence the timing of lake-level change. During pluvial maxima, each of the lake systems consisted of a closed basin composed of one or more subbasins, separated from each other by sills (the lowest point on the divide separating adjoining basins or a series of separate basins connected by overflow). Today, some of these subbasins are fed by perennial streams and still contain lakes that rise and fall with changes in climate. Thus, it is the climate in the watershed area of the bordering mountain range that controls surface inflow to a subbasin, not the climate of the basin floor area. The climate of the basin floor area, however, determines the free-surface evaporation rate. The climates of both areas are influenced by the pronounced orographic effect that the Sierra Nevada and other high mountain ranges have on precipitation in the Great Basin (Maxey and Eakin, 1949).

The topography of a basin influences lake-level fluctuations. Lakes in small basins fed by perennial streams will change these levels rapidly in response to changes in moisture storage (influx minus evapotranspiration). Such lake systems are potentially excellent recorders of high-frequency, low-amplitude change in climate on the subregional scale. Lakes that lie in basins with large surface areas fed by perennial streams respond more slowly to changes in moisture storage. Also, the initiation of lake-level rise in a basin that receives surface inflow only when an adjoining basin overflows

will lag the onset of lake-level expansion in that adjoining basin by the amount of time it takes the source basin to fill to its sill level. Thus, change in the size (elevation, volume, surface area) of a lake in any particular subbasin or series of basins can be a complicated function of intra-basin climate, extrabasin climate, and basin topography. In settings where adjoining basins coalesce at some level, the combined lakes act as a single body of water responding in an integrated manner to changes in the regional climate.

The records of the four paleolake systems that have been intensively studied are discussed in the following paragraphs.

Lake Bonneville system. Lake Bonneville, located in western Utah (Figure 5-15), had a surface area at its highest stage of about 52,000 km², a volume of 7,500 km³, and a maximum depth of 335 m (Smith and Street-Perrott, 1983); it occupied two main subbasins: Salt Lake subbasin and the Sevier subbasin. Today, runoff contributes about 66 percent, direct precipitation about 31 percent, and ground water about 3 percent of the average influx to Great Salt Lake, the largest present-day manifestation of Lake Bonneville.

Numerous workers have contributed to the knowledge of Lake Bonneville since the classic study of Gilbert (1890). Until recently, the interpretations of Morrison (1965) and Morrison and Frye (1965) were probably cited most frequently, although Eardley et al. (1957), Broecker and Orr (1958), and Broecker and Kaufman (1965) were cited in more complete reviews of the literature. Within the last few years, stratigraphic and chronologic investigations have made extensive use of dating techniques that were not available to earlier studies of Lake Bonneville. These recent studies have given rise to a substantially modified interpretation of fluctuations in the level of Lake Bonneville. The Lake Bonneville chronologies of Currey and Oviatt (1985), Spencer et al. (1984), Scott et al. (1983), and Morrison (1965) are shown together in Figure 5-16. Because of the large distances separating study localities, physically tracing and correlating units is difficult because of disconformities, abrupt facies changes, and similarities in the appearances of deposits from different lake cycles. Scott et al. (1983) relied mainly on analyses of amino acid in shells and radiocarbon dating of wood for correlation, and samples from deposits of uncertain origin were assigned to a specific lake cycle on the basis of the similarity of their alloisoleucine to isoleucine ratios relative to ratios of samples whose relative ages had been determined by other means.

Lake Lahontan system. The Lahontan basin consists of seven subbasins separated by sills of differing altitudes (Figure 5-17). Of the six rivers that terminate in Lahontan subbasins, four (Truckee, Carson, Walker, and Humboldt) presently contribute 96 percent of the total gaged surface inflow (Benson, 1986). The Truckee, Carson, and Walker rivers have their headwaters in the Sierra Nevada, west of the Lahontan basin. The annual flows of these three rivers are highly correlated with each other (Benson, 1986). The Humboldt River has its headwaters in mountain ranges east of the basin. Lake Lahontan, at its highest stage 13,000 yr ago, had a surface area of about 22,300 km², a volume of 2,130 km³ (Benson, 1978), and a maximum depth of 280 m in the Pyramid Lake subbasin (Smith and Street-Perrott, 1983).

Figure 5-15. Lake Bonneville subbasins. Source: Morrison (1965).

CONSULTATION DRAFT

Figure 5-16. The last cycle in the lake Bonneville Basin as interpreted by (a) Currey and Oviatt (1985), (b) Scott and others (1983), (c) Spencer and others (1984), and (d) Morrison (1965).

Figure 5-17. Surface extent of Lake Lahontan and geography of its basin 14,000 to 12,500 yr before present (modified after Benson and Mifflin 1986).

CONSULTATION DRAFT

The amount of subsurface inflow to present day surface-water bodies in the Lahontan basin is thought to be very small compared with the total amount of water reaching lakes, sinks, and playas in the form of runoff and precipitation (Everett and Rush, 1967). Surface flow accounts for about 85 percent of that total (Benson, 1986).

The first comprehensive study of Lake Lahontan was done by Russell (1885). Broecker and Orr (1958) and Broecker and Kaufman (1965) made the first systematic attempts to assign an absolute time scale to Lake Lahontan history. Morrison and Frye (1965) noted that certain tufa radiocarbon dates were reversed in relation to Morrison's (1964) stratigraphic assignments, which were developed without the benefit of radiocarbon dating. Benson (1978) developed a lake-level chronology for the Pyramid and Walker lake sub-basins that used a sample selection procedure that helped to eliminate problems caused by the introduction of secondary carbon into tufa samples. Thompson et al. (1986) presented a chronology for the last Pleistocene lake cycle in the central Lahontan Basin, based on radiocarbon dates and tephrachronology. Their data, together with the data of Born (1972) and Benson (1981), have been used to form the Lahontan lake-level chronology of Figure 5-18.

The age of older lacustrine deposits in the Lahontan Basin is not well known. Broecker and Kaufman (1965) dated several samples of gastropod shells and tufa from those deposits. Radiocarbon activities indicated ages in excess of 34,000 yr. Subsequent age determinations on gastropod shells containing less than 5 percent secondary calcite, using the thorium-230/uranium-234 method, resulted in ages near 120,000 yr. These ages are broadly consistent with thorium-230/uranium-234 ages of about 92,000 and older than 100,000 yr on gastropod shells from the Bonneville Basin (Broecker and Kaufman, 1965). Difficulties associated with applying the thorium-230/uranium-234 method to carbonate samples found in closed basin lake systems preclude strict reliance on age dates determined using this method. However, the data are sufficient to suggest the adoption of a working hypothesis that at least two lake cycles occurred more or less synchronously in both the Bonneville and Lahontan lake basins during the time intervals from 120,000 to 90,000 yr ago and from 25,000 to 12,500 yr ago.

Lakes Russell and Searles. The Owens River system in the late Pleistocene consisted of a chain of lakes occupying a succession of basins east of the Sierra Nevada in California (Figures 5-19 and 5-20). The Owens River supplied water to the lakes downstream from its terminus (Gale 1914; Smith, 1979). When Lake Russell in Mono Basin overflowed, its surplus flowed into the headwaters of the Owens River. Only two of the seven lakes in this system, Russell and Searles, have been studied in detail.

Mono Lake and its ancestral Lake Russell lies in a relatively steep-sided basin separated from Adobe Lake and the Owens River drainage by a sill. Because of its morphology, the lake responds in a sensitive manner to changes in water volume and is potentially an excellent recorder of high-frequency, low-amplitude climatic events that take place in its watershed. Lajoie (1963) made a detailed stratigraphic study of the lake sediments, the chronology of which has been supported by radiocarbon dates on ostracodes and tufa

Figure 5-18. Lake Lahontan chronology inferred from data of Born (1972), Benson (1981), and Thompson et al. (1986).

CONSULTATION DRAFT

Figure 5-19. Location of lakes in the Owens River system during pluvial periods in the late Pleistocene (modified after Smith, 1979).

Figure 5-20. Overflow sequence of lakes in the Owens River system (modified after Gale, 1914).

Figure 5-21. Chronologies of lakes in the Russell (Mono) and Searles basins (after Smith and Street-Perrott, 1983, Figure 10-5).

and by tephrachronology. Reconstruction of the lake-level history of Lake Russell for the past 20,000 yr is shown in Figure 5-21. The lake-level chronology depicted in Figure 5-21 is noteworthy in that it closely resembles the Lahontan chronology, Figure 5-18.

The other lake in the Owens River system that has been intensively studied is Searles Lake (Smith, 1979; Stuiver and Smith, 1979). During the late Pleistocene, Searles was third in a chain of lakes that received water from the Owens River. When Owens filled to a level of about 60 m, it overflowed into China Lake; when China Lake filled to a depth of 12 m, it overflowed into Searles Lake; and when Searles Lake reached a level of about 200 m it coalesced with China Lake and overflowed to create a relatively shallow lake in Panamint Valley. During pluvial periods of earlier Pleistocene times, Panamint Lake overflowed into Lake Manly (Death Valley), which was also the terminus of the Mojave and Amargosa rivers (Figure 5-20).

The chronologies of Searles Lake and Lake Russell for the past 20,000 yr are shown in Figure 5-21. The mismatch between the highest stands of Searles Lake and Lake Russell during the time period 20,000 to 11,000 yr ago might be explained in one of the following ways:

1. The data or the data interpretation of one or both of the lake chronologies may be in error.
2. The timing of lake-level fluctuations in the Searles Basin may have been delayed by the cumulative times required for basins upstream in the Owens River system to fill to overflowing or to cease flowing.
3. Both chronologies are basically correct, but there may have been a distinct north-south climate gradient between the watersheds that supply each of the lakes, with each watershed responding to the changing distribution of Pacific sea-surface temperatures in different ways and at different times.

The Searles Lake chronology for the past 30,000 yr, unlike the Russell, Lahontan, and Bonneville chronologies, is based about equally on radiocarbon analyses of samples from both outcropping sediments and core materials, but lacustrine materials of both types carry uncertainties about their carbon-14-based ages. An example from the subsurface record at Searles Lake (Stuiver and Smith, 1979) illustrates the difficulties inherent in all carbon-14 dates from materials deposited in lakes and why differences in chronologies of a few thousand years should not be necessarily viewed as significant. Of the four radiocarbon dates on disseminated organic carbon and wood from the overburden mud (Figure 5-22), Stuiver and Smith (1979) concluded that only the single date on wood was reliable. They reasoned that the unit was deposited in a much smaller lake than were most mud units in this basin, and large areas of older lake beds that contained carbon and carbonate minerals were being eroded, thus contaminating the units being deposited and creating the 3,000-yr discrepancy between the date on wood from this unit ($3,520 \pm 190$ yr) and the older date on detrital carbon from a nearby horizon. A discrepancy having the same magnitude and direction is found between the dates on wood and organic carbon in a mud layer in the Lower Salt (Figure 5-22).

CONSULTATION DRAFT

Figure 5-22. Summary of carbon-14 dates on subsurface samples composed of organic carbon and wood from Searles Lake (from Stuiver and Smith, 1979, Figure 30).

Figure 5-21 indicates that Lake Russell experienced a high lake cycle beginning at about 15,000 yr ago and ending about 12,000 yr ago. The lake peaked from about 14,000 to 13,000 yr ago and may or may not have spilled. The Searles record (Smith and Street-Perrott, 1983) shows a strong increase in lake size starting at about 12,500 yr ago. While carbon-14 dates from these areas could be in error, they appear to be substantially correct, and the presence of the brief period of high lake stands between about 12,500 and 10,500 yr ago in Searles is suggested by an episode and age assignment supported by extensive stratigraphic and radiometric dating evidence (Smith and Street-Perrott, 1983).

The broad agreement between the Bonneville, Lahontan, and Russell records--disregarding the differences of a few thousand years in peak stages--suggests that they all occupied a zone that shared a similar climatic history. The similarity between historical records from Great Salt, Walker, Pyramid, Mono, and Winnemucca lakes (Figure 5-14) supports this view. The strong evidence for a different history for lakes in the Owens River system, derived from Searles Lake, creates doubts as to the comparability that should be expected between the climatic histories of the two regions. The similarities in the latitudes (37°) of the Yucca Mountain site and the segment of the Sierra Nevada that contributed most of the runoff to the Owens River system and Searles Lake supports the notion that the Searles Lake history is the better of the pluvial lake records for the evaluation of future climatic change in the Yucca Mountain area.

Death Valley drainage system. During portions of the Pleistocene, Death Valley received surface water flow from the Owens River system (described previously), surface water from the Mojave River drainage, and surface flow and ground water from the Yucca Mountain region. Because of this latter hydrologic tie to the NTS, geologic records from Death Valley are relevant to site characterization activities in paleohydrology, as well as paleo-climatology. Additional discussions are found in Sections 3.7.4.1 and 3.7.4.5.

Because of the poor preservation of shoreline features, relatively little is known about the age and depth of the Pleistocene Lake Manly in Death Valley. Gravel and beach deposits of possible shorelines occur at heights more than 200 m above the basin floor; however, the estimates of the age of these sediments range from Pleistocene (Butler, 1986) to early Wisconsin (Hunt and Mabey, 1966). Radiocarbon dates on organic lake sediments from the last major lake cycle suggest that a deep water lake existed in Death Valley from 26,000 to 10,000 yr B.P. (Hooke, 1972).

In opposition to the above interpretation, Butler (1986) argued that terraces and uplifted alluvial fans in the elevational range from sea level to 120 m have never been covered by lake waters, and thus lake highstands have not occurred in Death Valley since the Pleistocene. However, even if Butler is correct, a lake of up to 86 m in depth could have been present (existing completely below sea level) in Death Valley during portions of the late Pleistocene. Craig (1984) used computer simulations to demonstrate the feasibility of past lake stands in Death Valley, given a range of past conditions of increased runoff or reduced evaporation or both. Dorn et al. (1987) interpreted isotopic evidence from rock varnish on alluvial fans in

Death Valley as indicating the past occurrence during the late Pleistocene of three periods of relatively more humid conditions separated by intervening arid intervals.

The Pleistocene Lake Mojave was present in what are now the Silver and Soda Lake playas in the Mojave River drainage, which drains into Death Valley from the south. Ore and Warren (1971) interpreted geomorphic features and radiocarbon dates on shell and tufa as indicating the following chronology:

1. A major lacustral episode, which reached the overflow level, ending approximately 14,500 yr B.P.
2. A second lacustral episode, which reached the same level, occurring between approximately 13,750 and 12,000 yr B.P.
3. A third lacustral episode, during which, the outlet was downcut, occurring between approximately 11,000 and 9,000 yr B.P.
4. A final lacustral episode, estimated to have occurred between approximately 8,500 and 7,500 yr B.P., was of lesser magnitude and did not overflow the outlet.

Further studies by Wells et al. (1984, 1987) provide additional radiocarbon dates that substantiated the late-Pleistocene portion of the chronology presented above, although these authors suggest that due to bedrock control at the outlet, relatively continuous highstand conditions (with only minor fluctuations in level) occurred between 15,500 and 10,500 yr B.P. These authors also question the dating of the lowest and youngest lake stand, and estimate its age to fall between 10,500 and 8,000 yr B.P. and suggest that it was a short-lived phenomenon (less than a few hundred years in duration). Wells et al. (1987) found evidence of larger than modern flood events in the early and middle Holocene and interpreted these deposits as indicative of intensified monsoonal conditions.

Banks of lake sediments up to 68 m thick are present in the Lake Tecopa basin in the Amargosa River drainage, which flows from the Yucca Mountain area to Death Valley (Sheppard and Gude, 1968). These lake beds are estimated to be of no greater than mid-Pleistocene age and were deposited in alkaline saline waters.

Marshes deposits in Southern Nevada. No evidence has been found of deep water lakes of late Wisconsin age in the Mojave Desert portion of southern Nevada (Mifflin and Wheat, 1979). Instead, the ground-water-fed marshes and wet meadows present in the early historical period were greatly expanded in distribution and in areal coverage. Stratigraphic studies at Tule Springs (Haynes, 1967) and Corn Creek Flat and Indian Springs (Quade, 1985) indicate that marshes and associated shallow lakes reached their maximal extent between 30,000 and 15,000 yr B.P. A progressive desiccation of the marsh-lake complexes occurred between 13,500 and 8,000 to 7,200 yr B.P. After this time perennial water was absent, and the sediments deposited under the moister Pleistocene and early Holocene conditions became dissected. At Corn Springs Flat, the modern water table in the valley center is more than 25 m below the level at 8,500 yr B.P., when it was near the ground surface.

Mehring and Warren (1976) constructed a radiocarbon-based chronology of changes in marsh coverage and dune migration over the last 5,300 yrs at Ash Meadows, in the Amargosa River drainage south of Yucca Mountain. Peat formation occurred between 5,300 and 4,500 yr B.P. on top of older dune sands. Dune sand covered the site from 4,500 to 4,000 yr B.P., followed by peat development between 3,000 and 2,000 yr B.P. Net deflation occurred between 2,000 and 400 yr B.P., when peat deposition resumed.

Ground water near Yucca Mountain. Radiocarbon dating of ground water in the Amargosa Desert (Claassen, 1985) and at Yucca Mountain (Benson et al., 1983; Benson and McKinley, 1985) indicates that the majority of the recharge that led to the accumulation of these waters occurred between 20,000 and 12,000 to 10,000 yr B.P. Some of the ground-water ages suggest that significant recharge may also have occurred between 35,000 and 30,000 yr B.P. and during the early Holocene. These radiometric dates provide evidence that the paleoclimatic conditions responsible for Pleistocene lake high stands also had significant effects on the hydrologic regime of Yucca Mountain.

Summary of paleolacustrine data

Available information indicates the presence of more than one lake cycle in the Great Basin during the past 150,000 yr. The last cycle that occurred during the late Pleistocene is well documented in the Bonneville, Lahontan, Mono, and Searles basins. Older cycles in both the Bonneville and Lahontan basins are less adequately documented (in terms of absolute age determinations), but one cycle may have occurred $115,000 \pm 15,000$ yr ago. In Searles basin, the nearly continuous existence of perennial bodies of water, over a period extending from 130,000 to 10,000 yr ago, interrupted by only brief periods of salt deposition, is documented by a wealth of subsurface data (Smith, 1979), supplemented by carbon-14 (Stuiver and Smith, 1979) and uranium-series dating (Bischoff et al., 1985). A large number of basins throughout the Great Basin that today are playas or dry lakes contain shoreline evidence of more than one former lake cycle. Radiocarbon analyses of material from these basins indicate that the youngest shoreline probably corresponds to a lake cycle that occurred in the late Pleistocene. The ages of the older shorelines are not known.

5.2.1.2.3 Time series of prehistorical vegetation change

Paleobotanical data provide the most widespread and continuous records of paleoclimatic change available in this region. Paleobotanical assemblages can be reliably dated, a necessary requirement for modeling. However, since radiocarbon dating is one of the only means of determining ages of most paleobotanical materials, this powerful tool can be applied widely only in studies of late Wisconsin climates (<40,000 yr). Spatial variations in modern climate in the Yucca Mountain region are reflected in the abundances and distributions of plant taxa. Adequate reconstruction of paleobotanical assemblages is, in part; dependent on understanding the variations in the abundance and distribution of modern plant taxa (Spaulding, 1983). Transfer function and response surface techniques will be used to produce quantitative estimates of past precipitation, temperature, and other climatic parameters.

CONSULTATION DRAFT

These estimates, in turn, will be used in the testing and validation of climatic models.

Plants react in various ways to variations in climate on many time scales. Annual or seasonal climatic fluctuations affect the growth rates of plants and may be recorded in the width of annual growth rings in mature trees. Dendroclimatological methods have been used to extract climatic data from tree-ring series covering the last 300 to 5,500 yr (LaMarche, 1974) in the western United States. Changes in the patterns of vegetation also reflect larger scale and longer term climatic changes. Pollen and pack rat midden studies have provided proxy paleoclimate data forming the basis for the preparation of maps or tables of vegetation types. These data imply certain ranges of paleoclimate. Longer term (and larger scale) climatic fluctuations can thus be deduced using radiocarbon-dated pollen profiles and pack rat (*Neotoma*) middens. Paleobotanical data are available from the region surrounding Yucca Mountain as plant macrofossil assemblages from pack rat middens from dry caves and rockshelters and as fossil pollen from lake and marsh sediments. The temporal resolution of the paleoclimatic reconstructions from these data will potentially be limited by the sampling interval for pollen data from lacustrine cores and by the availability of pack rat middens representing the time periods of interest. In the case of palynological data, studies of the pollen content of lake sediments in the Great Basin have had sampling intervals (in temporal terms) of 200 to 1,000 yr (refer to Thompson (1984) for examples). This sampling interval would permit the reconstruction of paleoclimatic fluctuations with a period of 200 to 1,000 yr. The temporal resolution for pack rat midden studies will probably be lower, and it will probably not be possible to confidently reconstruct fluctuations with periods of less than 1,000 yr using pack rat midden data.

Rationale

Studies of past changes in vegetation can be used in the following aspects of paleoclimatology:

1. To determine the nature, range, and frequency of past climatic variability.
2. To provide well-dated paleobotanical time series for the testing and validation of climatic models. These time series provide two types of necessary information:
 - a. Long continuous records reflecting low-frequency climatic changes that can be used to reconstruct the slowly varying aspects of climatic change.
 - b. A network of sites to test synoptic scale predictions of the more rapidly varying aspects of the models.

Modern vegetation and climate

The sensitivity of plant distributions to geographic and elevational variations in climatic parameters provides a basis for the interpretation of past climatic fluctuations from changes in vegetational records.

The distributions and associations of plant species in the region surrounding Yucca Mountain are greatly influenced by variations in climate, especially temperature and precipitation. For example, the geographic distribution of summer rainfall appears to influence the regional distributions of pinyon pine (Pinus monophylla), Utah juniper (Juniperus osteosperma), ponderosa pine (Pinus ponderosa), and other woodland and montane plants (Houghton et al., 1975). Temperature extremes may also limit the distributions of certain plants. For example, the geographic and elevational limits of many woodland species may be controlled by the extent or duration of winter frosts (Houghton et al., 1975).

On a more local level, plant taxa are segregated into elevational zones that reflect the effects of environmental lapse rates on the elevational distribution of temperature and precipitation (Houghton et al., 1975; Figure 5-23). In the Great Basin the general progression is from shadscale desert on the hot arid valley bottoms, through sagebrush steppe on bajadas and mountain slopes, pinyon-juniper woodland and upper sagebrush grassland at intermediate elevations, to subalpine conifer forest at or near the mountain summits. An exception to this pattern is found in mountain ranges of the northwest Great Basin, where there is even less summer rainfall. Here, sagebrush dominates the entire elevational range. The mountains of the eastern Great Basin that receive more summer precipitation than ranges farther west support ponderosa pine, douglas fir, and other montane plants at middle elevations.

A similar zonation of plant communities occurs in the Mojave Desert. The valley bottoms in this region are lower, hotter, and drier than those of the Great Basin. Mojave desert scrub communities, usually dominated by creosote bush (Larrea divaricata), joshua tree (Yucca brevifolia), and Mojave yucca (Yucca shidigera), occur in these arid settings.

Dendroclimatology

Past variations in tree-ring widths reflect interannual and even seasonal fluctuations in climate on the scale of 10 to 1,000 yr that can be precisely dated to the year or season of occurrence through dendrochronological techniques within a single climatic province (Brubaker and Cook, 1983). A network of 65 carefully selected tree-ring chronologies has been used to reconstruct synoptic scale variations in western climate for the period 1600 to 1960 (Brubaker and Cook, 1983). These reconstructions suggest that while other portions of the northern hemisphere were relatively cool (compared with 20th century means) from 1600 to 1900, the western United States, and in particular the Great Basin, was warmer during this earlier period than during this century (Brubaker and Cook, 1983). An important insight from this research is that the climatic variations in the West may not simply follow mean global or hemispheric trends.

LaMarche (1974) provides another example of paleoclimatic reconstruction from a tree-ring chronology of the bristlecone pines (Pinus longaeva) in the White Mountains of California. LaMarche reconstructed variations in warm-season temperatures over the last 5,500 yr. LaMarche (1974) describes climatic variations qualitatively (e.g., cooler and wetter) but does not quantify the variations from tree-ring data alone. Analysis of mean tree ring

Figure 5-23. Generalized relationships among altitude, precipitation, and the distribution of plant communities for the Great Basin portion of Nevada (modified from Houghton et al., 1975).

widths at the upper tree line suggests periods of relative warmth (compared with the long-term mean) from 3500 to 1300 BC and from 200 BC to 300 with an intervening period of cooler summers. Renewed cooling is suggested after 300 with a brief warm interval at 1200 followed by an extended cool period after 1200. In apparent contrast to the synoptic-scale reconstructions of western climate of Brubaker and Cook (1983), from 1600 to 1960, LaMarche's data from the White Mountains indicate that the last century has been warmer than the preceding few centuries. LaMarche and Mooney (1972) also used remnant tree stumps standing well above the modern tree line in the White Mountains, California, and the Snake Range, Nevada, to demonstrate that (during portions of the middle Holocene) summer temperatures were well above those of today.

Palynology and pack rat midden studies

Fossil pollen in stratigraphic deposits have been extensively used in eastern North America and northern Europe to reconstruct fluctuations in temperature and precipitation over the last 15,000 yr (Webb, 1985). The applicability of the method has been somewhat limited by the relative scarcity of bogs and small lakes, the optimal settings for most types of pollen studies. Nevertheless, substantial paleoclimatological insights have been gained from studies of the fossil pollen contained in sediments deposited in western pluvial lakes, alpine lakes, caves, rockshelters, and alluvium. Pollen records, in general, provide quasi-continuous records of regional vegetational change. While the majority of these studies are restricted to the Holocene, palynological studies of pluvial lake sediments can provide information for the last 40,000 yr.

Pack rat middens are restricted to the semiarid and arid deserts of western North America. In the study of middens, past vegetation is reconstructed from plant remains deposited in rock cavities by rodents of the genus Neotoma (pack rats, wood rats, or trade rats). The plant remains can often be identified to the species level, and the assemblage is believed, based on the contents of modern middens, to represent vegetation growing within a foraging radius of 30 m at the time of accumulation (Spaulding, 1985). Selected plant macrofossils or other materials from each midden are submitted for radiocarbon analysis. The degree of precision of radiocarbon dates for plant macrofossils and other materials from pack rat middens, as with wood and other organic materials, generally increases with decreasing sample age and also with the length of time allowed for the analysis. Because older samples possess less carbon-14 radioactivity than younger samples, a longer period of radioactivity counting is required to gauge the carbon-14 content of an older sample with the same precision that would result from a shorter counting period with a younger sample. The precision of radiocarbon dates for wood and plant macrofossils is high because the potential for contamination by carbon from outside the sample system is low. The energy required for the replacement of carbon in organic compounds is high. Where more precision in dating is required or where the accuracy of a given date is in question, multiple samples from the same midden are submitted for radiocarbon dating. The paleobotanical data from the individual midden assemblages are compiled to create time series of vegetational change from a given area and vegetational setting. Replication of pack rat chronologies from sites with similar settings ensures that site-specific phenomena do not introduce bias into the interpretation of the data set.

Figure 5-24. Locations of study sites for pollen and pack rat midden investigations discussed in text.

Palynological and pack rat midden studies can be used together to construct a regional network of time series of vegetational change spanning the last 45,000 yr. Figure 5-24 shows the locations where palynological and pack rat midden data have been collected. Table 5-13 compiles the results of these studies of vegetational assemblages and inferred paleoclimate conditions for the Great Basin, Mojave Desert, and Sierra Nevada at different times in the Pleistocene and Holocene. The table summarizes some of the interpretations resulting from paleoclimate research in these areas.

Table 5-14 is a compilation of vegetative cover and inferred climatic conditions in the Yucca Mountain area over the past 45,000 yr. It illustrates a correlation between variation of species with elevation over time and variation in temperature and precipitation.

The climatic inferences presented in Table 5-14 are slightly different from those discussed in Table 5-13 for some of the site locations. These differences are due to slightly different interpretations by the various workers. As more data are collected during site characterization, these differences are expected to be reduced.

Summary of paleovegetation changes

During the Wisconsin glacial period, greater than 40,000 to 11,000 yr ago, sagebrush steppe ranged over larger elevation and area than today in the Great Basin. It expanded into the Sierra Nevada and Wasatch mountains and descended southward into the modern Mojave Desert. Shadscale and other xerophytic steppe plants were common in this regime, particularly in the Mojave region. The modern Sierran forests, Great Basin pinyon-juniper woodlands, and southwestern ponderosa pine forests were absent or restricted in extent. In the Great Basin region, xerophytic subalpine conifers grew on rocky substrates as low as the local current base level, surrounded by steppe vegetation on fine valley-bottom soils. In the Mojave Desert, subalpine conifers grew on the lower mountain slopes, where pinyon-juniper and juniper woodlands now grow. Patches of treeless desert scrub were present in the most xeric settings in and near Death Valley. This was not modern Mojave Desert vegetation of the region, though some Mojave plants, such as Joshua tree, were present. Creosote bush was absent, and shadscale, sagebrush, and other Great Basin plants grew in its place.

The transition to Holocene conditions occurred after 11,000 yr ago. The lower elevation limits of junipers and sagebrush were below those of today until 9,000 to 8,000 yr ago. By 7,000 to 6,000 yr ago, the lower boundary of these plants moved upslope, creosote bush arrived within its modern range in the Mojave Desert, and pinyon pines dispersed into the Great Basin. While minor changes in vegetation occurred after 6,000 yr ago, the modern communities were in place by this time.

The late Wisconsin vegetation was characterized by a greater coverage (relative to today) of cold-dry steppe species (sagebrush and shadscale) and the elevational depression of cold-tolerant xerophytic conifers (bristlecone pine, limber pine, Utah juniper). This suite of plants reflects cooler temperatures than present and low to moderate rainfall with a pronounced dominance of cool-season precipitation.

Table 5-13. Summary of palynological and pack rat midden data from the Great Basin, Mojave Desert, and Sierra Nevada with inferred climatic changes (page 1 of 8)

Location ^a	Evidence	Age	Vegetational assemblage	Inferred climatic conditions
WESTERN GREAT BASIN				
Lake Lahontan basin (11,35)	Pack rat middens	12.1 to 11.5 thousand yr ago	Western juniper, sagebrush, and other Great Basin shrubs at lower than modern elevation limits (Thompson et.al., 1986).	Semiarid, with dominance of cool-season precipitation and little or no summer precipitation. Summer temperatures cooler than present, but winter temperatures may not have been much colder (Thompson, 1984).
CENTRAL GREAT BASIN				
Toquima Range (30)	Pollen, pack rat middens	6 thousand yr ago to present	Early Holocene sagebrush steppe with aspen replaced by pinyon-juniper woodland (Thompson and Hattori, 1983).	Early Holocene assemblages may reflect dry conditions with cooler than modern summer temperatures. Differences between middle Holocene and modern woodland plant associations reflect moisture conditions or higher proportion of summer rainfall (or both) from 6 to 3 thousand yr ago.
Ruby Marshes (20)	Pollen	35 to 8 thousand yr ago	Expansion of sagebrush steppe in the valleys; reduction of tree cover (Thompson, 1984).	Probably reflects relatively cold and dry conditions with strong predominance of winter precipitation.
		7 to 5 thousand yr ago	Expansion of shadscale steppe; establishment of pinyon-juniper woodland.	Suggests warmer and perhaps drier than modern conditions. (Thompson, 1984).

Table 5-13. Summary of palynological and pack rat midden data from the Great Basin, Mojave Desert, and Sierra Nevada with inferred climatic changes (page 2 of 8)

Location ^a	Evidence	Age	Vegetational assemblage	Inferred climatic conditions
EASTERN GREAT BASIN				
Mountain ranges of east-central Nevada and west-central Utah: Schell Creek (21), Snake (25), Deep Creek (6), Confusion (4), and Wah Wah (33) ranges.	Pollen, pack rat middens	> 40 to 11 thousand yr ago (late Wisconsin)	Absence of many woodland plants common in modern assemblage. Occurrence of relatively xeric subalpine conifers, including bristlecone pine (<u>Pinus longaera</u>), limber pine (<u>Pinus flexilis</u>), Englemann spruce (<u>Picea englemanni</u>), and prostrate juniper (<u>Juniperis communis</u>), at elevations as much as 1,000 m below their modern limits (Thompson and Mead, 1982; Thompson, 1984).	Suggests late Wisconsin climate colder than modern, and relatively dry with little summer precipitation.
		11 thousand yr ago to present	Subalpine conifers (except for limber pine) reduced at low elevation xeric sites. Quaking aspen (<u>Populus tremuloides</u>), Rocky Mountain maple (<u>Acer glabrum</u>), and other mesophytic montane shrubs increase in abundance. Woodland junipers well established by 7.4 thousand years ago, pinyon-juniper woodland by 6.1 thousand years ago.	Indicates early Holocene climate warmer than late Wisconsin, but still cooler than modern climate. No evidence for major fluctuations in assemblage over last 6.0 thousand years, indicating prevalence of modern climatic conditions during this period.

Table 5-13. Summary of palynological and pack rat midden data from the Great Basin, Mojave Desert, and Sierra Nevada with inferred climatic changes (page 3 of 8)

Location ^a	Evidence	Age	Vegetational assemblage	Inferred climatic conditions
Meadow Valley Wash (15)	Pack rat middens	20.1 to 12.6 thousand yr ago	Montane conifers (such as limber pine and Douglas Fir (<i>Pseudotsuga menziesii</i>)) grew at relatively low elevations (Wells, 1983).	Paleoclimatic interpretation similar to Snake Range region, with cold-dry late Wisconsin conditions yielding to early Holocene cool-dry conditions. Middle and late Holocene conditions were relatively warm. Summer rainfall probably low during late Wisconsin and early Holocene, increasing to near-modern levels by 6 thousand yr ago.
		12.6 to 8.9 thousand yr ago	Montane conifers replaced by Utah juniper and Campbell's oak (Thompson and Mead, 1982).	
		6.6 thousand yr ago	First appearance of pinyon pine.	
Lake Bonneville (10)	Pollen	~20 to 15 thousand yr ago	Sagebrush, pine, and spruce important in regional vegetation during period of high lake levels.	Probably reflect relatively cold and dry climatic conditions.
Wasatch Mountains: Snowbird Bog (26)	Pollen	12.3 to 8 thousand yr ago	Coniferous forest restricted, arid sagebrush steppe expanded at high elevations; alpine meadows on moist canyon bottoms. Establishment of coniferous forest at about 8 thousand yr ago (Madsen and Curry, 1979).	Cooler than average temperatures and slightly lower than average precipitation (Madsen and Curry, 1979).
EASTERN MOJAVE DESERT				
Western Grand Canyon: Rampart Cave (19)	Pack rat middens	33.6 to 8.5 thousand yr ago	500-1,500 m xerophytic pinyon-juniper woodlands (Wells, 1983).	Not determined.
		10 to 9 thousand yr ago	Desert shrubs descended to 530 m (Wells and Berger, 1967).	Warm desert conditions.

Table 5-13. Summary of palynological and pack rat midden data from the Great Basin, Mojave Desert, and Sierra Nevada with inferred climatic changes (page 4 of 8)

Location ^a	Evidence	Age	Vegetational assemblage	Inferred climatic conditions
Colorado River Valley: Turtle Mountains (32) Marble Mountains (14)	Pack rat middens	16.9 to 9.9 thousand yr ago	Juniper woodlands descended as low as 258 m, persisting as low as 425 m to at least 9.9 thousand yr ago (Wells and Berger, 1967; Wells 1983).	Indicates late Wisconsin and early Holocene climates of this region cooler and moister than modern conditions. Trend toward increasing aridity, warmth, or both, began by 14 thousand yr ago and accelerated after 11 thousand yr ago.
Newberry Mountains (17) (southernmost Nevada)	Pack rat middens	Late Wisconsin	Descent of pinyon-juniper-oak woodlands to 730-850 m.	Not determined.
Las Vegas Valley: Tule Springs (31)	Pollen	> 30 to modern	Presence of sagebrush and junipers as low as 700 m (Mehringer, 1967). Sagebrush pollen decreased in abundance after 11 thousand yr ago, and Mojave Desert vegetation in place by 7 thousand yr ago.	Wisconsin climate relatively cold and perhaps not much wetter than present, with little summer rainfall. Increase of warmth and aridity before 11 thousand yr ago, but conditions cooler and wetter than modern until about 9 thousand yr ago.
Clark Mountain (3), Sheep (24) and Spring (28) Ranges, Muddy Mountains (16)	Pack rat middens	Late Wisconsin	Presence of bristlecone pine and other subalpine plants as low as 1,500 to 1,700 m; pinyon-juniper woodlands below subalpine assemblage, with pinyon pine (<u>Pinus monophylla</u>) as low as 1,600 m and juniper (<u>Juniperus osteosperma</u>) below 1,900 m. Shadscale up to 320 m above present limits (Spaulding, 1981). Woodland plants at low elevations in Sheep Range until at least 9.4 thousand yr ago.	Late Wisconsin middens indicate higher than modern precipitation levels (Mehringer and Ferguson, 1969).

Table 5-13. Summary of palynological and pack rat midden data from the Great Basin, Mojave Desert, and Sierra Nevada with inferred climatic changes (page 5 of 8)

Location ^a	Evidence	Age	Vegetational assemblage	Inferred climatic conditions
Amargosa Desert (1)	Pack rat middens	Wisconsin to 9 thousand yr ago	Modern Mojave desert scrub preceded by mixture of Great Basin, Mojave, and woodland plants during Wisconsin; assemblages ranged from treeless Great Basin desert scrub to juniper to (rare) pinyon pine. Junipers absent by 9.3 thousand yr ago, and cacti and other succulents became abundant. Arrival of creosote bush (<u>Larrea divaricata</u>) by 9.3 thousand yr ago. Persistence of other plants not present today: Utah agave (<u>Agave Utahensis</u>) and beavertail cactus (<u>Opuntia basilaris</u>).	High abundance of succulents in latest Wisconsin and early Holocene middens may indicate higher than modern summer rainfall levels.
Frenchman Flat (9)	Pack rat middens	40 to 10 thousand yr ago	Abundance of juniper and pine between 1,200 and 1,500 m (Wells and Jorgenson, 1964).	Climate significantly less arid than at present (Wells and Jorgenson, 1964).
		10 to 9 thousand yr ago	Descent of xerophilous juniper woodlands to 1,100 m (600 m below present woodland limit). Coexistence with desert or semidesert shrubs throughout lowered elevation range (Wells and Berger, 1967).	Climate relatively cooler and not much wetter than present during period of 10 to 9 thousand yr ago.

Table 5-13. Summary of palynological and pack rat midden data from the Great Basin, Mojave Desert, and Sierra Nevada with inferred climatic changes (page 6 of 8)

Location ^a	Evidence	Age	Vegetational assemblage	Inferred climatic conditions
Eleana Range (7)	Pack rat middens	17.1 to 13.2 thousand yr ago	Limber-pine (<u>Pinus flexilis</u>) woodland with Great Basin Desert and woodland shrubs as important understory species.	Later glacial trend toward effectively drier conditions, starting at about 16 thousand yr ago and well under way at 13 thousand yr ago. Change to drier climate between 13.2 and 11.7 thousand yr ago.
		11.7 to 10.6 thousand yr ago	Shift to woodland dominated by juniper, pinyon pine, and prickly pear species present in modern nearby woodland. (Spaulding, 1985).	
Specter Range (27)	Pack rat middens	32 to 18.7 thousand yr ago	Middle Wisconsin juniper or juniper-shadscale woodland. Pinyon pine-juniper woodland during Wisconsin maximum. (Spaulding, 1985).	At 30 thousand yr ago, average annual precipitation higher than present, temperature relatively lower.
WESTERN MOJAVE DESERT				
Death Valley (5)	Pack rat middens	19.6 to 13.1 thousand yr ago	Utah juniper grew from 1,200 to 1,500 m below present limits from 19.6 to 13.1 thousand yr ago. Lowest elevations from this period have traces of juniper mixed with Joshua tree (<u>Yucca brevifolia</u>) and Whipple's Yucca (<u>Yucca whipplei</u>). Latter not found at present in Death Valley. Junipers absent from lowest elevations by 11.2 thousand yr ago but persisted below present limits into early Holocene.	Past occurrence of Whipple's Yucca may reflect very mild winter temperatures and substantial winter rain. Junipers receded from lowest elevations by 11.2 thousand yr ago, presaging trend to progressive warmth and aridity from 11.2 to 9.1 thousand yr ago.
		11.2 to 9.1 thousand yr ago		

Table 5-13. Summary of palynological and pack rat midden data from the Great Basin, Mojave Desert, and Sierra Nevada with inferred climatic changes (page 7 of 8)

Location ^a	Evidence	Age	Vegetational assemblage.	Inferred climatic conditions
Eureka Valley, CA (8)	Pack rat middens	14.7 to 5.4 thousand yr ago	Present creosote bush and shadscale (<u>Atriplex confertifolia</u>) area occupied by Utah juniper and limber pine at 14.7 thousand yr ago; also present then were sagebrush, shadscale, and other Great Basin shrubs. Limber pine apparently absent by 10.7 thousand yr ago, while juniper continued to persist well below present limits. Woodland species absent by 8.3 thousand yr ago. Creosote bush and other modern Mojave Desert species began to appear by 5.4 thousand yr ago, and were abundant by 3.9 thousand yr ago.	Changes in plant assemblages apparently reflect a trend toward increasing aridity from 10.7 to 5.4 thousand yr ago.
Far western and southwestern Mojave Desert: Lucerne Valley (13), Lake Searles (23), Scodie Mountains (22)	Pollen, pack rat middens	Late Wisconsin to 7.8 thousand yr ago	Woodland and steppe preceded desertscrub at low elevations during late Wisconsin and early Holocene (Leopold, 1967; Wells and Berger, 1967; Van Devender and Spaulding, 1970). Pinyon-juniper woodland present in Lucerne Valley until at least 11.8 thousand yr ago, while juniper persisted below its modern limits until 7.8 thousand yr ago. Creosote bush and other modern desertscrub species established by 5.9 thousand yr ago.	Late Wisconsin colder than modern conditions; warming and aridification began after 11.8 thousand yr ago and reached modern conditions between 7.8 and 5.9 thousand yr ago.

Table 5-13. Summary of palynological and pack rat midden data from the Great Basin, Mojave Desert, and Sierra Nevada with inferred climatic changes (page 8 of 8)

Location ^a	Evidence	Age	Vegetational assemblage	Inferred climatic conditions
SIERRA NEVADA				
Sierra Nevada: Osgood Swamp (18), Clear Lake (29), Balsam Meadows (2)	Pollen	Late Wisconsin	Sierran montane forests restricted in distribution; replaced in modern time by sagebrush and other steppe species in association with scattered pines and perhaps junipers (Adam, 1967; Davis et al., 1985).	Climate passed from glacial conditions into warmer postglacial conditions about 10 thousand yr ago and cooling within the last 2,900 yr (Adam, 1967).
		10.5 thousand yr ago to present	Steppe replaced by forest by 10.5 to 9.5 thousand yr ago at Osgood Swamp and Swamp Lake, while it may have persisted as late as 7.0 thousand yr ago at Balsam Meadows (Davis et al., 1985). Pine forests were more widespread than today in the Sierra Nevada from 7.0 to 3.0 thousand yr ago (Davis et al., 1985).	Dry from 10 to 7 thousand yr ago and cool and moister from 3 thousand yr ago to present (Davis et al., 1985).
Kings Canyon (12)	Pack rat middens	45 to 12.5 thousand yr ago	Modern oak and chaparral vegetation preceded by more xerophytic and cold-adapted species, including single-needle pinyon pine (<i>Pinus monophylla</i>), western juniper, and ponderosa pine (<i>Pinus ponderosa</i>) (Cole, 1983).	Pleistocene conditions in southern Sierra Nevada were colder and drier than today. Precipitation during the late Wisconsin not much greater than at present (Cole, 1983).

^aNumbers in parentheses following location names correspond to those on Figure 5-24.

Table 5-14. Vegetative cover and inferred climatic regime of the Yucca Mountain region^a

Yr B.P. ^b	Vegetation			Inferred climatic regime						Remarks
	Lower elevations (790-1200 m ^c)	Intermediate elevations (1200-1800 m)	Higher elevations (1800-2100 m)	Temperature (°) (compared to modern)			Precipitation (% change from modern)			
				Winter	Summer	Annual	Winter	Summer	Annual	
5,000	Modern desert scrub	Modern desert scrub	Woodland	-- ^d	--	--	--	--	--	Minor fluctuations
10,000	Amargosa: Utah agave, juniper creosote	Desert scrub expansion	Woodland dominated by juniper, pinyon pine, prickly pear	-1 to -2	+1 to +2	0	0	+50	+10 to +20	Trend toward drier conditions and increasing temperatures
18,000	Specter Range: pinyon-juniper woodland	Sheep Range: transition to subalpine-conifer woodland	Sheep Range: subalpine conifer woodland	-6	-7 to -8	-6 to -7	+60 to +70	-40 to -50	+30 to +40	Global glacial maximum at 18,000 ± 3,000 yr BP
30,000	Specter Range: open juniper-shad scale woodland	Sheep Range: Utah juniper	Sheep Range: juniper woodland with prickly pear, sage, mountain mahogany	--	--	-3 to -6	--	--	+10 to +25	
37,800	No record	No record	No record	--	--	-5	--	--	+20	Cooling trend
38,700	No record	No record	Increase in juniper decline of steppe shrubs	--	--	-1 to -2	+25 to +50	-40	+10 to +20	
45,000	No record	Sheep Range: juniper woodland	Elena Range: open juniper woodland	-2 to -3	--	-1 to -3	+20	-60	0	Seasonality of rainfall different, but annual amount approximately the same

^aSpaulding (1983) and Spaulding et al. (1984).^bB.P. = before present.^cm = meter(s)^d-- = No data.

There is little information on Holocene fluctuations in vegetation across most of this region. It is possible that warmer than modern temperatures or more abundant summer precipitation occurred during some portions of the last 10,000 yr (Antevs, 1948; Martin, 1963; Spaulding et al., 1984; Spaulding and Graumlich, 1986). This will be examined during site characterization as more data are collected within the Great Basin.

Paleoclimatic applications of paleobotanical data

The interpretation of palynological and pack rat midden records in climatic terms requires information on the modern relationships between vegetation and climate. Given such relationships, under certain assumptions it is possible to infer the nature of the climatic variations responsible for the vegetation changes recorded by the fossil evidence. The general assumption underlying the drawing of these inferences is "that past temporal changes in the abundances of a set of pollen types reflect climatic changes equivalent to those responsible for spatial variation in the abundances of the same taxa on the contemporary landscape" (Howe and Webb, 1983). This general assumption can be expanded to more-specific ecological and statistical assumptions.

The specific ecological assumptions are as follows (Bartlein et al., 1984):

1. The modern vegetation data is in equilibrium with the modern climate at the temporal and spatial scales of interest. Given this assumption, the modern vegetation will adequately reflect the modern climate.
2. The variations in the paleoecological record being interpreted are ultimately attributable to climatic variations.

It is not possible to test either assumption in practice, but by proper selection of the temporal and spatial scales of interest, violations of the assumptions can be minimized (Prentice, 1983). The scales of interest for the Yucca Mountain repository will imply an assumption that the modern vegetation-climate relationships can be extrapolated back through Quaternary time over the Great Basin.

The specific statistical assumptions are (1) the statistical model used is specified correctly and (2) the parameters of that model are estimated in an appropriate fashion.

Two data bases are thus required for paleoclimatic reconstruction: (1) a data base of modern combined vegetation and climatic data and (2) a data base of fossil vegetation data. The first data base is used to construct baseline relationships between vegetation and climate to determine, for example, the environmental requirements of individual taxa. The relationships are then applied to the fossil data, either to interpret them in climatic terms, or to permit comparison of the observed fossil record with that simulated by differing paleoclimatic scenarios.

Relationships between modern vegetation and climate are also required for the paleobotanic validation of climate simulation models (Webb, 1980).

To judge the ability of a model to correctly simulate the past climate, modern vegetation-climate relationships are used to transform model simulations of past climates into estimates of past vegetation, which can then be compared with the observed fossil record (Webb, 1980). The collection and the analysis of both modern and fossil vegetation data therefore contribute to both climatic reconstruction and the validation of climate models that may be applied for predicting future climates.

It is possible to construct relationships between modern climate and vegetation in both a qualitative and a quantitative fashion. The following sections discuss these two methods. Both methods rest on the two basic ecological assumptions listed previously.

Qualitative relationships. Thus far, paleoclimatic estimates from vegetation data from the arid West have been based on qualitative comparisons of modern plant distributions with climatic parameters. These methods involved

1. Comparisons of maps of species distributions with maps of climatic parameters. Past changes in the species distribution are assumed to reflect changes in the subjectively correlated climatic variable. Modern plant distributions are poorly mapped in the West, limiting the applicability of this method. Additionally, without means of assessing whether there is a mechanism linking the distribution of a given species with a certain climatic parameter, these mapped comparisons may be misleading.
2. Identifying analog sites that have modern vegetation similar to the fossil vegetation assemblage. The modern climate of the analog site is then assumed to be representative of the climate at the time that the fossil assemblage was deposited (Spaulding, 1985). The utility of this method is limited in that a wide array of modern climatic sites may be suitable analogs, depending upon the criteria for matching the fossil and modern vegetation assemblages.
3. Calculating the amount of elevational depression of a given plant species relative to its modern limits and using the modern environmental lapse rates to estimate the amount of climatic difference represented by this depression (Spaulding, 1981). This method is difficult to apply in that different species experienced widely varying amounts of elevational depression in the past. More importantly, as discussed in the previous section on lake levels, it is very difficult to independently estimate past lapse rates.

Quantitative approaches. There are two related approaches for constructing statistical relationships between modern vegetation and climate data: (1) a transfer function approach (Webb and Bryson, 1972; Bartlein et al., 1984), in which individual climate variables are expressed as a function several vegetation predictor variables (e.g., the percentages of different pollen types) and (2) a response function approach (Bartlein et al., 1985), in which the relative abundance or presence/absence of individual taxa are expressed as nonlinear functions of one or more environmental variables. In addition, forest simulation models have been used to generate vegetational assemblages consistent with a particular set of values of controlling climate

variables (Solomon and Webb, 1985). Such models incorporate the development of a response function for vegetation-climate relations.

The transfer function approach usually makes use of a multiple regression analysis to construct a relationship between a particular climate variable and a number of predictor variables--generally, the percentages of individual pollen types (Webb, 1980; Bartlein et al., 1984). In practice, a data set of paired observations of modern surface pollen samples and climate is required for calibration of the multiple regression equation. The resulting equation is then applied to fossil pollen data to interpret them in climatic terms.

Several statistical assumptions (in addition to the basic assumptions described earlier) underlie the application of regression analysis; these are described by Howe and Webb (1983) and Bartlein and Webb (1985). Comparisons among some of the different approaches to constructing the statistical relationships have been described by Kay and Andrews (1983). Strategies for identifying and minimizing the sources of uncertainty in the paleoclimatic estimates are given in Bartlein et al. (1984) and Bartlein and Webb (1985). Bartlein and Webb (1985) show how to determine the applicability of the methods in time and space.

In the response function approach, the relative abundance or probability of occurrence of different taxa are expressed as a nonlinear (usually polynomial) function of one or more environmental variables. The resulting functions are usually displayed as response surfaces that show how the abundance or probabilities vary in the space defined by the environmental variables (Bartlein et al., 1985). For the relative abundance data, the response functions are fit using linear regression (Bartlein et al., 1985), while for the presence or absence data, the surfaces are fit using logistic regression (Ter Braak and Barendregt, 1986).

Because response functions illustrate the environmental preferences of different taxa, they can provide guidance in the qualitative interpretation of fossil data in climatic terms. Quantitative reconstruction is possible as well, using response functions of several taxa to determine the environmental conditions necessary to give rise to a particular fossil assemblage (Ter Braak and Barendregt, 1986).

Applications of response functions for climate model validation are relatively straightforward. The relative abundance of different taxa can be estimated by applying response functions to the values of climate variables simulated by climate models. The simulated vegetation variables can then be compared with the observed values. In a typical example, the climate model would be set up to simulate some specific paleoclimate, and the simulated vegetation would be compared with the observed fossil record (Webb et al., 1985).

5.2.1.2.4 Glacial and periglacial records and their relation to lake levels

Isotopic records from deep-sea cores indicate that there have been at least seven high-latitude glacial and seven interglacial periods during the

last 730,000 yr (Hays et al., 1976). On land, successive glaciations have obliterated much of the evidence of earlier ice advances and, therefore, the record is fragmentary and heavily weighted toward the last glaciation. Although alpine glaciers did form nearby in the higher portions of the Sierra Nevada and in other Great Basin mountain ranges, there is no evidence suggesting that glacial activity ever occurred on Yucca Mountain. Nivation basins, and other periglacial features of assumed late Wisconsin age have been identified in the mountains near the NTS as far south as the Spring Range (Blackwelder, 1931; Sharp, 1938; Porter et al., 1983; Dohrenwend, 1984). The chronology of the formation of periglacial features is unknown.

Deposits from at least four glaciations of Pleistocene age have been found in the Sierra Nevada, with the last assumed to be correlative with the late Wisconsin in northeastern North America (Porter et al., 1983). Lake deposits impounded behind a natural dam composed of outwash from Sierran glaciers at Tulare Lake, California (Atwater et al., 1986) suggest that this latest Pleistocene glaciation began 26,000 yr ago and ended between 13,000 and 10,000 yr ago. The age of the maximal extent of the Tioga advance is unknown, although Porter et al. (1983) suggested that it was correlative with the last high lake stand at Lake Russell (Mono Lake). The Lake Russell maximum has been estimated to have occurred 14,000 to 13,000 yr ago (Figure 5-21). A radiocarbon date of 10,000 yr ago from lake sediments impounded behind a Tioga-age terminal moraine at Osgood Swamp (Adam, 1967), indicates that deglaciation occurred before this time. Mountain glaciers in eastern Nevada appear to have disappeared by about 13,000 yr ago (Wayne, 1984). A similar age of deglaciation is indicated for the central Wasatch Range of Utah (Madsen and Currey, 1979).

The stratigraphic and radiometric data on lake-glacial correlations are ambiguous, although the weight of the evidence points to glacial recessions occurring well before the Pleistocene lakes reached their highest levels (Benson and Thompson, 1987). The lag times between these phenomena cannot yet be assessed nor can the differences in climate (if any) between the "full-glacial" and "full-lake" phases be deduced.

Over the last 10,000 yr (Holocene time), mountain glaciers in the Sierra Nevada advanced at least four times (Burke and Birkeland, 1983). All were of much smaller magnitude than the Pleistocene glaciers. The oldest of these Holocene glaciations is thought to have occurred between 10,000 and 9,000 yr ago, while the other three advances appear to postdate 4,000 yr before present (Burke and Birkeland, 1983).

5.2.1.2.5 Regional paleoclimatic hypotheses

The various sources of paleoclimatological proxy data reviewed in the preceding sections have been used to formulate hypotheses concerning the nature of past climatic variations in the west. Theories of the nature of Wisconsin-age climates implied by the various data sets are wide-ranging. Galloway (1970), Brakenridge (1978), and Dohrenwend (1984) have argued that the past distributions of cryogenic features imply that full glacial mean annual temperatures in the southwest were from 7 to 11 Celsius degrees colder than those of today, with little or no increase in mean annual precipitation.

Others, such as Snyder and Langbein (1962) and Mifflin and Wheat (1979), have suggested that lake-level reconstructions provide evidence for increased rainfall with a more moderate decrease in mean annual temperatures. Benson (1981) modeled the lake budget for Pleistocene Lake Lahontan in western Nevada and concluded that a reduction in mean annual temperature, coupled with an increase in sky cover, were needed to create and maintain these lakes. Smith and Street-Perrott (1983) calculated streamflow volumes that would offset the estimated evaporation from the maximum cumulative areas of late-Pleistocene lakes in the Owens River system ($1,963 \text{ km}^2$), and found inflow volumes to be in the range of 3.5 to 6.5 times present volumes, depending on whether a 5 or 10 Celsius degree lowering of temperatures was assumed. Van Devender and Spaulding (1979) interpreted pack rat midden data as reflecting late-Wisconsin climates characterized by increased precipitation, cool summers, and mild winters. Spaulding (1985) presents evidence that the pack rat midden data from southern Nevada indicates temperatures about 7 Celsius degrees below modern mean annual temperature and more abundant winter rainfall. Wells (1979) interpreted pack rat midden from the late Wisconsin as reflecting increased summer rainfall, with little reduction in mean annual temperature.

Opinions on Holocene climates are equally diverse. Antevs (1948), largely on the basis of studies in the northern Great Basin, argued that the middle Holocene was characterized by hot and dry conditions across the entire western United States. Martin (1963) hypothesized that this was instead a period of enhanced monsoonal summer precipitation throughout much of the southwest. In the NTS region, various studies have supported both of these seemingly conflicting hypotheses. Before construction of the Los Angeles Aqueduct, Owens Lake, west of the Yucca Mountain site, was the only perennial lake supported by the Owens River. Though reduced to a body of water only 10 m deep in the late 1800s and supported by a small fraction of maximum pluvial inflow, it apparently never desiccated to form a bed of salts during Holocene time or any other part of the last million or more years (Smith and Pratt, 1957), implying that no period in the Pleistocene has been significantly more arid than the present. To the north, lakes in the Great Basin were at lower than modern levels during the middle Holocene (Davis et al., 1976; Benson, 1978; Currey, 1980; Thompson, 1984) and upper treelines advanced upslope (LaMarche and Mooney, 1972). Both of these phenomena may indicate relatively high summer temperatures. During the same period in this region, there were changes in plant communities that may indicate higher than modern levels of summer precipitation (Thompson, 1984; Spaulding and Graumlich, 1986).

What is clear from this collection of differing paleoclimatic hypotheses is that the paleoenvironmental data, as they now exist, are inadequate to provide critical tests. Plans to develop a more adequate data base are discussed in Section 8.3.1.5.1.

5.2.2 FUTURE CLIMATIC VARIATION

The paleoenvironmental record for the Quaternary reveals that significant climatic variations have occurred in the past, and similar variations

can be expected in the future. It seems clear, for example, that the repeated fluctuations between glacial and interglacial climates during the Quaternary have occurred in response to changes in the seasonal and latitudinal receipt of solar radiation as determined by variations of the earth's orbital elements (Hays et al., 1976). Since the orbital variations will continue in the future, it is very likely that the present interglacial climate will eventually give way to another glacial one.

Deciding whether such variations can significantly affect a repository will require a prediction of the times and intensities at which they will occur in the future. The prediction need not be exact; the decision will rest largely on calculations that, for conservatism, will be based on the most extreme ranges that climatic variations can reasonably be expected to exhibit in the future. The paleoclimatic record can supply the needed data on the range of past variations. Nevertheless, the ranges of past variations can be used with greatly increased confidence if the future variations can actually be predicted from an understanding of the causes of climatic variations as well as the record of their past occurrence.

Study of past climatic variations can provide only the general outlines of future variations on the time scale of 1,000 to 100,000 yr. The orbital variations can explain a considerable portion of the total variance of such large-scale climatic variables as global ice volume, but some variance still remains unexplained (Imbrie, 1985). Further, as will be described below, the link between global-scale variations and local ones is not clear. Moreover, for the near future (the next 10 to 1,000 yr) the large inputs of anthropogenic carbon dioxide and other trace gases into the climate system have the potential of greatly altering future climate. The potential effects of carbon dioxide and other trace gases have recently been reviewed by MacCracken and Luther (1985) and Wang et al. (1986), respectively. Specific forecasts of future climatic variations at a particular location should consider the effect of aerosols, anthropogenic carbon dioxide, and other trace gases, as well as the geologic record.

Methods of predicting climatic variations on the time scale of 1,000 to 100,000 yr do exist. In the following sections, the nature and causes of climatic variations that must be considered in furthering the development of methods for predicting future climates are reviewed, followed by a description of the necessary procedures for predicting future climates.

5.2.2.1 Components of the climate system

The major components of the climate system are the atmosphere, hydrosphere, cryosphere, surface lithosphere, and biosphere; these are linked by flows of mass and energy. The major components may be subdivided further. The hydrosphere, for example, can be subdivided into the oceans, atmospheric vapor, and continental water, while the cryosphere consists of sea ice as well as land ice and snow. The ocean can be subdivided further into the mixed layer and deeper water. Individual components have different thermal characteristics and response times. The atmosphere varies relatively rapidly, the mixed layer of the oceans less rapidly, and the deep ocean and continental ice sheets extremely slowly (Saltzman, 1985).

The state of the climate system at any given time can be described by a large and varied set of climate variables that are observed over temporal scales ranging from 10 yr (in the case of annual values of standard meteorological variables) to 10 million yr (the position of the continents), and over spatial scales from local to global. Individual variables can be classified (Saltzman, 1985) as those that describe (1) the boundary conditions or external controls; (2) the slowly varying components, such as the ice sheets and deep oceans; and (3) the fast-response components, such as the atmosphere.

Boundary conditions include those variables that influence the climate system, but in turn are not influenced by it (Imbrie, 1985). The boundary conditions thus include the incoming solar radiation, the position and size of the continents, and inputs of dust and aerosols from volcanic eruptions, and the inputs of dust and carbon dioxide from human activities.

Depending on the particular temporal and spatial scale of the climatic variation under consideration, additional components of the climate system can be added to the list of boundary conditions. At the scale of 10,000 to 1 million yr, the size (volume, area, and elevation) of the ice sheets and the temperature of the oceans are internal variables of the climate system that respond to the orbital variations. At shorter time scales (i.e., shorter than 10,000 yr), the ice sheets and ocean may be regarded as external controls because their response times are 10,000 yr or longer. At short time scales, the preindustrial concentration of carbon dioxide in the atmosphere and the loading of dust from wind erosion in areas of naturally sparse vegetation can be regarded as external boundary conditions. At longer time scales these variables are more properly regarded as internal variables.

Slowly varying components of the climate system include the ice sheets, ice shelves, tectonic activity, and deep layers of the ocean (Saltzman, 1985). These slowly varying components at a particular time serve as additional boundary condition controls of the fast-response components. The fast-response components include the atmosphere, the mixed-surface layer of the ocean, seasonal snow cover, and the land-surface temperature. These components generate the variations of climate at a particular location on time scales shorter than 1,000 yr.

Since the mid-1800s, the concentration of carbon dioxide in the atmosphere has increased as a result of the burning of fossil fuels and large-scale deforestation. The nature of this increase and its potential climatic effects have been reviewed recently in a series of reports published by the DOE (MacCracken and Luther, 1985). Because this increase in carbon dioxide concentration may be of sufficient magnitude to alter the way in which the climate system responds to changes in other boundary conditions, the effects of this increase may be significant in the prediction of future climates.

5.2.2.2 Climatic variations

It is possible to define three kinds of climatic variations (Imbrie, 1985): (1) forced variations produced by changes in the external controls, (2) free variations generated internally within the climate system, and

(3) resonant interactions that are a combination of the two. Climatic variations thus have two ultimate sources: the external controls of the climate system (or boundary conditions) and the climate system itself.

The specific response of the climate system to changes in its external controls (forced variations) depends on both the nature of the variations in the controls and the state of the climate system itself (Imbrie, 1985). For example, a constant change in solar radiation inputs will likely produce different responses depending on the presence or absence of continental ice sheets (Kutzbach and Guetter, 1984).

Free variations of the climate system can arise from the existence of many pathways for the flow of mass and energy within the climate system, and hence from the existence of many feedback mechanisms. It is not unlikely that significant variations in global climate could be generated by such feedback even under fixed boundary conditions. For example, Saltzman and Sutera (1984) described a climate model that considers the continental ice mass, the marine ice mass, and temperature of the ocean. With no external forcing, the model generated a variation of the ice and ocean with many characteristics of the geologic record, for instance, a dominant period of variation of about 100,000 yr. Such free oscillations of the climate system have been obtained using a number of different climate models (Saltzman, 1985). However, the phase relationships between the continental ice mass and ocean temperature in the model were not consistent with those observable in the real climate system. When the model was forced with realistic variations of solar radiation (such as those produced by the orbital variations), the correct relationships among the variables were obtained (Saltzman et al., 1984). This latter result provides an example of climatic variations produced by resonant interactions between forced and free variations of the climate system represented by the model.

The existence of temporal and spatial hierarchies in both the components of the climate system and also the sources of the climatic variations makes it difficult to assign past climatic change at a particular location to a specific ultimate cause, and consequently difficult to predict future changes. For example, variations of the fast-response components may be generated by variations of the slowly varying components (and by internal variations), and in turn, variations of the slowly varying components may be generated by variations of the boundary conditions (and by internal variations).

Similarly, local climate variations of the kind, for example, that make one winter different from the last, are embedded in hemispheric-scale circulation anomalies (Namias, 1975). Moreover, local variations are indeterminate with respect to larger scale variations. The climate of a particular location has two components: (1) a locational component, governed mainly by the latitude of the place and its position on the continent, which determines the mean values of temperature and precipitation and (2) an advective component, governed by the location of the place relative to the traveling weather systems, which determines the day-to-day variations of the weather. Thus, a variety of atmospheric circulation patterns could be envisaged that would lead to the same local climate variations.

Each level of the temporal and spatial hierarchy contributes additional uncertainty about the ultimate causes of past climate variations at a particular place, and the likely consequences of future variations in the controls of the climate system. The development of climate prediction methods for the time scales of interest must therefore proceed in a top-down fashion, from the larger-spatial, longer-temporal scale controls to the local responses to variations of those controls. Such a hierarchical methodology has not yet been developed.

5.2.2.3 Climate prediction methods

Prediction of future climatic variations on the time scale of 1,000 to 100,000 yr will require a multistep procedure that corresponds to the hierarchy of controls and responses in the climate system, and incorporates studies to validate model simulations. Climate prediction over this time scale can not be said to be operational, however. Thus, an important element of any prediction effort will be the selection and evaluation of the individual models.

As outlined previously, an overall approach to climate prediction would proceed from predictions of variations in the external controls to predictions of the slowly varying components, and then to the fast-response components, and from the global to the local scale. Steps in such a prediction approach include

1. Prediction of boundary conditions.
2. Prediction of the slow-response components.
3. Prediction of the fast-response components.
4. Disaggregation of the predictions of the slow- and fast-response components at the appropriate spatial scale.
5. Model validation studies.

Climate simulation models are typically classified into two categories: (1) statistical-dynamical models (Sellers, 1983) and (2) explicit-dynamical models (Kutzbach, 1985; and references therein). Models in the first category attempt to simulate the long-term, spatially averaged variations of the climate, while those in the second category attempt to simulate day-to-day variations of the weather. Other classifications of climate models are possible (Schneider and Dickinson, 1974; Meehl, 1984).

5.2.2.3.1 Prediction of boundary conditions

The key boundary condition for predicting future climatic variations over the next 100,000 yr is the latitudinal and seasonal distributions of solar radiation. Fortunately, these quantities are predictable over the required time interval. The concentration of carbon dioxide and the dust loading of the atmosphere are also important boundary conditions. The past record of these variables, and their potential predictability over the next

100,000 yr is not well known. Past estimates of volcanic loadings might be made stochastically, based on past summaries of volcanic-loading estimates such as those given by Hirschboeck (1980). The prediction of future loadings from natural (e.g., volcanos, wind erosion) and human-induced (desertification) sources is not likely to be easily implemented. Similarly, the source of past variations of the concentration of carbon dioxide in the atmosphere is not clear at present (Shackleton et al., 1983), and so this variable too may be inherently unpredictable, although it is clear that the concentration of carbon dioxide due to human activities will continue to increase at least into the next century.

While the specific course of future variations of climatic variables may not be predictable, the range of their past variations as deduced from the paleoclimatic record can serve to limit the range of variations expectable in the future.

5.2.2.3.2 Prediction of slowly varying components

The prediction of the slowly varying components of the climate system, such as the global volume of glacial ice and deep ocean temperature, will require the application of models from the statistical-dynamical family of climate models. Saltzman (1985) recognizes two groups of models of the slowly varying components: (1) a quasi-deductive group of models that derives models of, for example, the ice sheets from basic physical principles and (2) an inductive group of models that derives models consistent with observed paleoclimatic records that are required to be only physically reasonable. In other words, the first group proceeds from physical principles to a model of observed paleoclimatic variations, while the second starts with the geologic record and derives a physically reasonable model based on it. The distinction between the two groups arises from the inability of purely physically based models to represent all the facets of the records of the slowly varying components, hence the need for an alternative approach.

Recently, paleoclimatologists have recognized the importance of periodic variations of the earth's orbital motion and of insolation as important controls of the climate during the Quaternary, while admitting the potential for internal variations of the climate system to contribute to overall variations of climate (Saltzman, 1985). The quasi-deductive group of models is represented by a series of models that seek to describe the response of the ice sheets, ice shelves, sea ice, and the deformable bedrock (Oerlemans 1982; Pollard, 1983) to variations in solar radiation.

The inductive group of models is represented by a series of models that attempt to reproduce the characteristics of the observed paleoclimatic record using a geologically consistent, but physically reasonable model (Imbrie and Imbrie, 1980; Saltzman and Sutera, 1984; Saltzman et al., 1984; Imbrie, 1985). The Imbrie and Imbrie (1980) model is often cited as an example of a future climate prediction for its extrapolation of the global ice volume series to 100,000 yr into the future. The intent of the authors cited above in displaying this information clearly seems to have been to illustrate the nature of the temporal response of the output of their model rather than to provide an experimental prediction.

Estimates of ice sheet geometry can be produced with numerical computer models based upon a solution of the two-dimensional, time varying, ice flow law. Examples of such models are now in existence. One has been applied to the reconstruction of the dynamics of the Laurentide ice sheet for the last 100,000 yr (Budd and Smith, 1981) and was based on orbitally induced variations of insolation over the ice sheet. Other reconstructions of continental-scale ice sheets for times during the Pleistocene have been produced by Mahaffey (1976) and Sugden (1977).

The information about glacier configurations at various times have been used as input to model studies of the general circulation of the atmosphere. Such studies have been done for reconstructions of past climates using one-dimensional (flow-line) models of ice dynamics (Hughes et al., 1981; Stuiver et al., 1981). The global ice sheet reconstructions of Hughes et al. (1981) were used to model the atmospheric circulation during the time of the last glaciation and more recent periods in the past 18,000 yr. Other atmospheric general circulation model studies of global climate during an ice age have been provided by Williams et al. (1974), Gates (1976a,b), and Manabe and Hahn (1977).

There are several areas in which further development of models of the slowly varying components of the climate system is required. Most of the models thus far have focused on the ice sheets, and models of the deep ocean are relatively primitive (Saltzman, 1985). The models simulate global climate systems; methods must be sought for disaggregating this output to provide, for example, maps of the detailed spatial patterns of the ice sheets.

5.2.2.3.3 Prediction of fast-response components

Models of the fast-response components of the climate system can be drawn mainly from the explicit-dynamical family of climate models and consist mainly of general circulation models. Atmospheric general circulation models attempt to represent the three-dimensional structure of the atmosphere and to simulate the day-to-day variations of the weather. This is done by solving the basic equations governing the motion of the atmosphere over a three-dimensional grid (Schneider and Dickinson, 1974; Meehl, 1984). In practice, general circulation models provide snapshot views of the atmospheric components (wind, temperature, pressure, precipitation) of the climate system under a fixed, or seasonally varying set of boundary conditions. The potential of these models in climate prediction over the time scale of 100,000 yr lies in their ability to disaggregate global and hemispheric scale variations of the boundary conditions and slowly varying components of the climate system.

Applications of general circulation models for simulating past climatic variations include those by Gates (1976b), Manabe and Hahn (1977), Manabe and Broccoli (1985), and Kutzbach and Guetter (1986). In these applications, the state of the boundary conditions and slowly varying components for ice age conditions were used to initialize the models. General circulation models have also been used to investigate the response of the climate system to anthropogenic increases in the concentration of carbon dioxide in the atmosphere. An extensive review of these models is given by Schlesinger (1984).

An important element of general circulation models in need of further development is the manner in which the oceans are treated. In most paleoclimatic applications, ocean surface temperatures were prescribed, fixed at the CLIMAP Project Members (1981) values for 18,000 yr ago, for example. In Manabe and Broccoli (1985) a mixed-layer ocean was used, but such a model still does not represent heat transport by the oceans (Manabe and Wetherald, 1985). Even with increasing computer resources, fully coupled ocean-atmosphere general circulation models are still too inefficient for routine application (Meehl, 1984).

5.2.2.3.4 Spatial disaggregation of predictions

Although general circulation models are capable of providing three-dimensional simulations of the atmosphere, their spatial resolution is too coarse to provide meaningful results for specific sites (Gates, 1985). The National Center for Atmospheric Research Community Climate Model used by Kutzbach and Guetter (1984), for example, has a spatial resolution of 4.4° latitude by 7.5° longitude. Topographic features that may be quite important mediators of the local climate are thus greatly smoothed.

There are two distinct approaches that may be applicable for disaggregating the simulations produced by general circulation models: (1) an empirical approach based on statistical analysis of observed climate data and (2) a modeling approach using mesoscale meteorological models.

Two different strategies have been proposed for statistically based disaggregation of general circulation model output. Kim et al. (1984) described an approach based on a principal components analysis of time series of climate data observed over a dense network (relative to the resolution of a general circulation model). Often, the first several components will represent the general anomalies over a network. Correlations between those components and model output at a specific grid point could then be sought, and the spatial structure of the components could then be used to disaggregate the grid-point output.

The second statistically based approach involves the construction of regression equations that relate the long-term averages of individual climate variables observed over a calibration network to a set of predictor variables that describe the location and elevation of individual stations in the network (Craig, 1984). Location is described, for example, in terms of the distance from the sea coast, and the nature of upwind topographic barriers. In practice, the equations could be applied to locations within the calibration network where climate station data are not available to (1) interpolate the observed climate values or (2) to estimate values of the climate variables consistent with the general circulation models output using wind patterns as simulated by a general circulation model.

While computationally straightforward, the empirical approaches have several drawbacks. The empirical models do not represent the physical processes that generate local variations of climate in any explicit fashion. For example, they do not explicitly represent the control for the occurrence and abundance of precipitation by the large-scale circulation. The empirical

basis, and lack of explicit physics, may limit the usefulness of the statistical models when extrapolation is required, as it is for both paleoclimatic simulations or future climate predictions.

The second model-based approach for the spatial disaggregation of general circulation model simulations makes use of mesoscale meteorological models (Anthes, 1983). Mesoscale meteorological models like general circulation models attempt to explicitly represent the physical processes that govern atmospheric circulation and the surface energy and hydrologic balances. The spatial resolution of regional mesoscale models is generally much finer than that of general circulation models and the representation of the surface energy and hydrologic balances is usually more detailed, as is the representation of topography. In practice, mesoscale meteorological models could be used to spatially disaggregate the output of a general circulation model. This could be accomplished by using the general circulation models output to initialize the mesoscale meteorological model, in much the same way that widely distributed meteorological observations are used to initialize standard numerical weather prediction models. The greater topographic detail, and more realistic representation of the surface processes in the mesoscale meteorological models, permit the simulation of regional and local-scale snapshots of climate variables consistent with the larger-scale patterns of those variables simulated by a general circulation model. One limitation of mesoscale meteorological models, however, is that they use the broadly defined boundary conditions of the general circulation model. If there are assumptions and approximations made in developing these boundary conditions, these assumptions are transferred to the mesoscale meteorological models.

Like general circulation models, mesoscale meteorological models require considerable computing resources. This disadvantage is somewhat outweighed, however, by the physical basis of the models, which should make them more robust than the empirically based models when applied in the context of either paleoclimatic reconstruction or climatic prediction.

In making a choice between using a general circulation model-mesoscale meteorological model approach versus an empirically based approach, these factors as well as others (e.g., amount of data available and requirements made by the hydrologic investigators) need to be considered. This consideration is discussed briefly in Section 8.3.1.5.1 and will be a continuing process throughout site characterization.

5.2.2.3.5 Model validation

An important step in the overall modeling effort will be to measure the performance of the model in order to estimate the uncertainties inherent in the eventual climate predictions. The standard criterion used is how well a model simulates the modern climate. While such measures are useful, the danger always exists that a model may be tuned to perform well with respect to the modern climate.

In the hierarchy of climate simulation methods just described, there are two important sources of uncertainty. The first arises in the simulation, on

the basis of boundary conditions, of the slowly varying components, and in turn, simulations of the fast-response components of the climate system. The second source of uncertainty arises in the spatial disaggregation of the fast-response components. These uncertainties may be posed as two questions. As the boundary conditions and slowly varying components evolve through time, will the snapshots simulated by the models of the fast-response components adequately track the actual climate? At a specific time, will the spatial patterns of climate be adequately portrayed by the approach used to disaggregate the simulations of the fast-response components? The paleoclimatic record, containing as it does the output of the ideal climate model, can provide the answers to these questions.

Two model validation exercises can be envisaged, one involving long time series of paleoclimatic data at key locations, and the second using dense networks of data at key times. The first exercise would test the ability of the model to track the actual climate as the boundary conditions and slowly varying components change over the next 100,000 yr. Ideally, continuous paleoclimatic records from key locations extending back to the previous interglacial would be used in this exercise. The second exercise would test the ability of the model to simulate correctly the spatial patterns of climatic variations at an appropriate scale. For this exercise, paleoclimatic evidence will be collected to describe spatial patterns of climatic variations at such key times as the full-glacial, late-glacial, early Holocene, etc. For both exercises, methods such as those described in Section 5.2.1.2.2 would be used to transform the model output into simulations of the various paleoclimatic indicators that would then be compared with the observed fossil record. Plans for investigations to develop and validate climate models to satisfy the needs just discussed are specified in Section 8.3.1.5.1.

The individual elements of this procedure for predicting future climates follow the hierarchical nature of climatic variations in a particular region--from global-scale variations of boundary conditions and the slowly varying components of the climate system, to regional-scale variations of the fast-response components. While individual climate simulation models applicable to this approach exist, no comprehensive attempt at predicting future climatic variations has yet been made. For this reason, model validation experiments must be carried out to judge the reliability of the predictions. In addition to providing insight into the nature and extent of past climatic variations, the paleoclimatic record provides the information required for the climate model validation exercises. The use of paleoclimatic data for these purposes involves three broad tasks:

1. A data collection task in which the necessary paleoclimatic data, both geologic and biologic, are collected over appropriate networks where possible (i.e., long records at key locations and dense networks at key times).
2. A modeling task in which relationships are developed to link the various climatic indicators to their climatic controls (e.g., lake hydrologic models (Benson, 1986) and modern vegetation-climate relationships (Bartlein et al., 1985)).
3. An interpretation task in which (1) the geologic or paleoecologic record is interpreted in climatic terms, (2) the nature of past

climatic variations is inferred, and (3) the observed paleoclimatic record is compared with that generated by climate-simulation models.

Successful completion of these three tasks is required to document the range of climate variations experienced in the past, to simulate probable future variations, and to provide input for estimating the impact of the effect of those variations on the surface and ground-water systems in the vicinity of Yucca Mountain.

5.2.3 SITE PALEOCLIMATIC INVESTIGATIONS

Historical and prehistorical climatic, limnologic, hydrologic, and vegetational data obtained from near Yucca Mountain and the surrounding region will be used to increase and supplement the paleoclimatic data base used for characterization of the Yucca Mountain site. Detailed plans designed to obtain these data are presented in Section 8.3.1.5.1 for the following activities.

5.2.3.1 Synoptic characterization of regional climate

Historical meteorological data from the Great Basin will be compiled and summarized to provide synoptic-scale information. These data will be used in calibration activities for vegetation-climate and lake-climate models, as well as in the development of a regional climate model. Information on the stable isotope "signatures" from dated ground water and from modern precipitation networks will be obtained from the geohydrology (Section 8.3.1.2) and meteorology (Section 8.3.1.12) programs used in this activity.

5.2.3.2 Collection and analysis of pollen and pack rat midden macrofossil assemblages from the Great Basin

Recent pollen and vegetational assemblages are being collected and mathematically correlated with climatological data (temperature, amount of precipitation, etc.) using transfer functions and response surface techniques. Pollen and macrofossil suites from lake sediments and pack rat middens are being collected and isotopically dated. The transfer and response surface will be used to extract quantitative estimates of climate variation from the vegetational time series data.

5.2.3.3 Coring and paleontological, sedimentological, chemical, and chronological analyses of lacustrine and marsh deposits

Sediment cores will be obtained from playas, paleomarshes, and lakes near Yucca Mountain and in the surrounding region. The paleontological and sedimentological changes recorded in these cores will provide dated time

series of (1) changes in size, chemistry, and productivity of lakes and (2) wet phases in now-dry playas and marshes.

5.2.3.4 Development, validation, and application of a generalized model of late-Quaternary climatic variations to predict future climate change at the Yucca Mountain site

A generalized model of late-Quaternary climatic variations in the southern Great Basin will be developed, upon which estimates of the range of future climatic conditions can be based. This generalized model has two elements: (1) the reconstruction of past climatic variations to illustrate the range of potential regional climatic variations, and (2) the development of climate simulation models to estimate the response of the regional climate to future changes in the boundary conditions. Reconstruction of past climatic variations will be based on the paleoecological record (pollen and plant macrofossil), supplemented by paleohydrologic, sedimentologic, and geomorphic data sets. Climate simulation models will be validated with the same data sets to measure the ability of the models to correctly simulate past climatic variations, thereby gauging their potential for simulating future variations. These activities will require data on vegetative cover, lake size, lake temperature, glacier size, stream discharge, and infiltration rate. The validation exercises will demonstrate the degree to which the model can simulate climate change without recourse to arbitrary parametric adjustment.

5.3 SUMMARY

The data discussed thus far in this chapter have established the existing climatic information and discussed why data are needed and identified areas where data are lacking or inadequate. This section provides a synopsis of these points and relates them to performance objectives, conceptual models and bounding conditions, the need for supplemental data, and the quality of the presently available data.

5.3.1 SUMMARY OF SIGNIFICANT RESULTS

The existing climate in the vicinity of Yucca Mountain site, as defined by meteorological data available from sites around Yucca Mountain, is classified as midlatitude desert. Variations of a given meteorological parameter, however, are strongly influenced by elevation, giving rise to wide ranges for parameters within the climatological classification. The most notable general meteorological conditions expected at Yucca Mountain are temperature extremes, particularly during summer months, approaching 50°C; temperature ranges between maximum and minimum of nearly 25 Celsius degrees; and low annual precipitation amounts, less than about 6 in. (150 mm) annually.

Skies are predominantly clear throughout the year, and the average relative humidity is low. Precipitation in the area is limited by the blocking and subsequent rain shadow effect of the Sierra Nevada during winter months,

although monthly average precipitation reaches its annual peak during January or February. Precipitation is also generated in the area by summer thunderstorm activity, representing a secondary monthly average peak in July or August. Despite the existence of this well-defined annual cycle, precipitation averages only about 150 mm/yr. Winds from the north dominate during fall, winter, and into early spring, but shift to predominately south to southwesterly in late spring and early summer. Terrain influences somewhat disrupt this annual average cycle, with upgradient winds occurring during daylight hours in almost all months.

Meteorological data are also needed as input to infiltration studies and rainfall-runoff modeling. The existing data set is not site specific enough for these investigations, and does not provide detailed information on precipitation intensity, temporal, or areal variability that is necessary for hydrologic investigations (see Chapter 3 and Section 8.3.1.2 for more on hydrologic studies). Therefore, a site monitoring program that will provide the required data has been implemented and will be operated throughout site characterization.

A secondary purpose of collecting site-specific meteorological data, although not directly related to site characterization, is to fulfill permitting and licensing requirements. The environmental monitoring and mitigation plan and the environmental regulatory compliance plan will contain detailed information on the need and use of the meteorological data in permitting and licensing activities. These activities will ensure that the facilities can be built and operated within regulatory constraints.

These short-term considerations, however, are only one aspect of the repository system-climate interaction that must be considered to ensure long-term waste isolation. Long-term climatological evaluations are summarized in the following paragraphs.

The currently available information indicates the presence of two major lake cycles in the Great Basin during the past 150,000 yr. The last cycle, which occurred during the late Pleistocene, is well documented in the Bonneville, Lahontan, and Mono basins. The older cycle is poorly documented by absolute age determination in both the Bonneville and Lahontan basins, but seems to have occurred 120,000 to 90,000 yr ago. A large number of basins throughout the Great Basin, which today are playas or dry lakes, contain shoreline evidence of two former lake cycles. Radiocarbon analyses of material from the basins indicate that the youngest shoreline probably corresponds to a lake cycle that occurred in the late Pleistocene. The age of the oldest shoreline is not known. In addition to these two cycles there are indications that other lacustrine high stands may have occurred during the middle Wisconsin (Forester, 1987) and that in some lake basins these lesser known events may have been of larger magnitude than the late Wisconsin high stand. Future needs include more detailed analyses of sediment cores that extend further back in time. Studies of cores dating back to the oxygen isotope stage 5d-5e (Shackleton et al., 1983) are necessary to evaluate glacial-interglacial climatic transitions.

The correlation between prehistoric lake cycles and periods of increased precipitation in the region around Yucca Mountain is not well understood, although the last period of significantly increased precipitation in southern

Nevada may correlate with the timing of the last lake cycle in the northern portion of the Great Basin. Moreover, the relationship between increased precipitation and recharge in the region is not well understood. For this reason the available paleoclimatic data offer insufficient insight into the history of ground-water recharge there. Other data, however, have contributed to the current understanding of recharge; they are discussed in Section 3.7.

Information on the advance and retreat of mountain glaciers within the Great Basin is scant. The few data that do exist indicate that late-Wisconsin mountain glaciation may have been roughly synchronous with the last lake cycle.

Pack rat midden and palynological studies indicate that the sagebrush steppe and other vegetation assemblages that adapted to relatively cold, dry conditions were more widespread, during the period from at least 40,000 to 11,000 yr ago, than today; with the period from approximately 20,000 to 14,000 yr ago representing the maximal difference from modern conditions. Plant species and associations that adapted to warmer temperatures and higher levels of summer rainfall were absent or rare during this period. The Wisconsin-age vegetation appears to reflect reduced temperatures rather than great enhancements of pluvial precipitation. By 11,000 yr ago sagebrush and xerophytic subalpine conifers began to move to higher elevations, reflecting increasing warmth and perhaps aridity. However, in much of the Great Basin and Mojave Desert region it appears that conditions between 11,000 and 7,000 yr ago remained cooler than those of today. The pack rat midden data have been interpreted as indicating that monsoonal summer rainfall reached southern Nevada some time between 11,000 and 5,000 yr ago, and may have been at higher than modern levels for all or part of the period.

The repeated fluctuations between glacial and interglacial climates that occurred during the Quaternary are fundamentally related to variations in the seasonal and latitudinal receipt of solar radiation, as determined by variations of the earth's orbital elements. The orbital variations, which explain a considerable portion of the total variance of large-scale climatic phenomena, will continue to force large-scale climate change. The task that remains is to relate the response of climatic variation in the southern Great Basin to large-scale climatic phenomena. Prediction of future climatic variation over the next 100,000 yr requires a multistep procedure that incorporates the hierarchy of controls and responses in the climate system. The development of climate prediction methods must, therefore, proceed in a top-down fashion, from the larger-spatial, longer-temporal scale controls to the local responses to variation in these controls. The process involves the correlation of present-day vegetational and meteorological data and the application of the correlations to paleobotanic records to extract the prehistoric record of climate variation. The process also uses the paleolacustrine, glacial, and paleobotanic records for model validation.

For the near-future (the next 10 to 1,000 yr) predictions of climate, the effect of large inputs of anthropogenic aerosols, carbon dioxide, and other trace gases into the climate system must be taken into account.

The quality of published data needed for input to the developmental validation of a future-climate model is poor to fair depending on the areal

and temporal scales of interest. Climatic, stream-flow, and vegetational data for the region during the past 50 yr are of fair quality and are probably sufficient for modeling applications. Lake-level and vegetational time-series data exist for a limited number of sites in the northern and western parts of the region for approximately the past 20,000 yr, but they are insufficient in the southern Great Basin, particularly in the vicinity of Yucca Mountain. Vegetational, hydrologic, and geologic time-series data are poor to inadequate for times before 20,000 yr ago; the acquisition of this older time-series data will be necessary for future climate modeling efforts.

5.3.2 RELATION TO DESIGN

Design of surface facilities must account for extremely high temperatures in summer months that drop significantly at night. Freezing temperatures must be considered, and the possibility of snow cannot be ignored. The components of the surface facilities that extend above ground must be able to withstand occasional high-speed gusts of wind. Aside from providing human comfort and safety needs, the only other meteorological factor that must be considered in the design of the surface facilities is the assurance that flood potential has been considered. Paleoclimatic influences are not pertinent to design.

5.3.3 IDENTIFICATION OF INVESTIGATIONS AND INFORMATION NEEDS

The primary investigation relative to assessing repository performance in terms of meteorological influences is a comprehensive, site-specific base of meteorological data. The data should include hourly averages of wind speed, wind direction, temperature, and parameters for the determination of atmospheric stability. Ideally, the data should cover as long a time period as possible and should include parameters for comparison with less site-specific but longer-term climatological records (precipitation and relative humidity). The Meteorological Monitoring Plan (SAIC, 1985) (Section 8.3.1.12) provides a means by which this information need can be satisfied.

Additional paleoclimatic data are also needed for site characterization. Continuous paleoclimatic records from key locations extending back to the previous interglacial (about 125,000 yr ago) are needed to test the ability of the climate model to track the actual climate as the boundary conditions and slowly varying components change over the time scale of interest. A more dense network of paleoclimatic records extending back about 20,000 yr is needed to describe spatial patterns of climatic variation for such key times as full glacial, late glacial, and early Holocene. Climatic data at a single key location extending back several hundred thousand years are also desirable for determination of the nature (magnitude, frequency) of climate response to astronomic forcing. Models of future local and global climatic variability are needed to establish an important boundary condition (recharge) of the ground-water system.

Underlying these planned and ongoing studies and, indeed, all paleoclimatic studies in southern Nevada generally and at the Yucca Mountain site

specifically, is the need for integrating results and interpretations to produce a coherent picture of paleoenvironmental change. The interpretive, integrative, and data gathering processes must all equally progress in time because the insights gained by this effort will direct future research most efficiently.

5.3.4 RELATION TO REGULATORY GUIDE 4.17

A comparison of the information in Chapter 5 with Regulatory Guide 4.17 (NRC, 1987) shows that none of the information required by the guide has been omitted from Chapter 5, but the chapter does include additional information as follows:

1. An introduction that discusses the general information included in the chapter.
2. A summary (Section 5.3) that provides a link between the data presented in the chapter and the plans described in Chapter 8.

Nuclear Waste Policy Act
(Section 113)

REFERENCES

Consultation Draft

***Site Characterization
Plan***

***Yucca Mountain Site, Nevada Research
and Development Area, Nevada***

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