Repository Site Definition in Basalt: Pasco Basin, Washington

MASTER

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Sandia National Laboratories

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Abstract

Pasco Basin is a structural and topographic basin of approximately 2000 mi² (5180 km²) located within the Yakima Fold Belt Subprovince of the Columbia Plateau. The stratigraphic sequence within the basin consists of an undetermined thickness of lower Miocene and younger flood basalts with interbedded and overlying sedimentary units. This sequence rests upon a basement of probably diverse rock types that may range in age from Precambrian through early Tertiary.

The basin is bounded on the north, south and west by east-west-trending Yakima folds which, along with related folds within the basin, plunge to the east. Subsurface structures may be related to the intersection of both east-west-trending and northwest-trending structural features. Faults have been proposed to explain structural relationships at Wallula Gap and along the northeastern flank of Rattlesnake Hills. Two faults are present at Gable Mountain.

Although a large amount of information is available on the hydrology of the unconfined aquifer system, ground-water flow within the basin is, in general, poorly understood. Recharge areas for the Mabton interbed and the Saddle Mountains Formation are the highlands surrounding the basin with the flow for these units toward Gable Butte - Gable Mountain and Lake Wallula. Gable Butte - Gable Mountain probably is a ground-water sink, although the vertical flow direction in this zone is uncertain. The amount of upward vertical leakage from the Saddle Mountains Formation into the overlying sediments or to the Columbia River is unknown. Units underlying the Mabton interbed may have a flow scheme similar to those higher units or a flow scheme dominated by interbasin flow. Upward vertical leakage either throughout the basin, dominantly to the Columbia River, or dominantly to Lake Wallula has been proposed for the discharge of the lower units. None of these proposals is verified.

The lateral and vertical distribution of major and minor ions in solution, Eh and pH, and ion exchange between basalt and ground-water are not well defined for the basin. Changes in the redox potential from the level of the subsurface facility to the higher

stratigraphic levels along with the numerous other factors influencing K_d , result in a poor understanding of the retardation process.

Experimental determinations of mechanical properties of Columbia River basalts are not available in sufficient quantity or quality to adequately define the properties as a function of pressure, temperature, mineralogy, and porosity. The data set for thermal properties is more complete.

Contents

	<u>Page</u>
INTRODUCTION	1
CHAPTER 1 - REGIONAL SETTING	2
1.1 Tectonic Setting	2
1.2 Seismicity	6
1.3 Igneous Activity	7
1.4 Surface Geologic Processes	9
1.5 Natural Resources	9
CHAPTER 2 - REFERENCE SITE CHARACTERISTICS	12
2.1 Geology	12
2.1.1 Stratigraphy	14
2.1.2 Geologic Structure	22
2.1.3 Spatial Variation in Thickness of Stratigraphic Units	25
2.1.4 Intraflow Structures in the Grande Ronde Basalt	26
2.2 Hydrogeology	29
2.2.1 Hydraulic Head	38
2.2.2 Hydraulic Conductivity	50
2.3 Geochemistry	56
2.3.1 Ranges of Distribution Coefficient (K_d)	56
2.3.2 Solubility of Waste Nuclides ...	57
2.3.3 Hydrogeochemical Parameters	59
2.4 Other Physical Parameters of Basalt ...	61
CHAPTER 3 - RECOMMENDATIONS FOR FUTURE WORK	63
REFERENCES	66
APPENDIX A - Hydraulic Conductivity	A-1
APPENDIX B - Transmissivity	B-1

	<u>Page</u>
APPENDIX C - Geothermal Gradient	C-1
APPENDIX D - Density	D-1
APPENDIX E - Porosity	E-1
APPENDIX F - Thermal Conductivity	F-1
F.1 Theoretical Considerations	F-1
F.1.1 Mineralogy	F-1
F.1.2 Porosity and Saturating Fluid ..	F-3
F.1.3 Pressure	F-8
F.1.4 Temperature	F-10
F.2 Discussion	F-13
APPENDIX G - Specific Heat	G-1
APPENDIX H - Thermal Expansion	H-1
APPENDIX I - Thermal Diffusivity	I-1
APPENDIX J - Compressive Strength	J-1
APPENDIX K - Tensile Strength	K-1
APPENDIX L - Mechanical Properties of Joints or Fractures	L-1
Appendix M - Elastic Constants	M-1
Appendix N - Geochemistry	N-1
N.1 Discussion of K_d	N-1
N.1.1 Definitions	N-1
N.1.2 Factors Influencing K_d	N-2
N.1.2.1 Surface Area	N-3
N.1.2.2 Ion Exchange Capacity & Redox Conditions .	N-3
N.1.2.3 Solution pH	N-3
N.1.2.4 Temperature	N-4
N.1.2.5 Chemical Composition	N-4

	<u>Page</u>
N.1.2.6 Radionuclide Concentration	N-4
N.1.2.7 Particulate Transport	N-4
N.1.2.8 A Treatment of These Factors	N-5
N.1.3 Experimental K_d Values in Basalt Environments	N-5
N.2 Hydrogeochemical Condition of the Repository Environment	N-12
N.2.1 Dissolved Species	N-12
N.2.1.1 Major Ions	N-12
N.2.1.2 Trace Elements	N-12
N.2.2 Redox Conditions	N-15
APPENDIX O - Hydrologic Modeling of the Reference Site.....	O-1
APPENDIX P - Glossary	P-1

LIST OF FIGURES

<u>Figure</u>		<u>Page</u>
1-1	Location Map of Columbia Plateau and Pasco Basin	3
1-2	Geologic Subprovinces of Columbia Plateau and Contiguous Areas	4
1-3	Location of Present and Planned Shell Oil Company Drill Holes in Columbia Plateau	11
2-1	Location of Major Structural Elements Within and Adjacent to the Pasco Basin	13
2-2	Pasco Basin Stratigraphic Nomenclature	15
2-3	Structural Features of a Basaltic Lava Flow	27
2-4	Estimated 1944 Water Table Map for Hanford Site, Pasco Basin, Washington	32
2-5	Water Table Rise Beneath the Hanford Site, 1944-1978	33
2-6	Water Potentials for the Mabton Interbed, Hanford Site	34
2-7	Location of Selected Drill Holes in Pasco Basin	40
2-8A&B	Hydraulic Head Measurements Within Wanapum and Grande Ronde Formations at Drill Hole DC-1	41-42
2-9	Hydraulic Head Measurements Within Grande Ronde Formation at Drill Hole DC-2	43
2-10	Hydraulic Head Measurements Within Grande Ronde Formation at Drill Hole DC-6	44

<u>Figure</u>		<u>Page</u>
2-11	Hydraulic Head Measurements Within Wanapum and Grande Ronde Formations at Drill Hole DC-6	45
2-12	Hydraulic Head Measurements Within Wanapum and Grande Ronde Formations at Drill Hole DC-8	46
2-13	Hydraulic Head Measurements Within Wanapum and Grande Ronde Formations at Drill Hole DC-12	47
2-14	Hydraulic Head Measurements Within the Saddle Mountains and Wanapum Formations at Drill Hole DC-14	48
2-15	Hydraulic Head Measurements Within Saddle Mountains, Wanapum, and Grande Ronde Formations at Drill Hole DC-15	49
2-16	Hydraulic Head Measurements Within Grande Ronde Formation at Drill Hole RSH-1	51
D-1	Histogram of Bulk Density of Basalt -- World-wide and Columbia River Basalt	D-3
D-2	Histogram of Grain Density of Basalt -- World-wide and Columbia River Basalt	D-4
E-1	Histogram of Total Porosity of Basalt ...	E-2
E-2	Histogram of Effective Porosity of Basalt	E-3
E-3	Histogram of Effective Porosity/Total Porosity Ratio for Basalt	E-4
F-1	Thermal Conductivity of Basalt Versus Temperature	F-2
G-1	Specific Heat of Basalt Versus Temperature	G-2

<u>Figure</u>		<u>Page</u>
H-1	Coefficient of Linear Thermal Expansion of Basalt	H-2
I-1	Thermal Diffusivity of Basalt as a Function of Temperature	I-2
J-1	Histogram of Uniaxial Compressive Strength of Intact and Fractured Basalt	J-2
J-2	Relationship Between Compressive Strength and Confining Pressure for Columbia River Basalts	J-3
J-3	Histogram Showing Relationship Between Compressive Strength and Temperature for Pomona and Umtanum Flows	J-7
J-4	Relationship Between Compressive Strength and Total Porosity for Basalt	J-8
J-5	Uniaxial Compressive Strength as a Function of Moisture Content for Basalt	J-10
J-6	Relationship Between Mean Uniaxial Compressive Strength and Strain Rate for Umtanum Flow	J-11
J-7	Relationship Between Mean Uniaxial Compressive Strength and Strain Rates for Umtanum Flow Samples With Bulk Density Greater Than 2.60 g/cm^3	J-13
K-1	Relationship Between Tensile Strength, Loading Rate, and Bulk Density for Columbia River Basalts	K-2
K-2	Histogram Showing Relationship Between Tensile Strength and Bulk Density for Basalt	K-3
M-1	Histogram of Dynamic and Static Young's Modulus for Basalt	M-3
M-2	Histogram of Dynamic and Static Poisson's Ratio for Basalt	M-4

<u>Figure</u>		<u>Page</u>
M-3	Histogram of Dynamic Shear Modulus for Basalt	M-5
M-4	Histogram of Dynamic Bulk Modulus for Basalt	M-6
M-5	Relationship Between Mean Poisson's Ratio, Mean Young's Modulus, and Confining Pressure for Pomona Flow ...	M-7
M-6	Relationship Between Mean Poisson's Ratio, Mean Young's Modulus, and Confining Pressure for Umtanum Flow ..	M-8
M-7	Relationship Between Mean Poisson's Ratio, Young's Modulus, and Temperature for Pomona Flow	M-11
M-8	Relationship Between Mean Poisson's Ratio, Young's Modulus, and Temperature for Umtanum Flow	M-12
M-9	Relationship Between Poisson's Ratio and Total Porosity for Pomona Flow	M-14
N-1	Piper (trilinear) Diagram of Major Ion Composition of Various Ground Waters Associated with the Hanford Site	N-13
N-2	Hanford Site Repository Conditions and Constraints in General Eh-pH Space ...	N-18

LIST OF TABLES

<u>Table</u>		<u>Page</u>
1-1	Age and Distribution of Basalt Flows in the Pasco Basin	8
2-1	Ranges of Hydraulic Conductivities for Stratigraphic Units in Pasco Basin ...	53
2-2	K_d Values for the Basalt Environment at the Hanford Reservation	58
2-3	Ranges of Solubilities of Selected Elements in Bedded Salt Repository Site Conditions	60
2-4	Suggested Ranges for Physical Parameters of Hanford Basalts	62
A-1	Hydraulic Conductivity Values for Stratigraphic Units in Pasco Basin ...	A-3
B-1	Transmissivity Values for Stratigraphic Units in Pasco Basin	B-2
C-1	Geothermal Gradient Measured at Drill Hole RSH-1	C-2
C-2	Geothermal Gradient Measurements in Washington State	C-3
C-3	Mean Worldwide Geothermal Gradient	C-4
D-1	Comparison of Basalt Density Distribution With Distribution of Basalt Lithologic Types at Hanford	D-2
D-2	Density of Basalt	D-5
E-1	Porosity of Basalt	E-5
F-1	Mineralogical Constituents of Umtanum and Pomona Flows	F-5
F-2	Thermal Conductivity of Basalt Minerals at 23-27°C	F-6

<u>Table</u>		<u>Page</u>
F-3	Calculated Thermal Conductivity; Initial Conditions	F-7
F-4	Calculated Thermal Conductivity of the Pomona Flow; Effects of Varying Mineral Compositions	F-7
F-5	Calculated Thermal Conductivity of the Pomona Flow; Effects of Varying Porosity	F-9
F-6	Calculated Thermal Conductivity of the Pomona Flow; Effects of Varying Pressure	F-9
F-7	Estimated Relationships of Temperature and Basalt Mineral Thermal Conductivities	F-11
F-8	Calculated Thermal Conductivity of the Pomona Flow; Effects of Varying Temperature	F-12
F-9	Calculated Thermal Conductivity of Umtanum Entablature; Effects of Varying Temperature	F-12
F-10	Thermal Conductivity of Basalt	F-15
G-1	Calculated Specific Heats of Pomona and Umtanum Flows	G-4
G-2	Specific Heat of Basalt	G-5
H-1	Linear Thermal Expansion of Basalt	H-4
I-1	Thermal Diffusivity Ranges Calculated From Bulk Density, Specific Heat and Thermal Conductivity	I-3
I-2	Thermal Diffusivity of Basalt	I-4
J-1	Values of Cohesion and Angles of Internal Friction for Basalt	J-5
J-2	Compressive Strength of Basalt	J-15

<u>Table</u>	<u>Page</u>
K-1 Tensile Strength of Basalt	K-5
L-1 Friction Factors for the Umtanum Flow ...	L-2
L-2 Joint Strengths for the Umtanum Flow	L-2
M-1 Relationships of Young's Modulus and Poisson's Ratio to Pressure for Hanford Basalts	M-9
M-2 Relationships of Young's Modulus and Poisson's Ratio to Temperature for Hanford Basalts ($\rho \geq 2.60 \text{ g/cm}^3$)	M-13
M-3 Elastic Constants of Basalt	M-17
N-1 Synthetic Ground-water Formulations	N-7
N-2 Chemical Characterization of Geomedia in the Hanford Basalts	N-9
N-3 Average Mineralogy of Basalts in the Hanford Reservation	N-10
N-4 Physical and Exchange Characterization of Geomedia in the Hanford Reservation ..	N-11
N-5 Trace Element Concentrations in Ground Water at the Hanford Site	N-14

INTRODUCTION

This report documents work performed for the U.S. Nuclear Regulatory Commission (NRC) by Sandia National Laboratories in Albuquerque (SNLA) on a repository site definition for a high-level nuclear waste repository in basalt. Repository site definitions are an integral part of the NRC program at SNLA to develop risk assessment methodologies for the geological disposal of high-level waste. Developing these assessment methods requires an understanding of the natural system to be assessed and a data base with which to exercise and demonstrate the assessment methodology. By preparing repository site definitions, SNLA develops an understanding of the natural systems and a familiarity with the data to be used in risk assessment.

The repository site definition for basalt is based primarily on the Pasco Basin in Washington State, because one of the three sites for which the NRC expects to receive a Site Characterization Report from the U.S. Department of Energy is located in this basin. When the published reports on the Pasco Basin did not supply sufficient data for a particular parameter, the data base was supplemented with data from outside the basin. This procedure allows the NRC's risk assessment methodology development program to proceed in advance of the acquisition of a more complete data base to be supplied by the Department of Energy.

Discussions of the regional setting, geology, hydrogeology, and geochemistry are included in this report. Hydrologic properties and other physical parameters are defined for either each of the stratigraphic units or each general rock type. Detailed documentation and description of the parameters are in the Appendices.

Because of the technical nature of the material included in repository site definition, the readers of this report are assumed to be familiar with basic geological, hydrological, thermomechanical, and geochemical terminology and concepts. A glossary is included as an appendix of this report to provide precise definitions of selected terms. Readers are advised to review the maps available in Myers and others (1979) and Gephart and others (1979) prior to reading this report and to have these maps available for reference during reading. Both the size and the number of maps within these two reports preclude their inclusion here.

CHAPTER 1

REGIONAL SETTING

The Pasco Basin is situated within the Columbia Plateau in the northwestern United States (Figure 1-1). In order to better understand potential risk from future geologic processes, the regional setting of the Basin and the Plateau must be examined.

1.1 Tectonic Setting

The Columbia Plateau is situated east of the northwest-trending zone along which the Juan de Fuca plate may be descending beneath the American plate (Savage and others, 1981). This subduction presumably is involved in the generation of the volcanic activity in the Cascade Range. No igneous activity is presently occurring on the Plateau itself, although sporadic ash falls are deposited from explosive volcanism in the Cascades.

Seismic and magnetotelluric studies indicate that a relatively thin crust (less than 14 miles [22.4 km]) underlies the Columbia Plateau (Myers and others 1979). Heat flow data (Blackwell, 1978) show the plateau to be an area of relatively low to moderate heat flow with an average value of 1.0 to 1.5 heat flow units ($1.0 - 1.5 \mu\text{cal}/\text{cm}^2 \text{ sec}^{-1}$). This range is consistent with the overall average for the crust, although less than the average value of 1.8 heat flow units for Mesozoic/Cenozoic orogenic regions characterized by much thicker crust.

The majority of the deformation in the Columbia Plateau in Washington has been confined to the Yakima Fold Belt subprovince (Figure 1-2). This region is characterized by west- to northwest-trending folds, with the west-trending folds transected in some places by northwest- to north-trending shear zones and minor folds. Deformation began approximately 15 million years (my) ago and continued at least through the Pliocene (Myers and others, 1979).

Within the Pasco Basin, folding and subsidence is not recognized prior to approximately 14.5 my ago (Myers and others, 1979). Folding and faulting occurred first on northwest trends and later (from approximately 14 my

Figure 1-1. Location Map of Columbia Plateau and Pasco Basin (Atlantic Richfield Hanford Company, 1976).

Figure 1-2. Geologic Subprovinces of Columbia Plateau and Contiguous Areas (from Myers and others, 1979).

ago) on both west and northwest trends. This deformation was most intense from approximately 10.5 my ago into the Pliocene. Most of the existing faults are thought to have developed during this period of more intense folding (Myers and others, 1979).

Pliocene folding and faulting of the Ringold Formation, the oldest suprabasalt sedimentary unit, is observed both inside and outside the Pasco Basin. The degree of deformation decreased from early to late Ringold time (Myers and others, 1979).

With certain exceptions, sediments younger than the Ringold Formation do not seem to have been involved in the deformation, although this possibility has not been discounted (Myers and others, 1979). The exceptions mentioned are some faults along which movement may have taken place as recently as 12,000 to 40,000 years ago (Myers and others, 1979), such as the Saddle Mountains fault with 2000 feet of maximum displacement.

A recent study by Campbell and Bentley (1981) suggests that faulting of tectonic origin has occurred as recently as 500 to 13,000 years ago along the Satus Peak anticlinal portion of Toppenish Ridge, an area approximately 45 miles (75 km) west of the Hanford Reservation. Surface scarps show up to 13 ft (4 m) of relative vertical displacement. The authors point out that none of the other anticlines of the Yakima fold belt show faulting of similar age or style.

A number of studies have attempted to measure amounts of relative vertical movement in the Columbia Plateau region. Tillson (1970) performed an analysis of recent triangulation and leveling data for the Columbia Plateau, and reached the conclusion that any vertical movement is less than 0.04 in/yr (1 mm/yr), and thus is nearly insignificant relative to potential measurement errors. In addition, no evidence was found that any significant displacement is associated with known structural features. Reidel and others (1980) compared relative thicknesses of basalt flows of different ages and calculated minimum and maximum average uplift rates during the Miocene of 1.6×10^{-3} and 2.8×10^{-3} in/yr (3.9×10^{-2} and 7.0×10^{-2} mm/yr), respectively. Measurement of sediment thickness in the deepest part of the Pasco Basin provided an estimate of 1.0×10^{-3} in/yr (2.5×10^{-2} mm/yr) of subsidence during the Miocene (Reidel and others, 1980).

Tillson (1970) stated that triangulation data show no evidence of horizontal displacement in southeastern Washington. A recent study of geodetic strain measurements by Savage and others (1981) found that strain is accumulating at the Hanford site in the following amounts and orientations: $0.02 \pm 0.01 \mu\epsilon/\text{yr}^*$, N36°W (compression) and $0.04 \pm 0.01 \mu\epsilon/\text{yr}$, N54°E (compression). These values may or may not be significant. Systematic error of 0.01 to 0.02 $\mu\epsilon/\text{yr}$ may negate the measured strain rates. The rates are consistent with subduction of the Juan de Fuca plate beneath the American plate and slow accumulation of strain along the boundary between the two plates (Savage and others, 1981).

1.2 Seismicity

Good-quality earthquake records for the Columbia Plateau and surrounding regions are not available prior to the 1960s. Historical information of varying quality goes back to the mid-1800s, but is limited in usefulness to felt events of magnitude 5.0 or greater (Myers and others, 1979). In general, earthquake events throughout Washington are shallow (less than 43 miles [68.8 km]) and confined to the crust. Events in eastern Washington are, on the average, shallower than those in western Washington.

The Columbia Plateau is an area of relatively low seismicity. The largest known earthquake occurred near Milton-Freewater, Oregon in 1936, and is reported to have had a maximum modified Mercalli intensity of VII, equivalent to a shallow earthquake with a magnitude of approximately 5.7 (Myers and others, 1979).

Earthquakes characteristic of the Columbia Plateau are shallow (usually less than 1.8 miles [2.9 km]) swarms of low magnitude (less than 4.0, with most being less than 2.0). These earthquakes occur in groups of several hundred to over 1000 individual events in a relatively restricted volume of rock (less than 30 cubic miles [123 cubic km]). No regular foreshock/aftershock sequence occurs. The events increase slowly in magnitude to a maximum and then die off. Epicenters and hypocenters generally show no preferred alignments (Rothe, 1978; Myers and others, 1979).

*1 $\mu\epsilon/\text{yr} = 1 \times 10^{-6} \text{ in/in/yr}$.

Because the earthquake magnitudes are so low, focal plane solutions cannot be made for single events. However, composite focal plane solutions for groups of similar events suggest predominantly dip-slip movement on approximately west-trending planes with near vertical dip (Rothe, 1978).

Based on data from earthquakes in the Pasco Basin, Rothe (1978) has postulated stress release to occur by slip on columnar joints in dense basalts. However, more seismic data are necessary before other mechanisms are discounted. Campbell and Bentley (1981) describe an earthquake of magnitude 3.8 with an epicenter approximately 50 miles (83 km) west of the Hanford Reservation which occurred February 2, 1981. The mechanism is interpreted as right-lateral strike-slip movement. Similar mechanisms may also be operating within the Pasco Basin.

The earthquake activity below the Columbia Plateau basalts (deeper than 5.6 miles [9 km]) is more uniformly scattered and is not obviously associated with any known structures (Rothe, 1978).

1.3 Igneous Activity

The Columbia Plateau contains one of the world's largest accumulations of flood basalt. These basalts were erupted over a time period extending from earlier than 16.5 my ago until approximately 8.5 my ago (Myers and others, 1979). In the Pasco Basin, the youngest basalts are approximately 6 my old. The only evidence of more recent igneous activity in the Basin is airfall ash layers presumably derived from the Cascade volcanoes to the west (Myers and others, 1979). The most recent of these airfall ash layers is sporadically distributed on the surface of the Pasco Basin and is derived from the eruption of Mt. St. Helens during 1980. Table 1-1 contains information on the ages and inferred distributions of some of the basalt flows found in the Pasco Basin. For more detailed information, see Myers and others (1979).

Because of low heat flow, lack of igneous activity during the Quaternary, and relative isolation from extensive contemporary igneous provinces, the probability of any igneous activity occurring within the Pasco Basin during the next one million years is low.

Table 1-1. Age and Distribution of Basalt Flows
in the Pasco Basin

Formation	Member	Distribution	Age (my)
Grande Ronde	---	Entire Pasco Basin	14 - 16.5
Wanapum	Frenchman Springs	Entire Pasco Basin	---
	Roza	Most of Pasco Basin except for Tri- Cities area and area just south of Sentinel Gap.	---
	Priest Rapids	Same as Roza	---
Saddle Mountains	Umatilla	Entire Pasco Basin south of the Columbia River	---
	Wilbur Creek	Narrow E-W trending zone along and south of E-W por- tion of Columbia River	---
	Asotin	Same as Wilbur Creek, but area twice as wide (N-S)	12.6 $\pm$ 0.4*
	Esquatzel	Central part of Pasco Basin south of Columbia River, E-W trending zone	---
	Pomona	Entire Pasco Basin	12
	Elephant Mountain	Entire Pasco Basin	10.5
	Ice Harbor	Tri-Cities area only	8.5

*Ward (1976)

1.4 Surface Geologic Processes

Estimated rates of regional erosion have been calculated for the United States by Judson and Ritter (1964). These calculations are based on measured values and best estimates of suspended, dissolved and bed load sediment moved by major streams and rivers. For the Pacific Northwest drainage basin, including the Columbia and Snake Rivers and their tributaries, the erosion rate is calculated to be 1.5 in/1000 yrs (3.8 cm/1000 yrs). This is the lowest regional rate in the United States.

Past glaciations have not covered the central part of the Columbia Plateau, and glacial deposits occur only in the northern and in the far western portions of the plateau. These deposits are due to the advance of the Okanogan lobe of the Cordilleran ice sheet onto the Columbia Plateau. This advance occurred during the Pinedale glaciation, between 25,000 and 10,000 years ago (Richmond and others, 1965).

Bull (1980) analyzed the potential for future glaciation for all proposed geologic repository sites, including the Hanford site. He estimated that the chance that ice would cover Hanford during the next 1 my is small but finite. The presence of large thicknesses of glacial ice above a repository possibly could alter a pre-existing ground-water flow regime. Analysis of such an effect on radionuclide release times is not practical at this time due to lack of definition of the present flow regime. Stresses due to loading and unloading of the surface during advance and retreat of ice sheets could cause fracturing of the bedrock. Bull (1980) considered such fracturing to be likely in the Columbia Plateau during the next 1 my.

Associated with past glaciations of the Pacific Northwest have been glacial lakes caused by ice damming of rivers, and catastrophic floods as these lakes breached the dams. If the maximum erosive effects of these floods may be estimated by the depth of the coulees in the eastern part of the basin (about 200 feet, [61 m]), the probability of breaching a deep geologic repository is extremely small.

1.5 Natural Resources

The Columbia River basalts and other tholeiitic flood-basalt provinces throughout the world are notable

for a lack of mineral resources (Rockwell Hanford Operations, 1981). Any high-unit-value minerals that might be present in the rocks beneath the basalts are masked and are probably uneconomic due to the considerable thickness of the basalt. In the Columbia River Plateau, the total thickness of basalts may exceed 10,000 feet (3000 m).

In the 1920's and early 1930's, gas was produced from an interbed in the Columbia River basalts located to the south of the Hanford Site. No other production is recorded for the Columbia Plateau. Shell Oil Company began drilling a 20,000 foot (7000 m) wildcat well north of Yakima in 1980, and has filed for drilling permits for three more wells (McCaslin, 1981). Each of the drill sites is located on an anticlinal structure (Figure 1-3). The planned well closest to the Hanford Site is to be located in the Saddle Mountains immediately north of the site.

Flood-basalt provinces are not known as areas of geothermal resources. Heat flow in the Pasco Basin is approximately 1-1.5 heat flow units (Rockwell Hanford Operations, 1981). Because the average heat flow for the crust is 1.5 heat flow units, the Pasco Basin does not have anomalously high flow and is not a geothermal resource area.

The Hanford Site is not a major source of groundwater in view of supplies elsewhere in the region. Although the unconfined aquifers contain the most readily available ground water for future use, confined aquifers down to and including the Mabton interbed may be potential sources of ground water. Whereas neither the subsurface facility nor the ground-water flow at the depth of the facility is likely to be affected by the pumping of shallow aquifers, the pumping of the confined aquifers down-gradient from the facility may provide a pathway for released radionuclides to the accessible environment.

Figure 1-3. Location of Present and Planned Shell Oil Company Drill Holes in Columbia Plateau.

CHAPTER 2

REFERENCE SITE CHARACTERISTICS

2.1 Geology

The Pasco Basin in south-central Washington is one of several structural and topographic basins within the Columbia Plateau. It contains the Hanford Site and has been selected as the basis for repository site definition. The basin perimeter is partially defined by anticlinal ridges of the Yakima Fold Belt (Figure 2-1). The Saddle Mountains form the northern boundary; the Umtanum Ridge and Yakima Ridge comprise the western boundary; the Rattlesnake Hills and Horse Heaven Hills form the southern boundary; and a broad zone of gradually increasing structural westerly dip comprises the somewhat arbitrarily placed eastern boundary. Structural relief on some stratigraphic units approaches 3000 to 4000 feet (914 to 1219 meters) from the center of the basin to the surrounding mountain ranges (Myers and others, 1979, Plate III-4b).

The basin, which occupies about 2000 mi² (5180 km²), is a semi-arid steppe with sparse vegetative cover. Average precipitation is 6.25 in/yr (15.88 cm/yr). The average maximum temperatures in January and July are 38.8°F and 91.8°F (3.8°C and 33.2°C), respectively (Stone and others, 1972). The average minimum temperatures for the same months are 22.1°F and 61.0°F (-5.5°C and 16.1°C). Stone and others (1972) observed that 37 percent of the annual precipitation falls during November, December, and January, and that only 10 percent falls during July, August, and September. Almost half of all precipitation that falls during the months of December through February is snow.

The topographic highs that surround the basin are outcrops of basaltic rocks covered in places by a thin veneer of wind-deposited silt. The Rattlesnake Hills along the southwestern margin of the basin are the highest of these outcrops with an elevation of approximately 3600 feet (1097 meters). The Saddle Mountains to the north reach elevations of about 2600 feet (792 meters). In the central portion of the basin, 0 to 700 feet (0 to 213 m) of fluvial, glaciofluvial, and lacustrine sediments overlie the basalts. The general surface topography of these sediments forms a broad undulating

EXPLANATION

- ANTICLINAL AXIS: ARROW IN DIRECTION OF PLUNGE
- SYNCLINAL AXIS : DASHED WHERE INFERRED:
- ARROW IN DIRECTION OF PLUNGE
- INFERRED FAULT

Figure 2-1. Location of Major Structural Elements Within and Adjacent to the Pasco Basin (after Myers and others, 1979).

plain ranging in elevation between 400 and 900 feet (122 to 274 meters). The Columbia River flows through the central portion of the basin, and the Snake and Yakima Rivers join the Columbia in the southern part of the basin.

2.1.1 Stratigraphy

Figure 2-2 shows the detailed stratigraphic nomenclature for rocks and unconsolidated deposits in the Pasco Basin.* The basin is underlain by three major rock units: pre-Columbia River Basalt Group rocks, Columbia River Basalt Group rocks, and the overlying Pliocene, Pleistocene, and Holocene sediments. Pre-Columbia River Basalt Group rocks have not been recognized in deep borings in the Pasco Basin and are not discussed further.

Basalts of the Columbia River Basalt Group are identified and correlated on the basis of stratigraphic position, chemistry, paleomagnetic properties, and borehole geophysical log character. These criteria have proven useful for correlation of major units over long distances. Geophysical logs are particularly useful for correlation of the thinner units over distances of a few miles between boreholes. Detailed discussion of the chemistry and paleomagnetic properties of major stratigraphic units in the Pasco Basin is given in Myers and others (1979).

Within the basin, the Columbia River Basalt Group is represented by the Grande Ronde, Wanapum, and Saddle Mountains Formations. More recent geophysical, chemical, and paleomagnetic data probably are available for the Grande Ronde than for the other formations because this unit is being considered as the repository unit. The greater detail for the Wanapum and Saddle Mountains Formations in Figure 2-2 reflects their more extensive outcrop patterns and longer history of study within the basin.

*Unless otherwise noted, all data in the sections on stratigraphy, geologic structure and spatial variation are taken from Myers and others (1979)

		AGE			FORMATION (x10 ⁶ YRS)	MEMBER OR SEQUENCE	FLOW OR BED		
		.0086					MAZAMA ASH		
		.0125					GLACIER PEAK ASH		
							ST. HELENS ASH		
							TOUCHET BEDS		
		.04					PASCO GRAVELS		
							PALOUSE SOIL		
							UPPER RINGOLD		
							MIDDLE RINGOLD		
							LOWER RINGOLD		
							BASAL RINGOLD		
COLUMBIA RIVER BASALT GROUP	YAKIMA BASALT SUBGROUP	SADDLE MOUNTAINS BASALT	8.5	ICE HARBOR MEMBER	GOOSE ISLAND FLOW				ELLENBURG FORMATION (INFORMAL NAME FOR ALL SEDIMENTS IN BASALT SEQUENCE)
					INDIAN MEMORIAL FLOW				
			MARTINDALE FLOW						
			BASIN CITY FLOW						
			LEVEY						
			10.5	ELEPHANT MOUNTAIN MEMBER	ELEPHANT MTN. UPPER FLOW				
					ELEPHANT MTN. LOWER FLOW				
			RATTLESNAKE RIDGE						
			12.0	POMONA MEMBER	POMONA UPPER FLOW				
					POMONA LOWER FLOW				
			SELAH						
				ESQUATZEL MEMBER	GABLE MTN. UPPER FLOW				
					GABLE MTN. BED				
					GABLE MTN. LOWER FLOW				
	COLD CREEK								
		ASOTIN MEMBER	HUNTZINGER FLOW						
		WILBUR CREEK MEMBER	WAHLUKE FLOW						
	13.6	UMATILLA MEMBER	SILLUSI FLOW						
			UMATILLA FLOW						
			MABTON						
			PRIEST RAPIDS MEMBER	LOLO FLOW (1 FLOW)					
				ROSALIA FLOW (1-4 FLOWS)					
	14.5	ROZA MEMBER	QUINCY						
ROZA UPPER FLOW									
ROZA LOWER FLOW									
SQUAW CREEK									
	FRENCHMAN SPRING MEMBER	SENTINEL GAP FLOW							
		SAND HOLLOW FLOW							
		GINKGO FLOW							
GRANDE RONDE BASALT	SENTINEL BLUFFS FLOW	VANTAGE							
		MUSEUM FLOW							
		ROCKY COULEE FLOW							
15.4	SCHWANA FLOWS	6-10 FLOWS							
		UMTANUM FLOW							
		1-3 HIGH MG FLOWS							
						LOW K ₂ O FLOW			
						AT LEAST 20 FLOWS			
PRE-COLUMBIA RIVER BASALT GROUP ROCKS									

Figure 2-2. Pasco Basin Stratigraphic Nomenclature (from Myers and others, 1979, Rockwell Hanford Operations, 1981).

The Grande Ronde Formation consists of at least 40 separate flows. These flows are grouped into a basal low-MgO sequence and an upper high-MgO sequence, known as the Schwana and Sentinel Bluffs sequences, respectively. The thickness of the Schwana sequence is not known. The thickness of the Sentinel Bluffs sequence is reported to be 900-1000 feet (274-305 m), 830-944 feet (253-288 m), or 726-975 feet (221-297 m) with an average of 895 feet (272 m), depending upon which section of Myers and others (1979) is consulted. Local, thin, discontinuous interbeds are present within the sequence.

The Umtanum flow at the top of the Schwana sequence is the likely host unit for a nuclear waste repository. The flow is a thick (131 to 285 feet, [40 to 87 meters]) widespread flow, the top of which occurs at depths ranging from 3600 feet (1097 meters) in the center of the basin to surface exposures in the northwestern part of the basin. The dense flow interior of the Umtanum is non-vesicular and averages about 150 feet (46 meters) thick. In its upper part, the Umtanum contains a thick (27 to 126 feet, [8.2 to 38.4 meters]) flow-top breccia that is present in all borings that penetrate the Umtanum. This zone usually is described as fractured, vesicular and brecciated basalt. Some of the vugs and fractures that occur in the flow top are filled with products of weathering or subsurface mineralogic alteration.

The Grande Ronde Formation is overlain by the Vantage Member of the Ellensburg Formation. The Vantage is relatively thin (0-26 feet [0-7 m]), with a lithology which varies from clay, clayey sandstone and loosely consolidated sandstone in boreholes DC-4 and DH-5 (northwest part of Pasco Basin) to a black clay in the central and southern parts of the basin. Locally, the Vantage is absent and the top of the Grande Ronde is represented by saprolite (Swanson and others, 1979).

Above the Vantage is the Wanapum Basalt Formation. The formation contains three members. The lowest unit is the Frenchman Springs Member, which is composed of up to ten flows and has a total thickness of 650-780 feet (198-238 m) in the northern part of the basin to 820 feet (250 m) at Wallula Gap.

Fenix and Scisson (1978c) report 2.5 feet (0.8 m) of banded tuff and ash above the Frenchman Springs Member in borehole DC-8. The interval is assigned to the Squaw Creek interbed.

The Roza Member overlies the Frenchman Springs Member. The Roza contains at least two flows throughout the Pasco Basin, and may consist of three flows in the Rattlesnake Hills (Bond and others, 1978). The thickness of the Roza varies from 0 to 100-200 feet (0 to 30-61 m), with the thickest section in the central basin with thinning to the north and south, especially on present topographic highs. The Roza is absent at Wallula Gap.

Locally, an unnamed interbed occurs between the Roza Member and the overlying Priest Rapids Member. The interbed is predominantly tuffaceous silt, with locally abundant petrified wood. Where present, thicknesses vary from 16 feet (5 m) on Yakima Ridge to 26 feet (8 m) in the southeastern part of the basin. In borehole DC-6, this interval is reported to be the Quincy interbed, represented by 2 feet (0.6 m) of black clay (Fenix and Scisson, 1978b).

The uppermost unit of Wanapum Basalt is the Priest Rapids Member. This unit is usually composed of one of two flows, but up to four flows have been distinguished locally. Individual flows may be up to 200 feet (61 m) thick, while the total thickness of the Priest Rapids Member ranges from 0 to 160-260 feet (0 to 49-79 m). The Priest Rapids is absent at Wallula Gap.

Separating the Wanapum Basalt Formation and the Saddle Mountains Basalt Formation is the Mabton interbed. The area in which the Mabton has its greatest thickness is in the center of the Basin, approximately coincident with the Cold Creek syncline, where up to 170 feet (52 m) of sediments are present. Thinning occurs outward from this area, but the Mabton is present everywhere in the Basin. The lithology consistently shows three units: (1) a lower poorly consolidated sand or sandstone with interbedded diatomite or volcanoclastic sediments, (2) a middle very fine-grained tuffaceous clayey sandstone, and (3) an upper well-indurated lapilli tuffstone. Both Bond and others (1978) and Myers and others (1979) mention the usual presence of silicified layers near the top of the interbed.

The oldest member of the Saddle Mountains Basalt Formation is the Umatilla Member. One of two flows are present within the Pasco Basin, with total thickness ranging from 0 to about 300 feet (0-91 m). The Umatilla pinches out between the Saddle Mountains and the Gable Mountain - Gable Butte structure.

The next youngest unit is the Wilbur Creek Member, which is represented by the Wahluke flow in the Pasco Basin. The Wahluke is confined to a narrow east-west band in the area between the Saddle Mountains and Gable Mountain, and is interpreted to be an intracanyon flow. The total thickness within the Pasco Basin is reported to be 0 to 100 feet (0-30 m).

The Asotin Member overlies the Wahluke flow and is also restricted to the northern part of the Basin. The Asotin is represented by the Huntzinger flow, which has a total thickness of 0 to 200 feet (0-61 m) within the Basin.

Overlying the Umatilla and Asotin Members in the Cold Creek syncline area of the Basin is the Cold Creek interbed. The interbed is a tuffaceous to arkosic sandstone, with a conglomerate facies near the middle of the unit along the northern edge of the Cold Creek syncline. Total thickness varies from 0 to 100 feet (0-30 m), with the thickest portion near the axis of the syncline and thinning to zero thickness near the Rattlesnake Hills and Gable Mountain.

The next youngest basalt unit is the Esquatzel Member. One to two flows are present, with distribution restricted to the central part of the Basin. Where two flows are present, 2 to 6 feet (1-2 m) of vitric tuff locally occurs between them. Total thickness of the basalt ranges from 0 to more than 100 feet (0->30 m), with 119 feet (36 m) reported in borehole DC-4. Bond and others (1978) report an estimate of 12 m (39 feet) of Esquatzel basalt in the Horse Heaven Hills, while Myers and others (1979) show the Esquatzel pinching out north of the Horse Heaven Hills (see Myers and others, 1979, Figure III-21).

Above the Esquatzel Member is the Selah interbed, a vitric tuff, the top of which has been locally fused by overlying basalts. The Selah is present in most of the Basin except for the eastern and western edges, with the area of greatest thickness occurring near the axis

of the Cold Creek syncline. Myers and others (1979) on their Figure III-22 imply that the Selah does not extend south of the Rattlesnake Hills. However, Bond and others (1978) report 12.5 m (41 feet) in the Horse Heaven Hills. Total thickness within the Basin is 0 to 75 feet (0-23 m).

The Pomona Member Basalt overlies the Selah. In most of the Basin, the Pomona is only one flow. A second flow is present along the eastern portion of the Yakima River and between Gable Mountain and the Saddle Mountains. Myers and others (1979) mention a thin third flow found only in core holes on Gable Mountain. However, in a description of the geology of the Gable Mountain - Gable Butte area, Fecht (1978) does not mention this third flow. Where two flows are present, they are separated locally by a tuffaceous interbed up to 0.6 m (2 feet) thick (Fecht, 1978).

The lower flow of the Pomona (disregarding the possibility of a third flow) is found throughout the Pasco Basin except for the western portion of Umtanum Ridge and a small part of the eastern end of the Horse Heaven Hills. The flow is thinner across the crest of the Rattlesnake Hills. The upper flow is more restricted, as described in the previous paragraph.

Myers and others (1979) give total thicknesses as follows: lower flow 0 to 200 feet (0-61 m), upper flow 0 to 65 feet (0-20 m). The thickest section described for the lower unit by Myers and others (1979) is 192 feet (59 m) in borehole DC-1, while Fecht (1978) describes only 11.6 m (38 feet) of lower flow in DC-11 on Gable Mountain. The upper flow on the western part of Gable Mountain is assigned a thickness of approximately 44 m (144 feet) by Fecht (1978), a significantly thicker section than suggested by the range given by Myers and others (1979). The discrepancy cannot be resolved at this time.

Above the Pomona is the Rattlesnake Ridge interbed. The interbed is found throughout most of the Basin, but is absent on Gable Butte and Umtanum Ridge, and is reported to pinch out on top of Rattlesnake Mountain (Bond and others, 1978). Total thickness varies from 0 to 113 feet (0-34 m), with the thickest section near the axis of the Cold Creek syncline south of Gable Butte. In the northern part of the syncline, the interbed may contain up to 50 feet (15 m) of basalt cobble

conglomerate. More typical lithology is tuffaceous siltstone, with some arkosic sands. Fecht (1978) reports an upper silty vitric tuff in the Gable Mountain area.

The next youngest basalt unit is the Elephant Mountain Member. It consists of two flows in much of the Basin. The lower flow is the more extensive of the two, being absent only from the tops of the Saddle Mountains and Umtanum Ridge. Total thickness of the lower flow ranges from 0 to 115 feet (0-35 m), with thinner sections over the present topographic highs.

The upper flow of the Elephant Mountain Member has a somewhat more restricted distribution. It is absent over most of the topographic highs and much of the western part of the Basin. Total thickness varies from 0 to 66 feet (0-20 m). The thickness of the two flows together ranges from 0 to 143 feet (0-44 m) as reported for a number of boreholes in Myers and others (1979).

In the southern part of the Basin, the Elephant Mountain Member is overlain by the Levey interbed. Bond and others (1978) report a thickness of 6 m (20 feet) on the southwest flank of Red Mountain (inside the horn of the Yakima River) with thinning in every direction to less than 1 m (3 feet) within 2.5 km (4 miles). However, Myers and others (1979) report 15 feet (5 m) in borehole DDH-3. The Levey is tuffaceous silt or siltstone locally overlain by volcanic ash.

The youngest basalts in the Pasco Basin belong to the Ice Harbor Member, which is found only in the southeastern part of the Basin. Three thickness values are reported: Myers and others (1979) have 98 feet (30 m) in borehole DDH-3, while Bond and others (1978) report 9.5 m (31 feet) in the horn of the Yakima River and more than 25 m (82 feet) on a knoll southwest of West Richland.

Above the Columbia River basalts is the Ringold Formation, a thick sequence of sediments divisible into four major units. The basal unit is a gravel with coarse- to fine-grained sand matrix and lenses of fine-grained sand and silt. The unit is found in the Cold Creek and Pasco synclines, with thicknesses ranging from 0 to 160 feet (0-49 m) in the Cold Creek syncline (Routson and Fecht, 1979) with much thinner sections in the Pasco syncline (Myers and others, 1979).

The lower Ringold is composed of silty medium- to coarse-grained sand to sandy silt with frequent 0-1 foot (0-0.3 m) stringers of fine pebbles. The unit is present through most of the central part of the Basin, pinching out against the topographic highs. Thicknesses range from 0 to more than 350 feet (0 - >107 m) (Tallman and others, 1979).

The middle Ringold is a silty sandy gravel, very similar to the basal Ringold. The distribution is similar to the basal Ringold, which underlies the middle Ringold in some places. Sand and silt lenses are common, with a few having thicknesses up to 15 feet (5 m). Total thickness of this unit ranges from 0 to 350 feet (0-107 m) (Tallman and others, 1979).

The upper Ringold is composed of well-sorted sand and silt with minor pebble lenses. The unit is thin or absent in much of the basin, but is more than 400 feet (122 m) thick at White Bluffs along the Columbia River (Tallman and others, 1979).

Tallman and others (1979) report that the overall thickness of the Ringold Formation is up to 1200 feet (366 m).

Some fanglomerates found along the flanks of the anticlinal ridges within the Basin are judged to be time equivalents of the Ringold Formation (Myers and others, 1979). The unit is discontinuous, with maximum thicknesses of more than 50 feet (15 m).

The youngest unit within the Pasco Basin with formation status is the Hanford Formation. The lower part of the formation is the Pasco Gravel, which is up to 350 feet (107 m) thick in parts of the central Pasco Basin. Overlying the gravels are the Touchet Beds, a group of silts to fine sands with stringers of coarse sand and gravel. These beds reach thicknesses of more than 25 feet (>8 m).

Overlying all of the units described previously are various types of recent sediments, including sand dunes, alluvium, talus, loess and landslide deposits. All of these sediments have limited distributions.

2.1.2 Geologic Structure

The Saddle Mountains structure along the northern boundary of the basin (Figure 2-1) is an asymmetric east-plunging anticline, with a nearly vertical northern limb. Minor folds and faults are present on the limbs of the anticline. The central portion of the structure is a tight boxfold, whereas the western and eastern parts are more open. The axial trace follows a sinuous east-west path, and bifurcates in the western segment. Structural relief is nearly 2000 feet (610 meters) in the west, and diminishes to about 500 feet (152 meters) in the east, where the anticline is covered by the Ringold Formation. To the north of the western part of the anticline is the Saddle Mountains fault. The fault has been interpreted as a high-angle reverse fault, which becomes a low angle thrust farther west (see Myers and others, 1979 for a discussion of the interpretations of the fault). Maximum stratigraphic displacement is estimated at 1800 feet (549 meters). Along the Saddle Mountains structure are outcrops of the Saddle Mountains and Wanapum Formations and the upper part of the Sentinel Bluffs sequence of the Grande Ronde Formation. The Schwana sequence of the Grande Ronde outcrops in a small region in Sentinel Gap where the Columbia River cuts through the Saddle Mountains anticline.

The Umtanum Ridge - Gable Mountain structure extends east-west across the west central portion of the basin. Umtanum Ridge forms the western part of this structure, and is an east plunging (5°) asymmetric noncylindrical anticline with a vertical to slightly overturned northern limb. The southern limb dips uniformly from 5° to 12° . Structural relief does not exceed 2900 feet (884 meters). Imbricate thrust faults occur in the hinge area. This thrust zone causes discordance in the stratigraphic sequence in the northern limb relative to the southern limb. Gable Mountain and Gable Butte form the eastern extension of Umtanum Ridge. The major anticlinal hinge of the structure here contains minor folds which trend approximately west and northwest. The southern limb of the anticline has a shallow dip (2°) to the south, whereas the northern limb dips somewhat more steeply (11°) to the north.

Two faults are located in the Gable Mountain structure. At the west end is a north-trending, reverse fault dipping 30° to the east along which 75 feet (23 meters)

of displacement has occurred. A northeast trending escarpment marks the second fault in the central portion of the mountain. Undeformed sediments of greater than 40,000 years of age overlie the fault. Rocks of the Wanapum and Grande Ronde Formations outcrop in the Umtanum Ridge near the western margin of the basin. In the Gable Mountain - Gable Butte structures, only the Saddle Mountains Formation outcrops.

The Umtanum Ridge - Gable Mountain structure is bordered on the north by the Wahluke Syncline, and on the south by the asymmetrical Cold Creek Syncline. The southern limb of the Cold Creek Syncline dips less than 40° to the north, while the northern limb dips less than 10° to the south. This syncline plunges to the east at 3° to 5° .

South of the Cold Creek Syncline is the Yakima Ridge structure, which is a series of ridges with the greatest development lying to the west of the basin. The ridges contain both anticlines and monoclines, and are cut by a number of faults. The eastern part of the Yakima structure is comprised of small, en echelon, doubly plunging, somewhat asymmetric anticlines and synclines. Magnetic anomalies indicate that the structure continues into the subsurface in the central part of the basin.

South of the Yakima structure, and separated from it by a broad asymmetric syncline, is the Rattlesnake Hills structure, which trends northwest-southeast and extends past the basin in either direction. The western part of the structure, the Rattlesnake Hills segment, consists of two major asymmetrical anticlines and associated minor folds and faults. The more northerly of the two anticlines is northeast-trending with a 15° to 45° northerly dipping northern limb and a 5° to 23° southerly dipping southern limb, and plunges gently towards the basin. The anticline just to the southeast, the Rattlesnake Hills anticline, trends northwest and has a northern limb dipping 12° to 15° to the north, and a southern limb dipping 3° to 16° to the south.

The southeastern portion of the Rattlesnake Hills structure has a general northwest trend and consists of doubly plunging, locally faulted, en echelon anticlines. Rattlesnake Mountain, the northeastern portion of this segment, is bounded on the northeast by the Rattlesnake Mountain fault, which is tentatively identified as a reverse fault (Myers and others, 1979).

The northwest-trending fault zone is about 328 feet (100 meters) wide and about 4 miles (6.4 km) long and dies out to the southeast.

Only rocks of the Saddle Mountains and upper Wanapum Formations are exposed in the Rattlesnake Hills structure, except in the Rattlesnake Mountain fault zone where the Frenchman Springs Member is exposed.

The Horse Heaven Hills structure occurs south of the Rattlesnake Hills structure, and forms the southern boundary of the Pasco Basin. The western segment of the structure trends northeast, and the eastern segment trends northwest. Wanapum and Saddle Mountains basalt are exposed in this structure, with Grande Ronde basalt outcropping in a small area of the Wallula Gap and in the Horse Heaven Hills along the southeastern and southwestern boundaries of the basin. The northwest-trending segment is composed of a number of sub-parallel asymmetrical anticlines, monoclines, and faults, whereas the northeast-trending segment is formed primarily of faults, with some folding probably being involved. The intersection of the two segments of the Horse Heaven Hills structure is an area of complex faulting and folding.

A southwest-dipping (1°) homocline forms the eastern structural boundary of the Pasco Basin. This homocline contains low amplitude folds having a north-northwesterly trend. No large faults have been found in the area. A northwest-trending dike system is present in the eastern basin, and represents the vent system for the Ice Harbor flows (Myers and others, 1979).

The central part of the Pasco Basin mainly is sediment covered, and structural interpretation is difficult. Borehole data suggest a continuation of the surface structure to depth, at least for Saddle Mountains and Wanapum basalts. The major structural trends are east-west in the northern part of the basin, and northwest-southeast in the western and southern portions. A possible interpretation of magnetotelluric surveys indicates that the deep subsurface structure, i.e., below the Columbia River Basalts, has a predominant northeast trend, presumably dating back to an earlier tectonic phase distinct from that which caused the present structure.

The broad and low amplitude Pasco Syncline occurs in the southeast part of the basin. North of this syncline are several small domes that are aligned parallel to the east-west segment of the syncline. Some aeromagnetic anomalies are visible elsewhere in the basin, but structural interpretation has not been completed. Other dome-like structures may be hidden beneath the sediments thereby accounting for these anomalies.

The anticlines exposed along the basin margins in general are broad, gentle folds. These folds are slightly asymmetrical with the northern limbs of the anticlines having steeper dips than the southern limbs. This orientation suggests a northerly component of movement. The east-west trending faults found in the region occur along the northern limbs of anticlines, and if these faults are reverse faults, a north-south compressional scheme would be indicated.

2.1.3 Spatial Variation in Thickness of Stratigraphic Units

Isopach maps of stratigraphic units in the Pasco Basin are presented in Myers and others (1979). These maps show the distribution and thickness of the various interbeds within the Saddle Mountains basalts but do not depict the distribution or thickness of individual basalt flows, except in the rare instance where member-rank stratigraphic units consist of a single flow. As a result, the continuity of individual flows or their associated interflow zones on a basin-wide scale cannot be determined. In the center of the basin individual flows can be traced on a local scale using geophysical logs.

The isopach maps show a tendency for the Grande Ronde and Wanapum Formations to be thickest in the southwestern parts of the basin and to become thinner to the north and east away from the region of the Rattlesnake Hills. The Saddle Mountains Formation and its members, in contrast, tend to be thickest in the west-central part of the basin and to become thinner toward the basin perimeter. Sedimentary interbeds between basalt flows in the Saddle Mountains Formation also tend to be thickest in the center of the basin and to become thinner toward the basin margin. The Cold Creek and Selah interbeds probably pinch out within the basin and have no significant outcrops in the surrounding hills. The Mabton and Rattlesnake

Ridge interbeds both are thinner along the basin periphery, but have outcrops in the Rattlesnake Hills and Saddle Mountains.

2.1.4 Intraflow Structures in the Grande Ronde Basalt

Grande Ronde basalt is primarily aphyric, with tiny microphenocrysts of plagioclase, and has intergranular to intersertal textures. The predominant phases are plagioclase (An_{48} to An_{65}), clinopyroxene (augite, pigeonite), and glass, with titanomagnetite and ilmenite as accessory minerals, and rare orthopyroxene and olivine. Alteration products include nontronite and other smectites, palagonite, calcite, opal, chlorophaeite, clinoptilolite, thompsonite and other zeolites.

Drilling operations within the Pasco Basin have determined the Grande Ronde Formation to be in excess of 3000 feet (914 meters) thick, and to be divisible into at least 35 flows. Flow thickness ranges from 12 to 450 feet (3.6 to 137 meters), with an average of 102 feet (31 meters). Intraflow structures form within flows during emplacement and cooling, and include fractures, vesicles, and flow top breccia. Vesicles typically occur in the upper 1/4 to 1/3 of a flow, with vesicularity increasing towards the flow top, and a thin vesicular layer may also be present at the bottom of a flow (Myers and others, 1979). A thin zone of great vesicularity and rubbly flow top breccia most commonly marks the top of a flow. Cooling fractures commonly create elongate columns bounded by fracture surfaces. Each flow may contain one or more of the following subunits (Figure 2-3): lower colonnade, upper colonnade, and entablature. The colonnades are comprised of relatively large columns, while the entablature consists of smaller, less regular columns (Long, 1978).

Detailed observations of the intraflow structures in the Grande Ronde Basalt were made at Sentinel Gap, in the Saddle Mountains (Long, 1978). Long (1978) divided intraflow structure occurrences into three basic types. The first type, occurring in relatively thin flows (35 to 100 feet [10 to 30 meters]), contains large (3 to 6 feet [1 to 2 meters] diameter) irregular, tapering columns. These columns are four to five sided and generally occur across the entire thickness of the flow.

Figure 2-3. Structural Features of a Basaltic Lava Flow
(after Atlantic Richfield Hanford Company, 1976).

Vesicle cylinders occur from 0.5 feet (0.2 meters) above the flow base to about mid-flow. Vesicles also occur in the upper 1/3 of the flow, with larger diameter vesicles in the thin flow top.

Type two intraflow structures consist of thick flows (150 to 250 feet, [46 to 76 meters]) that often exhibit a multiple column structure in the entablature region of the flow. At the base of the flow is a thinner colonnade portion that contains stubby tiered columns. A clinker flow top is common and vesicles are abundant in the upper 1/3 of the flow. Local fanning of columns is noted. Knobby, platy zones of basalt, formed by closely spaced parallel fractures at high angle to columns, occur between column tiers. In the upper portion where the vesicles first occur, the entablature breaks down into areas of "brickbat" jointing. Pillow basalt locally occurs in the thick flow top.

Type three includes moderately thick flows 95 to 200 feet (29 to 61 meters), with a well-defined break between the colonnade and entablature characterized by a change in fracture abundance and column size. The entablature is characterized by smaller columns and more abundant fractures. The lower colonnade commonly exhibits pinch-and-swell structure, where the fractures alternately curve toward and away from the column center, and the entablature is formed of multiple radiating columns. A poorly developed upper colonnade often is included.

Lateral gradation between these three types of flow morphology does not appear to occur, although morphologies intermediate between these types have been noted (Myers and others, 1979).

Fractures encountered during drilling are generally of narrow aperture (<0.05 inches [<1 mm]) and a large proportion (83 percent) are at least partially filled with clay (Long, 1978). Unfilled fractures generally are less than 0.02 inches (0.5 mm), and there are rare occurrences of unfilled fractures of 0.16 to 0.4 inches (4 to 10 mm) aperture (Long, 1978). A comparison of fracture frequency with core depth indicates a possible decrease in frequency near the base of flows (Long, 1978).

Basalt texture generally is correlative with position in the flow. The percentage of glass and the number and size of crystallites increases in the entablature, relative to the colonnade (Long, 1978).

In the Umtanum flow, both an entablature and a lower colonnade generally are present. A 25 foot (7.6 meters) thick upper colonnade has been inferred in core hole DH-5, which is located south of the Saddle Mountains, approximately half way between Sentinel Gap and the northwestern boundary of Hanford Reservation. Flow-top breccia of the flow has a maximum thickness of more than 100 feet (30 meters). The upper 30 to 75 feet (9 to 23 meters) of the flow, including any flow-top breccia, are vesicular. According to the classification scheme of Long (1978), the Umtanum flow probably is a type three flow (Myers and others, 1979).

2.2 Hydrogeology

The paucity of hydrologic data in the Pasco Basin for units beneath the Mabton interbed results in a poorly defined ground-water flow system. In an attempt to supplement the available ground-water data, water budget calculations based on precipitation, evapotranspiration, and runoff can be used. Each of these three factors in the water budget can either be calculated or measured, but inherent errors in each determination may be large enough to invalidate the entire method.

A study in the central part of the Pasco Basin (Jones, 1978) determined that precipitation to the surface did not penetrate the soil to a depth greater than 39 feet (12 meters) and that this depth of penetration did not significantly change over the length of the year. This limited penetration indicates that the recharge for the underlying basalts and interbeds must either originate in the hills surrounding the Pasco Basin or originate outside of the basin and enter the basin by interbasin flow. Based on regional maps of evapotranspiration and precipitation (Gephart and others, 1979; Leonhart, 1979; Tanaka and others, 1979), precipitation in the recharge areas surrounding the Pasco Basin is not sufficient to supply all of the water in the ground-water system, thereby necessitating an influx of ground water from adjacent basins. By this method, only the Mabton Interbed and younger units comprise a local flow system confined to the Pasco Basin (Gephart and others, 1979). Units older

than the Mabton are part of a regional system. A problem with using regional maps is that the scale of the maps does not account for regionally small-scale but locally important increases in precipitation and decreases in evapotranspiration with increasing altitude (Summers and others, 1978). In areas bordering the Pasco Basin where the altitude is in excess of 3000 feet, (915 meters) precipitation exceeds evapotranspiration and recharge is possible (Summers and others, 1978). The amount of this local recharge relative to the amount of ground water in the system is uncertain, and as a result, regional interbasin ground-water flow can neither be excluded from consideration nor used as the sole source of ground water for any of the units.

Other problems in calculating a water budget also contribute to the uncertainties of the method. Evapotranspiration is calculated by techniques (Thornwaite and Mather, 1957) that may be inaccurate (Freeze and Cherry, 1979). Errors inherent to the techniques for determining stream discharge may be sizable under the most favorable conditions (Smoot and Novak, 1969). The error in stream discharge calculations for the Pasco Basin (Leonhart, 1979) is in the range of 5 to 15 percent. More accuracy in the water budget could be obtained if a gauging station existed at Sentinel Gap where possible recharge to deep aquifers may be occurring. Statistically meaningful data would require too much time to collect for such a station to be established and to contribute to the evaluation of the Pasco Basin as a possible repository site.

In addition to the water budget of the basin, the structural relief in the region may be an indication as to whether interbasin flow is possible. The existence of laterally widespread aquifers in an area, even in the presence of moderate structural relief, may allow regional ground-water flow (Freeze and Witherspoon, 1966, 1967, 1968). Widespread aquifers exist in the Columbia Plateau, including the Pasco Basin. Major faulting has taken place in the Rattlesnake Mountains on the southwestern edge of the basin and in the Saddle Mountains on the north of the basin (Tanaka and others, 1979; Myers and others, 1979). Whether, and to what degree, faulting in these regions has juxtaposed high- and low-permeability units to the extent that the potential pathways for regional flow are disrupted is not known.

Most of the holes drilled in the Pasco Basin and outside of the Hanford site are water wells. With the relatively abundant water supply of the unconfined aquifers, no need generally existed to drill into the basalts or interbeds or to measure all of the hydraulic properties of the units penetrated, especially the basalts. Approximately one-third of the field-checked drill holes outside of the Hanford site either could not be found or were mislocated in the records (Summers and others, 1978). The nonuniform distribution of drill holes in the basin, the inaccurately recorded location and elevation of many drill holes resulting in inaccurate head measurements, poor or meager descriptions of testing methods, and poorly chosen sample intervals in the drill holes result in considerable uncertainty in using this information in any description of the flow system.

In the Pasco Basin, the water table has undergone considerable change in recent times as the result of extensive irrigation north and northeast of the Columbia River and the construction of waste ponds in the Hanford site. Although the undisturbed water table in the Hanford site was not recorded, data collected from the slightly disturbed water table between 1948 and 1952 has been extrapolated backward in time to construct an approximately undisturbed system (Figure 2-4) (Gephart and others, 1979). The changes superimposed on the water table as of 1978 are shown in Figure 2-5.

A paleochannel of the Columbia River is present between Gable Butte and Gable Mountain and extends to the southeast (Ledgerwood and Deju, 1976). This channel was incised into the Saddle Mountains Formation to an undetermined depth. The 20 foot contour line including and extending to the southeast of B-Pond in Figure 2-5 and the flexure in the 380 foot contour line to the south of Gable Mountain in Figure 2-4 are indications of the presence of somewhat different sediments associated with the paleochannel. Ground water flow in the unconfined aquifers (Hanford and Ringold Formations) and the upper portions of the Saddle Mountains basalts will be affected by the presence of these channel deposits.

In the stratigraphic sequence beneath the Ringold Formation, only the Mabton interbed has sufficient data to allow the construction of an equipotential surface for most of the Hanford site (Figure 2-6). A problem with

Figure 2-4. Estimated 1944 Water Table Map for Hanford Site, Pasco Basin, Washington (Gephart and others, 1979).

Figure 2-5. Water Table Rise Beneath the Hanford Site, 1944-1978 (Rockwell Hanford Operations, 1978).

Figure 2-6. Water Potentials for the Mabton Interbed, Hanford Site (Rockwell Hanford Operations, 1981).

this flow pattern is that recharge from the Rattlesnake Hills is not taken into account. In a regional map of the Saddle Mountains Formation, Tanaka and others (1979, Plate 6) indicate that the Rattlesnake Hills are a recharge area for this unit in the Pasco Basin. Assuming that the Mabton interbed receives at least some recharge from the Rattlesnake Hills, the potentiometric contours should curve and become asymptotic to the Rattlesnake Hills in the vicinity of Yakima River.

The flow in the basalts overlying the Mabton interbed will not have a small-scale flow pattern identical to that of the Mabton. Differences are caused by overlapping lava flows resulting in the truncation of interbeds and interflows, thereby creating a more tortuous flow path in the basalts than in the comparatively homogeneous sediments of the Mabton. On a larger scale, at least the Mabton and the Saddle Mountains Formation are part of a local flow system and therefore should have a similar flow pattern (Gephart and others, 1979). Flow in the upper portion of the Saddle Mountains Formation will be somewhat disrupted because of the Columbia River paleochannel incised into the unit, but the amount of this disruption is uncertain.

An area of uncertain flow characteristics exists in the Hanford site that may have a considerable effect on the ground-water flow. This area is a zone of folding and faulting extending along and including Gable Butte and Gable Mountain. If the head value for the Mabton interbed reported for the location along the Columbia River north of Gable Mountain (Figure 2-6) is accurate, the ground-water flow in at least the Saddle Mountains basalts and the Mabton interbed converges from both north and south on the Gable Butte - Gable Mountain zone. Faults and fractures in this zone must act as vertical conduits for the converging water. Once the water reaches this zone three possible flow paths exist:

1. The ground water flows upward into the Ringold and possibly the Hanford Formations where the flow is then to the Columbia River.
2. The ground water flows downward and is a source of recharge for the Wanapum, Grande Ronde, and possibly pre-Grande Ronde rocks.
3. A combination of 1 and 2.

No data are available to support or discredit any of these three flow schemes. Exclusively upward flow in this zone would result only if the head values for the units below the Mabton interbed equaled or exceeded the head values of the Saddle Mountains Formation. In this case, the Gable Butte - Gable Mountain zone also would be a discharge area for the Wanapum and Grande Ronde Formations. For exclusively downward flow, the Gable Butte - Gable Mountain zone would be a recharge area for the lower formations and interbeds. Upon reaching these lower units, the ground-water flow would be to the north and south away from this zone. If these lower units are part of a regional flow system, the water moving away from this zone would be diverted to the east or southeast as part of the regional flow toward Lake Wallula. If the lower units are part of the local flow system, the water from this recharge area would flow outward (to the north and south in response to the higher pressures along this east-west zone) until an equal pressure exerted by water from another recharge area (such as Rattlesnake Hills) is reached, at which point the flow direction would be in the direction of the local or regional ground-water flow system, whichever flow scheme controls the flow in the lower units. The result of this latter type of flow scheme is that the direction of ground-water flow in the units beneath the Mabton interbed immediately south of the Gable Butte - Gable Mountain zone would be nearly opposite to the flow direction in the Mabton and overlying Saddle Mountains Formation.

Both upward and downward flow is possible in the Gable Butte - Gable Mountain zone. Interbeds and interflows in the Saddle Mountains Formation may have a higher transmissivity than the Wanapum and Grande Ronde Formations. As long as the amount of ground water supplied to the Gable Butte - Gable Mountain zone is greater than the amount that can be transmitted downward along faults and fractures and into the lower units, the water table of the aquifer(s) supplying the water will increase until the hydraulic head is greater than that of the overlying units. This greater hydraulic head will result in upward flow in addition to the downward flow.

As part of the local flow system, the portion of the ground water that remains in the Saddle Mountains Formation and the Mabton interbed is inferred to discharge into the Columbia River and Lake Wallula (Gephart

and others, 1979). The discharge areas for the units below the Mabton are less certain. Based on the assumed existence of a regional flow system, those units below the Mabton could discharge into Lake Wallula by any one of three mechanisms (Gephart and others, 1979):

1. The Wallula Gap fault, which is located beneath Lake Wallula, is a zone of high vertical hydraulic conductivity, thus providing a pathway for the ground water to reach the lake.
2. The ground water from both the Wanapum and Grande Ronde Formations leaks into the Saddle Mountains Formation through fractures, and the Saddle Mountains discharges into the lake, and
3. Both the Wanapum and the Grande Ronde Formations discharge directly into Lake Wallula.

Direct discharge into the lake probably is of little importance considering the limited outcrop area of these two formations in the Lake Wallula area. Extensive leakage along fractures provides a mechanism by which water from the lower units gains access to the Saddle Mountains Formation, which has extensive exposure under and around Lake Wallula. As in the Gable Butte - Gable Mountain zone where faults probably are conduits for vertical ground-water movement, a fault under Lake Wallula could provide the necessary route for discharge of the Wanapum and Grande Ronde ground water into the lake, although the existence of a fault at this location has been questioned by Brown (1968).

To have both the Wanapum and Grande Ronde Formation as part of a local flow system within the Pasco Basin requires recharge in the hills surrounding the basin and discharge into the Columbia River. The area of recharge for these units is uncertain, and for discharge into the river, considerable upward leakage through fractures beneath the river must occur. In areas where the hydraulic heads indicate a tendency for upward flow, water will migrate from the Wanapum and Grande Ronde Formations into the Saddle Mountains Formation along fractures. The amount of water migrating along the fractures is dependent upon the head potential, the fracture density, and the aperture

of the fractures. Because the rocks have an inherent vertical interstitial permeability, some water movement will occur in the presence of an upward hydraulic gradient even if the fractures are completely sealed. Artesian flow at well DC-6, which is along the river, indicates potential for flow rather than proving extensive upward ground-water migration. Although the location of DC-6 is to the north of the Gable Butte - Gable Mountain zone, and as such does not necessarily indicate potential upward flow along the river to the south of this zone, a major zone of upward ground-water migration along the length of the Columbia River cannot be eliminated from consideration.

The flow system for the Pasco Basin is not known at this time. Flow may be relatively simple in that all of the units have potentiometric surfaces similar to that of the Mabton interbed (Figure 2-6). In this instance, the water that reaches the Gable Butte - Gable Mountain zone would flow upward along fault zones and fractures to the unconfined aquifers, and then would discharge into the Columbia River. Ground water not reaching this zone would discharge into the Columbia River by way of fractures beneath the river. A much more complicated flow scheme would result if the confined aquifers above and including the Mabton interbed have a different potentiometric surfaces than those units beneath the Mabton. The Gable Butte - Gable Mountain zone could act as a recharge area for the lower units receiving water from the Saddle Mountains Formation. Discharge for the lower units could be partly into the Columbia River by way of leakage into the Saddle Mountains Formation and partly into Lake Wallula. Any gradation between these two schemes could be occurring in the Pasco Basin.

2.2.1 Hydraulic Head

In addition to indicating the tendency for horizontal ground-water flow direction within particular units, hydraulic head values can be used to determine direction and magnitude of the potential for vertical flow. Problems in interpretation occur when the measurements are made before equilibrium in the test interval is reached or the gradient between test intervals is less than the experimental error of the measurements.

For drill holes DC-1 and DC-2 (Figure 2-7), the data indicate a decrease in head with depth for at least the Schwana member of the Grande Ronde Formation (Figures 2-8A, 2-8B, and 2-9). Above the Schwana Member, the head values have no consistent pattern and the reliability of the measurements is questionable, especially the DC-1 results listed by Summers and others (1978) (Figure 2-8A). A similar pattern has been recorded for DC-6 (Figure 2-10), although other reported results are considerably less reliable (Figure 2-11). Drill hole DC-8 has head measurements for a relatively small interval in the upper half of the Wanapum Formation. Although the change in head values with depth suggest a slight downward gradient (Figure 2-12), the variation in values could be the result of experimental error.

More recent experiments have identified the stratigraphic intervals being tested, and the reported hydraulic heads seem to be less erratic than in earlier tests. In DC-12 (Figure 2-13), measurements through the Wanapum Formation and into the top of the Grande Ronde Formation indicate a uniform hydraulic head throughout the interval. This uniformity suggests lateral flow only, although a slight upward or downward gradient within the experimental error cannot be discounted.

Drill hole DC-14 (Figure 2-14) indicates a marked upward flow gradient from the upper portion of the Wanapum Formation to the upper portion of the Saddle Mountains Formation. The gradient is uniform to slightly downward through the Wanapum and into the Grande Ronde Formation. Head values for DC-15 (Figure 2-15) are more erratic in the Saddle Mountains Formation, but are nearly uniform through the Wanapum Formation. In the Grande Ronde Formation, the values are uniform within experimental error.

In the Pasco Basin, the vertical direction of hydraulic head gradient tends to be related to depth. Head values in the Saddle Mountain Formation tend to either be somewhat erratic or increase with depth. Within the Wanapum Formation, the values are either uniform or decrease with depth. The Grande Ronde Formation has head values that decrease with depth. Only in DC-15 do the values for the Grande Ronde remain uniform to slightly increase with depth, although this

Figure 2-7. Location of Selected Drill Holes in Pasco Basin (Deju, 1980e).

Figure 2-8A. Hydraulic Head Measurements Within Wanapum and Grande Ronde Formations at Drill Hole DC-1 (Summers, Weber and Schwab, 1978).

Figure 2-8B. Hydraulic Head Measurements Within Wanapum and Grande Ronde Formations at Drill Hole DC-1 (Deju and Fecht, 1979).

Figure 2-9. Hydraulic Head Measurements Within Grande Ronde Formation at Drill Hole DC-2 (Deju and Fecht, 1979).

Figure 2-10. Hydraulic Head Measurements Within Grande Ronde Formation at Drill Hole DC-6 (Deju and Fecht, 1979).

Figure 2-11. Hydraulic Head Measurements Within Wanapum and Grande Ronde Formations at Drill Hole DC-6 (Summers, Weber and Schwab, 1978).

Figure 2-12. Hydraulic Head Measurements Within Wanapum and Grande Ronde Formations at Drill Hole DC-8 (Deju and Fecht, 1979).

Figure 2-13. Hydraulic Head Measurements Within Wanapum and Grande Ronde Formations at Drill Hole DC-12 (Deju, 1980c).

Figure 2-14. Hydraulic Head Measurements Within the Saddle Mountains and Wanapum Formations at Drill Hole DC-14 (Deju, 1980d).

Figure 2-15. Hydraulic Head Measurements Within Saddle Mountains, Wanapum, and Grande Ronde Formations at Drill Hole DC-15 (Deju, 1981).

increase is within experimental error. Based on the available data, radionuclides that may be released from a subsurface facility located in the Umtanum member of the Grande Ronde Formation will remain in this formation or possibly migrate downward. A similar pattern of head data is indicated for the Grande Ronde at the edge of the basin by drill hole RSH-1 (Figure 2-16). Additional evidence suggesting downward flow was discovered by Noonan and others (1981) who found that the flow top breccia of the Umtanum Member in DC-2 is highly altered, whereas the rock beneath the flow top is nearly unaltered and fractures are filled with material that would have been derived from the flow top breccia.

2.2.2 Hydraulic Conductivity

Rocks in the Pasco Basin older than the Ringold Formation generally are divided into three major categories. Dense basalt refers to the interior portion of a lava flow that consists of the colonnade and entablature zones. Because of a low effective porosity, the hydraulic conductivity of the unfractured basalt is low. Fractures formed during the cooling of the basalt are present throughout the flow including the dense interiors. The contribution of the fractures to the conductivity of the rock depends on the fracture density, the degree of interconnection, fracture aperture, and the extent of fracture filling.

Interflows are formed as the top of a flow cools while the flow is moving. The movement causes fracturing and the accumulation of blocks of basalt. Near the front of the flow, the cooling crust is broken-up and pulled down the front of the flow, to be overridden as the flow advances. As a result of this process, lava flows may have highly fractured and brecciated tops and bases in which the fractures provide zones of high hydraulic conductivity. Weathering, secondary mineralization, chemical alteration, and infilling of voids by wind and stream deposition will decrease the conductivity of these zones. The older the flow, the more time is available to fill in the voids. Some interflows have conductivities that are the same or only slightly higher than dense basalt, whereas other interflows act as major aquifers.

Figure 2-16. Hydraulic Head Measurements Within Grande Ronde Formation at Drill Hole RSH-1 (Deju and Fecht, 1979).

Interbeds consist of sediments deposited on top of a lava flow during periods of igneous quiescence. The sediments may be wind-blown, deposited by streams, or deposited in lakes. Grain size is dependent on the energy level of the depositional environment. Lake deposits tend to be fine-grained and conductivities are low. Stream-channel deposits may have high conductivities. The hydraulic conductivity of any sedimentary deposit is dependent upon grain size, grain size distribution, clay content, and degree of cementation. Although variation in both vertical and horizontal conductivity is dependent on the depositional history of each interbed, the vertical conductivity is less than the horizontal. The thickness of the interbed is dependent on the time available for deposition, the presence of source material to be deposited, and the existence of a depositing agent.

One of the factors that limits the usefulness of hydraulic conductivity measurements, especially those measurements prior to about 1978, is the thickness of the zone isolated for testing. Because of the possibly marked differences in the conductivities of the various rock types, isolated drill hole intervals containing a mixture of dense interiors, interflows, and interbeds yield conductivities that are not representative of any of the rock types present. These values often were reported as being for dense basalt with or without mentioning the presence of the other rock types in the measured interval. Other factors that affect conductivity measurements are described in Appendix A.

To date, no measurements of vertical hydraulic conductivity have been reported for any of the stratigraphic units in the Pasco Basin. As a result, only the horizontal conductivities are considered in this report. Table 2-1 lists the stratigraphic units in the basin and indicates a range of conductivity values for each unit. Those units or portions of units for which no measurements are reported are assigned a range of values derived from generic rock type values for each formation as cited in the literature. The data used to compile Table 2-1 are included in Appendix A.

A general decrease in the hydraulic conductivities of dense basalt flow interiors with an increase in depth probably results from a combination of an increase in confining pressure with depth that closes the fractures

Table 2-1. Ranges of Hydraulic Conductivities for Stratigraphic Units in Pasco Basin.

Units		Hydraulic Conductivity (ft/d*)	Representative Hydraulic Conductivities (ft/d*)
Hanford		250 - 20,000	
Ringold			
	Upper	1 - 200	
	Middle	1 - 640	
	Lower	0.1 - 7	
Saddle Mountains Basalt			
Ice Harbor Member	interflow	no data	0.1 - 40
	interior	no data	1×10^{-1} - 1×10^{-2}
Levey Interbed		9.8	
Elephant Mountain Member	interflow	2.83 - 2040	
	interior	no data	1×10^{-1} - 1×10^{-2}
Rattlesnake Ridge Interbed		0.1 - 100	
Pomona Member	interflow	no data	0.1 - 40
	interior	2.83×10^{-2}	
Selah Interbed		1.3 - 23	
Esquatzel Member	interflow	no data	0.1 - 40
	interior	2.83×10^{-3}	
Cold Creek Interbed		1 - 180	
Asotin Member	interflow	no data	0.1 - 40
	interior	no data	1×10^{-1} - 1×10^{-2}

* 1 ft/d = 0.301 m/d.

Table 2.1 (Continued)

Units		Hydraulic Conductivity (ft/d*)	Representative Hydraulic Conductivities (ft/d*)
Wilbur Creek Member	interflow	no data	0.1 - 40
	interior	no data	1×10^{-1} - 1×10^{-2}
Umatilla Member	interflow	no data	0.1 - 40
	interior	2.83×10^{-4}	
Mabton Interbed		1.4 - 158	
Wanapum Basalts			
Interbeds		0.1 - 40	
Priest Rapids Member	interflow	2.83×10^{-4} - 1.3×10^4	
	interior	no data	1×10^{-2} - 1×10^{-3}
Roza Member	interflow	no data	1×10^{-2} - 10
	interior	4.26×10^{-2}	
Frenchman Springs Member	interflow	3×10^{-2} - 3×10^{-6}	
	interior	2.83×10^{-5}	
Vantage Interbed		3.02×10^{-1}	
Grande Ronde Basalts			
Interbeds		3×10^{-3}	
Sentinel Bluffs Member			
Upper	interflow	no data	2.8×10^{-10} - 3.0×10^2
	interior	3×10^{-5} - 3×10^{-8}	

*1 ft/d = 0.301 m/d.

Table 2.1 (Continued)

<u>Units</u>		<u>Hydraulic Conductivity (ft/d*)</u>
Middle	interflow	2×10^{-2} - 3×10^{-7}
	interior	3×10^{-3} - 3×10^{-8}
Lower	interflow	3×10^{-1} - 3×10^{-6}
	interior	3×10^{-5} - 3×10^{-7}
Schwana Basalts		
Umtanum	interflow	3×10^{-3} - 3×10^{-7}
	interior	1.3×10^{-5} - 2.8×10^{-9}
Low K ₂ O Flow	interflow	2.0×10^{-1}
	interior	2.8×10^{-5} - 2.8×10^{-10}
Below Low K ₂ O Flow	combined interflows	5.8×10^{-3} - 2.7×10^{-4}
	and interiors	
Pre-Grande Ronde Basalts	combined interflows	5.3×10^{-2} - 2.7×10^{-5}
	and interiors	

* $1 \text{ ft/d} = 0.301 \text{ m/d}$.

and the older basalt having had a longer time for chemical alteration and fracture filling. The interflows do not have this same trend because most of the processes that fill the voids take place while the zones are exposed at the surface, and the subsurface processes take longer to fill in the larger voids present in the interflows.

2.3 Geochemistry

As stated in the Introduction, a principal objective of this report is to compile a reliable data base which contains the information and parameters necessary for input to the risk assessment methodology for radioactive waste disposal by the Nuclear Regulatory Commission. Currently the methodology uses geochemical information in two ways: (1) to determine maximum aqueous concentrations in leach limited release scenarios, and (2) to calculate transport velocities of waste radionuclides from ground-water velocities via the radionuclide retardation factor.

The retardation factor is a function of the distribution coefficient K_d . The use of K_d alone to model radionuclide migration in ground water has been shown to be simplistic (see review by Reardon, 1981), especially in fractured media. A discussion of K_d is presented in Appendix N, along with a discussion of the volumetric utilization factor, this factor makes the use of K_d in the transition from porous to fractured media somewhat more plausible.

At present the methodology is unable to deal with the more complex aqueous/solid interactions which contribute to the retardation of radionuclides in basalt. It is for this reason that individual retardation mechanisms are not dealt with separately in this report. The most significant retardation mechanisms in basalts are ion exchange between ground water and secondary minerals and diffusion into the solid phase. It is anticipated that these mechanisms soon will be incorporated into the methodology, thus permitting a more realistic treatment of radionuclide transport.

2.3.1 Ranges of Distribution Coefficient (K_d)

The ranges of K_d values in the Hanford Reservation basalts are summarized under three cases in Table 2-2. The cases are:

1. Fresh basalt surfaces under reducing conditions expected in the near field once the repository is resaturated after closure.
2. Fresh basalt surfaces under oxidizing conditions, expected in some areas of the intermediate and far field in which ground-water movement is sufficiently rapid to eliminate reducing conditions.
3. Altered basalts, interbed materials and secondary minerals produced by the alteration of basalt along fractures under oxidizing conditions, expected along the principle pathways to the environment.

Although desirable, insufficient data currently exist to prepare a similar set of ranges for the materials in case 3 under reducing conditions. The few points available for reducing conditions fall within the ranges in case 3. The basis for the ranges in Table 2-2 appears in Appendix N, along with descriptions of the ground-water compositions and of the oxidation conditions involved in K_d determinations.

2.3.2 Solubility of Waste Nuclides

Solubility is the mass of a substance contained in aqueous solution that is in chemical equilibrium with an excess of that substance in a solid phase. Solubilities are conventionally expressed as saturation concentrations at a given temperature. The solubility of an element is independent of the isotopes present.* It is obtained from the sum of the stoichiometric concentrations of all dissolved species containing that element. Accurate solubility calculations thus require a knowledge of all species contributing to the observed chemical composition and of the relationship between the concentration of those species and their

*Rigorously, isotopic fractionation effects influence all phase changes, but at high mass numbers these differences are negligible even in theoretical treatments.

Table 2-2. K_d Values for the Basalt Environment at the Hanford Reservation

Geomedium: Conditions: Ground-water:	Fresh Basalt Oxidizing, 60°C GR-2		Fresh Basalt Reducing, Composite °T Composite GW		Altered Basalt, Interbeds and Secondary Minerals Oxidizing, Composite °T GR-1	
	K_d range (ml/g)	median	K_d range (ml/g)	median	K_d range (ml/g)	median
Cm, Am, Pa, Th, Ac, Po, Bi, Sm, Nb, Zr	50-200	100	50-200	100	190-251,000	500
Pu	5-500	25	3-5000	100	110-5000	200
Np	4-80	20	0-450	30	9-180	10
U	3-70	5	38-52	30	12-3030	20
Ra, Pb	50-150	100	90-240	100	50-2000	200
Tc	0-10	0	2.5-4500	50	1-6	1
Sr	60-200	100	100-400	100	230-480	300
Cs	100-1000	300	100-1000	300	550-225,000	2000
I	0	0	0	0	1-4	1
Se	2-10	3	1-20	3	0-10	1

chemical activities. This, in turn, requires a detailed knowledge of the aqueous geochemistry of the element in question. Even then, experimental verification is desirable. Theoretically calculated equilibrium solubility concentrations will tend to be underestimated because some contribution from species not treated always will be present (Garrels and Christ, 1965).

An evaluation of the speciation of an element in solution is essential to the calculation of solubilities as well as for evaluating K_d values (Appendix N). Speciation is a function of pH, Eh, the aqueous chemical activities of the element being considered, the aqueous chemical activities of species of elements with which the given element may combine or compete in solution, temperature and pressure, and the Gibbs free energies and enthalpies of each solid and aqueous specie. In all but the most simple solutions, speciation is cumbersome to calculate. Extensive chemical equilibrium computer codes have been developed for this purpose. The chemical equilibrium assumption made in these codes is not a liability, as it is in some other areas of geochemical interpretation since solubility values are sought, by definition, at equilibrium. Several of the most widely known of these codes recently have been compared and evaluated (Nordstrom and others, 1979).

In the absence of direct solubility measurements, and until detailed calculations can be made, the solubility ranges employed in the characterization of the bedded salt reference repository environment will be used for the basalt environment. Table 2-3 presents a summary of these values (Muller and others, 1981) and of values obtained subsequently for Pb and Pa (Cranwell and others, 1981).

2.3.3 Hydrogeochemical Parameters

The ranges of major and trace elements in solution in the ground waters of the Hanford Site are presented in Appendix N, as is a discussion of the redox conditions. The redox and thermal conditions at the repository level are believed to fall in the ranges:

pH: 9.4 - 10

Eh: -.40 - .55

T: 30° - 50°C

Table 2-3. Ranges of Solubilities of Selected Elements
in Bedded Salt* Repository Site Conditions

<u>Element</u>	<u>Log Molality</u>	<u>Log Mass Fraction</u>
Se	----- No Limit -----	
Sr	- 3.0 $\pm$ 0.5	- 4.1 $\pm$ 0.5
Zr	- 9.6 $\pm$ 0.6	- 10.6 $\pm$ 0.6
Tc	----- Low -----	
Pd	----- Low -----	
I	----- No Limit -----	
Sb	- 8 $\pm$ 2	- 9 $\pm$ 2
Sn	- 9 $\pm$ 2	-10 $\pm$ 2
Cs	----- No Limit -----	
Ra	- 7 $\pm$ 1	- 8 $\pm$ 1
Th	- 6.5 $\pm$ 0.6	- 7.1 $\pm$ 0.6
U	- 4.1 $\pm$ 1	- 4.7 $\pm$ 1
Np	- 15.0 $\pm$ 3	- 15.6 $\pm$ 3
Pu	- 9 $\pm$ 2	- 9.6 $\pm$ 2
Am	----- No Limit -----	
Cm	----- No Limit -----	
Pb	- 6.4 $\pm$ 1	- 7.5 $\pm$ 1
Pa	- 4 $\pm$ 0.6	- 5 $\pm$ 0.6

*See p. 59.

2.4 Other Physical Parameters of Basalt

Table 2-4 contains suggested ranges for a number of physical parameters of basalt. The ranges are derived from data gathered both from Hanford and from other areas of the world. The degree to which suggested data are generic varies among parameters, and depends upon the amount and reliability of data available for Hanford basalts. Further detailed discussion for each parameter can be found in Appendices D through L.

In general, values for the physical parameters of Hanford basalts show wide variation. Although the probable cause of these variations is the inherent spatial variability of the flows themselves, some may be due to inadequate experimental precautions and/or a failure to report all of the relevant experimental conditions. Future data gathering may enable further refinement of the ranges in Table 2-4.

Caution should be used in applying the ranges for parameters in Table 2-4. Except where specifically stated, all ranges are derived from data collected in the laboratory. Extrapolation of these ranges to in-situ conditions in which discontinuities become important is not a simple process. Such an extrapolation requires a combination of laboratory measurements with in-situ measurements and expert judgment. This statement is particularly true for the following parameters: thermal conductivity, thermal expansion, thermal diffusivity, compressive and tensile strengths and elastic constants.

Table 2-4. Suggested Ranges for Physical Parameters
of Hanford Basalts¹

Bulk density (ρ)	1.9 - 3.1 g/cm ³ (118.6 - 193.5 lb/ft ³)
Total porosity (ϕ_T)	0 - 14%
Effective porosity (ϕ_E)	0 - 2.5%
Thermal conductivity (k)	2.0 - 6.0 x 10 ⁻³ cal/cm sec°C (1.34 - 4.03 x 10 ⁻⁴ Btu/ft sec°F)
Specific heat (C_p)	0.175 - 0.280 cal/g °C (0.175 - 0.280 Btu/lb°F)
Thermal expansion (α)	4.0 - 7.4 x 10 ⁻⁶ /°C (2.2 - 4.1 x 10 ⁻⁶ /°F)
Thermal diffusivity (α^*)	$\alpha^* = k/\rho C_p$
Compressive Strength	Intact: 5400 - 60,000 psi (0-455 MPa) Fractured: 0 - 44,000 psi (0-303 MPa)
Tensile Strength	Intact: 1000 - 3500 psi (0-24 MPa) Fractured: 0 - 1000 psi (0-6.9 MPa)
Young's modulus (E) measured	8.0 - 14.0 x 10 ⁶ psi (55160 - 96530 MPa)
est. in-situ	0.8 - 1.4 x 10 ⁶ psi (5516 - 9653 MPa)
Poisson's ratio (ν)	0.15 - 0.35

¹Values are for intact basalt except for estimate of in-situ E or where noted.

CHAPTER 3

RECOMMENDATIONS FOR FUTURE WORK

The Hanford Site is one of the most intensively studied candidates for a geologic repository. A large volume of data has been generated for the region with extensive data collection continuing. As of this writing, a number of gaps in the data remain. The more important programs necessary to fill these are listed below.

Geologic Structure:

- o Determine the vertical and horizontal continuity of faults and fracture zones that might provide hydrologic connection between the Grande Ronde Formation and overlying units on the surface. Examples of these features are the Wallula Gap Fault and the faults associated with Gable Mountain - Gable Butte.

Hydraulic Conductivity:

- o Vertical hydraulic conductivities must be measured. These measurements either should be for each major basalt flow and interbed or should be sufficient to associate a conductivity with a particular rock type.
- o Perform a full-scale aquifer test consisting of a large pumping well and surrounding observation wells to place an adequate stress on the ground-water system for measuring bulk values of horizontal and vertical hydraulic conductivity and specific storage.

Hydraulic Heads:

- o Measure head values for the basalts and interbeds at Gable Mountain - Gable Butte in an attempt to determine what happens to the ground water that converges on this structure.

- o Measure head values in the vicinity of the repository site to determine an upward or downward flow gradient.
- o More measurements, and water samples, along the river to determine whether the river is a line sink.
- o More measurements throughout the Pasco Basin for the Grande Ronde Formation to determine whether the flow pattern in this unit is similar to or markedly different from the overlying units.

Porosity:

- o Measure effective porosities of the interbeds and interflows. In view of the pervasive fracturing of rocks in the Pasco Basin, such measurements will require field testing to obtain data.

Thermal Properties:

- o Determine the relationships between thermal conductivity of basalt and pressure, porosity, saturation state, presence or absence of fractures and temperature.

Mechanical Properties:

- o Obtain in-situ experimental data which can aid in the extrapolation of laboratory data to large-scale in-situ rock mass.
- o Determine in-situ stresses at or near the proposed repository site.

Elastic Constants:

- o Determine how, or whether, laboratory measurements can be extrapolated to in-situ conditions.

Geochemistry:

- o Obtain data set and rationale on waste nuclide solubilities and speciation in solution, either from experimental work or geochemical calculations.
- o Identify explicit mechanisms of retardation which are most important in basalt environments and substitute these for K_d in predicting the retardation of radionuclides.

Hydrologic Modeling:

- o Hydrologic modeling should be three-dimensional until more data are available to determine the usefulness of two-dimensional modeling.
- o The data currently available within the Pasco Basin are insufficient to define boundary conditions for modeling ground-water flow, so use should be made of hydrologic data available for the Columbia Plateau as a whole where possible.
- o Modeling could be used to identify input parameters which are inadequately characterized and are of importance in predicting radionuclide release to the accessible environment. Modeling could be used to help understand the ground-water flow system.

General Comments:

- o Experimental methods used in the determination of any parameter should be described in detail, including all other variables that might affect the results of the experiments.
- o Relationships must be established between data collected in the laboratory and data which are applicable at in-situ conditions.

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APPENDIX A

Hydraulic Conductivity

The hydraulic conductivity of the numerous lava flows and interbeds in the Pasco Basin is difficult to determine because of the inherently complex morphology and the distribution of the various rock types associated with each flow. In addition to flow morphology, erosion, alteration, and the infilling of fractures and other voids result in conductivity values that are inconsistent for a particular stratigraphic unit both laterally and vertically, as well as accounting for variation in the conductivity for similar features in different flows.

Some of the variations in hydraulic conductivity are the result of the methods used both in measuring and calculating these values. The results of pumping tests in a drill hole will vary depending on the drilling method used, the quality of the well completion, the length of the interval isolated in the hole, and the storage coefficient of the rock being tested. For this last factor, a low permeability rock may require days or weeks of monitoring before accurate measurements can be made, and these time intervals may not be available. The mathematical technique used to convert the field measurements to conductivities also will produce a variety of values.

Another source of inconsistent values is the possible presence of laboratory tests in the literature. The nature of the tests requires competent samples recovered during coring operations. As a result, the effects of fracturing on the conductivity cannot be measured by these techniques.

Table A-1 contains most of the hydraulic conductivity values reported in the literature for the Pasco Basin. The stratigraphic sequence has been divided into unconfined aquifers, major basalt sequences, and major interbeds. Each value or range of values has a reference and an indication as to whether an original source for the data is cited in the reference and whether the testing method is mentioned. Those values that are not reported in this appendix include those measurements taken over long drill hole intervals such that the values could not be assigned to a particular

formation, member, or rock type, and those values that are substantially out of the range of other reported values.

Until relatively recently, one of the more commonly referenced sources of hydrologic data was LaSala and Doty (1971). The conductivity values in this reference consist of estimates for various rock types in drill hole DC-1. Values for several rock types often were determined from a single, large-interval, packer test, and some conductivities were assigned to stratigraphic levels for which no tests were conducted (see Figure 10, LaSala and Doty, 1971). For completeness, these conductivity values are included in this appendix. The authors do not endorse the use of these values.

Table 2-1 provides the recommended range of hydraulic conductivity for each of the stratigraphic units.

Table A-1. Hydraulic Conductivity Values for
Stratigraphic Units in Pasco Basin

	Hydraulic Conductivity (ft/day ^a)
<u>Hanford (Glaciofluvial)</u>	1700 (Kipp & Mudd, 1973) NTM ^b 9100 (Deju, 1974) NTM 500-20,000 (Atlantic Richfield Hanford Company, 1976) NOS, NTM 3000 gravel (Summers and others, 1978) NOS? 250 sand (Summers and others, 1978) NOS? 1200-12,000 (Newcomb and others, 1972)
<u>Upper Ringold</u>	2 (Summers and others, 1978) NOS? 1-200 (Newcomb and others, 1972)
<u>Middle Ringold</u>	17 (Bierschenk, 1959) 640 (Deju, 1974) NTM 50 (Summers and others, 1978) NOS? 1-200 (Newcomb and others, 1972) 20-600 (Gephart and others, 1979) NOS, NTM
<u>Lower Ringold</u>	.11 (Deju, 1974) NTM 1.3 (Bierschenk, 1959) .5 (Summers and others, 1978) NOS? .1-7, av. 2 (Deju & Fecht, 1979) NOS, NTM .11-7 (Veatch, 1980) .11-10 (Gephart and others, 1979) NOS, NTM
<u>Saddle Mountains Basalt</u> <u>Interbeds</u>	
Levey	9.8 (Deju, 1980b) NTM
Rattlesnake Ridge	.39 - 15.6 (Deju, 1980b) NTM .13-100 (Gephart and others, 1979) NOS, NTM 3 (Deju, 1980b) NTM .1-103 (Gephart and others, 1979), NOS, NTM .08-9.1 (Ledgerwood & Deju, 1976) NTM .5-30 (Deju, 1974) NTM .25-30 (Veatch, 1980)
Selah	2 (Deju, 1980b) NTM 1.3-13.5 (Gephart and others, 1979) NOS, NTM 2.2-23.1 (Gephart and others, 1979) NOS, NTM

Saddle Mountains Basalt
Cold Creek

2 (Deju, 1980b) NTM
1.1-126 (Gephart and others, 1979) NOS, NTM
1.5-181 (Gephart and others, 1979) NOS, NTM
3 (Deju, 1980b) NTM

Interbeds
General

1.34 (LaSala and Doty, 1971) est
1.88 (LaSala and Doty, 1971) est
3.35 (LaSala and Doty, 1971) est

Basalt Members
Ice Harbor

No data

Elephant Mountain

2040 (Gephart and others, 1979) NOS, NTM interflow?

2.83 (RHO,^c 1981) NOS,
NTM, est.

Pomona

2.83×10^{-2} (RHO, 1981) col/ent
NOS, NTM, est.

Esquatzel

2.83×10^{-3} (RHO, 1981) col/ent
NOS, NTM, est.

Asotin

No data

Wilbur Creek

No data

Umatilla

2.83×10^{-4} (RHO, 1981) col/ent
NOS, NTM, est.

Basalt-General
Interflows

.5-30 (Deju, 1974) NTM

.1-40 (Summers and others, 1978)
NOS, NTM

Av 2 (Deju and Fecht, 1979),
NOS, NTM

2.9×10^{-2} (LaSala and Doty, 1971) est

Dense

.01-.1 (Summers and others, 1978) NOS, NTM

Av .030 (Deju and Fecht, 1979) NOS, NTM

2.9×10^{-2} (LaSala and Doty, 1971) est

Mabton Interbed

20 (RHO, 1981) NOS, NTM, est.

1.4-125 (Gephart and others, 1979) NOS, NTM

2-158 (Gephart and others, 1979) NOS, NTM

20-60 (Veatch, 1980) NOS, NTM

20-60 (Summers and others, 1978) NOS, NTM

Wanapum Basalt
Interbeds

0.1-40 (Summers and others, 1978) NOS, NTM

Av 1 (Deju and Fecht, 1979) NOS, NTM

.19 (LaSala and Doty, 1971) est

.012 (LaSala and Doty, 1971) est

1.9 (LaSala and Doty, 1971) est

Basalt Members
Priest Rapids

2.83×10^{-4} (RHO, 1981) NOS, NTM est interflow

2.9×10^{-3} - 1.3×10^4 (Gephart and
others, 1979) NOS, NTM interflow

Roza

4.26×10^{-2} (RHO, 1981) NOS, NTM, est col/ent

Frenchman
Springs

2.83×10^{-5} (RHO, 1981) NOS, NTM, est col/ent

3×10^{-2} - 3×10^{-6} (Deju, 1980b) NTM 2 interflows

Basalt-General
Interflows

1×10^{-2} - 10 (Summers and others, 1978)
NOS, NTM

8×10^{-2} (Deju and Fecht, 1979) NOS, NTM

.19 (LaSala and Doty, 1971) est

.012 (LaSala and Doty, 1971) est

Dense

1×10^{-2} - 1×10^{-3} (Summers and others,
1978) NOS, NTM

Av 2×10^{-2} (Deju and Fecht, 1979) NOS, NTM

.012 (LaSala and Doty, 1971) est

Vantage Interbed

3.02×10^{-1} (RHO, 1981) NOS, NTM, est

Grande Ronde Basalt
Interbeds

3×10^{-3} - ? (Summers and others, 1978) NOS, NTM
Av 3×10^{-3} (Deju and Fecht, 1979) NOS, NTM
 3.22×10^{-3} (LaSala and Doty, 1971) est
2.04 (LaSala and Doty, 1971) est

Basalt Members
Upper Sentinel
Bluffs

3×10^{-5} - 3×10^{-8} (RHO, 1981) NOS, NTM, est

col/ent

Middle Sentinel
Bluffs

3×10^{-4} - 3×10^{-7} (RHO, 1981) NOS, NTM, est
 3×10^{-5} - 3×10^{-8} (RHO, 1981) NOS, NTM, est
 3×10^{-3} - 3×10^{-7} (RHO, 1981) NOS, NTM, est
 1.5×10^{-2} - 3.1×10^{-4} (SAI, 1978)
 5.7×10^{-3} - 1.3×10^{-4} (SAI, 1978)
 2.0×10^{-2} (Raymond and Tillson, 1968)

interflow
col/ent
col/ent
fract. basalt
fract. basalt
mostly
interflow

Lower Sentinel
Bluffs

3×10^{-3} - 3×10^{-6} (RHO, 1981) NOS, NTM, est
 3×10^{-1} - 3×10^{-3} (RHO, 1981) NOS, NTM, est
 3×10^{-1} - 3×10^{-3} (RHO, 1981) NOS, NTM, est
 3×10^{-5} - 3×10^{-7} (RHO, 1981) NOS, NTM, est
 3×10^{-5} - 3×10^{-7} (RHO, 1981) NOS, NTM, est

interflow
interflow
interflow
col/ent
col/ent

Umtanum

3×10^{-3} - 3×10^{-7} (RHO, 1978) NTM
 7.1×10^{-4} (Deju and Fecht, 1979)¹
 3.9×10^{-4} - 1.7×10^{-3} (Deju and Fecht, 1979)²
 4.8×10^{-7} (Deju and Fecht, 1979)³
 1.6×10^{-4} (Raymond and Tillson, 1968)

interflow
interflow
interflow
interflow
mostly
interflow

1.7×10^{-6} - 1.5×10^{-8} (Gephart and others, 1979) NOS, NTM

fract. dense

1.8×10^{-3} - 4.0×10^{-4} (SAI, 1978)

fract. dense

7.4×10^{-7} - 1.5×10^{-8} (SAI, 1978)

fract. dense

1.7×10^{-6} - 4.2×10^{-8} (SAI, 1978)

fract. dense

2.8×10^{-5} - 2.8×10^{-7} (Gephart and others, 1979) NOS, NTM

col/ent

1.3×10^{-5} - 4.2×10^{-5} (Gephart and others, 1979) NOS, NTM

high density

2.8×10^{-5} - 2.8×10^{-9} (RHO, 1978) NTM

col

7.3×10^{-7} (Deju and Fecht, 1979)²

ent

1.6×10^{-6} (Deju and Fecht, 1979)²

col/ent

8.2×10^{-7} (Deju and Fecht, 1979)²

col

1.5×10^{-8} (Deju and Fecht, 1979)³

ent

4.2×10^{-8} (Deju and Fecht, 1979)³

col/ent

3.9×10^{-8} (Deju and Fecht, 1979)³

col

1.7×10^{-5} - 1.3×10^{-5} (Gephart and others, 1979) NOS, NTM dense
 4.2×10^{-5} (Gephart and others, 1979) NOS, NTM dense
 4.0×10^{-5} - 3.1×10^{-5} (Gephart and others, 1979) NOS, NTM dense

.032 (LaSala and Doty, 1971) est dense

Low K₂O Flow

2.0×10^{-1} (Apps and others, 1979) interflow
 8.2×10^{-7} - 4.0×10^{-8} (SAI, 1978) fract. dense
 2.8×10^{-5} - 2.8×10^{-10} (RHO, 1981) NTM col/ent

Below Low K₂O

5.8×10^{-3} (Raymond and Tillson, 1968) combined
 2.7×10^{-4} (Raymond and Tillson, 1968) dense and
 5.5×10^{-4} (Raymond and Tillson, 1968) interflows

.012 (LaSala and Doty, 1971) est interflow
1.58 (LaSala and Doty, 1971) est interflow
.012 (LaSala and Doty, 1971) est dense
 3.08×10^{-3} (LaSala and Doty, 1971) est dense

Basalt-General Interflows

1.1×10^{-5} - 300 (Gephart and others, 1979) NOS, NTM
 2.8×10^{-10} - 2.8×10^3 (RHO, 1978) NTM
 1×10^{-5} - 1×10^2 (Deju, 1980a) NTM
Av 1×10^{-2} (Deju and Fecht, 1979) NOS, NTM
 1.61×10^{-3} (LaSala and Doty, 1971) est interflow

Dense

2.8×10^{-3} - 2.8×10^{-6} (RHO, 1978) NTM
 1×10^{-3} - 1×10^{-8} (Deju, 1980a) NTM
Av 3×10^{-3} (Deju and Fecht, 1979) NOS, NTM
 1.61×10^{-3} (LaSala and Doty, 1971) est dense
 3.68×10^{-5} (LaSala and Doty, 1971) dense
 5.95×10^{-5} (LaSala and Doty, 1971) dense
 3.11×10^{-5} (LaSala and Doty, 1971) dense
 1.89×10^{-5} (LaSala and Doty, 1971) dense
 3.96×10^{-5} (LaSala and Doty, 1971) dense

Pre-Grande Ronde Basalt

5.3×10^{-2} (Raymond and Tillson, 1968) combined
 2.7×10^{-5} (Raymond and Tillson, 1968) dense and
interflows

General Rock Types Interflow

8.88×10^{-5} (Raymond and Tillson, 1968) magnitude only
 5.92×10^{-4} (Raymond and Tillson, 1968)
 2.37×10^{-2} (Raymond and Tillson, 1968)
 2.96×10^{-4} (Raymond and Tillson, 1968)
 5.33×10^{-2} (Raymond and Tillson, 1968)

Pre-Grande Ronde Basalt (Cont'd)

	10^{-9} (Summers and others, 1978) NOS, NTM
	1.5×10^{-2} (Deju and Fecht, 1979) ¹
	4.5×10^{-4} - 7.9×10^{-3} (Deju and Fecht, 1979) ²
	3.1×10^{-4} - 6.5×10^{-3} (Deju and Fecht, 1979) ²
	10^{-4} (Deju and Fecht, 1979) NOS, NTM
	10^{-10} (Deju and Fecht, 1979) NOS, NTM
	1 (Deju and Fecht, 1979) NOS, NTM
	10^{-4} (Deju and Fecht, 1979) NOS, NTM
	1.2×10^{-5} (Deju and Fecht, 1979) ³
Vesicular Basalt	3.12×10^{-5} - 5.96×10^{-5} (Deju and Fecht, 1979) NOS, NTM
	1.07×10^{-5} - 4.82×10^{-4} (Gephart and others, 1979) NOS, NTM
Dense Basalt	1.90×10^{-5} (Deju and Fecht, 1979) NOS, NTM
	10^{-4} (Deju and Fecht, 1979) NOS, NTM
	6×10^{-4} (Summers and others, 1978) NOS, NTM
	1.33×10^{-5} - 4.25×10^{-5} (Gephart and others, 1979) NOS, NTM

^a1 ft/day = 0.301 m/day

^bAbbreviations:

NTM = no test method
NOS = no original source
est = estimated value

col = colonnade portion of flow
ent = entablature portion of flow
fract. = fractured

^cRockwell Hanford Operations

^dScience Applications, Inc.

1,2,3 Mathematical technique used to calculate hydraulic conductivity

¹Papadapulos and Others, 1973

²U.S. Bureau of Reclamation, no source cited

³Hvorslev, 1951

APPENDIX B

Transmissivity

Transmissivity data for the Pasco Basin are mostly limited to the unconfined aquifers. Among the reasons for this limited amount of data are the method of well construction, the large interval of drill hole tested, the heterogeneity of the stratigraphic units, poor record keeping, and the lack of interest in the confined aquifers as a source of water. Recent work in the basin (Deju, 1980c) has been directed at alleviating this lack of data.

Table B-1 lists the transmissivity values in the literature that are recorded for specific formations or specific members of formation. Those values that were recorded for intervals of more than one member or formation have not been included. Data from outside of the Pasco Basin cannot be assumed to represent correlative units within the basin, and as a result, only Pasco Basin data are listed.

Table B-1. Transmissivity Values for Stratigraphic Units in Pasco Basin.

	Transmissivity (ft ² /d*)
<u>Hanford</u>	594,000 (Deju, 1974) 50,000 (Kipp & Mudd, 1973) 296,000 (Deju, 1974) 64,000 (Bierschenk, 1959)
<u>Middle Ringold</u>	4500 (Kipp & Mudd, 1973) 29,000 (Kipp & Mudd, 1973) 10,000 (Bierschenk, 1959) 25,000 (Kipp & Mudd, 1973) 35,000 (Deju, 1974) 1000 (Kipp & Mudd, 1973) 750 (Bierschenk, 1959) 30,000 (Kipp & Mudd, 1973) 9500 (Deju, 1974) 14,000 (Kipp & Mudd, 1973) 57,000 (Kipp & Mudd, 1973) 21,000 (Kipp & Mudd, 1973) 11,000 (Kipp & Mudd, 1973) 2800 (Deju, 1974) 28,000 (Kipp & Mudd, 1973) 19,000 (Kipp & Mudd, 1973) 3300 (Deju, 1974) 1600 (Kipp & Mudd, 1973) 13,000 (Kipp & Mudd, 1973) 4500 (Deju, 1974) 9100 (Kipp & Mudd, 1973)
<u>Lower Ringold</u>	8 (Deju, 1974) 5 (Deju, 1974) 210 (Bierschenk, 1959) 4 (Deju, 1974)
<u>Ringold & Hanford</u>	40,000 (Deju & Fecht, 1979) NOS, NTM
<u>Saddle Mountains Basalt</u>	1443 (Deju and Fecht, 1979) NOS, NTM
Interbeds Levey	No data
Rattlesnake Ridge	No data
Selah BH-16	1.1 - 1.1x10 ¹ (Deju, 1980c)
Cold Creek	No data

*1 ft²/d = 9.26 x 10⁻² m²/d.

Saddle Mountains Basalt (Cont'd)Basalt Members

Ice Harbor	No data
Elephant Mountain	No data
Pomona	No data
Esquatzel	No data
Asotin	No data
Wilbur Creek	No data
Umatilla	No data

Mabton Interbed DC-15 $1.1 \times 10^1 - 1.1 \times 10^2$ (Deju, 1980c)

Wanapum Basalt 99 (Deju and Fecht, 1979) NOS, NTM

Basalt-Members

Priest Rapids	DC-14	$1.1 \times 10^2 - 1.1 \times 10^3$ (Deju, 1980c)	interflow
	DC-14	1.1×10^3 (Deju, 1980c)	interflow
	DC-15	1.1×10^{-2} (Deju, 1980c)	interflow
Priest Rapids- Roza	DC-14	$1.1 \times 10^2 - 1.1 \times 10^3$ (Deju, 1980c)	interflow
	DC-15	$1.1 \times 10^3 - 1.1 \times 10^4$ (Deju, 1980c)	interflow
Roza	DC-15	1.1×10^3 (Deju, 1980c)	interflow
Roza-Frenchman Springs	DC-14	1.1×10^2 (Deju, 1980c)	interflow
Frenchman Springs	DC-12	$1.1 \times 10^1 - 1.1 \times 10^2$ (Deju, 1980c)	interflow
	DC-14	$1.1 \times 10^2 - 1.1 \times 10^3$ (Deju, 1980d)	interflow
	DC-14	1.1×10^2 (Deju, 1980d)	interflow
	DC-14	$1.1 \times 10^2 - 1.1 \times 10^3$ (Deju, 1980d)	interflow
	DC-14	$1.1 \times 10^2 - 1.1 \times 10^3$ (Deju, 1980d)	interflow
	DC-14	1.1×10^3 (Deju, 1980d)	interflow
	DC-15	1.1×10^3 (Deju, 1980d)	interflow
	DC-15	$1.1 \times 10^2 - 1.1 \times 10^3$ (Deju, 1980d)	interflow
	DC-15	$1.1 \times 10^1 - 1.1 \times 10^2$ (Deju, 1980d)	interflow
	DC-15	1.1×10^3 (Deju, 1980d)	interflow
	DC-15	1.1×10^0 (Deju, 1980d)	interflow
	DC-15	1.1×10^0 (Deju, 1980d)	interflow
<u>Vantage Interbed</u>	DB-15	1.1×10^{-5} (Deju, 1980c)	
<u>Vantage Interbed</u> <u>Grande Ronde</u>	DC-14	1.1×10^2 (Deju, 1980d)	mostly inter- flow?

Grande Ronde Basalt

	47 (Deju and Fecht, 1979) NOS, NTM	
DC-6	$1.1 \times 10^2 - 1.1 \times 10^3$ (Deju, 1981)	interflow
DC-7	1.1 (Deju, 1980c)	interflow
DC-7	1.1 - 1.1×10^1 (Deju, 1980d) composite of unit	
DC-6	$1.1 \times 10^{-2} - 1.1 \times 10^{-1}$ (Deju, 1980c)	interflow
DC-6	$1.1 \times 10^1 - 1.1 \times 10^2$ (Deju, 1980d)	interflow
DC-6	1.1×10^2 (Deju, 1980d) composite of unit	
DC-7	$1.1 \times 10^{-2} - 1.1 \times 10^{-1}$ (Deju, 1980d)	interflow
DC-7	1.1 - 1.1×10^1 (Deju, 1980d)	interflow
DC-7	$1.1 \times 10^{-3} - 1.1 \times 10^{-2}$ (Deju, 1981)	interflow
DC-12	1.1 (Deju, 1980d)	interflow
DC-12	1.1 (Deju, 1980d)	interflow
DC-12	1.1×10^3 (Deju, 1980d)	interflow
DC-12	1.1×10^3 (Deju, 1981)	interflow
DC-12	$1.1 \times 10^{-1} - 1.1$ (Deju, 1981)	interflow
DC-12	$1.1 \times 10^{-3} - 1.1 \times 10^{-1}$ (Deju, 1981)	interflow
DC-14	1.1×10^{-1} (Deju, 1980d)	interflow
DC-14	1.1×10^1 (Deju, 1980d)	interflow
DC-14	1.1 - 1.1×10^1 (Deju, 1981)	interflow
DC-14	$1.1 \times 10^{-2} - 1.1 \times 10^{-1}$ (Deju, 1981)	interflow
DC-14	$1.1 \times 10^{-1} - 1.1$ (Deju, 1981)	interflow
DC-15	1.1 - 1.1×10^1 (Deju, 1981)	interflow
DC-15	1.1×10^3 (Deju, 1981)	interflow
DC-15	1.1 - 1.1×10^1 (Deju, 1981)	interflow
DC-15	1.1×10^2 (Deju, 1981)	interflow

APPENDIX C

Geothermal Gradient

The geothermal gradients determined for the Pasco Basin (Table C-1 and Table C-2) are closer to values for tectonically active areas than to those for tectonically stable areas (Table C-3). Because of the volcanic activity as recently as 1 my ago (Meyers and others, 1979), the region would be expected to have a higher than average geothermal gradient.

Table C-1. Geothermal Gradient Measured at Drill Hole RSH-1. (After Raymond and Tillson, 1968.)

Temperature gradients not thought to be in equilibrium.

<u>Depth (ft)</u>	<u>Gradient (°F/100 ft)*</u>	<u>Depth (ft)</u>	<u>Gradient (°F/100 ft)</u>
600-800	0.8	5000-5200	1.7
800-1000	0.8	5200-5400	1.8
1000-1200	1.1	5400-5600	1.8
1200-1400	1.2	5600-5800	1.8
1400-1600	1.4	5800-6000	1.2
1600-1800	1.5	6000-6200	1.9
1800-2000	1.3	6200-6400	1.6
2000-2200	1.3	6400-6600	1.9
2200-2400	1.6	6600-6800	2.0
2400-2600	1.6	6800-7000	1.6
2600-2800	1.6	7000-7200	1.6
2800-3000	1.6	7400-7600	1.8
3000-3200	1.6	7400-7600	1.9
3200-3400	1.9	7600-7800	1.9
3400-3600	1.9	7800-8000	1.5
3600-3800	1.9	8000-8200	1.8
3800-4000	1.9	8200-8400	1.5
4000-4200	1.5	8400-8600	2.5
4200-4400	1.7	8600-8800	1.9
4400-4600	1.8	8800-9000	2.1
4600-4800	1.7	9000-9200	3.5
4800-5000	1.6	9200-9400	1.2

Calculated from above data

<u>Depth Interval (ft)</u>	<u>Gradient (°F/100 ft)</u>
600-1000	0.8
1000-2000	1.3
2000-3000	1.5
3000-4000	1.8
4000-5000	1.7
600-2000	1.2
600-3000	1.3
600-4000	1.5
600-5000	1.5

*(°F/100 ft)/(0.055) = °C/km

Table C-2. Geothermal Gradient Measurements in Washington State.
(After Sass and others, 1971.)

Locality	Latitude	Longitude	Elevation		Depth Range		Gradient	
	N	W	m	ft	m	ft	°C/km	°F/100 ft
Rattlesnake Hills								
RS-1	46° 26'	119° 47'	875	2870	900-2500	2952-8200	(34.8)	(1.91)
RS-2					58-119	190-390	28.0 ± 0.2	1.54 ± .01
Richland								
DH-3	46° 26'	119° 17'	120	394	305-608	1000-2000	38.95 ± 0.15	2.14 ± .01
					608-1079	2000-3540	34.58 ± 0.28	1.90 ± .02
Willa								
DH-1	46° 35'	119° 31'	168	550	53-183	174-600	37.2 ± 0.3	2.04 ± .02

Table C-3. Worldwide Mean Geothermal Gradient.
(After Verhoogen and others, 1970.)

Range, Worldwide

$8^{\circ}\text{K/km} - 40^{\circ}\text{K/km} = 0.45^{\circ}\text{F/100 ft} - 2.25^{\circ}\text{F/100ft}$

APPENDIX D

Density

Histograms of bulk and grain density for Hanford basalts and for all basalts as determined in the laboratory are given in Figures D-1 and D-2. The peaks for Hanford basalts as determined in the laboratory are at slightly higher density values than are the peaks for the composite of other basalts. A comparison of percentages of total samples in given bulk density intervals with percentages of basalt types in the Hanford stratigraphy is given in Table D-1. This comparison seems to indicate that the distribution of bulk densities for Hanford basalts is not significantly biased by sampling techniques.

Suggested Range: All Hanford basalts: 1.9-3.1,
skewed distribution

Dense Hanford basalts: 2.7-2.9.

Schmidt and others (1980) report a bulk density of 1.82 g/cm³ for the Vantage sandstone. No density values were found for other interbeds or interflows.

Table D-2 lists most of the values for basalt densities which were found in the literature during our investigations.

Table D-1. Comparison of Basalt Density Distribution
With Distribution of Basalt Lithologic
Types at Hanford

<u>Bulk Density Interval (g/cm³)</u>	<u>% of Total Samples</u>	<u>Basalt Type</u>	<u>% of Total Stratigraphic Thickness*</u>
1.9-2.1	1.7		
2.1-2.3	2.0	Breccia	11.8
2.3-2.5	4.9	Vesicular	12.2
2.5-2.7	8.1		
2.7-2.9	81.4	Dense	76.0
2.9-3.1	1.9		

*Derived from core logs for wells DC-4, DC-6, DC-8, published
in Fenix and Scisson, Inc. (1978 a,b,c).

Figure D-1. Histogram of Bulk Density of Basalt --
World-Wide and Columbia River Basalt.

D-4

Figure D-2. Histogram of Grain Density of Basalt --
World-wide and Columbia River Basalt.

Table D-2. Density of Basalt - Columbia Plateau

SOURCE	LOCALITY OR BASALT UNIT	BULK DENSITY (g/cm ³)	GRAIN DENSITY (g/cm ³)
Atlantic Richfield Hanford Co. (1976)	Columbia River basalt	2.66-2.96	-
Colorado School of Mines (1978)	Pomona flow	1.72-3.05	2.37-2.89 ^c
Coulson (1972)	Little Granite basalt	2.73	-
Deere and Miller (1966)	Columbia River basalt	2.73-2.87	-
Duvall and others (1978)	Pomona flow	2.69-2.87	2.85-2.89 ^c
Erikson and Krupka (1980)	Pomona flow	2.86-2.88	-
Foundation Sciences Inc. (1980a)	Pomona flow	2.81-2.88	2.91-3.11
Foundation Sciences Inc. (1980b)	Pomona flow	2.79-2.89	2.85-3.12
Foundation Sciences Inc. (1980c)	Pomona and Umtanum flows	2.75-2.89	2.80-3.60
Foundation Sciences Inc. (1981)	Umtanum flow	2.75-2.82	2.78-3.60
Hocking and St. John (1979)	Columbia River basalt	2.81	-
Johnston and Palmer (1981)	Survey, Hanford basalt	2.4-3.1	-
Kaser (1979)	Hanford basalt	2.84	-
Martinez-Baez and Amick (1978)	Pomona and Umtanum flows	2.78-2.87	-
Miller (1979a)	Umtanum flow	2.33-2.82	-
Miller (1979b)	Umtanum flow	1.91-2.81	-
Miller and Bishop (1979)	Umtanum flow	1.91-2.83	-
Podnieks and others (1972)	Columbia River basalt	2.84	-
Schmidt and others (1980)	Summary, Hanford basalt	2.72-2.85 ^a	3.00 ^a
Stephenson and Triandafilidis (1974)	Columbia River basalt	2.89	-
Taylor and others (1980)	Survey, Hanford basalt	NA	NA

Table D-2. (Continued)

Other Localities

SOURCE	LOCALITY OR BASALT UNIT	BULK DENSITY (g/cm ³)	GRAIN DENSITY (g/cm ³)
Aleksandrov and others (1969)	U.S.S.R	2.87	-
Anderson (1962)	Nevada Test Site	2.63-2.70	2.78-2.83
Avila (1966)	Brazil	2.44-2.97	-
Balakrishna (1967)	Deccan trap	2.29-2.85	2.73-3.01
Balashov and Volarovich (1971)	U.S.S.R	2.63	-
Bayuk and others (1975)	Indian Ocean	2.65-2.95	-
Berdinelli (1962)	Nevada Test Site	2.70-2.73	-
Blair (1956)	Eniwetok atoll	2.86	-
Bridgman (1924)	unknown	2.92	-
Carlson and Christensen (1977)	Deep Sea Drilling Project (DSDP)	2.73-2.88	-
Christensen (1973a)		2.56-2.84	-
Christensen (1973b)	DSDP	2.45-2.94	-
Christensen and others (1974)	DSDP	2.19-2.89	-
Farjallat (1974)	Brazil	2.32-2.92 ^a	-
Fox and Schreiber (1973)	DSDP	2.56-2.95	-
Fox and others (1972)	DSDP	2.25-2.58	-
Grady and Hollenbach (1979)	Dresser basalt	2.98	-
Hanley and others (1978)	Dresser basalt	2.97	-
Heezen and others (1973)	DSDP	2.13-2.86	-
Hughes and Maurette (1975)	Colorado	2.59	-
Hyndman and Drury (1976)	Mid-Ocean Ridge basalt	2.46-2.95	2.80-3.01

Table D-2. (Continued)

Other Localities

SOURCE	LOCALITY OR BASALT UNIT	BULK DENSITY (g/cm ³)	GRAIN DENSITY (g/cm ³)
Kaarsberg (1968)	unknown	2.97	-
Kawada (1964)	Japan (Kyushu)	2.58	-
Krech and others (1974)	Dresser basalt	3.01-3.02	-
Kulhawy (1975)	{ Nevada Test Site	2.70	2.83
	{ unknown	2.66	2.86
Kumar (1968)	Nevada Test Site	2.68	-
Lindroth and Krawza (1971)	Dresser basalt	2.97	-
Livingston (1961)	unknown	2.91	-
Lutton (1968)	Nevada Test Site	2.09-2.70	2.85 ^a
Marovelli and Veith (1965)	Dresser basalt	2.97	-
Morgan and West (1980)	Dresser basalt	3.02	-
Morrell and Larson (1974)	Dresser basalt	3.03	-
Nakagawa and Mori (1963)	Japan	2.84	-
Neprochnov and others (1974)	DSDP	2.58-2.89	-
Nordyke and Wray (1964)	Nevada Test Site	2.41-2.66	2.82-2.87
Nugent and Banks (1965)	Nevada Test Site	2.36-2.73	-
Overton and Moczygemba (1967)	Texas	2.43-3.18	-
Petrinin and others (1971)	U.S.S.R.	2.44-2.74	-
Ramana and Venkataharayana (1973)	Kolar	2.67-3.06	-
Ricketts and Goldsmith (1970)	unknown	2.72	-
Robertson and Peck (1974)	Hawaii	0.05-3.06	2.20-3.15

Table D-2. (Continued)

Other Localities

SOURCE	LOCALITY OR BASALT UNIT	BULK DENSITY (g/cm ³)	GRAIN DENSITY (g/cm ³)
Ruiz (1966)	Brazil	2.44-2.92	2.68-3.00
Salisbury and Francheteau (1977)	DSDP	2.5 -2.6*	-
Schreiber and others (1974)	DSDP	2.35-2.90	-
Sharp (1972)	Archatka Island	2.73 ^a	-
Somerton and others (1963)	Mchole Test Site	2.82	-
Stephens (1963)	Nevada Test Site	2.68	-
Stowe (1969)	Nevada Test Site	2.66-2.74	2.82-2.84
Tokuyama and Yoshida (1979)	Japan	2.63	-
Tomoshevskaya and Volodina (1976)	unknown	2.75	-
Vortman (1962)	Nevada Test Site	2.02-2.77	-
Washburn (1927)	{ Texas	3.19	-
	{ Washington	2.94	-
	{ Germany	3.06	-
Windes (1949)	Michigan	2.81-2.97	-
Windes (1950)	Michigan	2.04-2.80	-

*This value is a range determined from in-situ density logs.

Table D-2. (Continued)

Citations or Surveys of Other Sources

SOURCE	LOCALITY OR BASALT UNIT	BULK DENSITY (g/cm ³)	GRAIN DENSITY (g/cm ³)
Agapito and others (1977)	literature range	2.40-3.10	-
AVCO Everett Research Lab (1972)	Cites Marovelli and Veith (1965)	NA	NA
Clark and Caudle (1961)	literature range	2.04-3.01	-
Dames and Moore (1978)	Cites Deere and Miller (1966), Sharp (1972), Stave (1969), Vortman (1962)	NA	NA
Dubois and others (1979)	Cites Martinez-Baez and Amick (1978)	NA	NA
Lehnhoff and Scheller (1975)	Cites Lindroth and Krawza (1971)	NA	NA
Maurer (1968)	literature average (?)	2.8	-
Nafe and Drake (1967)	literature range	2.6 -3.2	-
Ratigan (1976)	literature average	2.56 ^a	-
Wingquist (1976)	Cites Lindroth and Krawza (1971)	NA	NA

a - average

c - calculated

NA - not applicable

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APPENDIX E

Porosity

Two types of porosity are commonly measured and reported. Total porosity is the volume percentage of void space in the rock. Effective porosity is the volume percentage of connected void space, or the void space through which flow could occur. Due to the presence of many isolated vesicles in basalts, the total porosity of intact basalt is usually much greater than the effective porosity. However, the in-situ effective porosity of basalt is enhanced by the presence of numerous fractures. Thus, effective porosity as described here is probably less than the effective porosity of basalt in situ.

A histogram of total porosity of Hanford and other basalts is given in Figure E-1; one of effective porosity is given in Figure E-2. Both histograms are compiled from data obtained in the laboratory. In addition, a histogram of the ratio (effective porosity)/(total porosity) is given in Figure E-3 in an attempt to delineate any strong correlation between the two porosity values. Although the ratio is usually less than 0.25, there are enough values greater than 0.25 to prevent any definitive statement at this time.

No attempt is made to establish relationships between porosity and other parameters due to the absence of correlative data for basalts. Table E-1 contains the values for basalt porosities considered.

<u>Suggested Range:</u>	Total porosity:	0-40%
	Effective porosity:	0-2.5%
	Effective/total:	0-0.4

Schmidt and others (1980) report the following values for the Vantage sandstone:

Total porosity:	40.67%
Effective porosity:	30.85%

No data were found for the porosities of other interbeds or interflows.

Figure E-1. Histogram of Total Porosity of Basalt.

Figure E-2. Histogram of Effective Porosity of Basalt.

Figure E-3. Histogram of Effective Porosity/Total Porosity Ratio for Basalt.

Table E-1. Porosity of Basalt - Columbia Plateau

SOURCE	LOCATION OR BASALT UNIT	TOTAL ϕ (%)	Effective ϕ (%)
Colorado School of Mines (1978)	Pomona flow	0.96-37.8	-
Deere and Miller (1966)	Columbia River basalt	-	0.75-1.92
Duvall and others (1978)	Pomona flow	-	1.60-2.39
Erikson and Krupka (1980)	Pomona flow	-	0.50-0.60
Foundation Sciences Inc. (1980a)	Pomona flow	2.71-8.14 ^c	0.5 -1.4
Foundation Sciences Inc. (1980b)	Pomona flow	1.0 -7.7 ^c	0.1 -0.6
Foundation Sciences Inc. (1980c)	Pomona Umtanum flows	0.71-9.68 ^c	0.19-1.85
Foundation Sciences Inc. (1981)	Umtanum flow	0.71-9.68 ^c	0.19-2.06
Nace and others (1959)	Columbia River basalt	3.8 -24.8	-
Podnieks and others (1972)	Columbia River basalt	-	2.0
Robertson (1970)	Columbia River basalt	18.5 ^c	-
Schmidt and others (1980)	Summary, Hanford basalts	-	1.5 -2.8 ^a
Stephenson and Triandafilidis (1974)	Columbia River basalt	2.2	-
White and Sarcia (1978)	Columbia River basalt	0.55-3.84	0.18-1.34

Table E-1. Other Localities (Cont'd)

SOURCE	LOCATION OR BASALT UNIT	TOTAL ϕ (%)	EFFECTIVE ϕ (%)
Aleksandrov and others (1969)	U.S.S.R	2.85	-
Balakrishna (1967)	Deccan trap	5.0-18.0	-
Brodsky and Spetzler (1979)	Ralston basalt	-	0.72
Burns and others (1973)	Deep Sea Drilling Project (DSDP)	≤ 25.0	-
Farjallat (1974)	Brazil	1.43-20.16	-
Friedman and others (1979)	New Mexico	5-8	-
Hanley and others (1978)	Dresser basalt	-	0.047
Hyndman and Drury (1976)	Mid-Ocean ridge basalt	1.7- 24.5	-
Johnson (1977)	DSDP	(Crack ϕ = 6%)	
Krech and others (1974)	Dresser basalt	-	0.15-0.25
Kulhawy (1975)	{ Nevada Test Site	4.6	-
	{ unknown	10.2	-
Nakagawa and Mori (1963)	Japan	-	1.1
Neprochnov and others (1974)	DSDP	-	2.93-11.87
Nordyke and Wray (1964)	Nevada Test Site	6.3-16.0	-
Nugent and Banks (1965)	Nevada Test Site	-	< 1.0
Petrinin and others (1971)	U.S.S.R	-	5.19-15.25
Ramana and Venkatanarayana (1975)	Kolar	7-9	-
Robertson and Peck (1974)	Hawaii	1.7-97.8	-
Ruiz (1966)	Brazil	2.7-18.7 ^c	1.4 - 11.6

Table E-1. Other Localities (Cont'd)

SOURCE	LOCATION OR BASALT UNIT	TOTAL ϕ (%)	EFFECTIVE ϕ (%)
Salisbury and Francheteau (1977)	DSDP	5-15*	-
Sharp (1972)	Amchitka Island	9.97 ^a	-
Somerton and others (1963)	Mohole Test Site	-	2.06
Stowe (1969)	Nevada Test Site	3.07-5.37 ^c	-
Stowe and Ainsworth (1972)	Nevada Test Site	4.5 ^a	-
Washburn (1927)	Germany	-	0.4-0.5
Waza and others (1980)	Japan	6.0	-
Windes (1949)	Michigan	-	0.4-1.5
Windes (1950)	Michigan	-	0.4-18.0

Citations or Surveys of Previous Literature

Clark and Caudle (1961)	Literature survey	0.4-24.8 ^a	-
Dames and Moore (1975)	Cites Deere and Miller (1966), Sharp (1972), Stowe (1969), Vortman (1962)	NA	NA
Davis (1969)	Literature Survey	0.8-11.4	-
Taylor and others (1980)	Cites Duvall and others (1978)	-	NA

a = average

c = calculated

NA = not applicable

*The range given is determined from in-situ geophysical logs.

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APPENDIX F

Thermal Conductivity

The data currently available for the thermal conductivity of Columbia River basalt are shown as a function of temperature in Figure F-1. All of the data were obtained from laboratory tests on intact samples. The majority of the data indicate that conductivity increases slightly with increasing temperature, though the data of Murase and McBirney (1973), Duvall and others (1978), and Martinez-Baez and Amick (1978) do not fit this general trend. The various experimental descriptions are insufficient to attempt an explanation of the discrepancies in the trends with temperature.

F.1 Theoretical Considerations

There are a number of parameters which can affect the thermal conductivity of rocks: mineralogy, porosity, the nature of the saturating fluid, pressure, and temperature. Each of these was examined separately in an attempt to quantify any effect on predicted conductivity. Intact basalt is assumed to be thermally isotropic. Supporting evidence for this is given by Tokuyama and Yoshida (1979) for a basalt from Japan. In-situ thermal conductivity will be a function of the degree and nature of fracturing, and may or may not be isotropic. More data are needed to relate laboratory and in-situ conductivity.

F.1.1 Mineralogy

There have been a number of equations proposed to relate the conductivity of a rock to the conductivities of its mineral constituents; perhaps the most thorough summary of the relationships is that of Robertson and Peck (1974). After examination of the equations and comparison with an extensive set of experimental data, Robertson and Peck (1974) concluded that the following equation closely matched the data*:

*This model matches the experimental data within an average of 0.3×10^{-3} cal/cm sec°C. Another model--the quadratic model--has similar or slightly better correlation; the parallel-series model is chosen for convenience.

Figure F-1. Thermal Conductivity of Basalt Versus Temperature.

DESCRIPTION OF SOURCES FOR THERMAL CONDUCTIVITY
DATA IN FIGURE F-1.

1. Pomona, Erikson and Krupka
2. Pomona, Foundation Sciences, Inc.
3. Umtanum, Foundation Sciences, Inc.
4. Picture Gorge, Murase and McBirney
5. Pomona, Duvall, et al.
6. Pomona, Colorado School of Mines
7. Pomona, Martinez-Baez and Amick
8. Umtanum, Martinez-Baez and Amick
9. Magnetite and Chlorite
10. Calculated "average"
11. Pomona, Foundation Sciences, Inc.

$$k_R = \frac{1}{2} \left[\sum_{i=1}^r n_i k_i + \frac{1}{\sum_{i=1}^r \frac{n_i}{k_i}} \right] \quad (1)$$

where

k_R = rock conductivity,

$n_i, n_{i+1}, \dots, n_r$ = volume fractions of minerals $\left(\sum_{i=1}^r n_i = 1 \right)$,

$k_i, k_{i+1}, \dots, k_r$ = conductivity of minerals.

F.1.2 Porosity and Saturating Fluid

Equation (1) can be used to estimate the zero-porosity conductivity of basalt. However, more physical significance can be achieved if porosity is included in the calculation. This is accomplished by assuming that the void space acts as a separate mineral constituent, or two constituents if the rock is not fully saturated. Thus equation (1) would become

$$k_R = \frac{1}{2} \left[(1-\phi_T) \sum_{i=1}^r n_i k_i + (\phi_T - \phi_E) k_{\text{air}} + \phi_E k_{\text{H}_2\text{O}} + \frac{1}{(1-\phi_T) \sum_{i=1}^r \frac{n_i}{k_i} + \frac{\phi_T - \phi_E}{k_{\text{air}}} + \frac{\phi_E}{k_{\text{H}_2\text{O}}}} \right] \quad (1a)$$

where ϕ_T = total porosity (volume fraction),

and ϕ_E = effective or connected porosity (volume fraction).

Equation (1a) assumes that all heat transfer within the rock is by conduction. This assumption probably is valid, in that radiation has been determined to have a negligible contribution below 500°C (Francl and Kingery, 1954), and convection is important only in the presence of significant flow channels and a relatively pronounced temperature gradient.

Representative modal analyses of the Pomona flow, the Umtanum upper colonnade and the Umtanum entablature are given in Table F-1. Individual mineral conductivities are listed in Table F-2; some of the values are representative of the mineral family rather than any specific solid solution composition.

Initial calculations of conductivity make the following assumptions: (1) total and effective porosity are equivalent and equal to 0.05, (2) the rock is fully saturated, (3) pressure is 0.1 MPa, (4) temperature is 25°C, (5) all Fe-Ti oxides are magnetite, and (6) opal is representative of all secondary minerals. The value of 0.05 for effective porosity in assumption (1) is larger than values measured for intact basalt. The larger value should account for at least a portion of the contribution of fractures in-situ.

The conductivities calculated for each of the three basalt types using these assumptions are given in Table F-3. Also in the table are conductivities calculated assuming that the rocks are completely dry.

The potential variation in conductivity due to changes in the percentages of major minerals in the rock is approximately 20 to 25 percent. The variation due to differences in water content may be at least as high as 30 percent, depending on the total and effective porosity.

In order to better understand the relative effects of changes in individual parameters, the conductivity of the Pomona flow was calculated for "average" conditions: pressure of 0.1 MPa, temperature of 25°C, total porosity of 0.06, and effective (i.e., water filled) porosity of 0.015. The resulting conductivity is 3.611×10^{-3} cal/cm sec°C.

Variations in this conductivity caused by mineralogical or compositional changes are shown in Table F-4. None of the changes result in a large variation in estimated conductivity.

Table F-1. Mineralogical Constituents of
Umtanum and Pomona Flow

(After Schmidt and others 1980)

<u>Mineral</u>	<u>Volume %</u>		
	<u>Umtanum colonnade</u>	<u>Umtanum entablature</u>	<u>Pomona</u>
Plagioclase ¹	46.4	32.1	39.9
Pyroxene (augite) ²	26.2	19.4	33.4
Opagues (ilmenite & magnetite)	5.1	4.1	2.6
Glass	19.0	40.7	14.2
Chlorophaeite	2.9	3.2	9.9
Apatite	0.1	--	--
Opal & Chalcedony	<u>0.3</u>	<u>0.5³</u>	<u>--</u>
Total	100.0	100.0	100.0

- 1: Plagioclase in basalt has a compositional range of 50 to 85 mole percent anorthite (Brown, 1967).
- 2: Augite in basalt has the compositional extremes of 10 to 88 mole percent enstatite, 12 to 50 mole percent ferrosillite, and 10 to 90 mole percent diopside (Brown, 1967).
- 3: Includes zeolites.

Table F-2. Thermal Conductivity of Basalt
Mines at 23-27°C

Mineral	Thermal Conductivity at 23-27°C (10^{-3} cal/cm sec°C)	Source
Plagioclase: An ₅₄ Ab ₄₆	3.65*	Horai & Simmons (1969)
	4.0	Robertson & Peck (1974)
Augite	9.13*	Horai & Simmons (1969)
Ilmenite	5.5	Robertson & Peck (1974)
	5.25 - 6.10	Horai & Simmons (1969)
Magnetite	11	Robertson & Peck (1974)
	12.18*	Horai & Simmons (1969)
	11.9	Clark (1966)
Apatite	3.28 - 3.31 (av. 3.295*)	Horai & Simmons (1969)
Chlorite ¹	10.38 - 14.77 (av. 12.575*)	Horai (1971)
	12.5	Clark (1966)
Opal	2.90*	Horai (1971)
Water	1.48 (av. 1.475*)	Robertson & Peck (1974)
	1.47	Clark (1966)
Air	0.063*	Robertson & Peck (1974)
Glass	3.0*	Robertson & Peck (1974)

*: Value used in calculations.

¹: Chlorite here refers to Mg₅ Al₂ Si₃ O₁₀ (OH)₈, while chlorophaeite is a mineraloid resembling chlorite and containing Fe and Ca as well; the conductivity value is taken as representative of the entire compositional range.

Table F-3. Calculated Thermal Conductivity;
Initial Conditions

<u>Basalt Unit</u>	<u>Saturated k at 25°C, 0.1 MPa (10⁻³ cal/cm sec°C)</u>	<u>Dry</u>
Umtanum colonnade	4.727	3.182
Umtanum entablature	4.277	2.898
Pomona	5.325	3.593

Table F-4. Calculated Thermal Conductivity of the
Pomona Flow; Effects of Varying Mineral
Compositions

<u>Variable</u>	<u>k at 25°C, 0.1 MPa (10⁻³ cal/cm sec°C)</u>
Plagioclase composition	
An ₅₄	3.611
An ₇₃	3.580 (-0.9%)*
An ₇₈	3.682 (+2.0%)
An ₈₄	3.716 (+2.9%)
Augite instead of augite & chlorite	3.449 (-4.5%)
Ilmenite instead of magnetite	3.527 (-2.3%)

*The values in parentheses in Tables F-4, F-5, F-6, F-8, and F-9 are the changes in the calculated conductivity relative to the conductivity at "average" conditions.

The effects of varying total and effective porosities are shown in Table F-5. Conductivity is calculated assuming that effective porosity is water-saturated, while the remaining void space is totally dry. As mentioned above, potential variation reaches at least 30 percent. Notice that the variation in effective porosity accounts for the possibility that laboratory measurements underestimate effective porosity by ignoring the contribution of fractures.

F.1.3 Pressure

Increasing pressure (P) is expected to increase thermal conductivity (k) because of the closure of cracks and ensuing increased thermal contact. This effect should decrease as pressure increases and fewer cracks remain open. An attempt was made to quantify this effect by fitting the data of Somerton and others (1963) for Mohole test site basalt with an equation of the form

$$\frac{1}{k} = a + bP . \quad (2)$$

The result for a temperature of 63°C is

$$\frac{1}{k} = 228.89 - 1.1716 P(\text{MPa}). \quad (2a)$$

In the absence of other data, and since the effective porosities of the Mohole test site basalt and the Pomona flow are similar, the slope is assumed to be the same for the Pomona flow at any given temperature. Also, the Pomona flow is assumed to have a conductivity of 4.65×10^{-3} cal/cm sec°C at 63°C and 0.014 MPa (Foundation Sciences, Inc., 1980a). Then

$$\frac{1}{k} = 215.07 - 1.1716 P(\text{MPa}) \quad (2b)$$

represents the relationship between conductivity and pressure for the Pomona. Using equation (2b), the

Table F-5. Calculated Thermal Conductivity of the Pomona Flow; Effects of Varying Porosity

Conditions	Variable	k(10 ⁻³ cal/cm sec°C)
i) $\phi_T = 6\%$ T = 25°C P = 0.1 MPa	$\phi_E = 1.0\%$	3.565 (-1.3%)
	1.5	3.611
	2.0	3.664 (+1.5%)
	2.5	3.726 (+3.1%)
ii) $\phi_E = 1.5\%$ T = 25°C P = 0.1 MPa	$\phi_T = 2.0\%$	4.845 (+34.2%)
	4.0	3.900 (+8.0%)
	6.0	3.611
	8.0	3.471 (-3.9%)
	10.0	3.388 (-6.2%)
	38.0	3.158 (-12.5%)
iii) $\phi_E/\phi_T = 0.25$ T = 25°C P = 0.1 MPa	$\phi_T = 2.0\%$	4.212 (+16.7%)
	4.0	3.802 (+5.3%)
	6.0	3.611
	8.0	3.500 (-3.1%)
	10.0	3.430 (-5.0%)
	38.0	3.236 (-10.4%)

Table F-6. Calculated Thermal Conductivity of the Pomona Flow; Effects of Varying Pressure

Conditions	Variable	k(10 ⁻³ cal/cm sec°C)
T = 63°C $\phi_E = 2\%$	P = 0.1 MPa	4.652 (+0.04%)
	10.0	4.918 (+5.8%)
	20.0	5.218 (+12.2%)
	27.5*	5.469 (+17.6%)

*: Maximum possible pressure, assuming that the repository is at 1000 meters in depth and that the average density of the overlying rock column is 2.8 g/cm³.

conductivities in Table F-6 are calculated; the maximum calculated deviation from the "average" conductivity is 18 percent.

F.1.4 Temperature

As seen in Figure F-1, conductivity has been found experimentally to vary in either direction with increasing temperature. In order to estimate basalt conductivity as a function of temperature, the conductivity of each constituent is estimated as a function of temperature, then equation (1a) is used to calculate the conductivity of the rock.

Birch and Clark (1940) found that resistivity $(1/k)$ showed a more consistent linear relationship to absolute temperature than did conductivity, so that empirical equations of the form

$$\frac{1}{k} = a + bT(^{\circ}K) \quad (3)$$

are used to fit experimental data for the minerals. Conductivities calculated using equation (3) will tend to underestimate the actual conductivity at higher temperatures (Birch and Clark, 1940). The equations for the individual minerals are given in Table F-7.

Using equations (3a) - (3f), the conductivity of the Pomona flow as a function of temperature was calculated and the results compiled in Table F-8. Two cases were examined: (1) all mineral conductivities variable with temperature, and (2) chlorite and magnetite conductivities constant with temperature. The second case was chosen because the lack of actual data for chlorite and magnetite prevents an accurate determination of the temperature effect on their conductivities. Of the two sets of values in Table F-8, the second is thought to be closer to reality, because the underestimation of basalt conductivity due to the assumption of the constancy of chlorite and magnetite conductivities will be at least partially balanced by the overestimation caused by the linear approximations for the other constituents (especially pyroxene).

Table F-7. Estimated Relationships of Temperature and Basalt Mineral Thermal Conductivities

Mineral	Equation	Temperature range of data (°C)	Source of Data
Glass	(3a) $\frac{1}{k} = 386.14 - .1673T(^{\circ}K)$	30 - 800	Murase & McBirney (1973)
Plagioclase (An ₆₀)	(3b) $\frac{1}{k} = 256.81 - .0517T(^{\circ}K)$	0 - 200	Birch & Clark (1940)
Augite ¹	(3c) $\frac{1}{k} = 62.481 + .1578T(^{\circ}K)$	0 - 200	Birch & Clark (1940)
Water ²	(3d) $\frac{1}{k} = 1137.197 - 1.4797T(^{\circ}K)$	0 - 80	Clark (1966)
Magnetite ³	(3e) $\frac{1}{k} = -500.722 + 1.9548T(^{\circ}K)$	35 - 70	Clark (1966)
Chlorite ⁴	(3f) $\frac{1}{k} = 30.695 + 0.1638T(^{\circ}K)$	0 - 400	Clark (1966)

- ¹: Birch and Clark (1940) give data for bronzite and hypersthene. These data were fit with equation (3), the slopes were averaged, and the new slope was combined with the conductivity of augite at 25°C to give equation (3c).
- ²: Strictly speaking, water will convert to steam at 100°C at atmospheric pressure. The conductivity of steam will be less than that of water, so equation (3d) will overestimate conductivity at high temperatures by a factor of 25-30 (for an exact representation of the conductivity of water and steam, see Kestin, 1978).
- ³: Clark (1966) gives data for MgAl₂O₄, which has a structure similar to that of magnetite. The data were fit with equation (3) and the derived slope was combined with the conductivity of magnetite at 25°C to give equation (3e).
- ⁴: Clark (1966) gives data for pyrophyllite, which has a structure somewhat similar to that of chlorite. The data were fit with equation (3) and the derived slope was combined with the conductivity of chlorite at 25°C to give equation (3f).

Table F-8. Calculated Thermal Conductivity of the Pomona Flow; Effects of Varying Temperature

Conditions	Variable	k(10 ⁻³ cal/cm sec°C)
i) P = 0.1 MPa Ø _T = 6% Ø _E = 1.5%	T = 25°C	3.611
	50	3.580 (-0.8%)
	100	3.415 (-5.4%)
	150	3.293 (-8.8%)
	200	3.195 (-11.5%)
ii) Chlorite and magnetite conductivities assumed constant		
P = 0.1 MPa Ø _T = 6% Ø _E = 1.5%	T = 25°C	3.611
	50	3.666 (+1.5%)
	100	3.591 (-0.5%)
	150	3.529 (-2.3%)
	200	3.477 (-3.7%)

Table F-9. Calculated Thermal Conductivity of Umtanum Entablature; Effects of Varying Temperature

Conditions	T(°C)	k(10 ⁻³ cal/cm sec°C)
P = 0.1 MPa Ø _T = 6% Ø _E = 1.5%	25	4.277
	50	4.298 (+0.5%)
	100	4.208 (-1.6%)
	150	4.169 (-2.5%)
	200	4.151 (-2.9%)

F.2 Discussion

The values in Table F-8 are plotted in Figure F-1. Neither of the empirical estimates of conductivity show an increase with temperature, although the predicted values show good agreement with the data of Duvall and others (1978) and Martinez-Baez and Amick (1978). Schmidt and others (1980) suggest a possible correlation between the volume fraction of glass and the behavior of conductivity with temperature. Calculations for the Umtanum entablature, which contains approximately 40 volume % glass, still show decreasing conductivity with temperature (Table F-9).

The lack of agreement between predicted and measured conductivities does not seem to support the use of empirical equations to predict thermal conductivity for the Hanford basalts. However, the discrepancies could have many causes, including inaccurate modal estimates or porosity measurements, failure to report axial pressures during conductivity tests, or incorrect characterization of the extent of saturation. All of these items are crucial to both the accurate determination and accurate prediction of thermal conductivity.

The benefit gained from the preceding analysis of parameters which can affect thermal conductivity is the insight into the relative effects each parameter can have. Of the parameters examined, the most important to properly characterize during the determination of thermal conductivity seem to be pressure, porosity and saturation state. As has been found experimentally (see Figure 1), temperature variation is predicted to be relatively unimportant to determination of conductivity.

In-situ conductivity will be a function of the conductivity of the intact basalt and the nature and geometry of the fracture system. The basalt conductivity is affected by all of the parameters described above. The effect of the fracture system will depend on fracture spacing and aperture and the nature of the material in the fracture (i.e., air, water, or secondary minerals). Open dry fractures will cause a sharp decrease in the thermal conductivity of the rock mass, with water causing a much smaller decrease and secondary minerals having relatively little effect. Section 2.1.4

contains a brief description of the nature of the fracture system in the Columbia River basalts.

Table F-10 summarizes the thermal conductivity data investigated. These data and those referenced in the preceding discussion are laboratory measurements on intact rock. The range suggested below is taken from these data, but is believed to be accurate for in-situ conditions as well, unless the site under examination contains numerous open dry fractures, in which case the in-situ conductivity may be even lower than 2.0×10^{-3} cal/cm sec°C.

No data have been found for the conductivity of interbeds or interflows. However, the choice of a thick flow interior for the location of a subsurface facility will make the contributions of units other than basalt to the thermal regime relatively unimportant.

Suggested Range: $2.0 - 6.0 \times 10^{-3}$ cal/cm sec°C;
no recommended function of temperature; rocks with higher (effective porosity)/(total porosity) ratio should have higher conductivity.

Table F-10. Thermal Conductivity of Basalt

Source	Columbia Plateau Location or Basalt Unit	$k(10^{-3} \text{ cal/cm sec}^\circ\text{C})$	T(°C)
Atlantic Richfield Hanford Co. (1976)	Columbia River basalt	$k = \frac{C_p^1}{p}$	
Colorado School of Mines (1978)	Pomona Flow	1.56 - 7.15	94-303
Duval and others (1978)	Pomona Flow	3.03 - 3.84	150-250
Erikson and Krupka (1980)	Pomona Flow	3.65 - 4.49	100-300
Foundation Sciences, Inc. (1980a)	Pomona Flow	4.30 - 6.07	31-300
Foundation Sciences, Inc. (1980b)	Pomona Flow	2.48 - 6.00	41-299
Foundation Sciences, Inc. (1981)	Umtanum Flow	4.35 - 6.01	56-305
Hocking and St. John (1979)	Hanford basalt	3.58	unknown
Johnston and Palmer (1981)	Survey, Hanford basalt	3.34 - 9.98	unknown
Kaser (1979)	Hanford basalt	3.56	unknown
Krupka (1974)	Columbia River basalt	2.58 - 5.42	13-985
Martinez-Baez and Amick (1978)	Pomona and Umtanum Flows	3.46 - 4.18	54-125
Murase and McBirney (1973)	Picture Gorge tholeiite	2.60 - 5.0	100-900
Schmidt and others (1980)	Summary, Hanford basalt	NA	NA
Skvarla and others (1981)	Hanford basalt	2.51 - 2.55	0-212
Taylor and others (1980)	Survey, Hanford basalt	NA	NA

Table F-10 (Continued)

Source	Location or Basalt Unit	$k(10^{-3} \text{ cal/cm sec}^\circ\text{C})$	$T(^{\circ}\text{C})$
	<u>Other Localities</u>		
Adams and Watts (1975)	Hawaii	3.64 - 4.88	unknown
Bayuk and others (1975)	Indian Ocean	3.70 - 4.75	unknown
Bridgman (1924)	Unknown	$k = f(P)$	30-75
Hyndman and Drury (1976)	Mid-ocean ridge basalt	3.97 ± 0.04	21
Kawada (1964)	Japan (Kyushu)	3.2 - 4.9	111-622
Krupka (1974)	Dresser basalt	3.58 - 7.17	100-900
Lehnhoff and Scheller (1975)	Dresser basalt	$k = .0089 - .0031 \ln T$ $+ .00091 \ln T^2$ $- .000083 \ln T^4$	
Marovelli and Veith (1965)	Dresser basalt	0.33 - 0.57	21-944
Morgan and West (1980)	Dresser basalt	6.47 - 7.58	46-267
Oxburgh and Griffiths (1976)	Deep Sea Drilling Project	2.52 - 3.73	23-27
Petrudin and others (1971)	U.S.S.R	3.35 - 8.90	77-977
Poole (1914)	Unknown	3.76 - 4.10	47-600
Robertson and Peck (1974)	Hawaii	0.21 - 5.75	25

Table F-10 (Continued)

Source	Location or Basalt Unit	$k(10^{-3} \text{ cal/cm sec}^\circ\text{C})$	T($^\circ\text{C}$)
	<u>Other Localities</u>		
Robie and others (1970)	Lunar basalt	4.0	unknown
Somerton and others (1963)	Mohole test site basalt	3.13 - 4.93	24-78
Stephens (1963)	Nevada Test Site	3.24 - 4.60	172-778
Tokuyama and Yoshida (1979)	Japan	4.28 - 4.44	20-23
Washburn (1927)	Japan	< 3.35	unknown
Watts (1975)	Hawaii	3.3 - 5.4	21-46
<u>Citations or surveys of previous literature</u>			
Agapito and others (1977)	Literature range	3.34 - 10.22	unknown
AVCO Everett Research Lab (1972)	Equation derived from data of Marovelli & Veith (1965)	$k = (1.45 \times 10^{-5})$	T($^\circ\text{K}$) - 15
Clark (1966)	Literature survey	4.1 - 9.0	20-75
Clark and Caudle (1961)	Cites Handbook of physical constants (1942)	NA	NA
Dames and Moore (1978)	Cites Marovelli and Veith (1965)	NA	NA
	Atlantic Richfield (1976)		

Table F-10 (Continued)

Source	Location or Basalt Unit	k(10 ⁻³ cal/cm sec°C)	T(°C)
<u>Citations or surveys of previous literature (Cont'd)</u>			
Dubois and others (1979)	Cites Martinez-Baez and Amick (1978)	NA	NA
Hasan (1978)	Literature range	0.3 - 7.4	unknown
Nafe and Drake (1968)	Literature range	4.1 - 6.1	0-100
Ratigan (1976)	Literature average	4.09 ^a	unknown
Walsh and Decker (1966)	Theoretical treatment	NA	NA
Woodside and Messmer (1961)	Theoretical treatment	NA	NA
<u>Other</u>			
Birch and Clark (1940)	Olivine	8.1 - 12.4	0-200
	Plagioclase, (An ₁₂₋₈₀)	4.1 - 4.8	0-200
	Pyroxene	8.0 - 12.0	0-200
Horai (1971)	Chlorite	10.38 - 14.77	25
	Opal	2.90	25
Horai and Simmons (1969)	Apatite	3.28 - 3.31	25
	Augite	9.13	25

Table F-10 (Continued)

Source	Location or Basalt Unit	$k(10^{-3} \text{ cal/cm sec}^\circ\text{C})$	$T(^{\circ}\text{C})$
<u>Citations or surveys of previous literature (Cont'd)</u>			
	Ilmenite	5.25 - 6.10	25
	Magnetite	12.18	25
	Plagioclase (An ₅₄)	3.65	25

a = average

NA = not applicable

¹ α = thermal diffusivity (cm^2/sec), ρ = density (g/cm^3), C_p = specific heat ($\text{cal/g } ^\circ\text{C}$)

APPENDIX G

Specific Heat

Values reported for the specific heat of intact Hanford basalts as a function of temperature are shown in Figure G-1. With the exception of the data of Duvall and others (1978) for the Pomona flow, the data are reasonably consistent. Duvall and others (1978) reported only the equations of regression lines which had been fit to experimental data. Since the measured data are not reported, no explanation of the discrepancy with all other reported values is possible.

Theoretical Considerations

Either specific heat or heat capacity may be calculated from the oxide composition of the rock. The equation is as follows:

$$C_p \text{ (rock)} = \sum_{i=1}^n X_i C_p(i) \quad (1)$$

where X_i = mole fraction of component i $\left(\sum_{i=1}^n X_i = 1 \right)$

and C_p = either specific heat or heat capacity.

This calculation was performed for four representative compositions of the Pomona and Umtanum flows reported in Schmidt and others (1980). The resulting curves also are plotted on Figure G-1. The data points used to plot the curves, together with the function derived from the C_p equations of the oxides, are given in Table G-1.

The degree of saturation of the test samples used to obtain the data in Figure G-1 is not specified. The specific heat of basalt with a fully saturated effective porosity of 1.5 percent may be 6 percent higher than the specific heat of dry basalt. A fully saturated total porosity of 6 percent may lead to a specific heat which is higher by 25 percent. Even these higher specific heats are in general similar

1. Duval and others (1978); Pomona Flow
2. Martinez-Baez and Amick (1978); Calculated Values
3. Values calculated in this study

Figure G-1. Specific Heat of Basalt Versus Temperature

to the calculated and measured values of C_p . Similar reasoning may be used to estimate the effect of fractures on in-situ specific heat.

No data have been found for the specific heat of interbeds or interflows. As described for thermal conductivity, this deficiency is not considered to be important if the subsurface facility is located in a thick flow interior.

<u>Suggested Range:</u>	<u>T(°C)</u>	<u>C_p (cal/g°C)</u>
	50	0.175 - 0.230
	100	0.185 - 0.250
	150	0.200 - 0.270
	200	0.205 - 0.280

Table G-1. Calculated Specific Heats of Pomona and Umtanum Flows

Specific heat (cal/g °C) calculated using equations below

<u>T(°C)</u>	<u>Pomona*</u>		<u>Umtanum*</u>	
50	0.1919	0.1920	0.1897	0.1898
100	0.2068	0.2068	0.2043	0.2045
150	0.2184	0.2184	0.2158	0.2161
200	0.2281	0.2281	0.2255	0.2258

Pomona flow: $C_p = 0.1975 + .0001029 T(^{\circ}\text{K}) - \frac{4056.9}{T^2}$

Umtanum flow: $C_p = 0.1931 + .0001057 T(^{\circ}\text{K}) - \frac{3916.1}{T^2}$

*Values plotted on Figure G-1 are averages of the two points for each flow at each temperature.

Table G-2. Specific Heat of Basalt

Source	Columbia Plateau Locality or Basalt Unit	C_p (cal/g°C)	T (°C)
Atlantic Richfield Hanford Company (1976)	Columbia River basalt	0.174-0.234	38-316
Duval and others (1978)	Pomona flow	0.288-0.585	100-300
Erikson and Krupka (1980)	Pomona flow	$C_p = 0.3062 - \frac{2.58 \times 10^{-5} T}{11533 - T}$	25-250
Foundation Sciences, Inc. (1980a)	Pomona flow	0.196-0.251	73-356
Foundation Sciences, Inc. (1980b)	Pomona flow	0.200-0.243	74-358
Foundation Sciences, Inc. (1981)	Umtanum flow	0.219-0.243	74-347
Johnston and Palmer (1981)	Survey, Hanford basalt	0.170-0.409	unknown
Kaser (1979)	Hanford basalt	$C_p = 0.21 + 0.000386 T$ (°C)	
Martinez-Báez and Amick (1978)	Pomona and Umtanum flows	0.217-0.296	50-300
Schmidt and others (1980)	Summary, Hanford basalts	$C_p = 0.2 + 0.0002T$	
Taylor and others (1980)	Survey, Hanford basalts	NA	NA
<u>Other Localities</u>			
Bayuk and others (1975)	Indian Ocean	0.19-0.24	unknown
Lehnhoff and Scheller (1975)	Dresser basalt	$C_p = 0.22 \exp [0.0011T + 0.00001T^2 - 0.22 \times 10^{-7}T^3 + 0.12 \times 10^{-10}T^4]$	unknown
Leonidov (1967)	U.S.S.R	0.208-0.281	100-800

Table G-2. (Cont'd)

Source	Locality or Basalt Unit	C_p (cal/g°C)	T (°C)
<u>Other Localities (Cont'd)</u>			
Lindroth and Krawza (1971)	Dresser basalt	(a) $C_p = 0.2176 + 0.000123T$ (b) $C_p = 0.2585$	(100-575) (575-1000)
Peck and others (1977)	Hawaii	$C_p = 0.2356 + 4.3635 \times 10^{-5} T(^{\circ}K) - 6344/T^2 (^{\circ}K)$	
Poole (1914)	unknown	$C_p = 0.188 + 0.265 \times 10^{-3} T(^{\circ}C) - 0.25 \times 10^{-6} T^2 (^{\circ}C)$	
Robie and others (1970)	Lunar basalt	0.1711-0.198	7-87

Citations or surveys of previous literature

Agapito and others (1977)	Literature range	0.227-0.251	unknown
Dames and Moore (1978)	Cites Atlantic Richfield (1976), Lindroth and Krawza (1971)	NA	NA
Dubois and others (1979)	Cites Martinez-Baez and Amick (1978)	NA	NA
Krupka (1974)	Cites Lindroth and Krawza (1971)	NA	NA
Maurer (1968)	Literature average (?)	0.24	unknown
Nafe and Drake (1968)	Literature range	0.170-0.201	0
Ratigan (1976)	Literature average	0.36	unknown

a = average

NA = not applicable

APPENDIX H

Thermal Expansion

Data reported for the linear thermal expansion coefficient (α) of intact Hanford basalts are plotted on Figure H-1. The range of α is broad, and α is relatively constant with increasing temperature. For any single sample, α has been found to increase, decrease, or remain constant with temperature.

In many cases, α is reported as an average value for the temperature range in which the experiment was performed; the maximum temperatures in these reports ranged from 357 to 640°C. Clark (1966) contains a brief discussion on the thermal expansion of rocks, one conclusion of which is that the measurement of α for temperatures greater than 100°C is of little value because the resulting data are complicated by the opening of microscopic fractures which would remain closed in environments having even a moderate confining pressure. If the data in Figure H-1 are examined with the 100° limitation in mind, the range of α measured for Hanford basalts becomes 4.0 to 7.4 x 10⁻⁶/°C, with no consistent trend with increasing temperature. Table H-1 lists the values for the thermal expansion coefficient found during our literature search.

Theoretical Considerations

It is suggested in Clark (1966) that α for temperatures above 100°C be calculated from the α value for the minerals in the rock. The equation would be:

$$\alpha_{\text{rock}}^{(T)} = \sum_{i=1}^n X_i \alpha_i (T) \quad (1)$$

where X_i = mole fraction of i^{th} mineral in the rock

$$\left(\sum_{i=1}^n X_i = 1 \right)$$

Figure H-1. Coefficient of Linear Thermal Expansion of Basalt.

and $\alpha(T) = \alpha$ at the temperature of interest. Unfortunately, no data exist for chlorite, clays, or basaltic glass, so that at present it is not possible to predict for basalt.

The data discussed in this Appendix have been collected in the laboratory. Any extrapolation to the in-situ rock mass should include consideration of the possible effects of large and small scale discontinuities.

No data have been found for the linear thermal expansion of interbeds and interflows. As discussed for thermal conductivity and specific heat, this deficiency is not considered to be important if the subsurface facility is located in a thick flow interior.

Suggested Range: 4.0 to $7.4 \times 10^{-6}/^{\circ}\text{C}$; no suggested trend with temperature.

Table H-1. Linear Thermal Expansion of Basalt

Source	Columbia Plateau Locality or Basalt Unit	$\alpha (10^{-6}/^{\circ}\text{C})$	T($^{\circ}\text{C}$)
Atlantic Richfield Hanford Company (1976)	Columbia River basalt	5.3 ± 1.3	NA
Colorado School of Mines (1978)	Pomona flow	5.23-10.65	unknown
Duvall and others (1978)	Pomona flow	6.5-6.7	20-300
Erikson & Krupka (1980)	Pomona flow	2.1-2.9	20-250
Foundation Sciences, Inc. (1980a)	Pomona flow	4.0-6.7	70-300
Foundation Sciences, Inc. (1980b)	Pomona flow	6.5-8.4	70-300
Foundation Sciences, Inc. (1981)	Umtanum flow	6.0-7.2	73-242
Hocking and St. John (1979)	Hanford basalt	6.5	unknown
Johnston and Palmer (1981)	Survey, Hanford basalt	2.9-12.0	unknown
Miller (1979a)	Umtanum flow	6.77-10.83	90-550
Miller (1979b)	Umtanum flow	7.34-11.96	98-497
Miller and Bishop (1979)	Umtanum flow	7.37-11.16	18-640
Schmidt and others (1980)	Summary, Hanford basalts	7.5 (unknown) 5.6 (Pomona)	NA NA
Taylor and others (1980)	Survey, Hanford basalt	NA	NA

Table H-1. (Continued)

<u>Source</u>	<u>Locality or Basalt Unit</u>	<u>$\alpha(10^{-6}/^{\circ}\text{C})$</u>	<u>T($^{\circ}\text{C}$)</u>
	<u>Other Localities</u>		
Griffin and Demou (1971)	Unknown	1.11-22.7	-137-941
Nakagawa and Mori (1963)	Japan	3.66-18.19	27-1027
Thirumalai (1970)	Dresser basalt	2.92-11.57	24-593

Citations or Surveys of Previous Literature

Agapito and others (1977)	Literature range	2.9-11.6	unknown
AVCO Everett Research Lab (1972)	Cites undefined source	NA	NA
Clark (1966)	Literature average	5.4+ 1 ^a	20-100
Dames and Moore (1978)	Cites Lehnhoff and Scheller (1975) Atlantic Richfield Hanford Company (1976)	NA	NA
Dubois and others (1979)	Cites Agapito and others (1978)	NA	NA
Lehnhoff and Scheller (1975)	Cites Thirumalai (1970)	NA	NA
Nafe and Drake (1968)	Literature range	4.2-6.3	unknown
Ratigan (1976)	Cites Clark (1966)	NA	NA

a = average

NA = not applicable

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APPENDIX I

Thermal Diffusivity

Experimentally determined values of diffusivity (α^*) of intact Hanford basalt as a function of temperature are given in Figure I-1. The data show no consistent trend. A broad range can be defined for temperatures below 150°C.

Theoretical Considerations

Thermal diffusivity, or the rate at which heat moves through a substance, is defined as

$$\alpha^* = k / \rho C_p \quad (1)$$

where k = thermal conductivity,

ρ = bulk density,

C_p = specific heat.

Because experimental data do not seem to produce any consistent trend, it is recommended that α^* be obtained through use of equation (1). Using the values of C_p for the Pomona and Umtanum flows and the ranges for ρ and k suggested in earlier appendices, the estimated values in Table I-1 are obtained. The assumption has been made that ρ and k are directly correlated, so that the lowest ρ and lowest k values are used together to calculate a value for α^* and so on. Table I-2 lists the entire range of values found in the literature.

Because thermal diffusivity and thermal conductivity are interrelated, the same caution as is described in Appendix F should be used in extrapolating thermal diffusivity values to in-situ conditions.

Suggested Range: 4.6 - 10.2 x 10⁻³ cm²/sec.

Figure I-1. Thermal Diffusivity of Basalt as a Function of Temperature.

Table I-1. Thermal Diffusivity Ranges Calculated
From Bulk Density, Specific Heat, and
Thermal Conductivity

<u>T(°C)</u>	<u>α^* (10^{-3} cm²/sec)</u>	
	<u>Pomona</u>	<u>Umtanum</u>
50	5.5 - 10.1	5.6 - 10.2
100	5.1 - 9.4	5.2 - 9.5
150	4.8 - 8.9	4.9 - 9.0
200	4.6 - 8.5	4.7 - 8.6

Table I-2. Thermal Diffusivity of Basalt

Source	Columbia Plateau Locality or Basalt Unit	α^* ($10^3 \text{ cm}^2/\text{sec}$)	T($^{\circ}\text{C}$)
Atlantic Richfield Hanford Company (1976)	Columbia River basalt	4.41-5.32	38-316
Erikson and Krupka (1980)	Pomona flow	$\alpha = K/\rho C_p$	
Foundation Sciences, Inc. (1980a)	Pomona flow	5.25-7.61	20-320
Foundation Sciences, Inc. (1980b)	Pomona flow	3.85-7.88	20-320
Foundation Sciences, Inc. (1981)	Umtanum flow	3.29-7.96	24-303
Hocking and St. John (1979)	Survey, Hanford basalt	4.12	unknown
Krupka (1974)	Columbia River basalt	4.08-6.16	211-1164
Taylor and others (1980)	Survey, Hanford basalt	NA	NA
<u>Other Localities</u>			
Bayuk and others (1975)	Indian Ocean	6.0-7.56	unknown
Hanley and others (1978)	Dresser basalt	5.76-12.10	27-727
Kanamori and others (1969)	unknown	4.93-7.21	75-500
Krupka (1974)	Dresser basalt	3.77-9.80	127-1164
Washburn (1927)	Japan	7.00	unknown

Table I-2. (Cont'd)

<u>Source</u>	<u>Columbia Plateau Locality or Basalt Unit</u>	<u>α^* ($10^3 \text{ cm}^2/\text{sec}$)</u>	<u>T($^{\circ}\text{C}$)</u>
<u>Citations or Surveys of Previous Literature</u>			
Agapito and others (1977)	Literature range	5.2-8.0	unknown
Dubois and others (1979)	Cites Martinez-Baez and Amick (1978)	NA	NA
Ratigan (1976)	Literature average	4.4 ^a	NA

a = average

NA = not applicable

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APPENDIX J

Compressive Strength

Vutukuri and others (1974) present evidence that the following parameters can affect compressive strength: pressure, temperature, total porosity, moisture content, and strain rate. In addition, pore pressure, stress history, orientation and the presence or absence of fractures in test samples can have significant effects. A number of these parameters are discussed briefly below. In most cases, data are too sparse to draw definitive conclusions. Unless otherwise specified, all discussion refers to Hanford basalts.

A histogram of uniaxial* strengths of basalt at 20-25°C, unsaturated conditions, and low** strain rates is presented as Figure J-1. This figure also includes data for basalts from other areas of the world. No effort has been made to separate fractured and intact samples, so that Figure J-1 is a better representative of the distribution of strengths for an in-situ rock mass than for intact basalt.

The relationship between compressive strengths and confining pressures of Columbia River basalts as reported in the literature is shown in Figure J-2. No strong correlation exists between the two variables. The lack of information on sample porosities and densities precludes any attempt to improve the correlation. The only information to be gained from Figure J-2 is that fractured samples, in general, have lower compressive strengths than intact samples, and there seems to be an increase in the maximum strength with increasing confining pressure.

*In this and succeeding Appendices, a confining pressure of 14.7 psi represents atmospheric, or uniaxial conditions.

**"Low" here is meant to imply rates less than or equal to 1000 $\mu\epsilon/\text{min}$.

Figure J-1. Histogram of Uniaxial Compressive Strength of Intact and Fractured Basalt.

Figure J-2. Relationship Between Compressive Strength and Confining Pressure for Columbia River Basalts.

A number of failure criteria have been proposed to relate compressive strength, shear strength and confining pressure. Of these, the most commonly used criteria in the reports on Hanford basalts are the Coulomb criterion and a linear version of the Mohr criterion:

Coulomb:
$$\sigma_1 = C_0 + q\sigma_3$$

where

$$\begin{aligned} \sigma_1 &= \text{axial stress at failure,} \\ C_0 &= \text{uniaxial compressive strength,} \\ \sigma_3 &= \text{confining pressure,} \\ q &= [\tan^2 \phi + 1]^{1/2} + \tan \phi]^2, \text{ and} \\ \phi &= \text{angle of internal friction;} \end{aligned}$$

Mohr:
$$T = C + \sigma_3 \tan \phi$$

where

$$\begin{aligned} T &= \text{shearing stress, and} \\ C &= \text{cohesion.} \end{aligned}$$

Table J-1 summarizes the data available in the literature on Hanford basalts, as well as some data for other basalts. The Coulomb criterion is used by Foundation Science, Inc., (1980a, 1981) (FSI), but the data show so much scatter that the validity of a linear fit is questionable, and the use of the values for C and ϕ as derived in these papers is not recommended. In addition, a number of ϕ values are derived for situations in which data are inconsistent (e.g., Miller, 1979a). Such ϕ values are indicated in the table.

Although basalt is considered to be a brittle material, time dependent deformation is possible. No data exist for Hanford basalts. Sharp (1972) found creep rates of $1.9 \mu\epsilon/\text{hr}$ for dense basaltic flow breccia and $1.0 \mu\epsilon/\text{hr}$ for dense basalt from Amchitka Island. Brodsky and Spetzler (1979) observed that no creep deformation occurred until a differential stress of approximately 36,000 psi was reached or exceeded. Both of these studies were at room temperature. Higher temperatures would probably enhance creep rate, but more data are needed.

Table J-1. Values of Cohesion and Angles of Internal Friction for Basalt

Flow or Locality	Failure Criterion	C (psi)	ϕ (°)	Confining Pressure	Source
Pomona flow	M	NG	28.4	0-3000	Colorado School of Mines (1978)
Pomona flow	M	NG	49.0	0-5000	Colorado School of Mines (1978)
Pomona flow	M	NG	58.0	1000-3000	Colorado School of Mines (1978)
Pomona flow	M	NG	45.0	2000-5000	Colorado School of Mines (1978)
Pomona flow	M	NG	43.5	500-5000	Colorado School of Mines (1978)
Pomona flow	M	NG	45.0	2000-5000	Colorado School of Mines (1978)
Umtanum flow	M	NG	41.2 ^c	0-7500	Miller (1979a)
Umtanum flow	M	NG	52.3 ^c	0-5000	Miller (1979a)
Umtanum flow	M	NG	49.2 ^c	0-7500	Miller (1979a)
Umtanum flow	M	NG	46.6 ^c	0-10000	Miller (1979a)
Umtanum flow	M	NG	47.4 ^c	0-7500	Miller (1979a)
Umtanum flow	M	NG	49.4 ^c	0-10000	Miller (1979a)
Umtanum flow	M	NG	44.4	50-2500	Miller and Bishop (1979)
Umtanum flow	M	NG	38.7 ^c	50-7500	Miller and Bishop (1979)
Umtanum flow	M	NG	36.4 ^c	50-5000	Miller and Bishop (1979)
Pomona flow	Co	51900 ^a	53 ^a	0-2000	FSI (1980a)
Pomona flow	M	NG	52+3	0-2000	FSI (1980a)
Pomona flow	M	NG	54+3	0-2000	FSI (1980b)
Umtanum flow	Co	NG	41 ^a	0-2000	FSI (1981)
Buckboard Mesa basalt (dense)	Co	4520	45	700-4000	Nugent and Banks (1965)
Buckboard Mesa basalt (vesic.)	Co	1720	28	200-1200	Nugent and Banks (1965)
Buckboard Mesa basalt	M	NG	45	1000-4000	Lutton (1968)
NTS basalt	M-Co	4003	64.0	0-500	Stowe (1969)
NTS basalt	M-Co	9602	31.0	500-5000	Stowe (1969)
NTS basalt	M	3250-3900	50-55	1000 ^b	Stowe and Ainsworth (1972)

M = Mohr

Co = Coulomb

NG = not given

a: Values derived from linear regression of highly scattered data.

b: Confining pressure constant but variable strain rates.

c: Values derived from attempts to fit inconsistent data.

Nugent and Banks (1965) found that increased confining pressure increased the compressive strength of basalt from the Nevada Test Site. Similar results were obtained by Lutton (1968) and Stowe (1969) for different samples of the same basalt, by Smith and Forristall (1973) for basalt cores recovered during the Deep Sea Drilling Project, by Lindholm and others (1974) for Dresser basalt, and by Friedman and others (1979) for Cuerbio basalt from New Mexico.

The effect of temperature on compressive strength is unclear. Figure J-3 contains histograms of strengths at different temperatures. Increasing temperature causes either a slight decrease or no change in the mean strength. The scatter in strength values tends to decrease at higher temperatures. FSI (1981) states that the strength of fractured specimens is not affected by temperatures to 300°C and that for samples of intact Pomona, strength tends to decrease at higher temperatures. Based on the limited* data shown in Figure J-3, the mean strength of intact Pomona samples increases with temperature, while the mean strength of jointed samples decreases by 9 to 18 percent. More data are needed before definitive relationships can be established.

Podnieks and others (1972) measured compressive strengths of Columbia River basalt from Oregon and found that strength definitely decreased at higher temperatures. Three other investigations have found that the compressive strength of other basalts decreases with increasing temperature: Griggs and others (1960) for basalt from the Blairsden quadrangle of California; Lindholm and others (1974) for Dresser basalt; and Friedman and others (1979) for Cuerbio basalt from New Mexico.

Vutukuri and others (1974) suggest that compressive strength should decrease with increasing porosity. This general trend apparently holds true for the few data points available for Hanford basalts, as shown in Figure J-4. The figure also shows a decrease in the effect of porosity on strength as confining pressure increases.

*The data are limited, but as far as is known to the authors, the FSI (1981) conclusions were based upon the same data set.

Figure J-3. Histogram Showing Relationship Between Compressive Strength and Temperature for Pomona and Umtanum Flows.

Figure J-4.. Relationship Between Compressive Strength and Total Porosity for Basalt.

However, more data are needed to differentiate the effects of porosity from the possible contributions of joint behavior to strength variations.

Lutton (1968) found that compressive strength of basalt from the Nevada Test Site decreases with increasing total porosity for three separate confining pressures.

The effect of moisture content on the compressive strength of Hanford basalt is equivocal at best. Strengths of eighteen Umtanum samples were measured at 20°C and 50 psi (0.3 MPa), nine in dry and nine in saturated conditions (Miller, 1979a, Miller 1979b, Miller and Bishop, 1979). The mean strengths showed the saturated samples to be slightly stronger. However, when the sample strengths were averaged according to the locality from which the samples were obtained, two groups showed the dry samples to be stronger and only one group showed higher strengths under saturated conditions.

The comparison above does not account for either the possible effects of porosity or for the effects of strain rate and thus drained versus undrained conditions. When samples with bulk densities less than 2.60 g/cm³ are removed from consideration, mean dry strength is slightly higher than mean saturated strength. The difference is insignificant, especially considering the limited amount of data. Similarly ambiguous results were obtained by Podnieks and others (1972) on other Columbia River basalt samples (see Figure J-5). More data are needed to determine the effects of moisture content on strength.

Ruiz (1966) examined the effect of moisture content on the compressive strength of five basalts from Brazil. Four of the five showed lower strengths under saturated conditions. Tomashevskaya and Volodina (1976) state that water-saturated samples of a doleritic basalt from the U.S.S.R. had 20 to 25 percent lower strengths than dry samples; no supporting data are given.

The relationship between strain rate and compressive strength is shown in Figure J-6. There is no pattern which describes either the strain rate-strength relationship in general or the relationship as a function of temperature. As with other parameters, the lack of a pattern may result from a variety of porosity values.

J-10

Figure J-5. Uniaxial Compressive Strength as a Function of Moisture Content for Basalt.

Figure J-6. Relationship Between Mean Uniaxial Compressive Strength and Strain Rate for Umtanum Flow.

Figure J-7 shows the same diagram, this time using those samples with bulk densities greater than 2.60 g/cm^3 . Again, no pattern is apparent. Until more data are available which specify the presence or absence of fracturing and the total porosity, no conclusions are possible concerning the effect of strain rate on experimentally determined compressive strength.

Kumar (1968), Stowe (1969) and Stowe and Ainsworth (1972) found that the compressive strength of basalt from the Nevada Test Site increased as strain rate increased. Similar results were obtained by Lindholm and others (1974) for Dresser basalt.

One other effect mentioned by Vutukuri and others (1974) is that preheating or drying of samples before testing increases the compressive strength. However, limited data from Miller (1979a,b) and Miller and Bishop (1979) indicate that the mean strength of samples preheated to 500°C and cooled prior to testing is only 60 percent of the mean strength of unheated samples. This decrease in strength may be due to irreversible thermal microcracking with heating and resultant expansion, while the effect suggested by Vutukuri and others (1974) possibly is based on the idea that heating to low to moderate temperatures drives off water and dry rocks have higher strengths than rocks with a higher moisture content.

Maurer (1968) provides two compressive strengths for an unspecified basalt which show a decrease in strength with preheating. The maximum temperature is unknown.

The data discussed in this Appendix have been collected in the laboratory. Any extrapolation to the in-situ rock mass should include consideration of the possible effects of large and small scale discontinuities.

No data have been found for the compressive strength of interbeds or interflows.

* For samples with bulk densities greater than 2.60 g/cm^3 .

Figure J-7. Relationship Between Mean Uniaxial Compressive Strength and Strain Rates for Umtanum Flow Samples With Bulk Density Greater Than 2.60 g/cm^3 .

Suggested Range of Uniaxial Compressive Strength:

0-44,000 psi (0-303 MPa)	fractured
5400-60,000 psi (37-414 MPa)	intact

No further constraint on compressive strength is possible until more data are obtained on the specific effect(s) of each parameter discussed earlier. The data found in the literature are given in Table J-2.

Table J-2. Compressive Strength of Basalt
Columbia Plateau

<u>Source</u>	<u>Location or Basalt Unit</u>	<u>Strength (psi)*</u>	<u>T(°C)</u>	<u>Conf P(psi)</u>
Colorado School of Mines (1978)	Pomona flow	540-54850	Unknown	14.7
Coulson (1972)	Little Granite basalt	33000	Unknown	14.7
Deere and Miller (1976)	Columbia River basalt	33300-51500	Unknown	14.7
Duvall and others (1978)	Pomona flow	35533-54967	Unknown	14.7-2000
Foundation Sciences, Inc. (1980a)	Pomona flow	9500-72900	22-300	14.7-2000
Foundation Sciences, Inc. (1980b)	Pomona flow	9500-76200	23-300	14.7-2000
Foundation Sciences, Inc. (1980c)	Umtanum and Pomona flows	3100-54500	200	14.7-10000
Foundation Sciences, Inc. (1981)	Umtanum flow	8880-36500	23-300	14.7-2000
Johnston and Palmer (1981)	Survey, Hanford basalt	27991-58013	Unknown	14.7
Miller (1979a)	Umtanum flow	4710-83400	20-500	14.7-7500
Miller (1979b)	Umtanum flow	2660-91660	20-500	14.7-10000
Miller and Bishop (1979)	Umtanum flow	2980-127360	20-500	14.7-7500
Podnieks and others (1972)	Columbia River basalt	51875-85688	-157-100	14.7
Schmidt and others (1980)	Summary, Hanford basalts	1390 ^a -53000 ^a	20-500	14.7-5000
Stephenson and Triandafilidis (1974)	Columbia River basalt	32800-34920	Unknown	14.7
Taylor and others (1980)	Survey, Hanford basalt	NA	NA	NA
<u>Other Localities</u>				
Berdinelli (1962)	Nevada Test Site	20080-20270	Unknown	14.7
Blair (1956)	Eniwetok	28200	Unknown	14.7
Cohn and Ahrens (1978)	Ralston basalt	21755	Unknown	Unknown
Drennon and Handy (1975)	California	16797-25388	Unknown	14.7

*1 psi = .006895 MPa.

Table J-2 (Continued)

<u>Source</u>	<u>Location or Basalt Unit</u>	<u>Strength (psi)*</u>	<u>T(°C)</u>	<u>Conf P (psi)</u>
<u>Other Localities (Continued)</u>				
Friedman and others (1979)	New Mexico	9427-79768	25-1000	14.7-7252
Griggs and others (1960)	California	31865-240596	25-800	73500
Krech and others (1974)	Dresser basalt	59900-69500	Unknown	14.7
Kulhawy (1975)	Nevada Test Site	21465	Unknown	14.7
	Unknown	21189	Unknown	14.7
Kumar (1968)	Nevada Test Site	27500-67412	-196-25	14.7
Lindhohm and others (1974)	Dresser basalt	35000-201000	-193-1110	0-100000
Livingston (1961)	Unknown	10700	Unknown	14.7
Lutton (1968)	Nevada Test Site	6053-51622	Unknown	1000-5000
Morrell and Larson (1974)	Dresser basalt	63610	Unknown	14.7
Nordyke and Wray (1964)	Nevada Test Site	16783-34846	Unknown	14.7
Nugent and Banks (1965)	Nevada Test Site	3100-45960	Unknown	14.7-4000
Overton and Moczygemba (1967)	Texas	7550-61400	Unknown	Unknown
Price and Knill (1966)	Unknown	18945 ^a	Unknown	14.7
Ramana and Vekatanarayana (1973)	Kolar basalt	31250	Unknown	Unknown
Ricketts and Goldsmith (1970)	Unknown	37000-39400	Unknown	Unknown
Ruiz (1966)	Brazil	9245-32443	25	14.7
Sharp (1972)	Amchitka Island	10945-51671	Unknown	14.7
Smith and Forristall (1973)	Deep Sea Drilling Project	38100	Unknown	0-10000
Stowe (1969)	Nevada Test Site	18920-54000	Unknown	14.7-5000
Stowe and Ainsworth (1972)	Nevada Test Site	20960-34580	Unknown	14.7
Vortman (1962)	Nevada Test Site	5860-31970	Unknown	14.7
Washburn (1927)	Germany	28446-66848	Unknown	Unknown

Table J-2 (Continued)

<u>Source</u>	<u>Location or Basalt Unit</u>	<u>Strength (psi)*</u>	<u>T(°C)</u>	<u>Conf P (psi)</u>
<u>Other Localities (Continued)</u>				
Wawersik and Fairhurst (1970)	Unknown	51250	Unknown	14.7
Windes (1949)	Michigan	11800-52000	Unknown	14.7
Windes (1950)	Michigan	2490-49600	Unknown	14.7
<u>Citations or Surveys of Previous Literature</u>				
Agapito and others (1977)	Literature range	0-58013	Unknown	14.7
Clark (1966)	Literature survey	38514-80997	24	14.7-15141
Clark and Candle (1961)	Literature range	2390-49600	Unknown	Unknown
Dames and Moore (1978)	Cites, Deere and Miller (1966), Sharp (1972), Stephenson and Triandafilidis (1974), Vortman (1962), Stowe (1969)	NA	NA	NA
Lehnhoff and Scheller (1975)	Cites unknown source, Dresser basalt	42396	Unknown	14.7
Maurer (1968)	Literature average (?)	6685-22757	Unknown	Unknown
Ratigan (1976)	Literature average	25000 ^a	Unknown	14.7

a = average

NA = not applicable

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APPENDIX K

Tensile Strength

Figure K-1 is a plot of tensile strength as a function of bulk density and loading rate for Columbia River basalts. The data indicate a general increase of strength with density, but the correlation is weak. In addition, loading rate apparently has no effect on the measured strength. Figure K-2 reinforces the observations that bulk density and tensile strength are weakly correlated. The figure also shows that data available to date do not define any useful distribution of tensile strengths for Columbia River basalts.

Theoretical Observations

In their analysis of different methods of determining tensile strength, Vutukuri and others (1974) indicate that the following parameters affect the strength found by using the Brazilian test:* (1) ratio of sample thickness to sample diameter; (2) moisture content; and (3) total porosity. Thus any analysis of the validity of a set of tensile strengths should take these factors into account.

Of all the data available for Hanford basalts, only three measurements are supported by values for each of the parameters above. In general, total porosity has not been measured for tensile strength samples. The only possible conclusion at this time is that the samples with higher strengths probably have lower total porosity.

The lack of correlation between bulk density and tensile strength shown in Figures K-1 and K-2 indicates that other factors besides total porosity are also important. In general, however, little information is given concerning the characteristics of the test samples. More data are needed to establish relationships between tensile strength and other parameters.

*Any tensile strength test other than direct extension results in values which are higher than the actual strength. Thus all strengths for Hanford basalts are overestimated because they were determined by the Brazilian test.

Figure K-1. Relationship Between Tensile Strength, Loading Rate, and Bulk Density for Columbia River Basalts.

K-3

Figure K-2. Histogram Showing Relationship Between Tensile Strength and Bulk Density for Basalt.

Another parameter in tensile strength tests is the presence or absence of fractures or joints. Foundation Sciences, Inc. (1981) found that almost without exception failure occurred along a pre-existing plane of weakness (if one were present) and that fractured samples failed at significantly lower stresses than intact samples (see Figure K-1, Umtanum points). Thus the relevance of determining the tensile strength of intact basalt is not clear, as any failure due to tensile stresses apparently will occur along fractures. Table K-1 contains the values obtained from the literature.

The data discussed in this Appendix have been collected in the laboratory. Extrapolation to in-situ conditions should include consideration of the possible effects of large and small scale discontinuities. Specifically, any rock mass which contains an open fracture the dimensions of which are significant relative to the rock mass will have a tensile strength of zero. Thus caution is necessary in applying the data compiled here to an in-situ analysis.

No data have been found for interbeds or interflows.

Suggested Range

1000 - 3500 psi (6.9 - 24.1 MPa)	for intact basalt with bulk density of 2.30-3.10 g/cm ³ .
0 - 1000 psi (0 - 6.9 MPa)	for fractured basalt.

Table K-1. Tensile Strength of Basalt

Columbia Plateau

<u>Source</u>	<u>Location or Basalt Unit</u>	<u>Strength (psi)*</u>
Colorado School of Mines (1978)	Pomona flow	63-4190
Deere and Miller (1966)	Columbia River basalt	1610-2100
Duvall and others (1978)	Pomona flow	2175-4133
Foundation Sciences, Inc. (1980a)	Pomona flow	1950-3670
Foundation Sciences, Inc. (1980b)	Pomona flow	2470-3150
Foundation Sciences, Inc. (1981)	Pomona flow	80-1320
Johnston and Palmer (1981)	Survey, Hanford basalt	1798-3481
Schmidt and others (1980)	Summary, Hanford basalts	1210 ^a -3021 ^a
Taylor and others (1980)	Survey, Hanford basalt	NA

Other Locations

Berdinelli (1962)	Nevada Test Site	1690-1820
Grady and Hollenbach (1979)	Dresser basalt	18854
Griggs and others (1960)	California	17581
Krech and others (1974)	Dresser basalt	2560-3480
Kulhawy (1975)	Nevada Test Site	2625

*1 psi = .006895MPa

Table K-1 (Continued)

<u>Source</u>	<u>Location or Basalt Unit</u>	<u>Strength (psi)</u>
<u>Other Locations (Continued)</u>		
Morrell and Larson (1974)	Dresser basalt	1982
Nugent and Banks (1965)	Nevada Test Site	2369-3058
Price and Knill (1966)	Unknown	8331-9736
Ricketts and Goldsmith (1970)	Unknown	1190-1400
Sharp (1972)	Amchitka Island	951-3688
Stowe (1969)	Nevada Test Site	1790-2090
Stowe and Ainsworth (1972)	Nevada Test Site	1900 ^a

Citations or Surveys of Previous Literature

Agapito and others (1977)	Literature range	0-3336
Dames and Moore (1978)	Cites Berdinelli (1962), Deere and Miller (1966), Sharp (1972), Stowe (1969)	NA
Lehnhoff and Scheller (1975)	Cites unknown source, Dresser basalt	2195
Ratigan (1976)	Literature average	1665 ^a

a = average

NA = not applicable

APPENDIX L

Mechanical Properties of Joints or Fractures

As mentioned in previous appendices, the in-situ mechanical behavior of the basalts at Hanford may be dominated by the mechanical properties of the physical discontinuities. Such parameters as the compressive and tensile strengths of intact basalt will probably have little relevance to the problem of mechanical behavior.

Preliminary experimental work on the mechanical behavior of jointed samples from the Umtanum flow is reported in Miller (1979a,b) and Miller and Bishop (1979). Table L-1 summarizes ranges and mean values for strengths and function factors (ratio of joint shear stress to joint normal stress). All of the samples used in the testing were multiply jointed, and the orientations of joints which underwent slippage during the testing are not reported. Most joints were filled with secondary minerals. The frequency of occurrence of secondary mineral assemblages was as follows: montmorillonite > montmorillonite + opal + chalcedony > montmorillonite + chalcedony + zeolite(?) > montmorillonite + zeolite(?). One sample tested was apparently a flow-top breccia (sample TJ17 in Miller, 1979a). The data for the sample do not show any significant difference from the data for dense basalt.

Many factors can affect joint strength, including confining pressure, temperature, strain rate, presence or absence of secondary minerals, type of secondary minerals, porosity of fracture filling, saturation state of fracture filling and spatial continuity of joint surfaces. More data are needed before the relative importance of these factors can be evaluated for the Hanford site.

Colson (1972) performed a study of the properties of artificial joints (saw cuts) in various rock types. Two conclusions potentially are relevant to this study: (1) residual friction was higher than initial friction, indicating that the creation of surface damage and/or separate small particles along the joint surface reduced the potential for shearing along the joint, and (2) residual friction was higher for dry surfaces than for saturated surfaces, indicating that the presence of

Table L-1. Friction Factors for the Umtanum Flow

<u>Number of Samples</u>	<u>Confining P (psi)</u>	<u>T(°C)</u>	<u>Range</u>	<u>Mean</u>
9	1000	-	0.821-1.583	1.117
8	2000	-	0.834-1.638	1.127
8	3000	-	0.799-1.347	0.971
6	5000	-	0.678-1.140	0.906
<hr/>				
4	1000 ^s	-	0.685-3.029	1.532
3	2000 ^s	-	0.630-2.152	1.243
3	3000 ^s	-	0.633-1.934	1.114
3	5000 ^s	-	0.560-1.617	0.935
<hr/>				
13	-	20	0-3.185	1.318
12	-	150	0.169-2.062	1.008
12	-	300	0.169-1.815	0.966
12	-	500	0.175-1.391	0.948

s: joint surface water-saturated.

Sources of data: Miller (1979a,b), Miller and Bishop (1979)

Table L-2. Joint Strengths for the Umtanum Flow

<u>Number of Samples</u>	<u>Strength (psi)</u>	
	<u>Range</u>	<u>Mean</u>
21	0-1449	377
3 ^s	177-4021	2042

s: joint surface water-saturated.

Sources of data: Miller (1979a,b), Miller and Bishop (1979)

water may enhance the potential for shear movement along joints. Thus "clean" (i.e., unaltered basalt) joints may undergo shearing more easily than those joints which are filled with fine-grained material, although the applicability of this inference to joints filled with clay is uncertain. Also, tests performed on dry joints may not be conservative, since Coulson's study suggests that joints will undergo shearing more easily in the presence of water than when dry.

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APPENDIX M

Elastic Constants

In order to facilitate representation of the deformational behavior of rocks, many are assumed to undergo elastic deformation. The elastic behavior of rocks may be described by defining some number of elastic constants; the greater the anisotropy, the more constants are necessary. In most instances, rocks are assumed to be isotropic on a large scale, so that only two constants are necessary. The most commonly measured constants are Young's modulus (E) and Poisson's ratio (ν). Other frequently encountered constants are bulk modulus (K), shear modulus (G) and Lamé's constant (λ). These are related to E and ν by the following formulae:

$$K = E/3(1-2\nu) \quad (1)$$

$$G = E/2(1+\nu) \quad (2)$$

$$\lambda = \nu E/(1+\nu)(1-2\nu) \quad (3)$$

In most of the references examined, basalt has been assumed to obey equations (1)-(3).

Lama and Vutukuri (1978) state that the following parameters can affect the measured values of the listed elastic constants:

<u>Parameter</u>	<u>Constants Affected</u>
Pressure	E, ν , K
Temperature	E
Strain rate	E
Total porosity	ν

In fact, all of the elastic constants are functions of the four listed variables because of the constitutive relationships given above.

The effects of each of the parameters above is briefly discussed below. Unless stated otherwise, discussion concerns the behavior of intact Hanford basalts.

Figures M-1 through M-4 are histograms of static and dynamic values of E , ν , G , and K , respectively. The constants were determined under ambient conditions in the laboratory. Each constant seems to have a relatively well-defined range and distribution, with the possible exception of static E .^{*} In the cases of E and ν , the peak is at slightly higher values for the dynamic data.

Although the data are limited, Figures M-1 through M-4 show no well-defined distribution for any elastic constant for basalts from localities other than the Hanford site.

The effects of pressure on E and ν are represented in Figures M-5 and M-6 and Table M-1. The values of E and ν plotted in the figures are mean values. The number of samples at any given pressure varies from 3 to 6. For the Pomona flow, ν decreases with increasing pressure, as does E for fractured Pomona samples. However, no similar pattern is defined by data for the Umtanum flow. More measurements are necessary before any definite trend can be delineated.

Nugent and Banks (1965) and Lutton (1968) found that in general E for basalt from the Nevada Test Site increases with confining pressure, though the trends are not smooth. In addition, in comparing laboratory data with in-situ measurements, Nugent and Banks (1965) estimated that E in-situ will be roughly 5% of measured laboratory values due to the presence of large scale discontinuities.

Christensen (1973a,b), Christensen and others (1974) and Carlson and Christensen (1977) determined that increasing pressure caused an increase in E , K , and λ and a slight increase in ν and G for basalts recovered during the Deep Sea Drilling Project. Balashov and Volarovich (1971) demonstrated that K for a basalt from the U.S.S.R. increased with pressure, with most

^{*} Static refers to low strain rates ($< 1000 \mu\epsilon/\text{sec}$); dynamic values here are obtained using ultrasonic wave frequency measurements, not high strain rates.

Figure M-1. Histogram of Dynamic and Static Young's Modulus for Basalt.

Figure M-2. Histogram of Dynamic and Static Poisson's Ratio for Basalt.

Figure M-3. Histogram of Dynamic Shear Modulus for Basalt.

Figure M-4. Histogram of Dynamic Bulk Modulus for Basalt.

Figure M-5. Relationship Between Mean Poisson's Ratio, Mean Young's Modulus, and Confining Pressure for Pomona Flow.

Figure M-6. Relationship Between Mean Poisson's Ratio, Mean Young's Modulus, and Confining Pressure for Umtanum Flow.

Table M-1. Relationships of Young's Modulus and Poisson's Ratio to Pressure for Hanford Basalts

<u>Pomona Flow</u>						
<u>Intact Samples</u>						
<u>E (10⁶ psi)</u>			<u>ν</u>			<u>Conf. P (psi)</u>
<u>Mean</u>	<u>Range</u>	<u>Number Samples</u>	<u>Mean</u>	<u>Range</u>	<u>Number Samples</u>	<u>Sources</u>
10.09	3.91-13.90	95	.262	.131-.540	95	0 Duvall and others (1978), CSM (1978), FSI (1980a,b)
12.76	10.8-14.1	7	.298	.255-.323	7	100 FSJ (1980a,b)
12.49	9.99-13.7	17	.252	.212-.272	17	500 CSM (1978), FSI (1980a,b)
11.61	7.84-13.7	25	.249	.186-.289	25	1000 CSM (1978), FSI (1980a,b)
10.43	5.20-17.64	3	.259	.185-.385	3	1500 CSM (1978)
11.53	6.83-16.81	30	.260	.194-.513	30	2000 CSM (1978), FSI (1980a,b)
10.08	5.72-12.69	11	.261	.221-.302	11	3000 CSM (1978)
10.65	8.42-12.54	16	.249	.193-.327	16	5000 CSM (1978)
9.33	-	1	.193	-	1	7500 CSM (1978)
<u>Printed Samples</u>						
12.24	10.9-13.5	22	.249	.179-.277	22	0 FSI (1980a,b)
12.88	11.3-16.1	14	.318	.282-.404	14	100 FSI (1980a,b)
12.76	10.2-15.9	26	.257	.220-.334	26	500 FSI (1980a,b)
12.45	9.1-13.4	25	.255	.217-.290	25	1000 FSI (1980a,b)
12.01	7.1-13.6	25	.255	.217-.285	25	2000 FSI (1980a,b)
<u>Umtanum Flow</u>						
<u>Jointed Samples</u>						
11.37	10.4-12.3	7	.271	.206-.384	7	0 FSI (1981)
9.47	7.1-11.2	3	.257	.144-.348	3	100 FSI (1981)
12.29	11.2-13.6	9	.264	.213-.306	9	500 FSI (1981)
12.34	11.4-13.7	8	.266	.214-.310	8	1000 FSI (1981)
12.70	11.5-15.1	7	.261	.231-.311	7	2000 FSI (1981)

CSM: Colorado School of Mines

FSI: Foundation Sciences, Inc.

of the change occurring in the pressure interval at 0-28500 psi. Hughes and Maurette (1957) found that G, E and K all increased slightly with increasing pressure.

Figures M-7, M-8 and Table M-2 show the relationship between temperature and E and ν . In general, ν decreases with temperature, while E behaves erratically. Foundation Sciences Inc. (1980a,b, 1981) attribute the decrease in ν to stress cycling, not to the temperature increases. However, no definite conclusion seems warranted until more data are obtained.

Griggs and others (1960) found that E for basalt from the Blairsdien quadrangle in California decreased with temperature. Similar results were obtained by Hughes and Maurette (1957) for a basalt, the origin of which was unknown, by Nakagawa and Mori (1963) for a basalt from Japan, by Podnieks and others (1972) for Columbia River basalt from Oregon, and for Dresser basalt by Lehnhoff and Scheller (1975) and Wingquist (1976).

The behavior of other constants with temperature varies. Nakagawa and Mori (1963) found ν to increase, while Lehnhoff and Scheller (1975) and Wingquist (1976) found it to decrease. Decreases in G (Hughes and Maurette, 1957; Nakagawa and Mori, 1963; Wingquist, 1976) and K (Hughes and Maurette, 1957; Nakagawa and Mori, 1963) have been demonstrated, though Bridgman (1924) found no variation in compressibility ($1/K$) between 30° and 75°C.

No data exist relating strain rate and the elastic constants of Hanford basalts. Lama and Vutukuri (1978) suggest that E should increase as strain rate increases. This was demonstrated by Stowe and Ainsworth (1972) for basalt from the Nevada Test Site, but other investigators (Stephenson and Triandafilidis, 1974; Brodsky and Spetzler, 1979) found the behavior of E to be variable as strain rate was increased. In addition, Brodsky and Spetzler (1979) found no consistent trend of ν with strain rate.

Figure M-9 is a plot of ν versus total porosity. There seems to be a slight decrease in ν with increasing porosity for the intact samples, but no such trend is defined by the fractured specimens. Although Lama and

Figure M-7. Relationship Between Mean Poisson's Ratio, Young's Modulus, and Temperature for Pomona Flow

Figure M-8. Relationship Between Mean Poisson's Ratio, Young's Modulus, and Temperature for Umtanum Flow.

Table M-2. Relationships of Young's Modulus and Poisson's Ratio
to Pressure for Hanford Basalts ($p \geq 2.60 \text{ g/cm}^3$)

Pomona Flow

Intact Samples at 100 psi

<u>E (10^6 psi)</u>						<u>T ($^{\circ}\text{C}$)</u>	<u>Sources</u>
<u>Mean</u>	<u>Range</u>	<u>Number Samples</u>	<u>Mean</u>	<u>Range</u>	<u>Number Samples</u>		
12.77	10.8-14.1	7	.298	.255-.323	7	22-24	FSI (1980a,b)
13.06	10.8-16.3	7	.283	.241-.348	7	150	FSI (1980a,b)
12.90	12.1-14.9	7	.270	.216-.308	7	300	FSI (1980a,b)

Jointed Samples at 100 psi

12.88	11.3-16.1	14	.318	.282-.404	14	22-24	FSI (1980a,b)
12.31	11.0-14.2	14	.270	.248-.318	11	150	FSI (1980a,b)
12.78	11.4-14.9	12	.259	.211-.316	10	300	FSI (1980a,b)

Untanum Flow

Intact Samples at 50 psi

13.15	9.4-18.27	4	.289	.272-.309	3	20	M (1979a), MB (1979)
4.14	1.08-6.18	3	.281	.143-.419	2	150	M (1979a,b)
3.41	1.55-6.45	5	.236	.121-.346	5	300	M (1979a,b), MB (1979)
3.40	0.69-5.35	7	.258	.115-.652	6	500	M (1979a,b), MB (1979)

Jointed Samples

9.47	7.1-11.2	3	.257	.144-.348	3	24	FSI (1981)
9.90	7.9-11.7	3	.241	.149-.294	3	150	FSI (1981)
9.55	9.0-10.2	4	.302	.289-.316	4	300	FSI (1981)

M: Miller
MB: Miller and Bishop

FSI: Foundation Sciences, Inc.

Figure M-9. Relationship Between Poisson's Ratio and Total Porosity for Pomona Flow.

Vutukuri (1978) suggest that ν should decrease at higher porosity values, it is possible that such a trend is ill-defined for the low total porosities found in Hanford basalts.

Similarly ambiguous results are reported by Deere and Miller (1966) for other Columbia River basalts and by Ruiz (1966) for basalts from Brazil. However, Neprochnov and others (1974) examined basalts with a greater range of total porosity recovered during the Deep Sea Drilling Project. They substantiated the trend of decreasing ν with increasing porosity. In addition, both E and G were shown to decrease as porosity increased.

Related to the porosity effect(s) is the possible effect of moisture content or saturation state. Brodsky and Spetzler (1979) found that basalt from Ralston, Colorado had a higher value of E when partially saturated than when completely dry; the difference was greater at higher strain rates. The preceding observations are reasonable, since water-filled microcracks are stiffer than air-filled microcracks, especially at higher strain rates at which drainage becomes more difficult. No consistent relationship between ν and moisture content was found. Table M-1 contains the reported literature values for elastic constants.

The presence of large scale discontinuities in-situ will lower the elastic constants of the rock mass as a whole. Nugent and Banks (1965) estimated that in-situ values for basalt would be approximately 5 percent of laboratory values for intact basalt. Other investigators have estimated that in-situ values for other rock types will be 15-20 percent of laboratory values (Wilson, 1977; Kulhawy, 1979). An initial estimate of 10 percent is used in deriving the ranges suggested below. More data are needed before a more definitive relationship may be established.

Suggested Ranges:¹

E: 8.0-14.0 (10^6 psi³) [0.8-1.4 (10^6 psi)]²

ν : 0.15-0.35

G: 2.96-6.09 (10^6 psi) [0.3-0.6 (10^6 psi)]²

K: 3.8-15.6 (10^6 psi) [0.4-1.6 (10^6 psi)]²

λ : 1.5-12.1 (10^6 psi) [0.2-1.2 (10^6 psi)]²

Note: The values of G and K calculated using mean values of E and ν approximated from Figures M-1 and M-2 are almost exactly the values which would be chosen as means from Figures M-3 and M-4.

¹Ranges of G, K, λ derived from those for E, ν .

²Assuming in-situ values are ~ 10% of laboratory values.

³1 psi = .006895 MPa

Table M-3. Elastic Constants of Basalt

Columbia Plateau

Source	Location or Basalt Unit	E*(10 ⁶ psi)	σ *	G*(10 ⁶ psi)	K*(10 ⁶ psi)	λ *(10 ⁶ psi)
Colorado School of Mines (1978)	Pomona flow	0.12-17.64	.115-.554	NA	NA	NA
Coulson (1972)	Little Granite basalt	7.28	NA	NA	NA	NA
Deere and Miller (1966)	Columbia River basalt	7.28-12.16	0.24-0.29	NA	NA	NA
Duvall and others (1978)	Pomona flow	s: 9.15-12.62 d: 10.12-14.47	s: .191-.290 d: -.143-.257	4.12-7.70	2.80-8.12	-1.41-4.99
Foundation Sciences, Inc. (1980a)	Pomona flow	s: 7.1-16.3 d: 10.29-13.32	s: .216-.355 d: .215-.301	4.057-5.481	7.106-8.970	NA
Foundation Sciences, Inc. (1980b)	Pomona flow	s: 11.0-16.1 d: 10.31-12.64	s: .179-.348 d: .208-.319	3.907-5.113	6.977-10.110	NA
Foundation Sciences, Inc. (1980c)	Pomona and Umtanum flows	s: 6.0-15.8 d: 9.7-15.5	s: .149-.421 d: .146-.288	3.8-6.1	5.4-11.5	NA
Foundation Sciences, Inc. (1981)	Umtanum flow	s: 7.1-15.1 d: 4.8-14.8	s: .144-.384 d: .146-.330	1.8-6.1	4.7-10.3	NA
Hocking and St. John	Hanford basalt	6.96	0.25	NA	NA	NA
Johnston and Palmer (1981)	Survey, Hanford basalt	s: 7.98-15.95	0.22-0.30	NA	NA	NA
Miller (1979a)	Umtanum flow	0.69-18.86	.098-1.560	NA	NA	NA
Miller (1979b)	Umtanum flow	1.08-13.89	.024-1.547	NA	NA	NA
Miller and Bishop (1979)	Umtanum flow	0.91-22.66	.115-.581	NA	NA	NA
Podnieks and others (1972)	Columbia River basalt	7.09-10.06	NA	NA	NA	NA
Schmidt and others (1980)	Summary, Hanford basalts	2.25 ^a -12.81 ^a	0.202 ^a -0.323 ^a	NA	NA	NA
Stephenson and Triandafilidis (1974)	Columbia River basalt	s: 12.0 ^a d: 11.2 ^a	s: 0.30 ^a d: 0.30 ^a	NA NA	NA NA	NA NA
Taylor and others (1980)	Survey, Hanford basalt	NA	NA	NA	NA	NA

*Unless otherwise stated, all E and σ values are static, all G, K, λ values are dynamic.

s = static d = dynamic a = average NA = not applicable

Table M-3 (Continued)

Source	Location or Basalt Unit	$E^*(10^6 \text{ psi})$	σ^*	$G^*(10^6 \text{ psi})$	$K^*(10^6 \text{ psi})$	$\mu^*(10^6)$
<u>Other Localities</u>						
Aleksandrov and others (1969)	U.S.S.R.	d: 9.12	d: 0.26	4.55	7.99	NA
Avila (1966)	Brazil (in-situ)	d: up to 14.27	d: up to 0.409	0.54-5.91	1.31-8.15	NA
Balashov and Volarovich (1971)	U.S.S.R.	NA	NA	NA	2.05-6.94	NA
Berdinelli (1962)	Nevada Test Site	d: 8.9-9.6	d: 0.26-0.27	NA	NA	NA
Blair (1956)	Eniwetok	10.0	0.19	NA	NA	NA
Brodsky and Spetzler (1979)	Ralston basalt	9.35-11.89	0.241-0.307	NA	NA	NA
Carlson and Christensen (1977)	Deep Sea Drilling Project (DSDP)	d: 8.99-11.75	d: 0.28-0.32	3.48-4.50	7.11-10.30	4.79-7.25
Christensen (1973a)	Deep Sea Drilling Project (DSDP)	d: 6.03-11.91	d: 0.29-0.34	2.35-4.56	5.73-11.32	4.12-8.82
Christensen (1973b)	DSDP	d: 3.37-14.55	d: 0.28-0.36	1.47-5.59	4.26-11.32	3.38-7.64
Christensen and others (1974)	DSDP	d: 3.23-12.64	d: 0.23-0.31	1.18-4.85	2.50-10.73	1.47-7.35
Griggs and others (1960)	California	1.86-13.70	NA	NA	NA	NA
Hughes and Maurette (1957)	Colorado	d: 9.24-10.27	d: .233-.274	3.73-4.05	5.92-7.53	NA
Hyndman and Drury (1976)	Mid-ocean ridge basalt	NA	d: 0.26-0.33	NA	NA	NA
Iida and others (1960)	Japan	7.98-12.47	NA	NA	NA	NA
Kaarsberg (1968)	unknown basalt	d: 14.21 ^a	d: 0.282 ^a	5.54 ^a	NA	NA
Krech and others (1974)	Dresser basalt	s: 11.9-19.2 d: 15.62-16.34	0.261-0.266	6.15-6.49	NA	NA
Kulhawy (1975)	Nevada Test Site unknown basalt	5.06 5.63	0.32 0.16	NA NA	NA NA	NA NA
Kumar (1968)	Nevada Test Site	d: 22	NA	NA	NA	NA

*Unless otherwise stated, all E and σ values are static, all G, K, μ values are dynamic.

s = static d = dynamic a = average NA = not applicable

Table M-3 (Continued)

Source	Location or Basalt Unit	E*(10 ⁶ psi)	σ^*	G*(10 ⁶ psi)	K*(10 ⁶ psi)	λ *(10 ⁶ psi)
	<u>Other Localities</u>	(Continued)				
Lehnhoff and Scheller (1975)	Dresser basalt	0.001-14.60	0.10-0.24	NA	NA	NA
Livingston (1961)	unknown basalt	4.13	NA	NA	NA	NA
Lutton (1968)	Nevada Test Site	3.11-5.23	NA	NA	NA	NA
Morrell and Larson (1974)	Dresser basalt	s: 14.25 d: 15.54	0.273	6.14	NA	NA
Nakagawa and Mori (1963)	Japan	d: 10.24-12.36	d: 0.254-0.263	3.96-4.93	7.05-8.37	NA
Neprochnov and others (1974)	DSDP	d: 5.58-12.28	d: 0.24-0.28	2.26-4.89	NA	NA
Nordyke and Wray (1964)	Nevada Test Site	s: 3.24-7.11 d: 5.80-8.14	s: 0.08-0.25 d: .146-.225	s: .0018-.012 d: 2.47-3.33	NA	NA
Nugent and Banks (1965)	Nevada Test Site	1.35-17.91	0.14-0.28	NA	NA	NA
Overton and Moczygemba (1967)	Texas	s: 16.9-18.7 d: 5.75-17.4	d: 0.175-0.285	2.46-6.83	NA	NA
Ramana and Venkatanarayana (1973)	Kolar basalt	9.05-15.18	0.17-0.28	NA	4.58-12.76	NA
Ricketts and Goldsmith (1970)	unknown basalt	s: 4.65-6.5 d: 9.6	NA	NA	NA	NA
Ruiz (1966)	Brazil	s: 6.20-10.59 d: 5.58-9.95	0.16-0.21	NA	NA	NA
Sharp (1972)	Amchitka Island	2.08-13.33	0.17-0.21	3.12-4.01	3.97-7.53	NA
Somerton and others (1963)	Mohole test site basalt	d: 7.03	d: 0.384	NA	NA	NA
Stowe (1969)	Nevada Test Site	2.71-6.66	0.15-0.47	NA	NA	NA
Stowe and Ainsworth (1972)	Nevada Test Site	4.28-6.66	0.20-0.37	NA	NA	NA
Wawersik and Fairhurst (1970)	unknown basalt	14.85	NA	NA	NA	NA
Waza and others (1980)	Japan	4.15	0.29	1.61	NA	NA
Windes (1949)	Michigan	8.92-12.4	NA	s: 3.89-4.91	NA	NA

*Unless otherwise stated, all E and σ values are static, all G, K, λ values are dynamic.

s = static d = dynamic a = average NA = not applicable

Table M-3 (Continued)

Source	Location or Basalt Unit Other Localities	$E^*(10^6 \text{ psi})$ (Continued)	σ^*	$G^*(10^6 \text{ psi})$	$K^*(10^6 \text{ psi})$	$\lambda^*(10^6 \text{ psi})$
Windes (1950)	Michigan	s: 0.25-8.4 d: 0.91-5.9	s: 0.15-0.34 d: -0.04-0.09	NA	NA	NA
Wingquist (1976)	Dresser basalt	d: 6.46-15.0	d: 0.04-0.25	2.99-6.04	NA	NA
<u>Citations or Surveys of Previous Literature</u>						
Agapito and others (1977)	Literature range	10.15 ^a	0.26 ^a	NA	NA	NA
AVCO Everett Research Lab (1972)	Citation of unknown source, Dresser basalt	NA	0.24	NA	NA	NA
Clark (1966)	Literature survey	s: 8.97-16.39 d: 7.13-10.29	s: 0.22-0.276 d: 0.384	s: 3.97-5.00 d: 0.30	6.07-11.76	NA
Clark and Caudle (1961)	Literature survey	s: 0.38-8.7 d: 0.91-8.92	s: 0.01-0.22 d: -0.04-0.25	0.48-2.68	NA	NA
Dames and Moore (1978)	Cites Sharp (1972), Stephenson and Triandaifilidis (1974), Stowe (1969)	NA	NA	NA	NA	NA
Dubois and others (1979)	Cites Agapito and others (1978)	NA	NA	NA	NA	NA
Nafe and Drake (1968)	Literature range	9.41-17.6	0.12-0.35	4.12-6.47	4.59-16.33	NA
Ratigan (1976)	Literature average	4.5 ^a	0.29 ^a	NA	NA	NA

*Unless otherwise stated, all E and σ values are static, all G , K , λ values are dynamic.

s = static d = dynamic a = average NA = not applicable

APPENDIX N

Geochemistry

N.1 Discussion of K_d

N.1.1 Definitions

With the exception of the noble gases and some very stable fluorocarbons, aqueous species are not inert in the ground-water environment. They tend to be adsorbed, ion exchanged, filtered, precipitated, and redissolved or simply to participate in various chemical reactions. As a result of these processes, their transport time through geologic media is not the same as the water which bears them. Although most of the processes which contribute to the retardation of solutes with respect to the water solvent are well understood in principle and under laboratory conditions, they cannot yet be adequately modeled to predict their cumulative effects in the natural environment. Measurement of these effects is therefore necessary.

The results of such measurements are conventionally expressed as K_d values. The value indicates how much of a material being borne by the water can be withheld by the geomedium under a specific set of conditions. It is defined as the ratio between the concentration in the solid and liquid phases. It has the form of

$$K_d = \frac{C_{\text{solid}}}{C_{\text{liquid}}} \text{ (in ml/g)} \quad (1)$$

and is used as equilibrium constants in chemical equilibria or as a fractionation factor between the immobilized and mobile states. The mechanisms implied by such usage are not necessarily those causing the retardation of the water borne material.

The K_d parameter frequently is used to indicate the retention capacity of the geomedium. In porous media, where the aqueous species completely penetrate and surround the solid phase, the retardation factor R may be defined

$$R = \frac{t_{\text{waste}}}{t_{\text{water}}} = 1 + K_d \rho \frac{(1 - \phi)}{\phi} \quad (2)$$

where ρ = the bulk density of the geomedium (g/cm³)

ϕ = porosity of the geomedium

t_{waste} = waste material travel time

t_{water} = water travel time.

R represents how many times faster the water moves compared to the aqueous species, or waste. This is equivalent to a ratio of travel times or flow velocities. In sparsely fractured media, where the entire rock matrix is not penetrated by the species being transported and only a small surface layer is involved in the retardation mechanism, the retardation factor also will be a function of that penetration depth. This may be expressed by introducing the utilization factor Ψ , which represents the fraction of the rock influenced by the waste. Here

$$R = \frac{t_{\text{waste}}}{t_{\text{water}}} = 1 + \Psi K_d \rho \frac{(1 - \phi_f)}{\phi_f} \quad (3)$$

where ϕ_f = fissure porosity involved in waste transport (excluding dead fissure porosity)

Ψ = utilization factor, which is a function of contact time and fissure geometry

Both K_d and Ψ are strongly dependent on the type of geomedium and the aqueous species being transported (Fontes and others, in preparation).

N.1.2 Factors Influencing K_d

Currently, empirical K_d measurements are made by either static or dynamic tests on unconsolidated, crushed, intact or fractured samples of various geomedia. The tests are unable to distinguish between sorption/desorption, precipitation/dissolution, filtration or oxidation/reduction processes. These

processes, in turn, are dependent on temperature, surface area, ionic exchange capacity, and other variables. The distribution coefficient K_d is likely to vary with changes in any of these terms (Relyea, Serne and Rai, 1980).

N.1.2.1 Surface Area

Surface area primarily affects sorption/desorption on secondary minerals such as those found in basalt fractures. Under conditions where other factors remain equal, K_d tends to increase with surface area. Surface area itself is not constant, particularly in fractured media where secondary minerals are being formed. Surface area, or specific surface, is measured by ethylene glycol wetting methods.

N.1.2.2 Ion Exchange Capacity and Redox Conditions

The cation or anion exchange capacity (CEC or AEC) of a geomedium affects the sorption/desorption of charged aqueous waste species. The effect of ion exchange capacity is species-specific. For example, under the high pH, reducing environments expected in a basalt repository, the anion $U(OH)_5^-$ will be the dominant uranium species in solution. A geomedium of high AEC will tend to have a high K_d since $U(OH)_5^-$ will be easily sorbed. The same solution, with the same pH, but under more oxidizing conditions expected in upper interbeds would tend to contain uranium in the form of the cation $(UO_2)_3(OH)_5^+$ (if no cationic complexes can form). The K_d of uranium would be low because $(UO_2)_3(OH)_5^+$ retention is not influenced by AEC.

N.1.2.3 Solution pH

The hydronium ion activity of the solution, which is measured as pH, affects mineral stability, the solubility of the radionuclide and the speciation of the aqueous forms. In the previous example, had pH been increased due to an external reaction mechanism, $U(OH)_5^-$ would have successively transformed to $U(OH)_4^0$, $U(OH)_3^+$, $U(OH)_2^{2+}$, UOH^{3+} and finally U^{4+} . The sorption of each of these species is successively less dependent on AEC and more dependent on CEC. Further, the stability of uraninite ($UO_2(cr)$) is pH-dependent, as is the maximum amount of uranium which may be put in solution. The influence of pH on K_d may only be predicted by a detailed knowledge of the aqueous speciation and mineral stability of each radioelement in the waste.

N.1.2.4 Temperature

Temperature has some effect on waste ion speciation, but primarily increases reaction rates. Minerals tend to precipitate less easily, dissolve faster and solubilities tend to increase with higher temperatures.

The K_d of Cs tends to decrease with temperature, those of Sr, Se, Ra and U species tend to increase, and those of Tc, I, Np, Am and Pu do not appear to be greatly affected. Temperature will also affect CEC and AEC to some extent.

N.1.2.5 Chemical Composition

The chemical composition of both ground water and geomedium influence K_d through sorption, mineral stability, and complex formation. The formation of complexes tends to greatly increase the solubility of waste nuclides. These complexes, in general, do not tend to form in the reducing environments of the potential repository, but C, F, Si, P and N complexes may become important under the more oxidizing conditions that the waste may encounter as it approaches the biosphere.

N.1.2.6 Radionuclide Concentration

Generally, K_d decreases with increasing concentration of a waste radionuclide or of other waste nuclides of similar charge and structure. This results from sorption site competition at higher concentrations and clearly illustrates the dissimilarity of K_d and equilibrium constant behavior. Further, if no account is taken of solubility constraints, experiments involving high concentrations of relatively insoluble nuclides will yield erroneously high K_d values. It is frequently assumed that any decrease in radionuclide aqueous concentration is due to adsorption on solid matrix surfaces. Actually, under real experimental conditions with poorly soluble nuclides, high initial radionuclide aqueous concentrations, neutral or high pH values and the presence of competitive ions, may invalidate the K_d values which are obtained.

N.1.2.7 Particulate Transport

Radionuclides may migrate through a ground-water system adsorbed on small particles if the particle size

is less than the pore diameter of the geomedium. If the particle is larger than the pore diameter, the nuclide will be filtered from solution. Direct filtration of particles with large hydration spheres in ion selective media such as clays also may occur. Particle migration is a potentially serious problem in fractured media such as basalts, in which interactions with fracture surfaces is less than interactions in porous media. Nuclides transported in this manner are neither in solution nor immobilized by adsorption to a fixed solid surface, and therefore aberrant K_d measurements are obtained.

N.1.2.8 A Treatment of These Factors

In the absence of a coherent theoretical treatment of the many factors influencing and obscuring K_d , a statistical method has been employed by Mucciardi (1977, 1978) in which K_d 's are predicted from empirically derived functions of these factors. For example, this method can predict the K_d of Tc with a correlation coefficient of 0.96, using the Sorption Information Retrieval System at Battelle Pacific Northwest Laboratory (Hostetler and others, 1980). The early success in this area would suggest that if work were continued, results of great value to consequence modelers would be obtained.

Clearly, K_d values are not the best parameter for characterizing the retardation of radionuclides in ground-water transport. A more fundamental treatment seems required, in which many processes and variables are not lumped. It is this lumping characteristic of K_d which leads to the large ranges and uncertainties apparent in the following section, and which gives the impression that the geochemistry of the retardation process is understood poorly now and cannot be much better understood in the future.

N.1.3 Experimental K_d Values in Basalt Environments

The ranges of K_d values in the basalt at the Hanford Reservation can be attributed to three distinct material/condition pairs:

- 1) Fresh basalt surfaces under reducing conditions, expected in the near field after resaturation of the subsurface facility is complete.

- 2) Fresh basalt surfaces under oxidizing conditions, expected in some areas of the intermediate and far field.
- 3) Altered basalts, interbed material and secondary minerals produced by alteration of basalt along fractures, expected along the principal pathways to the environment.

The K_d ranges observed in these three cases, and the approximate median value are summarized in Table 2-2. Although desirable, there are insufficient data currently available to estimate ranges for altered basalts and alteration products under reducing conditions. The ranges in this table were compiled from approximately 1000 selected values in the SIRS at Battelle Pacific Northwest Laboratories and represents data appearing in the following primary sources: Ames (1978a,b), Ames and McGarrah (1980), Barney and Anderson (1978), Barney and Brown (1979, 1980), Barney and Grutzeck (1977), Deju (1979a,b,c,d; 1980a,b,c,d). This represents all of the reliable data appearing in open literature up to January 1, 1981. Data from poorly characterized experiments and data generated under ambiguous conditions have not been included.

The chemical composition of the synthetic ground waters used in the K_d experiments appears in Table N-1. Water GR-1 reflects the average composition of waters from boreholes which intersect the upper Wanapum/lower Saddle Mountain basalts. Water GR-2 is a similar composite from the proposed reference repository level in the Grande Ronde basalt (Ames and McGarrah, 1980). Notice that neither of these waters is identical to the Reference Synthetic Ground-Water Composition (RSGC) in the July 11, 1980 release of the BWIP data package item RSD-BWI-DP-007 (Alford, 1980a), although GR-2 is similar.

Various basalt types and samples from the Hanford Reservation were used by the different groups reporting the experiments.* Most of the work was done on well characterized reference basalts (cf. Ames, 1978a). In many cases the same sample was used by both groups; in others the same core or road-cut were sampled. Chemical,

*The two principal researchers whose data are represented in Table N-1 are Dr. Lloyd Ames (PNL) and Dr. Scott Barney (RHO).

Table N-1. Synthetic Ground-Water Formulations

Species	Composition (mg/l)		
	GR-1	GR-2 (Ames and McGarrah, 1980)	RSGC (Alford 1980a)
Na ⁺	30.7	225	263
K ⁺	9.0	2.5	1.9
Ca ²⁺	6.5	1.06	1.3
Mg ²⁺	1.0	0.07	0.04
Cl ⁻	14.4	331	148
CO ₃ ²⁻	0	59	27
HCO ₃ ⁻	81.5	75	70
F ⁻	0	29	37
SO ₄ ²⁻	11.1	72	109
SiO ₂ (aq)	25	108	121
pH	8.0	10.0	9.92

morphological and mineral differences have been identified between tops and bottoms of flows as well as among flows. Nevertheless, laboratory findings show that results of various basalt samples can be combined, often introducing no more uncertainty than would a replicate measurement on a single sample. No significant trends in K_d are apparent for any radionuclide for samples from various parts of the Umtanum, Flow E or Pomona basalts. The chemical and mineralogic characteristics of representative reference samples of Umtanum basalt (from an outcrop just below Priest Rapids Dam on Umtanum Ridge; Ames, 1978b), Flow E basalt (from a road cut on Washington State Highway 243 between Beverly and Mattawa, Washington; Ames and McGarrah, 1980) and Pomona basalt (from the excavation in the RHO Near-Surface Test Facility; Ames and McGarrah, 1980) are presented in Tables N-2 and N-3. There is every indication that the chemical composition and mineralogy of these fresh basalt samples is representative of the same properties at depth.

Three basic secondary minerals occurring in fractured basalt were examined. An adequate secondary mineral sample, composed primarily of smectite clay, was obtained from a fracture zone in the RHO Near-Surface Test Facility excavation (Ames and McGarrah, 1980; Barney, 1980). Heulandite zeolite, a common secondary mineral in fractured basalts, and nontronite clay, the predominant clay alteration product in basalt fractures, also were used in K_d measurements (Ames, 1978b). The chemical composition of these materials also appears in Table N-2. The interbed material in Table 2-2 is from tuffs within the Ellensburg Formation. The material, collected and analyzed by RHO, has not yet been described in the open literature. The K_d values of the various materials have been lumped together into one set of ranges in Table 2-2 because the variability among the data is less than the expected error intervals for the data of each material.

The experiments considered in Table 2-2 (see Section 2.3.1) were from 5 to 120 days duration and many of the experiments are still in progress. The experiments were conducted up to 300°C but most results were measured at 23 and 60°C. The composite temperatures in the table indicate measurements over this range. The temperature/ K_d relationship discussed earlier in the report was observed. The average representative ethylene glycol surface areas, sample mesh sizes, and CEC values are presented in Table N-4.

Table N-2. Chemical Characterization of Geomedia in the Hanford Basalts
(after Ames, 1978b)

Average Whole Basalt or Selected Secondary Mineral Chemical
Analysis on Multiple Replicates (in Wt %)

Constituent	Umtanum Basalt	Flow E Basalt	Pomona Basalt	Secondary Mineralization (smectite)	Heulandite	Nontronite
SiO ₂	55.23	53.99	52.34	44.3	63.7	39.92
Al ₂ O ₃	13.70	13.89	15.18	6.6	15.2	5.37
TiO ₂	2.07	1.94	1.58	0.2	---	29.28
FeO	12.86	12.28	10.94	9.5	---	---
MnO	0.20	0.20	0.19	---	---	---
CaO	7.23	8.42	10.03	1.1	8.8	2.46
MgO	3.30	4.35	7.01	12.6	---	0.93
K ₂ O	1.64	1.20	0.46	0.7	0.2	---
Na ₂ O	3.36	3.41	2.14	---	---	---
P ₂ O ₅	0.40	0.31	0.23	---	---	---
H ₂ O	0	0	0	24.9	11.8	21.38
	99.99	99.99	100.10	99.3	99.7	99.80

Table N-3. Average Mineralogy of Basalts in the Hanford
Reservation (in weight percent)
(after Ames and McGarrah, 1980)

Mineral Group	Umtanum Basalt	Flow E Basalt	Pomona Basalt
Plagioclase	26 (mostly labradorite)	44 (mostly andesine)	34 (mostly labradorite to andesine)
Clinopyroxene	21 (mostly augite)	32	32 (mostly augite)
Glassy Groundmass	37	13	29 (mostly tachylyte)
Olivine	---	---	2
Metal Oxides	7	6	4
Clays	9	5	----

Table N-4. Physical and Exchange Characterization
of Geomedia in the Hanford Reservation
(See discussion of Table 2-2 for sources)

Formation	Umtanum Basalt	Flow E Basalt	Pomona Basalt	Secondary Mineralization (smectite)	Heulandite	Nontronite
Mesh size	20-40	20-50	10-20	20-50	---	40-200 (< 2 m)
Surface area (m ² /g)	17.7 _{+3.75}	10.3 _{+1.0}	31.2 _{+1.4}	27.4 _{+0.7}	650 ₊₂₀	55.0 _{+3.4} 861 ₊₃₆
CEC (meg/l)	1.65 _{+0.1}	4.7 _{+0.2}	-----	-----	90-150	2.4 _{+0.1} 95 ₊₃

Experiments under oxidizing conditions were equilibrated with the atmosphere while those under reducing conditions were anoxic (< 2 ppm O_2).^{*} Although analyses were not made on all the isotopes considered by the Sandia Risk Assessment Methodology, certain chemical homologies exist based on similarities in solubility, valence and redox stability. Provisionally, the following equivalences can be made:

Am = Cm
Pa
Th
Ac
Po
Bi
Sm
Nb
Zr

Ra = Pb.

N.2 Hydrogeochemical Condition of the Repository Environment

N.2.1 Dissolved Species

N.2.1.1. Major Ions

The major ion geochemistry of the ground waters in the Páscó Basin basalts has been summarized in Section 2.4.4.2 of Smith and others, (1980). A Piper (trilinear) diagram showing the similarities of the major ion concentrations in the waters associated with the Hanford Site is presented in Figure N-1. These values were obtained from all available published sources. No further work will be presented here.

N.2.1.2 Trace Elements

Trace element distribution is also summarized by Smith and others, (1980). No description of the speciation of trace elements has appeared in the open literature. A summary of trace element composition in various formations at the Hanford Reservation appears in Table N-5.

^{*} Cf. Section on the redox conditions (N.2.2) at the proposed repository level.

Figure N-1. Piper (trilinear) Diagram of Major Ion Composition of Various Ground Waters Associated with the Hanford Site.

Table N-5. Trace Element Concentrations in Ground Water at the Hanford Site

Confined aquifers of Grande Ronde Formation							
	Gephart and others 1979		Priest Rapids Member of Upper Wanapum basalt, Gephart and others 1979		Mabton Interbeds Gephart and others 1979		Unconfined Ground water at Hanford Site Gephart and others 1979
	Apps and others 1979						
Ag	<0.010	--	--	--	--	--	<0.010 - 0.002
Al	--	--	0.11	<0.05	0.086	<0.020 - 2.170	<0.050 - 0.470
As	<0.002	0.001	--	--	--	--	0.001 - 0.014
B	--	--	1.39	0.10	0.013	<0.005 - 0.550	<0.009 - 0.150
Ba	0.150	<0.112	<0.005	0.027	0.053	<0.005 - 0.065	0.007 - 0.100
Br	0.201	0.285	--	--	--	--	--
Cd	0.007	0.089	<0.005	<0.005	<0.005	<0.005 - 0.009	<0.003 - 0.140
Co	<0.017	0.047	<0.020	<0.005	<0.005	<0.005	<0.002 - 0.010
Cr	<0.0002	<0.0004	<0.005	<0.005	<0.005	<0.005	<0.050 - 0.100
Cu	0.050	0.060	<0.005	0.005	<0.005	<0.005	<0.010 - 0.047
Fe	0.017	0.015	0.054	0.228	0.181	<0.005 - 4.700	<0.005 - 3.9
Mn	0.004	<0.009	<0.010	<0.100	<0.10	<0.010	<0.001 - 0.480
Mo	0.270	0.31	0.310	<0.020	<0.020	<0.010	<0.001 - 0.030
Ni	0.070	--	<0.005	<0.005	<0.005	<0.005 - 0.030	--
Sr	0.012	0.003	<0.005	--	--	0.009 - 0.111	--
Zn	0.260	0.240	0.096	<0.005	<0.015	<0.005 - 0.093	<0.005 - 1.6

N.2.2 Redox Conditions

The importance of characterizing the redox conditions at the proposed repository and along potential paths along which wastes could be released has been shown in Sections 1.0 and 2.0. Several attempts have been made to characterize the redox environment at the repository level and surrounding environment.

The waters of the Grande Ronde formation are basic. Measurements at 45°C show a pH near 10 (Gephant and others, 1979). Simulation experiments at higher temperatures show pH = 9.45 at 65°C (Barnes and Scheetz, 1979). The empirical relationship

$$\text{pH} = 1.64 + 2,640/T(^{\circ}\text{K})$$

was determined from these results. Nevertheless the (very similar) relationship

$$\text{pH} = 1.21 + 2,810/T(^{\circ}\text{K})$$

is cited in the BWIP Data Package (Alford, 1980b). Over the temperature range expected, pH is predicted in the range 9.4 to 9.9 (Smith and others, 1980). The pH is thought to be controlled by the silicic acid dissociation process which buffers the solution to about pH = 9.5. This mechanism is plausible in a basaltic environment where ground water is saturated with silica, as it is at Hanford. However, silica saturation alone is insufficient to postulate this buffer mechanism, since silica saturation is not uncommon in many environments. The presence of silica producing mineral weathering reactions in basalt alteration is also required.

In the basaltic environment the fugacity of oxygen, a primary parameter determining Eh, can be taken to vary between Nickel-Nickel Oxide (NNO) and Quartz-Fayalite-Magnetite (QFM) control. For further discussion of these mechanisms, see Huebner and Sato (1970) and Eugster and Wones (1962), respectively. Here

$$\log f_{\text{O}_2} = 9.36 - 24,930/T(^{\circ}\text{K}) \quad (\text{NNO}) \text{ and}$$

$$\log f_{\text{O}_2} = 9.0 - 25,080/T(^{\circ}\text{K}) \quad (\text{QFM}).$$

The relationship used to determine Eh from pH, T(°K) and log fO₂ is (Smith and others, 1980):

$$\begin{aligned} \text{Eh (volts)} = & 1.23 + 4.96 \times 10^{-5} T(^{\circ}\text{K}) \log f\text{O}_2 \\ & - 1.984 \times 10^{-4} T(^{\circ}\text{K}) \text{ pH} \\ & - 9.196 \times 10^{-4} (T(^{\circ}\text{K}) - 298). \end{aligned}$$

Figure N-2 is an Eh-pH diagram. Eh is presented in volts and pH = - log [H⁺] where [H⁺] is hydronium ion activity. The diagonal lines A and B bound the stability range of water. Water tends to decompose to O₂ above line A and to H₂ below line B (cf. e.g. Krauskopf, 1967). The large irregularly shaped field within these lines is the Eh-pH stability field of natural waters, as determined by Baas Becking, Kaplan and Moore (1960). The two closely spaced parallel lines C and D represent the Eh-pH relationship expressed by the equation above at 25°C and log fO₂ at NNO and QFM control, respectively. Note that since the data from which the NNO equilibrium is calculated were taken at higher temperatures, the location of line C is questionable and the assumption that the NNO system controls Eh needs further verification. The area EFGH represents the expected Eh-pH range after Smith and others (1980) which is -0.45 V to -0.55 V(Eh) and 9.4 to 9.9 pH. The smaller area IJKL represents the Eh-pH range reported in the current BWIP data package (Alford, 1980b) which is -0.50 V to -0.54 V(Eh) and 9.4 to 9.9 pH. Note that both of these areas have portions significantly below the Baas Becking field, indicating that these Eh-pH combinations have not been observed in the field and are unlikely. Nevertheless, they indicate that extremely reducing conditions are expected at the repository level. The area MNGH reflects the entire Eh variation expected as the water goes from anoxic to atmosphere equilibrated conditions (Smith and others, 1980). Here again a theoretical, rather than practical, extreme value was used. No Eh-pH field measurements for Hanford appear in the open literature. Point O represents an unpublished value (Eh = -0.46V, pH = 10) obtained by RHO at the proposed repository level. This falls within the Baas Becking field and very near the area suggested by Smith and others (1980).

Although field Eh measurements are difficult to make, the Eh of a ground-water environment may be estimated with confidence if the mineral assemblages present in that water are known. Although no laboratory experiments have yet been conducted using site-specific materials to determine the stability relations of secondary mineral assemblages, the minerals pyrite, magnetite and siderite have been identified in the basalts of the Hanford Site. These minerals can coexist under certain physiochemical conditions in the Eh range -0.36 to -0.41 V at pH around 10 (lines P and Q, Figure N-2). This association may support the Eh-pH ranges proposed by Smith and others (1980) in the area EFGH. However, the use of mineral assemblages to characterize the Eh-pH of a ground-water environment requires knowledge of the values of additional physiochemical parameters. Further solubility speciation modeling and mineralogy can lead to a better characterization of the Eh-pH environment at the proposed repository and along the flow path to the accessible environment.

Section 2.3 of the text contains the major results and conclusions of this Appendix.

Figure N-2. Hanford Site Repository Conditions and Constraints in General Eh-pH Space

APPENDIX O

Hydrologic Modeling of the Reference Site

Modeling of the hydrologic system in the Pasco Basin by Rockwell-Hanford Operations, Los Alamos Technical Associates and INTERA, Rockwell-Hanford Operations and Resource Management Associates, and Pacific Northwest Laboratory either are completed or are in progress. Each of these studies considers a release of undispersed tracer from the subsurface facility, and each calculates travel time along a pathway to either the Columbia River or the boundary of the model. Whereas none of these efforts are available in the open literature, the description of the details of these studies is not appropriate in this report. Because of the limited understanding of the ground-water flow in the Pasco Basin, these modeling efforts are more a development of the modeling methodology than an actual characterization of the flow system in the basin. Additional hydrologic data may be useful in refining these models.

As recommendations on how to proceed in the area of modeling, the authors suggest:

1. Following the Pacific Northwest Laboratory approach, the entire, or a sizable portion of, the Columbia Plateau should be modeled in three dimensions to take advantage of the hydrologic data outside of the Pasco Basin. The purpose of this effort is to establish as realistic a set of boundary conditions as possible for the Pasco Basin.
2. Once these boundary conditions are set, a relatively detailed three dimensional model should be constructed for the Pasco Basin. This effort will allow for the characterization of the ground-water flow scheme as realistically as the boundary conditions and the intrabasin hydrologic data will allow.
3. Depending on the flow scheme determined for the basin by three dimensional modeling, one or more curvilinear cross-sections may exist that can be modeled for use in risk or consequence analysis.

This recommended modeling effort will be useful in determining where and what type of additional data are needed to better understand ground-water flow in the Pasco Basin. In addition, certain modeling techniques presently in use may be found to be inadequate or to need varying degrees of modification. The main goal is to provide NRC with as realistic a flow scheme of the basin as possible for the evaluation of the results of potential radionuclide releases from the reference site.

APPENDIX P

Glossary

- ALLUVIAL. Deposited by a stream or running water.
- APHYRIC. The texture of a fine-grained igneous rock that lacks phenocrysts.
- ARKOSIC. A granular texture of sedimentary rocks formed by mechanical aggregation and composed of large grains of feldspar and quartz mixed together unequally, with mica and clay as minor constituents.
- AXIAL SURFACE. A surface that connects the fold hinge lines of the strata in a fold.
- AXIAL TRACE. The intersection of the axial surface of a fold with the surface of the earth or other given surface.
- BOX FOLD. A fold with the approximate profile of three sides of a rectangle.
- BRAZILIAN TENSILE STRENGTH TEST. A test in which load is applied to two diametrically opposed sections of a thin disk resulting in tensile failure at the center of the disk.
- CLINKER. A rough, jagged pyroclastic or autobrecciated fragment that resembles clinker or slag from a furnace.
- COLLUVIAL. A general term applied to any loose, heterogeneous and incoherent mass of soil material and rock fragments deposited by rainwash, sheetwash, or slow continuous downslope creep.
- COULEE. A long, steep-walled, trench-like gorge or valley representing an abandoned over-flow channel that temporarily carried meltwater from an ice sheet.
- CRYSTALLITE. A minute mineral form common in glassy volcanic rocks, usually not referable to any mineral species but marking the first step in crystallization.

DIATOMITE. A light, friable, siliceous material derived chiefly from the tests of diatoms.

DOLERITIC. Sufficiently fine-grained to be difficult to identify the constituent minerals megascopically. Usually used to describe dark, fined-grained igneous rocks that compositionally may be dolerite (diabase).

ENTHALPY. A thermodynamic quantity that is defined as the sum of a body's internal energy and the product of its volume and the pressure.

FANGLOMERATE. A sedimentary rock consisting of slightly water-worn heterogenous fragments of all sizes, deposited in an alluvial fan and later cemented into rock.

FOCAL PLANE. A plane defined by the foci of a number of earthquakes. (See hypocenter.)

FUGACITY. A thermodynamic function, f , defined by the equation $dG = RT d \ln f$, where G is the Gibbs free energy, R is the gas constant, and T is absolute temperature. Fugacity is expressed in units of pressure.

GEODETIC. Describing the determination of the size and shape of the earth and the precise location of points on the surface.

GIBBS FREE ENERGY. A thermodynamic function, G , defined by the equation $G = H - TS$, where H is enthalpy, T is absolute temperature, and S is entropy.

HOMOCLINE. A general term applied to strata that dip in one direction at a relatively uniform angle.

HOMOLOGY. Similarity in position, proportion, or structure.

HYDRATION SPHERE. The collection of water molecules attached to an ion in solution by electrical forces.

HYDRONIUM ION. The hydrated hydrogen in H_3O^+ .

HYPOCENTER. The initial rupture point of an earthquake where strain energy is first converted to elastic wave energy.

IMBRICATION. The overlapping of features dipping in the same direction, such as faults or pebbles.

INTERSERTAL. The texture of a porphyritic igneous rock in which the groundmass, composed of a glassy or partly crystalline material other than augite, occupies the interstices between unoriented feldspar baths and forms a relatively small proportion of the rock.

LACUSTRINE. Pertaining to, produced by, or formed in a lake or lakes.

MAGNETOTELLURIC. An electromagnetic method of surveying in which material electric and magnetic fields are measured.

MERCALLI SCALE. A scale of earthquake intensity based on degree of destruction of man-made structures.

MICROPHENOCRYSTS. A crystal that is easily visible only on a microscopic scale and also is large relative to the surrounding groundmass.

MONOCLINE. A local steepening of strata having an otherwise uniform, gentle dip.

NONCYLINDRICAL FOLD. Any fold having a surface that cannot be approximated by moving a line through space parallel to itself.

REPLICATE. One of several identical experiments, procedures, or samples.

SILICIC ACID. The aqueous species H_4SiO_4 .

SOLID SOLUTION. A single crystalline phase that may be varied in composition within finite limits without the appearance of an additional phase.

SPECIATION. When applied to aqueous geochemistry, the term refers to the distribution of an element into its various valences and complexes.

STOICHIOMETRIC. With reference to a compound or a phase, the term pertains to the exact proportions of the constituents specified by the chemical formula.

TACHYLYTE. A volcanic glass that may be black, green, or brown because of the abundance of crystallites. This glass is formed from basaltic magma and commonly is found as chilled margins of dikes, sills, or flows.

VITRIC. A descriptive term for pyroclastic material that is characteristically glassy (more than 75% glass).

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