ERDA-76-43

VOL. 5 OF 5

MASTER

Alternatives For Managing Wastes From Reactors And Post-Fission Operations In The LWR Fuel Cycle

**Volume 5:
Appendices**

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Alternatives For Managing Wastes From Reactors And Post-Fission Operations In The LWR Fuel Cycle

MAY 1976

United States
Energy Research & Development
Administration

Volume 5
(5 of 5 Volumes)

Appendices

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Report Coordinated By
Battelle, Pacific Northwest Laboratories
At The Request Of The
Division Of Nuclear Fuel Cycle And Production

FOREWORD

This report describes technical alternatives for managing wastes from the back end of the commercial LWR fuel cycle. It describes waste types, the alternative technologies that can be used to manage the wastes, and the state of availability of those technologies.

The report does not select preferred waste management technologies or make comparative assessments. Its purpose is to provide a comprehensive digest of technical information expected to serve as a basis for future decisions, assessments, and environmental impact statements. The technology is still evolving, and this report will be subjected to public review and comment. Its contents therefore serve as a baseline subject to future supplementation and updating.

There are four major waste management functions: treatment, interim storage, transportation, and final storage or disposal. This report is structured in five volumes that reflect these functions. Volume 1 summarizes the report, describes alternatives for the back end of the LWR cycle, and provides background information on waste types and quantities. Volume 2 describes alternatives for waste treatment. Volume 3 discusses interim storage and transportation. Alternatives for final storage and disposal are described in Volume 4. The Appendices, Volume 5, provide supplementary information, with emphasis on characteristics of geologic formations that might be used for final storage or disposal.

Preparation of this report was started in August of 1975, in response to a directive from ERDA's Nuclear Fuel Cycle and Production Division. The report was authored by personnel of ERDA's contractors, of industry, and of universities; coordination was provided by Battelle's Pacific Northwest Laboratories. The report was completed in March of 1976. The cutoff date for characterizing the status of the technologies was September 1, 1975.

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APPENDIX A

SELECTED GLOSSARY

APPENDIX A

SELECTED GLOSSARY

Back end of the fuel cycle: Includes reactors, spent fuel storage, fuel re-processing, mixed-oxide fuel fabrication, and waste management.

Burial grounds: Areas designated for storage of containers of treated radioactive wastes by near-surface burial in geologic media.

Concepts: The basic methods by which waste management operations can be achieved. An example of a waste treatment concept is solidification of liquid high-level waste.

Decommissioning: Preparation of worn out or obsolete nuclear facilities for retirement. Decommissioning operations remove facilities such as re-processing plants and burial grounds from service and reduce or stabilize radioactive contamination. Concepts include

- demolition and restoration to original conditions requiring no control
- partial demolition and fixation of residues
- minimal demolition followed by isolation and control of residues.

Deep continental geologic formations: Geologic media beneath the continents and isolated from biologic species and phenomena. Distinguished from ice sheets and sea floor geologic media.

Disposal: Operations designed to eliminate wastes from existence on earth or to permanently isolate them from mankind and his environs with no expectation of retrieval after emplacement. Isolation concepts include

- placement in subsurface geologic formations using technologies that offer no practical method for recovery
- emplacement into or beneath sea floors
- emplacement in ice sheets
- dispersion.

Elimination concepts include extraterrestrial disposal and transmutation.

Final storage: Storage operations for which a) no subsequent waste treatment or transportation operations are anticipated and b) conversion to disposal (i.e., termination of monitoring and human control) is considered possible.

The current concept for final storage is emplacement of wastes in geologic formations. The geologic formations may be on or near the earth's surface and augmented by engineering technology (e.g., burial grounds), or subsurface and augmented with engineering technology.

Head-end of the fuel cycle: Mining, milling, enrichment, and fabrication of UO_2 fuel.

High-level waste: The fission product waste resulting from reprocessing spent fuel to separate uranium and plutonium from the fission products.

A.2

Immobilization: Treatment and/or emplacement of the wastes so as to impede their movement.

Interim storage: Storage operations for which a) monitoring and human control are provided and b) subsequent action involving treatment, transportation, or final disposition is expected.

Concepts for interim storage include bulk and unitized storage of solid, liquid, and gaseous wastes.

Alternative interim storage technologies include:

- tank storage of liquids
- canister storage in air-cooled vaults
- spent fuel storage in water basins.

Intermediate-level waste: Waste requiring some kind of action to protect personnel from radiation.

Isolation: A term encompassing both final storage and/or disposal in geologic formations.

Long-lived nuclides: Radioactive isotopes with half-lives greater than about 30 years. Most nuclides of interest to waste management have half-lives on the order of thousands to millions of years (^{239}Pu - 24,400 years; ^{99}Tc - 2.1×10^5 years; ^{129}I - 1.6×10^7 years).

Low-level waste: Wastes containing types and concentrations of radioactivity such that shielding to prevent personnel exposure is not required.

Non-high-level waste: Intermediate- or low-level waste.

Operations: Broad classification of waste management activities in terms of their basic function (e.g., waste storage, treatment, transportation or disposal).

Overpack: Secondary (or additional) external containment for packaged nuclear waste.

Primary wastes: As-generated forms and quantities of wastes.

Repository: A location containing wastes in storage or disposal.

Retrievability: Capability to remove waste from its place in final storage. The method and rate of removal and the subsequent location of the waste must satisfy retrievability criteria.

Secondary wastes: Forms and quantities of all wastes that result from applying waste treatment technologies to primary wastes.

Short-lived nuclides: Radioactive isotopes with half-lives no greater than about 30 years, e.g., ^{137}Cs and ^{90}Sr .

A.3

Solidification: Conversion of radioactive waste to a dry, stable solid.

Technologies: The specific methods for implementing concepts. An example is calcination of liquid high-level waste by using a spray calciner.

Transmutation: Conversion of a radioactive nucleus to another isotope by bombarding it with radiation or nuclear particles.

Transportation: Movement of materials between sites. Intra-site movement is not considered. Includes alternative methods for packaging, handling, and transport of waste materials and plutonium compounds. Concepts include all conventional methods of land and water transport required by the waste management system.

Transuranic elements: Elements with mass number greater than 92. They include neptunium, plutonium, americium, and curium.

Transuranic waste: Any waste material measured or assumed to contain more than a specified concentration (e.g., presently proposed as 10 nanocuries of alpha emitters per gram of waste) of transuranic elements.

Treatment: Operations intended to benefit safety or economy by changing the waste characteristics. Four basic treatment concepts are defined:

- volume reduction
- immobilization of radioactivity
- change of composition
- removal of radioactivity from the waste.

APPENDIX B

CONVERSION FACTORS

Conversion Factors

<u>To Convert</u>	<u>Into</u>	<u>Multiply By</u>
calories, gram	Btu	3.9685×10^{-3}
centigrade	Fahrenheit	$(C^{\circ} \times 9/5) + 32$
centimeters	inches	0.3937
cubic meters	cubic feet	35.31
cubic meters	liter	1000.0
cubic meters	gallons (U.S.)	264.2
grams	pounds	2.205×10^{-3}
grams/liter	pounds/cubic feet	0.062427
hectares	acres	2.47
hectares	square kilometers	0.01
kilograms	pounds	2.205
kilometers	miles	0.6214
kilometers	feet	3,281
kilowatts	Btu/min	56.92
kilowatts	horsepower	1.341
kilowatt-hrs	Btu	3413
liters	cubic feet	0.03531
liters	gallons (U.S.)	0.2642
liters/min	cu. ft/sec	5.886×10^{-4}
meters	feet	3.281
meters/min	feet/min	3.2.81
millimeters	inches	0.03937
mills/kilowatt-hr	dollars/kilogram U	277.2
newtons	pounds	0.2248
square centimeters	sq. inches	0.1550
square kilometers	acres	247.1
square kilometers	square miles	0.3861
square meters	square feet	10.76
tons (metric)	kilograms	1000
tons (metric)	pounds	2205
watts	Btu/hr	3.413
watt-hours	Btu	3.413
watt-hours	ergs	3.60×10^{10}
watt-hours	foot-pounds	2,656
watt-hours	kilogram-calories	0.8598
watts/cm ² -°C	Btu/(hr-sq ft-°F)	1760.6
(watts/cm ²)(°C/cm)	(Btu/hr-sq ft)(°F/ft)	57.8

APPENDIX C

GEOLOGIC ISOLATION

APPENDIX C

GEOLOGIC ISOLATIONC.1 SITE SELECTION FACTORS FOR REPOSITORIES OF WASTES IN GEOLOGIC MEDIA

Sites for geologic isolation of high-level radioactive waste must be selected to assure that there will be no radiation exposure to the general public, either through direct contact or by indirect means involving other portions of man's ecosystem. Because geologic environments vary, it is conceivable that this fundamental requisite could be met in different ways at different locations according to the nature of the particular geologic formations and the local conditions.

The main natural mechanism which could move radionuclides from a deep underground emplacement into the biosphere is the action of groundwater. Hence, most of the geologic and hydrologic factors considered for each rock type are related to the protection of the waste from circulating groundwater. Due to the numerous types of geologic formations and their varying conditions, the extreme complexity of many geologic environments, and both the interrelated and independent nature of the rock features that govern the presence and circulation of groundwater, it is not feasible to derive strict criteria for selection of a geologic site for a radioactive waste repository. Each potential repository location must be evaluated according to its own unique geologic and hydrologic setting. However, the following general site selection guidelines are considered to facilitate the screening process:

Topography

In general, low relief and gently sloping terrain should characterize the topography at waste repository sites. Since rail and/or highway transportation must be provided for such facilities, it is essential that topography be amenable. In some instances, the surficial features of an area reflect the subsurface conditions. Thus, irregular topography may reflect complex geologic and hydrologic conditions in the subsurface while more regular surface terrain may indicate flat-lying, undisturbed rocks at depth. The screening process for thick bodies of salt should be limited to those with low to moderate relief at and near the sites. Excessive relief could create differential loadings on the salt formation that could trigger salt flow and diapirism.

Tectonics and Seismicity

Because a major crustal disturbance would adversely affect any type of rock, areas of low seismicity and tectonic stability are favored for waste disposal facilities. In addition, surface and shaft handling facilities or surface entrances into the waste repository chambers are susceptible to damage

during violent earthquakes. In extreme cases, these events could lead to temporary disruption in operations and even possibly to breaching the geologic containment for some types of rocks. A credible consequence could be flooding of the waste repository due to a shaft failure where it passes through an aquifer. On the other hand, there is no documented evidence of earthquakes causing extensive structural damage to deep mines.

In the long term, the major risk for the integrity of a repository would be faulting through the disposal zone. For the more plastic-behaving rocks, such as salt and possibly clay and shale, any fractures are healed and geologic containment is sustained. However, for brittle rocks, such as granite and limestone, fractures commonly lead to passageways for the free circulation of groundwater and thus to some reduction in the effectiveness of the rock to contain the emplaced wastes. In general, regions having a long geologic history of tectonic stability also exhibit a low frequency and intensity of seismic events. Conversely, those areas that have experienced mountain-building episodes in recent geologic time also are likely to have a high incidence of historical earthquake activity.

In assessing the tectonic state and seismicity of selected areas, regions can be classified by the occurrence and distribution of historic earthquakes and through specialized geologic studies. Microearthquake studies may be used to identify localized instabilities.

Stratigraphy

Depth of Disposal Zone

Due to weathering, most rocks lying within a hundred meters of the land surface contain numerous open fractures capable of transmitting water. These fractures, coupled with erosion, make it highly desirable that the screening process be limited to prospective rock zones at depths of at least 200 to 300 m. Thus, for some arid areas, most waste emplacement levels will be below the substratum of local water table aquifers. Because of a pronounced acceleration in the rate of plastic flow of rock salt as the pressure increases with depth, it is prudent to restrict mechanical mining operations in salt to depths of no more than about 1,500 m. However, solution-mined openings in salt could probably be maintained for waste disposal to maximum depths of 3,000 m. For shale rocks, it is expected that stable openings could not be formed below 1,500 m from the surface. The brittle rocks such as limestone, granite and basalt could undoubtedly be mined at depths considerably greater than 1,500 m. But the increased costs of testing, evaluating and actually developing disposal space at these great depths would appear to limit greatly their potential for repository use. Thus, it is clear that the most common depths for disposal would range from about 300 to 1,500 m; however, slightly shallower and somewhat deeper rocks may also be suitable under certain conditions.

Another factor that might relate to the desirable depths of disposal is the increase of ambient temperatures with depth in the earth's crust. In areas of normal geothermal gradient the temperature increases about 30°C every kilometer of depth. However, there are many areas where geothermal gradients significantly higher or lower have been observed. In exceptional cases, the geothermal gradient can be as high as 80 to 100°C/km of depth, but these cases always occur in areas of recent volcanism where waste repositories could not be situated anyway.

Formation Configuration - Thickness and Extent

In general, the formation must be of such vertical and lateral extent that any fractures coming from the immediate surroundings of the emplaced wastes will be absorbed or buffered so as not to reduce the containment effectiveness of the host rock. Also, the host rock must be large enough to provide for adequate heat dissipation. Generally, for plastic rocks like salt and shale a smaller volume of rock would be needed to absorb and heal fractures than for the brittle rocks like limestone, granite and basalt that deform elastically. Wherever brittle host rocks are completely surrounded by more plastic-behaving rocks, it is likely that the host rocks could be less extensive.

Consistency, Uniformity, Homogeneity or Purity

In general, rocks having a high degree of homogeneity or consistency can be considered superior to others for emplacement of radioactive wastes. However, the limits of acceptability of nonhomogeneity can not be stated unequivocally. They can vary greatly, depending on such factors as the type of host rock, the nature and characteristics of the "contaminating" rocks and minerals, the types of waste, and the design of underground waste emplacements. For sedimentary rocks like limestone, salt, and anhydrite, it is common to find rather homogeneous layers that are 30 cm or more in thickness separated by thin partings of shale and/or other minerals. The nature and location of these partings are important in mining; often, when widely spaced, they are used to accommodate a smooth and continuous roof for the workings. However, closely spaced shale partings may contribute to roof instability as the bulk of the rocks tends to separate along these partings and subsequently fails.

Hydrated minerals or other rock constituents that react adversely with waste products can be detrimental, especially for high-heat generating wastes. Heat dissipation may also be adversely affected in rocks that contain significant quantities of certain impurities, especially when they occur as discrete and continuous interbeds near the disposal horizons. Ore bodies or potentially mineable minerals associated with the host rock can detract from the usefulness of such rocks for waste emplacement, as would interbeds of differing rock types known to contain circulating groundwater.

Nature and Extent of Overlying, Underlying, and Flanking Beds

Although the primary geologic containment for radioactive waste emplacements in rock formations exists, for the most part, within the host rocks, it is apparent that additional protection and/or containment may be gained through impervious beds that might surround the host rocks. Thick beds of shale and/or other plastic-behaving rocks that are impermeable to circulation of groundwater are generally preferred as they may deform without fracturing. This may enable them to accommodate the subsidence and/or buckling, for instance, due to thermal expansion of the host rocks. Perhaps the least desirable surrounding rocks are those containing large quantities of circulating groundwater, especially in the overlying beds. These waters have the potential for flooding the waste-filled cavities should unexpected fracturing occur in the host rocks. Most rocks contain some open passageways for circulating water near to the land surface, but where separated by 300 m or more of impervious rock, the potential hazard of this water to the emplaced wastes should be minimal.

Structure

Dip or Inclination

In bedded sedimentary rocks, excepting salt diapirs, the preferred dip or inclination of strata for waste emplacement should not exceed a few degrees, as the design of underground workings and of subsurface waste transport operations favors low haulage grades and nearly horizontal mining levels. Also, to maintain consistency and uniformity, it is desirable to restrict excavations to specific stratigraphic horizons or beds that almost always parallel the dip of the strata. Sudden and frequent reversals in dip would therefore be disfavored. Steep dips indicate that the rocks have been subjected to severe tectonic stresses that commonly fracture brittle rock and create passageways for circulation of groundwater. Thus, even though underground openings could probably be designed to accommodate steeply dipping beds, their utility for waste disposal would be lessened by the increased potential for circulation of groundwater. Also, the "protective" covering of impermeable, overlying beds would not generally be present in steeply dipping formations. For plastic-behaving rocks, like salt and some argillaceous sediments, deformations accompanying moderate to steeply dipping beds would tend to have a minimal effect on water movement as fractures would be healed and thus not open to circulating groundwater. However, for brittle rocks like limestones and anhydrite, fractures would not heal readily but would enhance any movement of water in the formations.

Igneous and metamorphic rocks, because of their modes of origin, are generally more lithologically consistent than sedimentary rocks. Thus, the dips or inclinations of planar features, like faults, joints and other structures, may affect the suitability of these rocks for waste disposal more than would

lithologic variations. Nevertheless, the preferred orientation of these features can not be generalized but would be governed by the local and regional geologic settings of special study areas.

Faults and Joints

In general, faults and joints have a deleterious effect on rocks that are candidates for waste emplacement as these features cause physical discontinuities in rock masses. They can affect the initial mining as well as the stabilities of such openings once they are excavated. More importantly, they constitute the principal pathways for water circulation in many rocks like limestone, granite, and other crystalline rocks. If case faulting through the disposal zone were to occur after closure of the repository, plastic formations should be able to maintain their integrity whereas containment could be compromised in brittle rock.

Major faults that can be traced for hundreds of kilometers on the earth's surface generally disrupt rocks over broad zones at least several kilometers in width and most assuredly should be avoided entirely in selecting areas for waste repositories. On the other hand, some minor flexuring occurs in most rocks, and its significance in siting can be evaluated only in light of such factors as its nature and extent, rock types, and hydrology. In general, faults and joints are more open and thus freer to transport groundwater near the land surface than at depths of hundreds of meters. However, depth alone is not sufficient to guarantee "tightness." Independent appraisals must be made of the fault and joint systems at individual sites although these perturbations adversely affect the suitability of brittle rocks for waste utility more than the plastic-behaving ones.

In summary, favored areas for waste repositories are those having no or few faults or joints and if they are present, locatable, so that they can be avoided or circumvented in the excavations for emplacement.

Diapirism

In extreme cases where thick bodies of salt or other natural plastic-behaving rocks exist at depth under sufficient differential pressures, mass flowage of these rocks may produce diapiric structures like domes and anticlines. These types of structures may be especially suitable for waste disposal if it can be determined that they are stable and will not begin to flow again during the hazardous lifetime of the wastes. The energy from high-heat generating wastes affects too small a volume of the formation and persists over too short a time to initiate or reactivate diapirism. Nevertheless, for some salt and argillaceous deposits that are thick and deeply buried and exhibit extreme surface relief, geologic examinations must ascertain that diapirism is not incipient and that geologic processes during the next few hundreds of thousands of years will not create conditions conducive to such movements.

Physical and Chemical Properties

The physical and chemical properties of rocks are extremely important in establishing the utility of various types of geologic formations for the emplacement of radioactive wastes as they determine such things as the circulation of groundwater, the dissipation of radiodecay heat, the effects of radiation, the stability of mined openings, the waste form, and any manmade barriers used to contain the waste.

Permeability, Porosity and Dispersivity

Rocks that possess low permeabilities are favored for radioactive waste disposal. This is chiefly related to the preference for utilizing dry excavations for waste emplacement. In most cases, rocks having low permeabilities will also possess low porosities. However, argillaceous materials are commonly characterized by extremely low permeabilities while still having significant porosities. Pore porosity (intergranular voids) is generally low (<5%) for many rock types such as salt, limestone, and granite. Interconnecting pore space (primary permeability) generally is extremely low for these same rocks. These conditions commonly lead to dry excavations; however, should fractures exist in these rocks, their permeabilities (fracture permeability) may be increased greatly (excepting the self-healing rocks like salt and some shales) and thereby enhance the free circulation of groundwater. A more thorough discussion of the influence of permeability and porosity on water circulation is given in the section on Hydrology.

Inclusions of Gases and Liquids

Inclusions of small amounts of gases and liquids may occur in many evaporites. For high-level, heat-generating wastes, these inclusions can adversely affect repository operations through rock decrepitation and the migration of brine toward heat sources or waste containers. For example, in some rock salt deposits at temperatures below the decrepitation point of about 250°C, small brine-filled cavities (size ranging from a few millimeters down to microscopic) may migrate toward heat sources. The mechanism for cavity migration is the diffusion of sodium and chloride ions from the hot to the cold site of the cavity due to the slightly greater solubility of NaCl at the higher temperatures. Migration rates are directly proportional to temperature gradient and increase with increasing temperatures. When these cavities represent less than about 1% of the salt volume, the maximum inflow of brine around buried waste containers would be expected to be small and would probably cease after a few tens of years. Other than its effect on radiolysis and container corrosion, this moisture should be insignificant.

For most other rock types, migration of fluids by the foregoing mechanism would probably be inconsequential. However, localized fracturing of the rocks could occur should the rocks be heated beyond their decrepitation points.

Isolated brine or other fluid inclusions should not adversely affect the disposal of low-heat generating wastes in rock salt or any other rock types.

Frequently, in mining in sedimentary rock sequences, small pockets of gas are encountered that cause blowouts and/or floor and ceiling buckling. In most cases, rock pressures immediately around the openings can be relieved by drilling relief holes into the floor or ceiling. Once encountered, the gas, if it exists in small quantities, may be discharged through the normal ventilation system of the underground workings without deleterious effects. In extreme cases, where rich beds of carbonaceous materials are interbedded with rocks like shale and salt, seeps of methane or other hazardous gases may persist which would be extremely difficult to combat in operations requiring the use of men and equipment in the underground workings. Geophysical methods are available to locate gas-filled cavities prior to their being encountered by mining.

Rock Mechanical Behavior

For all rocks potentially suitable for emplacement of radioactive wastes, it must be established that the transient and permanent rock deformations (displacements, strains and stresses) induced in the rocks will not produce conditions leading to a breach of the integrity of the long-term containment. These deformations are produced by the designed closure of the mine openings and, in high-heat generating wastes, by the thermal expansion of the entire rock column. Plastic-behaving rocks like salt and some shales generally creep, especially over long periods of geologic time, to fill subsurface openings without rupturing or fracturing. The wastes contained in these types of rocks are thus eventually encapsulated by the host rock and, since open fractures do not occur, circulating groundwaters are prevented from contacting the wastes.

For the most part, subsurface excavations in brittle rocks may be expected to resist complete closure for the duration of the hazardous life of most radioactive waste products. However, some localized spalling and deformation of the walls and roofs of such openings would occur over long periods that would probably modify the shapes and sizes of the original cavities. Backfilling would minimize the extent of rock collapse or other rapid deformations; it is apparent that fracturing could proceed upward and in extreme cases might reach the land surface. This might lead to the circulation of groundwaters in the disposal zone and hence to some loss of geologic containment. In general, the deeper the excavations the greater the overburden pressure and hence the greater the deformations. Conversely, at shallow depths the rock loading and resulting deformations may be less, but there also would be less cover rock to absorb any fractures emanating from the disposal cavities.

Thermal Effects

The solidified high-level wastes may be cooled for some period in a storage facility to reduce the initially high incidence of fission product decay heat. After this period, though the heat generation rate would have lowered considerably, keeping the solidified waste thermally stable below certain temperatures, would still be significant heat generation (up to a few tens of watts per liter). This would be superimposed on the ambient temperature existing at the depth of the disposal.

Preferably, therefore, rocks to be utilized for the disposal of high-level, heat-generating waste should possess thermal properties that promote rapid dissipation of waste decay heat and whose stabilities are not adversely affected in the presence of elevated temperatures. Rock salt possesses rather high thermal conductivities compared to some other earth materials, but it also is known to exhibit accelerated rates of creep at elevated temperatures. The latter characteristic of accelerated creep is generally desirable in certain plastic-behaving rocks as this promotes sealing and healing of excavated cavities that contain the waste. Conversely, for brittle-behaving rocks, accelerated deformation at elevated temperatures would be detrimental as it might lead to fracturing and some loss of geologic containment. For argillaceous rocks, heat loading may cause significant and perhaps undesirable, changes in physical properties.

To determine the effects of imposed thermal loading on the temperature distributions and the structural stabilities of various host rocks, the appropriate thermal transport and mechanical strength related properties, including thermal conductivity, specific heat, coefficient of expansion, rupture stress and creep rate, must be defined and determined.

Retention Capacity

Although completely dry environments are preferred for emplacement of radioactive wastes, some water-saturated argillaceous rocks, like clay, mudstone, and some shale, that transmit water only at extremely slow rates may also be acceptable host rocks. These water-saturated host rocks must possess high adsorption capacity as this property associated with the extremely low transmissivity would assure the long-term containment of the waste. For competent rocks like limestone, granite, and basalt that transmit water only along fractures and hence possess extremely low effective adsorption capacities, it is imperative that the waste remain isolated from circulating groundwater. Nevertheless, adsorption could provide a secondary geologic barrier against dispersal of radioactivity should the primary geologic containment be accidentally or otherwise breached. In particular cases other phenomena such as precipitation and co-precipitation might contribute to restricting the geochemical mobility of certain radionuclides. Therefore, it is important that

the retention potential and groundwater flow be carefully examined at each individual site. Additionally, the effect of heating on the adsorption capacity of each host rock must be examined.

Mineral Sources of Water

Significant quantities of hydrated minerals may create undesirable conditions in rocks that are candidates for the emplacement of high-level, heat-generating wastes as these minerals release water at elevated temperatures. One of the most troublesome of these minerals is gypsum ($\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$), which occurs in varying quantities, especially at shallower depths in evaporate sequences. Gypsum is known to dehydrate at atmospheric pressure in the temperature range of 110 to 200°C under conditions of relatively rapid heating. Furthermore, under conditions of very slow heating, such as would occur in an actual disposal facility, it is expected that the gypsum would dehydrate at even lower temperatures or, at least, in the lower range of reported temperatures. Gypsum contains 21 wt% water. Hence, 1 m³ of pure gypsum would be capable of releasing about 480 liters of water. A number of other hydrated saline minerals have been recognized in evaporite sequences, including such minerals as polyhalite, carnallite, bloedite, and epsomite. Locally the quantities of such minerals may be significant. Most of these minerals dehydrate at higher temperatures than does gypsum. Nevertheless, should these minerals be present close to the disposal horizon, the quantities, distributions and characteristics of dewatering must be determined.

In addition to hydrated evaporite minerals, it is known that pore water and absorbed water held on clay minerals may be driven off at temperatures well under 200°C. When these minerals or other hydrated materials occur near the disposal zone, they must be considered as potential sources of water, and such factors as the rate of dewatering, the mechanism(s) and path(s) by which the freed water might escape or be recombined, must be evaluated.

Radiation Effects

Energy may be stored in irradiated solids due to the disorder induced by radiation damage. When the solid is heated following its irradiation, energy is released as the solid passes through various stages of reordering. In some cases, all of this energy can be released with a relatively low initiating temperature.

It is known that gamma-radiation energy can be stored in salt and perhaps other rocks. Furthermore, the amounts stored would depend not only upon dose but also upon the rate at which the dose is accumulated and upon the salt temperature. Thermal annealing is probably the most important factor in limiting the stored energy in rock salt exposed to high radiation doses and correspondingly high temperatures. Studies of irradiated salt indicate that annealing will completely suppress the formations of significant amounts of

stored energy at temperatures above about 150°C. Also, at temperatures appreciably less than 150°C where thermal annealing is small, the total dose would be relatively small. Thus, at the maximum, only a moderate amount of stored energy will accumulate in salt exposed to high-level radioactive wastes.

The problem of the radiation effects on the rocks does not occur for alpha-bearing wastes, and it has to be considered only for the disposal of high-level waste. As we have only few data on how the radiation affects rocks other than salt, it is thus essential to establish the identity and magnitude of radiation effects on the various types of rocks that would be otherwise potentially suitable for the disposal of high-level radioactive wastes.

Hydrology

Surface Waters

The mere presence of surface streams, lakes, ponds, etc. above otherwise suitable repository sites would not necessarily rule out their use. However, it must be ascertained that these waters would not interfere with the short-term operation of a facility or jeopardize the long-term geologic containment of any emplaced wastes. Flood plains or other locations vulnerable to flooding waters are especially difficult to develop as extreme conditions could lead to filling of underground excavations through water flow down open shafts, boreholes, etc., unless unique design precautions were utilized. Recreational, scenic, and even industrial interest in large lakes and streams may also create public resentment for competing waste disposal sites in such locations.

Over long periods of geologic time, surface streams may be expected to undergo radical changes in their flow regimes. Under extreme conditions of erosion, deep incisions may be cut into the rocks, while for extended periods of stream aggradation (deposition) thick beds of sediments may accumulate. In any event, it is clear that the future behavior of surface streams must be predicted to ensure that geologic containment can be maintained for the required period of time.

Groundwaters

Circulating groundwater poses the main real threat to containment of radioactive wastes placed in geologic formations. Thus the nature and characteristics of water-bearing formations close to potential disposal zones are critical elements in establishing the suitability of specific rock types and special study areas.

Nature and Occurrence. Generally, the host rock for wastes should be free of circulating groundwater although overlying and underlying formations may contain water-bearing zones. For plastic-behaving rocks like salt and shale, the vertical as well as the horizontal distances between circulating groundwater and waste emplacement zones can generally be less than those for brittle rocks. Thick bodies of rock salt lying at relatively shallow depths

commonly are in contact with circulating groundwater along their uppermost boundaries. In these instances, the disposal levels should lie at least ~100 m below the water zones and/or at such distances that the projected rates of dissolution of the salt, which incorporate projected possible hydrological and climatic changes, will not uncover the wastes during its hazardous period. Water-bearing zones are much less common at the base of thick salt formations; however, should they be present, waste emplacement zones should also lie ~100 m above these waters and/or at an appropriate distance to ensure long-term containment of the wastes.

Ideally, for hard, brittle rocks any overlying and/or underlying aquifers should be separated from the host rock by thick bodies of impermeable shales or other aquicludes. In some cases, it may be necessary to choose host rocks containing some water for waste repositories. However, under these conditions it would be imperative to demonstrate that the radionuclides, if mobilized, would move at such a slow rate that significant quantities of them would never reach the biosphere, being confined to some acceptable zone in the subsurface.

Direction, Velocity, and Volume of Groundwater Flow. To properly assess the hydrologic regimes at candidate sites for waste repositories, determinations must be made of the directions of water flow, linear velocities and volume rates of such movement, both in the host rocks, when they may be permeable, and in other rock types which may intervene between them and the biosphere. Thus, it will be necessary to determine, in the field, the three-dimensional geometry of the several rock types, the porosity and hydraulic conductivity of each type and the hydraulic gradients in and between them. Data that are representative in three dimensions should be sought, within and beyond any zone which might be dedicated to, or conceivably might be invaded by radionuclides.

Horizontal and vertical conductivities must be determined separately for rocks that may be anisotropic. Conventional methods of investigation to determine their hydraulic properties may be ineffective, especially when the rocks possess very low conductivities. Commonly, ad hoc test procedures will need to be adopted. These special tests might include such methods as pumped circulation between boreholes combined with an isotopic tracer, injecting water at less-than-critical pressure, pumping boreholes by depth zones isolated between packers, and so-called "slug" or injection tests.

In general, groundwater velocities in and around waste repositories cannot be measured directly but can be estimated since they are proportional to hydraulic conductivity to effective porosity, and to gradient. Actual velocity at a particular point may deviate considerably from the computed mean, according to variations of effective porosity and permeability within any one water-bearing medium. For instance, actual velocities in crystalline rocks can be expected to range widely because of variations in the geometry of the fractures.

Theoretically, any combination of conductivity and gradient leading to a hydrologic regime would be acceptable that promotes extremely slow groundwater velocities at a potential waste repository and/or prevents escape of radionuclides to the biosphere for the required periods of containment. However, errors and uncertainties are inherent in such determinations since future geologic processes, climatic changes and actions induced by man can alter drastically hydrologic regimes. It is therefore believed that a much greater reliability of containment would exist for disposal schemes using host rocks that are virtually impermeable than exists for water-bearing rock systems which depend predominantly on low groundwater velocities for limiting the dispersion of waste products.

Some argillaceous sediments that, for all practical purposes, are generally considered impermeable, behave in reality as semipermeable membranes. The energy needed to drive water through the membrane can be provided by compaction of the sediments, by hydraulic gradients, or even by thermal or electrical potential gradients. When water is transported by molecular diffusion through a semipermeable membrane, an ion-filtration mechanism operates leading to salinity gradients in the fluids. Consequently, osmotic pressures can exist in sequences of argillaceous sediments. With the mining of an opening in a mass of argillaceous rocks, a point sink could be created that might draw fluids into the opening by means of pressure differentials. Should the opening be filled with fluids, the point sink would ultimately diminish until the fluid movement would return to the conditions existing prior to the mining perturbation. In this case a phase discontinuity would exist between the cavity and the surrounding rock. Although the transport of radionuclides in argillaceous sediments would be expected to be extremely small, this would probably need to be demonstrated through laboratory and even in-situ tests in the host rocks of interest. It may also be pointed out that in certain cases osmotic phenomena can modify the local direction of groundwater flow.

In many areas, especially arid ones, groundwater is an extremely important commodity as it may be the principal source of water for municipalities, industry, and agriculture. Thus any actions that would threaten contamination of these waters (even remotely) would naturally be unfavorably received by the general public, especially in the affected regions. Therefore extensive deposits of fresh groundwater that lie above or below otherwise acceptable waste repository sites could adversely affect the availability of such rocks for waste emplacement.

Future Geologic Events

In general, geologic processes work slowly to form mountains and otherwise generate prominent physiographic changes on the earth's surface. However, some processes like faulting and related earthquakes, volcanic activity, and glaciation can cause dramatic changes in landforms, locally as well as regionally,

within geologically short periods. The future occurrence of these events must be considered in site selection investigations. Based on geologic history, the general location and maximum effects of these events can be predicted. For example, within the time span considered, it is reasonable to expect at least another glacial episode. During the last glacial age, ice sheets covered about half of North America and all of northern Europe. These areas underwent profound alteration as a consequence of scouring and deposition, and isostatic adjustment to loading and unloading. The areas not covered by glacial ice were affected by drastic climatic changes. Some large new lakes and streams were formed while others underwent profound modifications in deposition and erosion, occasionally in a catastrophic manner. Associated with these changes in the surficial water flows were corresponding changes in groundwater regimes.

The following sections are geologic descriptions of some of the occurrences of the rock types being investigated as potential repository sites. Wherever possible, tectonic history and hydrology are also described. This set of rock types and geologic/geophysical areas is not intended to be comprehensive. Instead it is meant to illustrate the types of information available and to give a cross section of the geologic investigations in progress.

C.2. ROCK TYPES - GEOLOGIC OCCURRENCE

The following sections are geologic descriptions of some of the occurrences of the rock types being investigated as potential repository sites. The types of rock considered include bedded and domed salt, shales, carbonates, metamorphic rocks, flood basalt, and talc-serpentine deposits. General areas containing the different rock types are shown in Figure C.1. Major bedded salt deposits are the Permian and Supai Basins, the Luke Salt Body, and those found in the Northeastern United States (Michigan, New York-Pennsylvania, and Ohio). Domed salt is located throughout the Gulf Coast states and the Paradox Basin. Potentially suitable carbonate sites are found near Barberton, Ohio, and Kansas City, Kansas. Metamorphic rocks considered are generally in the Canadian shield area while flood basalts are in the Northwestern U.S. The Appalachian Mountain Region, including the Piedmont, Blue Ridge, and New England Provinces, contain shales and talc-serpentine deposits. Shales are also found in the Interior Plains (especially the Michigan Basin) and the Southeast Triassic Basin.

Whenever possible the section describes tectonic history and hydrology. This appendix of rock types and geologic/geophysical areas is not intended to be a comprehensive list. Instead it is meant to only illustrate the types of information available and to present a cross section of the geologic investigations in progress.

FIGURE C.1. Rock Salt Deposits in the United States, Parts of Which are Potentially Suitable for Waste Disposal

C.2.1 Bedded Salt - Permian Basin

The Permian Basin is not a single structural basin. It comprises portions of many structural basins that contain Permian sediments of several ages from Wolfcampian to Ochoan.

The Permian Basin contains at least five areas where salt beds within a formation cumulatively total greater than 65 m and are overlain by between 310 and 1550 m of rock (Figure C.2). These salt beds are in the Colorado-Kansas, Kansas, Oklahoma-Texas, Clovis and Carlsbad segments of the basin.

A few other areas within the Permian Basin may contain salt beds collectively greater than 65 m thick, but published information about much of the remainder of the Permian Basin is scarce.

The very large Carlsbad area contains a great thickness of salt (Figure C.2). There are few data available for the Texas portions of the Clovis and Oklahoma-Texas areas and for the Colorado-Kansas area, especially for the Colorado portion.

The Carlsbad Area sediments were deposited in two basins, and two different major stratigraphic sequences developed, one in the Delaware Basin and the other in the Midland Basin. Figure C.3 shows the interrelation between these basins with carbonate reefs, such as the Capitan Reef having developed in shallow water around the peripheries of the basins. The generalized stratigraphic column for the Gnome site (Figure C.4) serves as a good example of Delaware Basin stratigraphy.

Deposition was thickest in the central part of the Delaware Basin, but some formations are thickest on the margins, as exemplified by Pennsylvania strata and some evaporite beds. Major lateral changes in sedimentary rock of the basin occur at the reefs. Shelf rocks close to the reefs consist of dolomite with subordinate amounts of limestone, shale, sandstone and a little anhydrite. Further away from the reefs onto the shelf area dolomite tends to give way to anhydrite. Deposits of the basin are typically shale, fine-grained sandstone, and dark-colored limestone.

Salt Bearing Formations

There are two formations that contain many thick salt beds in the Delaware Basin, Salado and Castile. The Midland Basin contains only the Salado. The Rustler Formation overlies the Salado in both basins, but the Rustler contains only a few thin beds of salt. Areas with more than 310 m of overlying strata and greater than 65 m of combined thickness of salt beds in the whole formation stratigraphic section are shown in Figure C.5. Several isopachus maps of salt bearing formations are available.^(1,2) Salt bed variability in thickness across the basins is shown in Figure C.6. From this cross section one

FIGURE C.2. Permian Basin: Areas with a 200 foot, or Greater, Thickness of Halite between 1,000 and 5,000 feet below Ground Surface

FIGURE C.3. Cross Section of Delaware and Midland Basins and Central Basin Platform. After Eardley, p. 229.

Generalized Stratigraphic Column at Gnome Site and Description of Lateral Variations. After Brokaw et al., 1972, p. 18.

FIGURE C.4.

FIGURE C.5. Cross Section of Salt Beds in Delaware and Midland Basins. After Pierce and Rich, p. 34

FIGURE C.6. Folds in Castile Formation. After Jones et al., 1973.

can see that beds in general are thicker and more numerous toward the center of the basins. Thickest salt beds occur in the Salado Formation.

The Salado Formation contains thick beds of salt and thin intervals of anhydrite, shale, and polyhalite. Locally the middle of the formation is rich in potash minerals. Salt of the Salado Formation is generally less pure than that in the Castile Formations. Exceptions are the very pure basal beds of the Salado Formation in the Delaware Basin. The thickest salt unit in the Carlsbad area of more than 525 m is on the east and north side of the Delaware Basin within the Salado Formation. In the center portion of the Midland Basin there is a salt unit in the Salado Formation more than 310 m thick. Overlying the Salado Formation is the Rustler Formation which contains much anhydrite. And the underlying rock type is, in most places, anhydrite. Anhydrite beds above and below the Salado Formation apparently act as barriers to water which might dissolve the salt. The Dewey Lake Red beds about 155 m above the salt beds of the Salado Formation probably contain permeable water-bearing zones. The Santa Rosa Sandstone about 310 m above salt beds of the Salado Formation is permeable and waterbearing.

The Castile Formation

The Castile Formation is restricted to the Delaware Basin. There are several beds of relatively pure rock salt up to 215 m thick. A maximum aggregate thickness is about 310 m. Salt beds are separated by anhydrite beds up to 155 m thick. Salt comprises no more than 60% of the formation thickness. Fine grained sandstone of the Bell Canyon Formation immediately underlies the Castile Formation in most areas. The sandstone is likely permeable and water-bearing; however, overlying and underlying salt beds there are beds of anhydrite which act as barriers to dissolution.

Overlying permeable water-bearing strata are the Santa Rose Sandstone and the Dewey Lake Red beds about 930 and 775 m, respectively, above the thick salt beds in the Castile Formation.

The strata dip gently inward toward the central areas of the two basins: the Midland Basin with a maximum of about 3100 m of strata and the Delaware Basin with a maximum of about 6200 m of strata. The basins are separated by an extremely gentle arch, the Central Basin Platform.

The Central Basin Platform has two main fault zones on either side of the elongate Platform which trends north-south. The faults are steep with the down side toward basins. Near the south end of the Platform there are several intersecting faults which strike northwest.

There are faults which almost encircle both the Midland and Delaware Basins.⁽³⁾ Many faults are along the west side of the Delaware Basin^(4,5) and Kelley⁽⁶⁾ has interpreted a prominent lineament as one of these fault traces; therefore, this fault may be of Holocene age. He has also mapped slight warps

in strata of the backslope of the Sacramento Mountains, and these structures may extend into the salt beds.⁽⁷⁾

At least three alkali syenite dikes cut the Castile Formation south of Carlsbad. They are 310 to 1240 m long, strike about N60°E, and dip is about 80°N⁷. And on the south side of the Delaware Basin there is a large field of Tertiary volcanic rocks, the Davis Mountain Volcanic Area.⁽³⁾

In the Midland Basin a few miles southwest of Midland is the Odessa Disturbance, covering about 4 hectares. In an area of almost horizontal strata there are dips up to 30° near the Odessa Crater. Many meteorites have been found in the immediate area, establishing a meteoritic origin for the feature. There are many solution depressions in the surrounding plain.⁽⁸⁾ There are many lineaments, especially with a northwest strike, over most of the Carlsbad area. A few have been attributed to deflation between longitudinal sand dunes.⁽⁷⁾ Olive⁽⁹⁾ has explained lineations in Eddy County, not far from Carlsbad, as due to solution-subsidence troughs along joint systems in gypsum. Small Holocone domes have been described in southeastern New Mexico by Vine.⁽¹⁰⁾

Hydrology

Surface Water. The Pecos River runs along the southwest side of the area. This is the only major perennial stream. It flows to the southeast. Most of the area drains east-southeastward across a plateau along subparallel temporary streams such as Sulphur Springs Creek. The area is very arid, and there are many temporary lakes and small playas. There are only a few small, permanent lakes, and some are manmade. One perennial lake is Salt Lake, east of Loving.

Groundwater. The main aquifers are the Culebra Dolomite Member, the basal solution breccia zone of the Rustler Formation, and the Santa Rose Sandstone. Locally, the Gatuna Formation is an aquifer. The basal solution breccia zone is the unit that is most significant in solution of rock salt in the upper part of the Salado Formation. Formations above the Salado seem to be interconnected and to constitute a single hydrologic system. A map of the water table⁽¹⁾ is useful in predicting flow within the formations above the Salado Formation. The water table is 65 to 435 m above rock salt in the Salado Formation. Near Carlsbad, groundwater above the Salado Formation moves generally southward and southwestward into the Pecos River near Malaga Bend.⁽¹⁾

Permeability of the Salado Formation is low. The dissolution front of the Salado Formation is estimated to have moved laterally at a rate of 6 to 6.7 km per million years.⁽⁷⁾

Aquifers below the Salado Formation are the Capitan Limestone, Yates, and Tansill Formations. Water in them is under artesian pressure.

Information about Areas

The Clovis area is in the Palo Duro Basin of Texas and Tucumcari Basin of New Mexico. These are bound by the Amarillo Uplift (a west-northwest extension of the Wichita Uplift) on the north, the Matador Uplift and Sierra Grande Arch to the west.

The Palo Duro Basin is a shallow foreland portion of the foreland basin of the Paleozoic Marathon Mountains orogeny to the south. The Palo Duro Basin was localized by bounding high areas during different times within the Paleozoic Era. One of these high areas, the Matador, is unusually narrow. It is bound on the south by faults and dikes. The Amarillo Uplift has faults on the south adjacent to the Palo Duro Basin. Part of the Ancestral Rockies of Pennsylvanian age, the Pedernal Uplift, was intruded by ultrabasic rocks in the Mesozoic Era. The area was tilted, moderately folded and faulted between Late Cretaceous and Early Tertiary time.⁽¹¹⁾

Several earthquake epicenters of small intensity have been located in and near the Clovis area,^(12,13) but the only concentration of epicenters which might be associated with known structures are those along the Amarillo Uplift.

The Upper and Lower Permian Artesia Group is composed of the Tansill and Yates Formation, Seven Rivers Formation, and Green and Grayburg Formations. The Tansill and Yates Formations are up to 65 m thick and composed of sandstone, siltstone, and minor amounts of dolomite, anhydrite, and rock salt. The Seven Rivers formation is 47 to 155 m thick and composed of rock salt, shale, and anhydrite. The Green and Yates Formations are 65 to 115 m thick and composed of sandstone, shale, and minor amounts of rock salt, anhydrite, and dolomite. The Artesia Group is a complex deposit of red beds and evaporite.

The Lower Permian San Andres Formation is below an unconformity. It is 200 to 390 m thick and composed of rock salt, anhydrite, shale, and dolomite. The formation thickens northward and eastward and is composed of lower and upper members. In recent years this formation has been an exploratory target for petroleum.

Conformably below the San Andres Formation is the Lower Permian Glorieta Sandstone, 28 to 48 m thick, composed of sandstone and minor dolomite and anhydrite. Halite may be in the sandstone as a cement in Curry and Roosevelt Counties. The sandstone is generally permeable and yields brackish water to wells in the western part of DeBaca County and salt water to wells in the eastern part of the county.⁽¹⁴⁾

The Abo and Yeso Formations consist of a red bed sequence of shale and sandstone and lesser amounts of anhydrite and dolomite. Rock salt is also present but data on the two formations is meager.

The salt beds are in a varied sequence of evaporites, red beds, and carbonate rocks known as the San Andres Formation and the Artesia Group.

"The San Andres is mostly salt and anhydrite interbedded with dolomite and shale, and is 660-1250 feet (200-380 m) thick. By contrast, the Artesia Group is chiefly shale and sandstone interbedded with rock salt and anhydrite and ranges in thickness from 335 feet (105 m) to 1,100 feet (335 m)."⁽¹¹⁾ The salt beds do not contain hydrous minerals. Salt beds are thicker in the San Andres and contain fewer interbedded sedimentary rocks.

San Andres Formation. Figure C.7 shows the aeral distribution and thickness of salt in the San Andres Formation. The further distribution outside the area of the mapped area was estimated for Figure C.2. Thickness of overlying rocks exceeds 465 m over most of the area. It is 310 m in the northwest corner and about 930 m in the vicinity of Portales in the southeast corner.

The San Andres Formation consists of an upper member and a lower member. The upper member is 125 to 310 m thick and contains seven major beds of rock salt from 3.6 to 46 m thick. The upper member is largely rock salt and anhydrite interbedded with shale and dolomite. The lower member is 65 to 92 m thick and is predominantly dolomite and anhydrite interbedded with shale.

Artesia Group. The Artesia Group unconformably overlies the San Andres Formation and consists of five formations in descending order: Tansill, Yates, Seven Rivers, Queen, and Grayburg. Rock salt and anhydrite are interbedded with the red beds of the group but are most plentiful in the middle part which includes the Seven Rivers Formation, much of the combined Queen and Grayburg Formations, and a small part combined Yates and Tansill Formations. Salt content of the stratigraphic section and thickness, 510 to 620 m, of overlying rocks is greatest in the vicinity of Clovis and Portales (Figure C.8). Salt beds are thickest, up to 18.5 m, in the Seven Rivers Formation.

The Clovis Permian Formations in general are flat lying. Different stratigraphic horizons may slope gently in different directions. The basement nonconformity dips toward the center of the Tucumcari Basin to the west and toward the Palo-Duro Basin Center to the east. The topmost, formerly continuous stratum, the base (major unconformity) of the Ogallala Formation of Pliocene and locally Miocene age, has a gentle homoclinal eastward dip over a very large area. The top of the San Andres Formation (unconformity) of Early Permian age has a gentle southeastward dip.⁽¹¹⁾

There are a few faults in and near the Clovis area. There are steep faults that strike north-northeast and are known to cut at least Permian through Triassic formations but are not believed to have affected the Ogallala Formation of Pliocene to Miocene age (Figure C.9). Their last movement is considered by Jones⁽¹¹⁾ to be Late Cretaceous or Early Tertiary, but he indicates their precise age is unknown. The Bonita Fault (Figure C.10) is Cretaceous or

FIGURE C.7. Thickness Map of Salt-bearing Interval in Upper Member of San Andres Formation. After Jones, 1974, Figure 3, p. 11.

FIGURE C.8. Thickness Map of Salt-bearing Interval in the Artesia Group. After Jones, 1974, p. 15.

FIGURE C.9. Cross Section Showing Faulting Has Not Off-set Ogallala Formation. After Jones, 1974, p. 19.

FIGURE C.10. Fault Map of the Permian Basin.
After Bachman et al., 1973, p. 30.

younger. It strikes northeast, and strata are down on the northwest side. An east-west striking fault is along the Matador Arch on the south side of the area. There are numerous northwest and a few northeast trending lineaments over most of the area. Their origin and age are unknown.

There are two east-west trending dikes southwest of the area about 16 to 32 km. One is about 9 m wide and 40 km long, and the other 23 to 32 m wide and 48 km long.

Most surface water drains eastward along a subparallel set of channels. The main one is Running Water Creek. A small portion of the area drains westward and enters the Pecos River, which flows southward. There are many closed basins of drainage into playas and sinkholes. Large amounts of sodium chloride are carried by the streams which drain the area.⁽⁷⁾ The sandstone beds of the Artesia Group are water-bearing,⁽¹¹⁾ and sandstone and shale make up about half of the group. The Glorietta Sandstone underlying the salt-bearing San Andres Formation contains permeable sandstone which yields brackish water in the western part of DeBaca County and saltwater in the eastern part.⁽¹⁴⁾ Throughout the area, aquifers are probably fewer to the south because there is less sandstone.

Dissolution of salt beds from outcrop has been taking place very slowly. The present dissolution front has been retreating for 230 million years.

The Oklahoma-Texas area is within the Anadarko Basin with the Wichita Uplift along the south side. To the east of the basin where salt beds thin to extinction is the Nemaha Ridge.

The north flank of the Anadarko basin is a long dip slope of Paleozoic rocks on top of the Precambrian basement. The basin descends to a depth of 6355 m. On the south side, the basin is fault bound, and dip of beds is very steep. The basin developed throughout Paleozoic time with occasional epeirogenic uplifts which produced unconformities.

Since the area was uplifted by epeirogenic forces toward the end of Paleozoic time, no tectonic forces are known to have affected the area. The Oklahoma-Texas area is within the mid-continent central stable region, and seismic activity is low. Docekal⁽¹³⁾ has plotted a MMI VII to VIII and several smaller intensity epicenters near and along the trend of the Wichita Uplift. This suggests activity of faults which extend into basement rocks between the Anadarko Basin and Wichita Uplift. Yet there is no substantiated historical evidence of offset on these faults.

The Nemaha Ridge has four MMI VII to VIII and several smaller intensity epicenters near and along the trend of the Nemaha Ridge. This suggests activity of faulting associated with the Nemaha Ridge. There is no substantiated historical movement on faults associated with either the Wichita Uplift or Nemaha Ridge. Yet, according to Smith⁽¹⁵⁾ there are prominent lineaments on

ERTS images which parallel, if not coincide with, fault traces associated with the Nemaha Ridge.

The Oklahoma-Texas Area is underlain by the Leonardian Clear Fork Group, which contains halite units interbedded with red shale, anhydrite, and some dolomite. The three formations of the group are from top to bottom: the Beckham Formation, Cimarron Formation, and the Wellington Formation. The Beckham Formation is divided into an upper member, the Yelton Salt Member, a middle member, the Blaine Anhydrite Member, and a lower member, the Flowerpot Salt Member.

The Yelton Salt Member is about 90 m thick. At that type of locality the Yelton Salt is 20 to 25% shale. Conformably below the Yelton Salt Member is the Blaine Anhydrite Member. It is about 65 m thick and composed of anhydrite with interbeds of halite and some shale.

The Flowerpot Salt Member is about 90 to 190 m thick. In northern Gray and west-central Wheeler County, Texas, the Flowerpot Salt Member lies about 375 m below the surface of the ground.

The Cimarron Formation is composed of four members: an upper shale member, the Upper Cimarron Salt Member, the Cimarron Anhydrite Member, and the Lower Cimarron Salt Member.

The upper shale member is about 90 m thick and contains discontinuous interbeds of halite and some gypsum and siltstone. This unit would probably be impermeable to downward movement of water toward a radioactive waste repository in the two members below.

The Upper Cimarron Salt Member is about 65 m thick and is coarse halite that is reddish brown due to small amounts of clay. There are interbeds of reddish brown shale and some anhydrite.

The Cimarron Anhydrite Member is about 32 m thick and contains thin interbeds of shale and dolomite.

The Lower Cimarron Salt Member is about 65 m thick with a maximum thickness of about 140 m in north central Wheeler County, Texas. This member is composed of coarse, clear halite. There are a few intercalations of moderately reddish-brown and greenish-gray shale. The halite is generally pure and some sections as much as 9.2 m thick contain 95% halite.

The Wellington Formation is composed of three members: an upper shale member, the Hutchinson Salt Member, and a lower anhydrite member. The upper shale member is about 140 m thick.

The Hutchinson Salt Member is about 120 m thick and is composed of halite and interbedded shale and anhydrite. Halite beds are not as thick as in the overlying two members and there is probably not a halite unit as thick as 65 m in the Oklahoma-Texas Area except in the north portion where the Oklahoma-Texas Area overlaps the Kansas Area.

The lower anhydrite member is about 46 m thick. This unit would probably serve as a barrier to upward movement of water.

The beds dip toward the lowest part of the Anadarko Basin in the southeast corner of the Oklahoma-Texas area. Beds dip gently except along the south side, where they dip steeply northward off the Wichita Uplift or they are faulted with the down side on the north toward the Anadarko Basin. This basin is one of the deepest. It has more than 6200 m of structural relief. There are no faults except those along the south side, and no folds. Yet there is much oil production, and many structures have probably been found during petroleum exploration.

Major streams form a subparallel pattern and flow generally eastward. The longest streams are the Canadian River, the North Canadian River, and the North Fork of the Red River, which runs along the southern edge of the area.

Little information is available on groundwater. Much of the movement of the groundwater is controlled by the deep Anadarko Basin structure. Extensive salt springs in and around the area indicate active solution of Permian halite beds.⁽⁷⁾ Chloride concentration ranges from 20,000 to 200,000 ppm. Nine major springs are estimated to dissolve 6300 tons per day.⁽¹⁶⁾

Mineral Resources

At the southeastern edge of the area, salt is solutioned near Sayre in Beckham County. The Upper Cimarron Salt Member is brined from a depth of 468 m.⁽¹⁶⁾ Salt production in Oklahoma is only about 10,000 tons per year.

Oil and gas are produced throughout much of the area and is concentrated in the southern part where the Anadarko gives way to the Wichita Uplift.⁽⁷⁾ Producing horizons occur throughout much of the Paleozoic section all the way down to granite wash on top of the Precambrian rocks.

The Kansas area is within the Hugoton Embayment bound on the east by the Nemaha Ridge and on the north by the Central Kansas Arch and on the west by the Las Animas Arch. The Hugoton Embayment is a broad, shallow northward extension of the Anadarko Basin. No more than a few hundred meters of sediments were deposited in the Hugoton Embayment.

The Hugoton Embayment began subsidence in the Cambrian Period and continued throughout most of Paleozoic time. There were several epeirogenic uplifts with resulting unconformities. Much of the arching of the Las Animas Arch developed in Pennsylvanian time probably due to subsidence of the basins on either side.⁽¹⁷⁾ The Nemaha Ridge associated faults have been active from Precambrian into the Permian Period and perhaps to the present if associated seismic activity is an indication of movement on the faults. The Central Kansas Arch and underlying Ellis Arch, as it is called in pre-Mississippian rocks, have been positive during Paleozoic time and possibly in Cretaceous time;⁽⁴⁾ so a symmetrical broad arch has developed with local folds having

axes parallel to the main arch axis trending northwest. Also faults parallel fold axes and lineaments visible on ERTS images.⁽¹⁵⁾ The presence of the lineaments suggests possible tectonic activity into very recent time, but the seismic record of epicenters is not yet developed sufficiently to make very positive statements.⁽¹³⁾ There are a couple of MMI VII to VIII intensity epicenters which might be associated with the arch, but the activity could be due to movement associated with the Nemaha Ridge since the epicenters are near the intersection of these two major tectonic features. There are several smaller intensity earthquake epicenters which might be due to movement associated with the Nemaha Ridge since the epicenters are near the intersection of these two major tectonic features. There are several smaller intensity earthquake epicenters which might be due to activity along the Central Kansas Arch. Most of this seismic activity does not have to be of tectonic origin. It may be due only to such smaller forces as erosion and isostatic adjustment.

The Kansas area is seismically very stable. It is in the mid-continent, and only a few earthquakes of less than MMI VII have been recorded. There do not seem to be any linear trends of epicenters through the area. The nearest earthquakes of MMI VII or larger have occurred far to the northeast along the north-south trending Nemaha Ridge and also far to the southeast along the west-northwest trending Wichita Uplift.

Salt Formations

The salt bearing formation which has a cumulative thickness of greater than 65 m and occurs at greater than 310 m below the surface of the ground is the Wellington Formation, as determined from Bachman et al.⁽⁷⁾ The purest unit in the Wellington Formation is the Hutchinson Salt Member. The Hutchinson Salt Member has very complex internal stratigraphy. Usually, anhydrite is missing in the upper part of the member, but both anhydrite and shale are interbedded with the lower part of the Hutchinson.⁽¹⁶⁾

The Hutchinson Salt Member is up to 130 m thick and is fairly pure, with the principal impurity being anhydrite. At Hutchinson, Kansas, the member consists of alternating clear and white coarse halite layers several inches thick, with thin laminae of silty shale, gypsum, and anhydrite. The Wellington Formation is divided by the Hutchinson Salt Member into three members.⁽¹⁷⁾ An unnamed shale member is above the Hutchinson. The shale is about 32 m thick and composed of gray and red shale. The lower member is about 47 m thick and consists of anhydrite interbedded with gray shale. The upper and lower members are probably impermeable and will likely keep circulating fluids from dissolving the halite of the Hutchinson Salt Member.

At the Kansas-Oklahoma border at least three formations, and maybe four, overlap with at least 185 m total halite thickness. The Cimarron Salt Members are mapped in Oklahoma extending this far north. At about the same stratigraphic position, the Stone Corral Formation halite beds are mapped in

Kansas⁽¹⁶⁾ and are indicated to extend south across the state line. Perhaps the Cimarron Salt Members and halite in the Stone Corral Formation are equivalent. Also, the Blaine Formation has been mapped in Kansas in the same general area along the state line as the Stone Corral.⁽¹⁶⁾

The Kansas area is underlain by the Prairie Plains' homoclinal surface structure with very gently westward dipping strata; however, subsurface strata do not share the westward dip. For example, the Precambrian surface dips very gently to the south.

The Meade Fault in Meade County is Pliocene or younger. Bayley and Muchlberger⁽¹⁸⁾ depict a fault near the Meade Fault, but with different strike which cuts basement rocks. There are two faults between Hutchinson and McPherson.⁽⁷⁾ Several other faults a few miles west of Wichita have been mapped as cutting basement rocks.⁽¹⁸⁾

There are several lineaments striking northwest near the Meade Fault.⁽⁷⁾ Many short lineaments and a few circular features are in the area, and to the east along the Central Kansas Uplift there are northwest striking lineaments which transect north-south lineaments along the Nemaha Uplift.⁽¹⁵⁾

Major streams run subparallel east-southeastward. The Cimarron River is the largest stream to run across the area. The Salt Fork of the Arkansas River, Bluff Creek, Rattlesnake Creek, and Medicine Creek originate within the area. The largest stream, the Arkansas River, runs along the south side.

Some very large sinks produced by dissolution of soluble rocks are in the western part of the area.⁽⁷⁾ Some of those sinks are so large they are visible on ERTS imagery.⁽¹⁵⁾ One large sink is manmade; i.e., it developed over a cavity made by the Meade Salt Well in Meade County, Kansas.⁽⁷⁾

Studies, including hydraulic testing, were made of three areas several miles north of the Kansas area in Lincoln County, Kansas, and several miles northwest in Wichita County, Kansas.⁽¹⁹⁾ In Lincoln County hydraulic tests on Permian rocks above and below the Hutchinson Salt Member show them to be impermeable. However, in Wichita County only 14 m above the Blaine-Cedar Hills salt there is sandstone with some permeability. Below the salt to the top of the Stone Corral Formation there is little permeability.⁽¹⁹⁾

Within 40 to 80 km north and east of the Kansas Area are eight salt mines, three of which are active.⁽²⁰⁾ Within the Kansas Area salt was produced from a brine well in Meade County.

There are a great number of oil fields in the eastern half of the area and only a few oil and gas fields in the western half. Producing horizons are both above and below the Permian salt beds.

The Colorado-Kansas Area is a forelandward extension of the Aughton Basin from the Paleozoic orogenic zone of the Marathon Mountains and Ouachita Mountains

to the south. The Colorado-Kansas Area is transected by the northeast trending Las Animas Arch along which Precambrian rocks may have been exposed above sea level in early and middle Pennsylvanian time. During Cretaceous and early Tertiary time this area was the foreland downwarp for the Denver Basin to the west, which developed during the Laramide Orogeny as the Front Range was folded, thrust faulted, and intruded by magma which created a great peripheral dike system.

There are no epicenters within the Colorado-Kansas area. The nearest large-intensity (MMI VII to VIII) earthquakes have occurred near Denver along the Front Range faults which strike principally north and therefore do not trend toward the Colorado-Kansas area. Recently some earthquakes near Denver have been shown to be associated with Front Range faulting by fluid injection into wells at the arsenal with the frequency of earthquakes. Docekal⁽¹³⁾ shows a weak concentration of much smaller intensity epicenters with an eastward trend from the Front Range generally toward the Colorado-Kansas area. However, these epicenters are probably associated with the Apishare Uplift, which extends along the southwest side of the Colorado-Kansas area.

The Permian salt deposits are not well known in the Colorado-Kansas area. The Blaine Formation contains halite beds up to 155 m thick. The map of Figure C.11 shows the maximum thickness of the Blaine Formation to be a little over 185 m thick. Therefore, the Formation must be almost all halite where the formation is thickest. Where the Formation is 130 m thick it probably contains at least 65 m of halite. The 130-m contour used to outline the Colorado-Kansas area in Colorado is estimated from the outline of the Permian Basin as published by Bachman et al.⁽⁷⁾

In Thomas County, Kansas, about 32 km north of the Colorado-Kansas area, the halite beds are 695 m below the surface of the ground. In the south part of the Colorado-Kansas Area, in Hamilton County, the halite beds are 310 m below the surface of the ground. It is, therefore, assumed that the Colorado-Kansas Area is all greater than 310 m below the ground surface.

The Colorado-Kansas area sediments parallel the extremely shallow dip of the Precambrian surface southeastward toward the deeper portion of the Hugoton Basin.⁽¹⁸⁾ There are no known faults in the area. The Las Animas Arch is a very broad flexure with a northeast trending axis. A few lineaments trend northwest just south of the Arkansas River.⁽⁷⁾ Smith,⁽¹⁵⁾ using ERTS imagery, found several lineaments in the area, especially east-west trending ones and a large circular feature at the prominent bend in the Arkansas River. All the lineaments are of unknown origin.

Drainage is eastward along subparallel streams. The major streams are the Arkansas River, White Woman Creek, and North Fork Ladder Creek. In the west corner of the area, Big Sandy Creek flows to the south and there are several large lakes.

FIGURE C.11. Thickness Map of Salt Beds in the Blaine Formation.
From Robert O. Kulstad, State Geological Survey of
Kansas.

There is very little published information available on groundwater in the Colorado-Kansas area. Sharon Springs occurs in the north corner of the area.

C.2.2 Bedded Salt - Northeastern U.S.

In the northeastern United States, salt occurs in great abundance in Paleozoic sedimentary beds. Devonian salt beds overlie Silurian salt beds, but the Devonian salt beds are not as extensive in area or as thick. Thus the area of study for the Silurian salt beds encompasses that for the Devonian salt beds. Figure C.12 delineates areas in which rock-salt beds of Silurian age, and separately those of Devonian age, have an aggregate thickness of at least 65 m and are not excessively deep beneath the land surface.

FIGURE C.12. Extent and Depth of Certain Rock Salt Deposits in Michigan, New York, and Ohio.

Michigan

Stratigraphy (Figure C.13.)

Both Silurian and Devonian rocks contain salt beds with a cumulative thickness of greater than 65 m and at depths between 310 and 930 m. The greatest thickness of salt unit drilled is in the north-northeast portion of the Michigan Basin.

Devonian salt is found within the Lucas Formation of the Detroit River group. The total thickness of salt exceeds 130 m and may include as many as eight different salt beds, the thickest exceeding 32 m in thickness. The salt lies at a depth of 400 m near the northeast edge of the deposit to over 1330 m near the south-central area of its extent.

According to Landes⁽²¹⁾ the evaporites in the Lucas Formation consist mainly of anhydrite and salt. Anhydrite is much more widely distributed, both vertically within the formation and laterally across the basin. The salt is confined to the upper half of the evaporite sequence and to the central part of the basin.

Most of the salt having an aggregate thickness exceeding 130 m is confined to the deepest part of the basin in Missaukee and Roscommon Counties where the Lucas is more than 310 m thick. There may be as many as eight distinct salt beds, and the thickest bed may exceed 32 m in thickness.

The Devonian beds are being solution mined since they are near the surface and are a proven commercial source of salt.

The Silurian evaporite sequence units A through G have been summarized by Pierce and Rich⁽²²⁾ as follows.

Unit A rests with apparent conformity on beds of Middle Silurian age. It ranges in thickness from about 32 m near the margin of the basin to more than 340 m in the center of the basin. This member consists of dolomite and shaly dolomite around the basin margin, but toward the center of the basin it contains increasing amounts of salt. One well near the center of the basin penetrated 280 m of unit A. In this well the salt sequence, which contains numerous paper-thin laminae of dolomite-anhydrite, is 270 m thick, with a medial sequence of dolomite 18 m thick.

Unit B is absent at the outer margin of the basin but has a relatively uniform thickness of about 75 to 85 m over most of the basin. This member, about 90 to nearly 100% salt, contains some laminae of dolomite. A 9 m sequence of dolomite is in the upper part of the member in the southeastern part of the basin.

Unit C ranges in thickness from 18.5 to 48 m and consists of greenish gray shale and shaly dolomite with some anhydrite. Buff dolomite is interbedded with the shale in the southern part of the Michigan Basin.

Unit D generally ranges from 7.5 to 20 m in thickness. It is nearly pure salt but locally may contain a thin bed of buff dolomite.

Unit E ranges in thickness from about 28 to 39 m except in the northern part of the basin, where it consists of about 9 m of red shale. Elsewhere it

is composed of shaly dolomite, anhydrite, and red or gray shale. In the southeastern part of the basin it grades into buff and gray dolomite with some anhydrite.

Unit F ranges in thickness from 0 to more than 375 m; it is thin or absent near the margins and thickest at the center of the basin. It consists of thick sequences of salt separated by thin sequences of shale, shaly dolomite, and dolomite. Some anhydrite generally is present, particularly in the shale beds. According to Landes⁽²¹⁾ the unit is 50 to nearly 100% salt; the percentage of salt in this unit increases outward from the center of the basin and is highest near the north edge of the basin.

Unit G is about 25 to 32 m thick over most of the Michigan Basin, but near the southern limit of the basin it thins to about 15 m in a short distance. The unit consists mainly of shaly dolomite and dolomite containing some anhydrite. Red and green shale appears in the sequence in the northern part of the basin. Unit G is overlain, apparently conformably, by Bass Island dolomite of latest Silurian age.

Structure

The only major structural feature in the southern peninsula of Michigan shown on a regional structural map (Figure C.14) is the nearly symmetrical Michigan Basin. It is the classical intracratonic basin or autogeosyncline with very shallow dipping strata. The Basement Rock Map of the United States⁽¹⁸⁾ shows the Grenville Front passing north-south along the east side of the Basin within the Precambrian basement rocks. This front has not been accurately located, but it probably does not pass under the Paleozoic rocks of the Michigan Basin. No known seismicity has been definitely associated with this proposed feature below the basin. Passing northwest near the center of the Basin is a gravity high which has been interpreted as due to basic rocks intruded into the crust along an incipient rift zone connecting at a triple junction under Lake Superior with the mid-continent gravity high rift zone. There is no seismic trend across the Michigan Basin to support this hypothesis, but the structural and depositional grain trend northwestward appears to confirm it. Perhaps these structural and sedimentary features owe their trends to activity along northwest trending faults. Major tectonic activity was apparently during or prior to deposition of the northwest sedimentary features.

There are two anticlines in the Basin, the Howell Anticline and the West Branch Anticline. Also there is a dome near the center of the basin and a slight uplift suggested between the dome and West Branch Anticline. There are no faults shown on the Tectonic Map of the United States⁽³⁾ in the southern peninsula of Michigan. But it would not be surprising to find the anticlines faulted at depth, since many folds are formed in the mid-continent by vertical uplift along faults.

FIGURE C.14. Sketch Map Showing the Regional Structural Features Associated with the Silurian Salt Deposits of the Northeastern States (modified from map compiled by R. L. Griggs).

Landes⁽²¹⁾ suggested that faulting had been caused by collapse above cavities produced by solution of the Silurian salt beds around the periphery of the Michigan Basin. This faulting extends toward the center of the basin a few miles across the following counties: Presque Isle, Cheboygan, Charlevoix, Antrim, Leelanau, and Benzie.

Hydrology

Surface Water. The Great Lakes surround three-fourths of the Michigan Basin and serve as areas of recharge for many of the Paleozoic formations including much of the Salina Formation or Group. Rate of recharge is extremely slow because recharge is probably governed by the rate at which people extract fluids from wells. Many small glacial lakes serve to recharge the drift which covers most of the southern peninsula of Michigan. The drift is no more than a few hundred feet thick.

Rivers are short because they drain radially toward the Great Lakes. Most streams and water wells are mainly charged by meteoric water percolating through glacial drift. Erosion rates, presently very slow, will not likely increase much due to increased stream length because they are restricted geographically by the Great Lakes.

Groundwater. Most presently utilized groundwater is from glacial drift which forms a surficial layer up to a few hundred feet thick. Consolidated sedimentary rock formations of Paleozoic age are in a basin shape, and some rocks are permeable; thus, if water is pumped from a bed in the basin, water from the surface will move down to take its place. Down dip water is salty due to entrapment of sea water during deposition of the sediments or solution of salt from sedimentary beds in the basin. Only relatively small amounts of fluid have been pumped from aquifers; therefore water circulation has been extremely slow. There are no deep wells in aquifers for fresh water. Brine aquifers are as follows: Saginaw Formation, Para Sandstone, Marshall Sandstone, Dundee Limestone, Detroit River Dolomite, Sylvania Sandstone, Bass Island Dolomite, Traverse Formation, and Berea Sandstone. Below the salt beds there is little specific data, but the St. Peter Sandstone is a very good aquifer in other areas of the country. Also the stratigraphic units which produce oil or gas are also likely aquifers.

All information available indicates that known aquifers are continuous from the ground surface to depths penetrated because pressure increases continually with increasing depth.⁽²³⁾ "Considerable withdrawal of natural brines for production of salt, sustained for nearly a century, is incontrovertible proof that transmissibility is at least moderately large over extensive areas."⁽²³⁾

Mineral Resources

Salt is mined below Detroit and brined from several well locations. Salt is taken from natural brines in Gratiot County, and from artificial brines in Manistee, Midland, Muskegon, St. Clair, Wayne, and Lapeer Counties. Most of the salt is used as industrial salt. Bromine is extracted from natural brines; magnesium, calcium, and potassium compounds and iodine are extracted from brine.⁽²³⁾

New York-Pennsylvania

Stratigraphy

The Salina Group in the New York-Pennsylvania area is divided upward into the Pittsford Shale; the Vernon Shale containing units A, B, and C; the Syracuse Formation containing units D, E, F₁, F₂, F₃, F₄, and F₅; the Camillus Formation, unit G; and the Bertie Limestone, unit H. These formations have been described by Pierce and Rich⁽²²⁾ as follows.

The Pittsford shale is of local extent in western New York where it rests conformably on the Lockport dolomite of Middle Silurian age. This member ranges in thickness from 0 to about 7 m and consists of dark gray to black shale.

The Vernon shale is a purplish red and grayish green shale which, near its eastern limit, contains some thin beds of gypsum. This member is continuous along strike in New York from the easternmost exposures of the Salina group westward to near Niagara Falls and overlies the Pittsford shale and the Lockport dolomite. It attains a maximum thickness of about 185 m near Syracuse, and thins to the east and west. Southward this member may grade into the Bloomsburg red beds of eastern Pennsylvania, Maryland, and West Virginia.

The Syracuse salt rests conformably on and possibly inter-fingers with the Vernon shale. It ranges in thickness from 0 near the eastern limit of the Salina group to about 775 m in a basin southwest of Syracuse. This member consists of beds of salt and interbedded gray shale and dolomite. As many as six beds of salt are present in places, and the aggregate thickness of salt beds reaches a maximum of about 275 m in the basin southwest of Syracuse.

The Camillus shale ranges from about 14 to 185 m in thickness. It rests conformably on the salt sequence in New York and consists of gray dolomitic and calcareous shale, shaly dolomite, and dolomitic limestone, gypsum, and anhydrite.

Above the Camillus shale is the Bertie limestone, the uppermost member of the Salina group in New York. It consists of fine grained, argillaceous, grayish buff, dolomitic limestone about 15 m thick.

Structure

Structure of the Allegheny synclinorium in the area of interest is a dominant very gentle dip of strata to the south-southeast. There are several warps with north-northeast trending axes and occasional faults which chiefly strike northeast. Fault displacement generally decreases upward and probably occurred during late Paleozoic mountain building.⁽²⁴⁾

Folds increase in number, size, and steepness of fold limbs toward the southeast. Faults increase in number and size toward the southeast, indicating a continuous increase and mergence into Appalachian mountain structures.

Salt beds have been faulted in the Ludlowville mine and Restof mine and in the axial regions of an anticline in the same mining area. There was very little disturbance of the salt beds by groundwater.⁽²⁵⁾ Salt units are greatly

thickened.⁽²⁵⁾ A few small peridotite dikes and a sill have been found intruding salt beds in the salt mines. There was very little disturbance of the salt, and the dikes are dated probably post-Appalachian.⁽²⁶⁾ Jacoby⁽²⁵⁾ implies that tectonic folding and igneous intrusion were coeval.

Hydrology

Surface Water. The area of interest is centered along a drainage divide with drainage northward into the Finger Lakes region of New York. Two of the Finger Lakes, Seneca Lake and Cayuga Lake, are partially in the north-central part of the area. Streams drain westward into the Allegheny River, and the Susquehanna River runs through the southeastern corner.

Groundwater. Most of the domestic water wells are in glacial drift. Nearly all the formations in the Paleozoic section contain sandstone, carbonate rock, or other rocks which might be permeable and contain water and therefore be considered an aquifer. Known aquifers, according to Piper⁽²³⁾ are as follows:

The Lockport Dolomite is the most productive and yields up to 500 gpm and averages 110 gpm. The water is very hard, with notably large contents of SO_4 and Cl. The Rochester Shale Member yields a maximum of 125 gpm and averages about 50 gpm. Its water is very soft and generally the least variable in total dissolved solids. The Salina Formation yields no more than 30 gpm; the water is generally very hard and contains much SO_4 . The Clinton Formation, Albion Sandstone, and Queenston Shale yield only about 10 gpm average but may yield as much as 30 gpm; the water is hard and may contain a large amount of NO_3 .

There are a few deep wells such as gas, oil, water wells and brine wells from which fluids are pumped and therefore cause movement of water within the strata; but there is no information on the directions and rates of movement other than the general statement that circulation of water will, in general, be down dip from the north. The quantities of fluids involved are extremely small so that the rate of movement is very slow, so slow that fresh water is not drawn down from meteoric water into any known water wells more than a few hundred feet. Below this level, water wells more than a few hundred feet have salty water.⁽²³⁾ For water to circulate near a storage site there would have to be pumping wells between the site and down dip toward the trough of the structural basin. The axial trace of the trough of the Allegheny synclinalorium trends east-northeast through Scranton, Pennsylvania.

Mineral Resources

Salt is mined and brined. Brine wells are in the area of greatest combined thickness of salt near Watkins Glen, and gas wells are also in the same area.

OhioStratigraphy

The areal extent and thickness of salt is less in Ohio than in New York or Michigan. The aggregate thickness of salt beds is about 92 m, and the maximum thickness of one bed is only about 15 m. As many as 19 beds have been recorded. Ohio salt beds can be correlated with those of the Michigan Basin, New York and Pennsylvania.⁽²⁶⁾

Salt Formation.

In Ohio the Salina consists mainly of carbonate rocks and salt. Some beds are shale, and a few thin beds are dolomitic shale. Stratigraphic subdivisions of the formation can be correlated with those in the Michigan Basin and in New York and Pennsylvania.⁽²⁷⁾

A-unit - There is no A-unit salt in Ohio. In Ohio it is a series of dolomites, anhydrite, and shales.

B-unit - (no. 4 salt) In Michigan and B-unit is practically pure salt, with the upper half containing numerous shale-carbonate stringers. In Ohio it is a sequence of interbedded shale and salt. It exceeds 65 m in thickness in Lake County.

C-unit - This unit is a series of shales and argillaceous dolomites, commonly grayish. The C-unit may represent a tongue of the Vernon shale of New York.

D-unit - (no. 3 salt) This unit in Ohio is an evaporite sequence between two thicker shale-carbonate sequences and a thin shale parting near its middle. It has a maximum thickness of 14.5 m.

E-unit - This unit contains a series of dolomite, anhydrite, salt, shale, and lutite. It is commonly brown and varies in thickness from 23 to 65 m.

There are four evaporite zones recognized in the F-unit of Ohio. It ranges in thickness from 18.5 m in areas of no salt deposition to more than 92 m in Portage, Mahoning, and Columbiana Counties.

Structure

The Precambrian rocks deformed by the Grenville Orogeny form a basement upon which Paleozoic sediments were deposited. The structural grain of the basement is probably north-northeast, but not enough is known about them to be sure. Some of the rocks are known to be felsic plutonic igneous rocks or felsic gneiss.

Paleozoic sediments dip very gently toward the southeast toward the Appalachian Geosyncline and away from Findlay Arch. No folds or faults are shown in the area on the Tectonic Map of the United States⁽³⁾ although some extremely gentle folds have probably been found due to the extensive drilling for oil and gas, and faults have been reported in a salt mine at Cleveland.⁽²⁵⁾

Hydrology

Short streams drain northward into Lake Erie about 32 km north of the area, where the salt beds are about 65 m thick. The headwaters of the Ohio River drain, in general, southeastward.

All aquifers dip gently southeastward into the Appalachian Geosyncline, and for formations above the salt beds there are no fluid outlets at elevations lower than the terrain in Ohio and northward to the area of outcrop of the Salina Group.

Water is trapped in these formations along with much of the salt from the ocean water in the formations developed since deposition of the sediments. Thousands of oil and gas wells have been drilled through the salt-bearing strata, and fluids have been pumped from aquifers and caused flow mainly down dip from the northwest. Rates of flow within the aquifers is unknown, but for all aquifers affected by wells the rate is very slow, as evidenced by fresh water extending only a few hundred feet below the ground surface in the recharge areas. Recharge is mainly from the blanket of drift covering almost entirely the eroded edges of the aquifers.

The majority of formations are aquifers, those which contain permeable stratigraphic zones. Gas- and oil-producing stratigraphic units are likely aquifers. Known aquifers are the Sharpsville Sandstone Member and locally the top part of the Berea Sandstone. The Pottsville Formation is very productive, but the Allegheny Formation yields only enough for individual homes. Rate and quantity of groundwater withdrawal are small. The water ranges from soft to hard, and from a small amount to too much total dissolved solids for ordinary use. The Berea Sandstone is salty at shallow depths.

Mineral Resources

By 1947, 3555 wells had been drilled for oil and gas to the Clinton Sand below the salt series.

Salt brine wells exist at Rittman in the northeast corner of Wayne Co., at Newburg, near Cleveland, and at Wadsworth in the southeastern corner of Medina County.⁽²⁸⁾

Six or seven wells have been drilled to recover salt at Kenmore.⁽²⁸⁾ Also at Barberton several wells have been drilled for salt used in the manufacture of soda-ash.⁽²⁸⁾

An underground mine exists under Cleveland and the Lake at Whisky Island near the mouth of Cuyahoga River. Also there is a mine under development near Painesville.⁽²⁰⁾

The location of salt mines can be summarized as being in three areas: at Cleveland, within 16 km of Barberton, and 32 km northeast of Cleveland.

C.2.3. Bedded Salt - Supai Salt Basin and Luke Salt Body

The Supai Salt Basin, or the Holbrook Basin of some reports (C of Figure C.15), is a marine evaporite basin of Permian age in the southernmost part of the Colorado Plateaus Province. The Luke Salt Body near the northwest edge of Phoenix Basin (E, ii of Figure C.15) is within the Basin and Range province. The Supai Salt Basin embraces an area of about 5900 km² (Figure C.16) in Navajo and Apache Counties, Arizona. This salt basin lies north of the Mogollon Rim, southeast of the Black Mesa Basin, and southwest of the Defiance Uplift. Paleozoic strata exposed along the Mogollon Rim dip to the north and northeast at low angles, forming a near-dip slope⁽²⁹⁾ commonly referred to as the Mogollon slope. The area occupied by the salt basin is underlain by thick, generally non-water-yielding, shaly rocks of the Chinle and Moenkopi Formations of Triassic age. The Kaibab Limestone and Coconino Sandstone of Permian age overlie the salt-bearing rocks but crop out only in the southwestern part of Navajo County.

Stratigraphy

The stratigraphic nomenclature for the region of the Supai Salt Basin is given in Table C.1. Figure C.17 indicates the thickness and extent of the major stratigraphic units. Rocks exposed in the area range from Permian to Quaternary in age. Rocks of Jurassic and Early Cretaceous age are not present in the region. The thickness and extent of the Supai Formation and rocks ranging from Pennsylvanian to Precambrian age are known only from subsurface information based on exploratory drilling. The depth to the Precambrian basement generally ranges from 900 to 1255 m below the surface.

An investigation of the Supai Formation by Winters⁽³⁰⁾ in the Fort Apache Indian Reservation shows that the contact between the Supai and underlying Naco Formation is apparently concordant, but that the thick sequence of limestone, siltstone, and mudstone of the Naco may possibly be disconformably overlain by the calcareous siltstone and sandstone of the Supai. In the absence of paleontological data, Winters⁽³⁰⁾ has arbitrarily placed the Pennsylvanian-Permian boundary in the Fort Apache Reservation at the Naco-Supai contact. Gerrard⁽³¹⁾ considers the basal part of the Supai to be Late Pennsylvanian but gives no evidence of his own to support this.

According to Gerrard⁽³²⁾ the contact between the Supai and overlying Coconino Sandstone, in both the surface and subsurface, is a transition zone of varied thickness, with red Supai-like siltstone and sandstone alternating with white Coconino-like sandstone. The Supai-Coconino contact in the subsurface, in the area of the Supai salt basin, is arbitrarily placed at the uppermost level of reddish-orange Supai-type clastic rocks, thereby including the transition zone in the upper Supai.

FIGURE C.15. Colorado Plateaus Province. Modified from Eaton and others, 1972, figure 15.

FIGURE C.16. Extent of Supai Salt Basin. Basin outline from Peirce and Gerrard, 1966, figure 4.

TABLE C.1. Stratigraphic Nomenclature for Area of Supai Salt Basin. Modified from Akers, 1964, table 2; Winters, 1963, figure 4; Elias, 1958, p. 65.

System	Series	Group or formation	Member
Quaternary ^(a)	Holocene and Pleistocene		
Tertiary ^(a)	Pliocene	Unconformity	
		Bidahochi	
Cretaceous ^(a)	Upper	Unconformity	
		Chinle	Petrified Forest
Triassic ^(a)	Upper	Unconformity	Mesa Redondo
			Shinarump
	Middle(?) and Lower	Moenkopi	Holbrook
		Unconformity	Moqui
Permian	Leonardian	Kaibab ^(a)	Wupatki
		Coconino	
	?	Supai	Upper
	Wolfcampian		Ft. Apache
Pennsylvanian	?		Lower
Mississippian	Meramecian-Kinderhookian	Naco	
		Unconformity	
Devonian	Upper	Redwall	
		Unconformity	
Cambrian	Middle	Martin	
		Unconformity	
Precambrian		Tonto	
		Unconformity	

a. Not everywhere present

FIGURE C.17. Generalized Geologic Cross Section from Mogollon Rim to Black Mesa Basin. Modified from Brown and Lauth, 1958, exhibit II.

The Supai Formation is widely distributed in northern and central Arizona, where its distinctive red color is a characteristic feature. The Supai Formation consists of interbedded sandstone, siltstone, mudstone, and evaporite beds containing anhydrite, gypsum, and halite. (The halite, in general, makes up 5 to 15% of the formation.) The formation is about 500 m thick near the Arizona-New Mexico State line, but thickens to more than 800 m southwest of Holbrook. The Supai Salt Basin, as reflected in the stratigraphic sequence of the Supai Formation, was a depositional basin during Early Permian time and represents the area of maximum thickening of the Permian sequence in Arizona. A relatively thick succession of evaporites intermixed with clastics was deposited in the basin from the shallow marine waters that invaded east-central Arizona from the south and southeast.

The Supai Formation in the area of the Supai Salt Basin has been divided by Pierce and Gerrard⁽³¹⁾ into lower and upper members separated by the Fort Apache Member. According to Gerrard,⁽³²⁾ the Fort Apache Member is a marine unit of variable lithologic composition, limited areally to east-central Arizona. In the Supai Salt Basin the Fort Apache Member consists chiefly of dolomite with interstratified and intergrown halite. The Fort Apache is as much as 140 m thick southeast of the Supai Salt Basin, but thins to the north and west. It is about 22 m in the vicinities of St. Johns, Snowflake, and Zeniff, about 14 m near Holbrook, and only about 1.5 m thick at Chambers, Arizona.

Although evaporites are present in the Supai Salt Basin below the Fort Apache Member, more than 80% are in the upper member of the Supai.⁽³²⁾ The upper member is about 155 m thick in the vicinity of the Mogollon Rim (Figure C.18) but thickens markedly in the Supai Salt Basin, attaining a thickness of nearly 400 m in an exploratory drill hole about 12 km southwest of Holbrook (No. 3 of Figure C.19). The upper member thins gradually to about 215 m near the Arizona-New Mexico State line. The deepest part of the basin during the deposition of the upper member of the Supai was most likely the area in Navajo County between Snowflake and Holbrook.

The general distribution of lithologic types constituting the upper member of the Supai in the area is shown in stratigraphic correlation diagrams (Figures C.19-20) that are based primarily on information from exploratory drill holes. On the basis of laterally persistent carbonate and sulfate units, Pierce and Gerrard⁽³¹⁾ have grouped the lithologies of the upper member of the Supai into four zones, numbers, I-IV, from bottom to top. The principal halite deposits of the upper member occur as discontinuous units in Zone I-III and are associated with anhydrite, dolomite, and clastic red beds. The total thickness of the evaporite-bearing stratigraphic sequence ranges from about 105 to 320 m. The top of that sequence, marked by anhydrite and gypsum, lies about 200 to 965 m below the surface in the Supai Salt Basin, the greatest depths being in the northeastern part of the basin.

FIGURE C.18. Isopach Map of Upper Member of Supai Formation. (Supai salt basin outlined by dotted line; Black Mesa basin and Defiance uplift outlined by heavy dashed lines.) Basin outline from Peirce and Gerrard, 1966, figure 4.

FIGURE C.19. Stratigraphic Correlation Diagram of Upper Member of Supai Formation along Section A-B; datum is base of Coconino Sandstone. Modified from Peirce and Gerrard, 1966, figure 5.

FIGURE C.20. Stratigraphic Correlation Diagram of Upper Member of Supai Formation along Section C-D; datum is base of Coconino Sandstone. Modified from Peirce and Gerrard, 1969, figure 6.

The halite units shown in the stratigraphic correlation diagrams (Figures C.19-20), however, are not pure halite but are composed of halite interbedded with anhydrite, siltstone, and claystone. Individual halite units, as interpreted from stratigraphic and gamma-ray logs, are commonly 3 to 6 m thick, and of unknown purity.

The thickest halite units shown in the correlation diagrams are in drill holes four and five (Figure C.19). About 315 m of rock overlies the halite in drill hole four and about 480 m in drill hole 5. The halite lenses of Zones I and II extend for the 22 km between drill holes five and six. A halite unit more than 77 m thick is indicated in Zone IIb in drill hole 10 (Figure C.20); the top of the unit is about 230 m below the surface.

Drill holes 2, 9, and 10 of section C-D (Figure C.20) are located near the axis of the Holbrook anticline, on the northern margin of the area of solution-collapse phenomena. Gerrard⁽³²⁾ suggests that the halite units in drill hole 9 are markedly thinner than in drill holes 2 and 10 because of sub-surface dissolution.

Pierce⁽²⁹⁾ reports that a northeast trending potash zone about 780 km² in area has been delineated by drill holes in the northern part of the Supai Salt Basin. He reports this potash zone to be within 46 m of the top of the evaporite section which, in most places, is marked by an anhydrite bed. Depths below the surface, to the top of the potash zone range from 215 to 620 m, shallower depths being to the southwest and greater depths to the northeast.

The Coconino Sandstone underlies the entire area and crops out between Holbrook and Snowflake. The sandstone is uniform in composition and consists of quartz grains that are well sorted and subangular to rounded. Outcrops characteristically display eolian-type, large-scale, wedge-trough cross-stratification. The thickness of the Coconino is about 65 to 77 m in eastern Apache County, 140 m in the vicinity of Holbrook, and about 600 feet near Snowflake.⁽³³⁾

Zone IV of the upper member of the Supai Formation ranges in thickness from slightly less than 32 m to at least 140 m (Figures C.19-20). The lithology of Zone IV is chiefly sandstone and siltstone.

Structure

The Supai Salt Basin lies between two structurally and topographically high areas, the Mogollon Rim to the southwest and the Defiance Uplift to the northeast. The Black Mesa Basin, a negative tectonic element, lies north of the Supai Salt Basin. Regional topography in general reflects regional structure (Figure C.21). According to Pierce and Gerrard,⁽³¹⁾ the distribution of the upper member of the Supai (Figure C.18) indicates that major tectonic features of the region existed during late Supai time. They note that the long axis of the Supai Salt Basin nearly parallels the existing Mogollon Rim and

FIGURE C.21. Stratigraphic Correlation Diagrams of Upper Member of Supai Formation along Sections E-F and G-B; datum is base of Coconino Sandstone. Modified from Peirce and Gerrard, 1969, figure 7.

that the upper member thins both to the southwest and to the northeast, suggesting the existence in Early Permian time of structural equivalents to the Mogollon slope and the Defiance Uplift.

The Mogollon slope, which extends northward from the Mogollon Rim to the southern edge of the Black Mesa Basin and Defiance Uplift, has a broad gentle dip to the northeast, modified in places by folds generally trending northwest (Figures C.21 and C.22). The folds are small compared to regional features such as the Defiance Uplift. The anticlines are fairly well defined, but the synclines form rather indefinite structural depressions with poorly developed axes. The northwest-trending anticlines commonly plunge to the northwest and are asymmetrical, with the southwest limb steeper than the northeast limb. The regional northeastward dip is only about 1° - 5° , but local reversals in regional dip range from 4° to 6° . Their plunge to the northwest is about 3° - 6° . Closures range from 23 to 55 m. The folds are considered to be Late Cretaceous to early Tertiary in age, probably having developed during the Laramide orogeny that produced the Black Mesa basin and the northeast-tilted Mogollon slope. Lower Tertiary and older rocks have been folded, but Miocene and younger rocks have not.⁽³³⁾

The most prominent fold in the area of the Supai Salt Basin is the Holbrook anticline (Figure C.22), which has its surface expression in a low escarpment northwest of Snowflake. Its axis can be mapped at the surface for more than 96 km.⁽³⁴⁾ The regional dip of strata marking its northeast flank averages about 2° . The south flank is formed by a rather sharp reversal of the regional dip; some dips are 50° or more, but representative dips average less than 15° .

The most striking single topographic feature of the Holbrook anticline and its immediate vicinity is the abundance of sinks and related solution-collapse structures, resulting in typical karst topography. The south flank of the anticline is so pockmarked with potholes and sinks that the structure of the exposed rocks is nearly chaotic.⁽³⁴⁾ The synclinal area just south of the Holbrook anticline is the result of subsurface solution and collapse. The walls of some open sinks expose alluvial-filled older sinks. The appearance of numerous deep sinks on aerial photographs taken in 1953 and absent from photos taken 17 years earlier⁽³⁴⁾ shows that solution at depth has been a continuing influence on the structural development of this area. According to Bahr,⁽³⁴⁾ the sinks and related features have resulted from the solution of the Permian evaporites which, with interbedded shales and siltstones, aggregate some 435 m in thickness. The evaporitic section, which includes both anhydrite and halite, probably attained a depositional thickness of 215 m but has been reduced considerably by solution thinning, as suggested by a comparison of drill hole nine with drill holes two and ten (Figure C.20). The solution-collapse process has probably been facilitated by percolating groundwater from the overlying Coconino Sandstone, the major aquifer in the region. Extensive

FIGURE C.22. Relationship of Topography to Regional Structure in Area of Supai Salt Basin (Supai salt basin outlined by dotted line). Contours modified from Peirce and Gerrard, 1966, figure 2.

jointing, as seen in outcrops of the Coconino, and partings along cross-bedding have increased the permeability of the Coconino.

Bahr⁽³⁴⁾ suggests that the Holbrook anticline (Figure C.22) may have originated from extensive subsurface evaporite solution along the Mogollon slope and that no subsurface anticlinal structure exists, or, if it does exist, the subsurface crest may lie some distance south of the surface anticlinal crest. Because dip directions on the southwest flank of the anticline are everywhere directly into sinks, Bahr points out that the reversal in dip of the south flank may result from solution-collapse rather than from tectonic forces. Although the solution-collapse phenomena are restricted to this area and are not known to occur to the south, he suggests that the south flank of the Holbrook anticline may mark the present position of a regionally retreating collapse front that originally developed near the Mogollon Rim when Permian evaporites were exposed there.

Faults are scarce in the area of the Supai Salt Basin. Akers⁽³³⁾ reports a few small faults of Late Cenozoic age in Apache County. They trend northwest and are downthrown 15 to 46 m to the southwest. Cooley⁽³⁵⁾ describes a regional pattern of closely spaced faults and joints trending northwest, north, and northeast. Figure C.23 shows the fault pattern.

Hydrology

Surface Water. The major drainage system in the area of the Supai Salt Basin is that of the Little Colorado River, which drains an area of about 56,240 km² of the Colorado Plateau.⁽³⁶⁾ In the upper reaches, the available streamflow has been developed extensively for irrigation purposes. Most of the significant runoff is from the area of the Mogollon slope adjoining the Mogollon Rim and the White Mountains to the southeast. The average annual runoff in the area ranges from less than 1.5 cm to more than 13 cm, the greatest runoff being from higher elevations.

Most of the streamflow is from tributary streams that drain the region on the south side of the Little Colorado River. Unit runoff of these streams per square kilometer varies from about 0.1 to 5 ha-m. The average annual runoff for most of the area north of the Little Colorado River is estimated to be less than 1.5. Unit runoff of the Puerco River is only 0.8 ha-m per square kilometer.

Luke Salt Body

The log shown in Figure C.24 is an interpretation of gamma ray-neutron logs of an exploratory drill hole⁽³⁷⁾ 1.6 km east of Luke Air Force Base. The lithology between the depths of 245 and 1740 m is interpreted as being chiefly halite with shale interbeds overlain by about 18.5 m of anhydrite caprock. In descending order the lithologic sequence includes: 1) about 155 m of halite with shale interbeds 0.6 to 2.4 m thick; 2) nearly 190 m of halite with argillaceous material; 3) about 155 m of alternate beds of halite and halite

FIGURE C.23. Distribution of Fold Axes in Area of Supai Salt Basin; Supai salt basin outlined by dotted line. Geology modified from Doeringsfeld, Amuedo, and Ivey, 1958; Bahr, 1962.

FIGURE C.24. Generalized Lithologic Log of Salt Mass Penetrated by Well 1 of Eaton, Peterson, and Schumann (1972, figure 10); interpretation of gamma ray-neutron logs by R. J. Hite (written communication, 1972); intervals 1-7 are described in text.

containing argillaceous material; 4) 420 m of halite with shale interbeds less than 0.3 to 1.8 m thick; 5) 66 m of halite; 6) 32 m of alternate beds of halite and halite with argillaceous material; and 7) nearly 110 m of halite with sparse shale interbeds less than 0.3 to 1.8 m thick.

C.2.4. Dome Salt - Gulf Coast

The Gulf Coast region of the United States is referred to here as the northern Gulf Coast salt-dome province to distinguish it from salt-dome provinces or basins that occur in other parts of the Gulf of Mexico region such as in Mexico and Cuba.⁽³⁸⁾

There are about 263 known and suspected onshore salt-domes in the northern Gulf Coast salt-dome province (Figure C.25). The province can be conveniently divided into interior and coastal subprovinces that can be further subdivided on the basis of the states or the geologic basins in which the domes occur. To conveniently describe general geologic and hydrologic factors that pertain to the domes, the subdivision on geologic basins is as follows:

Interior subprovince:

- Northeast Texas salt-dome basin
- North Louisiana salt-dome basin
- Mississippi salt-dome basin

Coastal subprovince:

- East Texas-south Louisiana salt-dome basin
- South Texas salt-dome basin

The outlines of these salt-dome basins are shown in Figure C.25. The east Texas-south Louisiana salt-dome includes a large number of offshore domes not considered in this report.

The Louann Salt, Jurassic in age, is generally considered to mark the horizon of bedded salt from which the domes are derived in both subprovinces-- interior and coastal. Whether or not salt was deposited in the domeless area between the subprovinces is not known, but if it was, its thickness was probably much less than in the depocenters represented by the salt-dome subprovinces. The domes of the interior subprovince were probably emplaced from Late Cretaceous to Oligocene time and the coastal domes from Miocene to Pleistocene time.⁽³⁹⁾

This difference in age has been schematically depicted by Hanna⁽⁴⁰⁾ in a group of time-sequential cross sections through the Gulf Coast geosyncline (Figure C.26). The cross sections show a four-stage depositional model for the development of the geosyncline and growth of the domes. Domes are supposedly formed by differences in static load of sediments at and ahead of a seaward migrating axis of principal deposition. It is emphasized that the model refers to the main episodes of diapirism and is not intended to preclude minor tectonic adjustments resulting from such processes as differential compaction or salt dissolution. Diapirism in the interior subprovince, according to this model, has been virtually inactive since Miocene time and is now most active along the offshore axis of the geosyncline under the Gulf of Mexico.

Gera⁽⁴¹⁾ has summarized reported cases of present salt movement related to diapirism. It is interesting to note that the three cases involving onshore

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FIGURE C.25. Tectonic Map of Northern Gulf Salt-Dome Province.

FIGURE C.26. Schematic Time-sequential Cross Sections Showing Development of Gulf Coast Geosyncline and Growth of Salt Domes. Modified from Hanna, 1959.

Gulf Coast domes are from the coastal subprovince (Belle Isle, Jefferson Island, and Hoskins Mound). Clark⁽⁴²⁾ states that many of the interior domes are known to be growing at present, but there is little evidence to support his statement. What evidence there is for present dome growth seems restricted to the coastal subprovince.

DeWitt Van Siclen (written communication, 1972) has studied Holocene terraces along the Brazos River and found that along the portion of the river course that crosses the coastal salt-dome subprovince (south of the Clay Creek and Millican domes), the terraces have been "warped both regionally and over salt domes; and displaced by zones of an echelon of normal faults which trend about parallel to regional strike, are mostly down-to-coast, and near the coast seem generally to connect salt domes." He notes that of the eight salt domes overlain by Holocene flood plain and deltaic deposits from the Raccoon Bend salt dome to the coast, five of the domes warp those deposits.

D. H. Kupfer and C. O. Durham (verbal communication, 1972) make a strong contrast between the inactive interior domes, many of which are blanketed by

undeformed Miocene and post-Miocene sediments, and the domes within the active, unstable coastal subprovince which Kupfer likens to a "shaking bowl of pudding" in which differential displacements related to several processes, including diapirism, are active. The apparent absence of surface indications of diapirism in the interior subprovince is, of course, no proof that diapirism is not occurring there. As Goldman⁽⁴³⁾ stated the problem, "Have solution and upward movement balanced each other or has there been practically none of either (in the recent geologic past)?" As far as diapirism is concerned, however, the interior domes are more stable.

East Texas-south Louisiana Salt-dome Basin

There are 141 reported onshore domes in the east Texas-south Louisiana salt-dome basin. Six of the shallower are in the western part of the basin west and southwest of the Trinity River-Galveston Bay area. The following summary pertains only to the Texas part of the east Texas-south Louisiana salt-dome basin--that part where the six shallow domes occur, and subsequent use of the term "basin" refers only to that western part of the entire basin.

Depth to salt in the six shallow domes ranges from 140 m for Hawkinsville to 375 m for Davis Hill. Thickness of caprock ranges from 87 m for Gulf and as much as 275 m for Hockley. Of the six domes, most is known about Hockley, which has been the site of salt mining and petroleum production.

Physiography

The land surface is nearly featureless, rising gently from sea level to about 65 m some 128 km inland. Near the coast much of this surface is depositional on Pleistocene and Holocene formations, but it becomes progressively more eroded inland. The chief inland relief is caused by small streams that have carved youthful valleys. Most of these streams flow to the Gulf, but some are tributary to the Brazos and Trinity Rivers, which flow across the basin in broad shallow valleys and deposit their sediment load in actively growing deltas. Extensive marshy areas occur in the bottom land of some of the major drainages. The coast is characterized by a system of barrier islands that protect numerous indentations and bays, some of which have areas of several hundred square kilometers. Some of the bays are only a few meters deep, and extensive marshy areas and tidal flats exist adjacent to and between them. The Colorado, Brazos, and Trinity Rivers are actively infilling the areas behind the barrier bars with deltaic sediments, and thus are producing additional marshy areas. Three of the unrejected domes (Gulf, Hawkinsville, and Hoskins Mound) are located in this coastal area.

Several of the salt domes are expressed at the surface by hills or mounds that rise several tens to more than 32 m above the flat coastal plain. Examples are Davis Hill, Hoskins Mound, High Island, Barbers Hill, and Damon Mound.

Geology

The east Texas-south Louisiana salt-dome basin, including its offshore portion in the Gulf of Mexico, embraces the active northern part of the Gulf Coast geosyncline where south- to southeast-flowing rivers are depositing huge quantities of sediment in deltas along the coast landward of the geosynclinal axis. In its simplest sense this sedimentary process has been active since Cretaceous time, and the axis of main sedimentation has shifted southward some 480 km to its present position since then.

Exposed strata are mostly continental deposits of fine to coarse sand and conglomerate of Pliocene age (Goliad Sand), and coastal and estuarine deposits of sand, gravel, silt, and clay of Pliocene(?) age (Willis Sand) and Pleistocene age (Bentley Formation, Montgomery Formation, Beaumont Clay, and Deweyville Formation in ascending order). The inland protrusion northwest of Houston includes exposures of rocks as old as the Jackson Group of Eocene age. Near the coast the cumulative thickness of Pliocene and Pliocene(?) strata is about 275 m and of Pleistocene strata about 930 m. All strata wedge out landward. The upper parts of the three domes located along the coast (Gulf, Hawkinsville, Huskins Mound) are in the unconsolidated clay, silt, sand, and gravel of Pleistocene age. However, the two inland domes, Hockley and Davis Hill, have their upper parts in clay and sand of the Miocene Fleming Formation and Oligocene(?) Frio Clay, respectively.

Normal faults that parallel the coast are related to the geosynclinal subsidence. They displace only the subsurface strata and generally have no surface expression. Faults related to salt domes are known but generally have little or no surface expression. It is possible, though, that the surface fractures resulting from subsidence related to groundwater withdrawal in the Houston area may be associated.

Factors Related to Depth

One of the most significant differences between the interior and the coastal subprovinces is the depth to the source bed of salt. In the interior the depth is generally less than 465 m whereas in the coastal subprovince it is generally considered to be a minimum of 9455 m, more likely 12,400 to 12,555 m, and possibly as much as 18,600 m deep in parts of coastal Louisiana.⁽³⁹⁾

Thus, the average piercement distance is much greater for the coastal domes. These have a much more complex structural history that includes much folding, refolding, and attenuation suggestive of plastic deformation and flow of fluids. One possible consequence of the greater piercement distance in the coastal domes is the refining or purification of the salt. Masses of halite have a lower bulk density and behave more plastically than mixtures of halite and anhydrite or masses of anhydrite. It seems reasonable that halite would rise preferentially in the diapiric column and that anhydrite would lag behind. Thus, domes with

the greatest piercement distance could consist of "purified" salt as a result of a combined rheological and gravitational refining process. The process may be aided by continual preferential recrystallization of the halite at progressively higher levels in the diapiric column. Meager data indicate that coastal domes consist of purer salt than interior domes.⁽³⁹⁾

Salt in interior domes tends to be more competent than in coastal domes. This difference could reflect any one or any combination of the following characteristics of interior domes:⁽³⁹⁾ 1) a longer period of structural stability, 2) a simpler tectonic history (less recrystallization), or 3) a greater content of impurities.

The more complex structural history of coastal domes could be an important hydrologic factor if it is learned that discontinuous lenticular masses of highly sheared clastic sediments, such as those reported at Belle Isle and Avery Island domes by Kupfer,⁽³⁹⁾ are common in the domes of the coastal subprovince. These masses of sediments are thought to be former wall rock incorporated into the main salt mass by continued large-scale upward movement. If they include permeable rock, these inclusions could provide channelways that would transmit water into the salt mass, appreciably increasing the probability that mined openings might reach saturated permeable rock within the central salt mass.

Hydrology

The east Texas-south Louisiana salt-dome basin area has a moist subhumid climate. Its eastward-increasing mean annual rainfall ranges from 100 cm along the western edge of Matagorda County, west of Gulf dome, to about 130 cm along a line connecting Davis Hill dome with the eastern part of Galveston Bay. The highest monthly precipitation is mostly from thunderstorms. Tropical storms also occasionally sweep in from the Gulf during those months and drop as much as 76 cm of rain in 24 hr. Flood hazards from these storms are severe. Higher water levels in poorly drained marshy areas along major stream courses and behind the barrier bars can persist for long periods following the storms.

Urban, industrial, and agricultural development in the basin have stimulated several comprehensive investigations of the groundwater hydrology. Groundwater-resources reports are available for Matagorda,⁽⁷⁵⁾ Harris,⁽⁵⁴⁾ Liberty,⁽⁷⁶⁾ and Galveston⁽⁷⁷⁾ Counties, and a report is in progress for Fort Bend County (J. B. Messelman, written communication, 1972). These reports contain data on the quantity and quality of usable water, water usage, and characterization of aquifers including their lithology, hydraulic properties, and potential for future development. Five of the six unrejected salt domes are located in the counties covered by these reports; the exception is Hoskins Mound in Brazoria County. Two regional hydrologic reconnaissance reports that include the basin area are also available.^(78,79)

The oldest strata that yield fresh groundwater near the six unrejected domes are sands in the Miocene Fleming Formation in parts of Harris County. These sands are more deeply buried coastward, where they are not flushed. The Pliocene Goliad Sand and overlying Pliocene(?) (or Pleistocene) Willis Sand are major fresh-water aquifers throughout a broad band 80 to 144 km wide that crosses the basin parallel to the coast. In the northwestern two-thirds of the broad band, the base of the Goliad almost coincides with the base of fresh groundwater which dips coastward. In the southeastern third of the band, the base of fresh water rises through the aquifers, marking the limit to which they are flushed. They do not contain fresh water along the coast.⁽⁷⁸⁾

The resulting configuration resembles an irregular fresh-water trough that parallels the coast with an axis 64 to 96 km inland and along which fresh to slightly saline water occurs to depths ranging from 500 to 930 m. Saratoga, South Liberty, and Pierce Junction domes lie approximately along this axis, whereas Long Point is on the south flank of the trough and Davis Hill on the north flank.

Prior to heavy pumping from wells in the fresh-water trough, the movement of groundwater was coastward down the dip of the aquifers, and discharge was upward through the confining clay beds of low permeability in the southeastern third of the trough.⁽⁵⁴⁾ A state of dynamic equilibrium probably existed between fresh and saline groundwater in the discharge area. That is, the landward-dipping fresh-saline water interface (actually a diffuse zone) probably had a fixed location. Heavy pumping in the Houston area⁽⁵⁴⁾ and along the northeasterly extension of the trough in southwestern Louisiana⁽⁸⁰⁾ has caused acceleration of coastward groundwater movement in the northwest flank of the trough and reversals of movement in the southeast flank of the trough, the discharge area. This condition has created concern about possible landward saline-water encroachment and also about upward movement into the lower parts of fresh-water aquifers along the axis of the trough. Hockley dome is situated in the area of accelerated groundwater movement toward the coast. Fresh water is apparently absent over Hockley dome although it occurs to depths of about 620 m surrounding the dome. Fresh and slightly saline water occurs over Davis Hill dome but at somewhat shallower depths than in the surrounding area where the interface with saline water occurs at about 525 m. At Long Point dome the depth to saline water decreases from about 570 m adjacent to the dome to less than 92 m on top of the dome.

Other important aquifers in the basin are in sands in the several Pleistocene formations noted above; in Holocene deposits along the coast; and in the delta area of the Colorado and Brazos Rivers.^(75,78) The area in which the Pleistocene strata contain fresh to slightly saline water generally extends inland about 96 km, overlapping the area in which Pliocene and Pliocene(?) strata contain fresh water. In general, fresh to slightly saline water is restricted to depths less than 310 m in that area, and the depths decrease toward the coast where

locally no fresh water is available. In the area of Gulf, Hawkinsville, and Hoskins Mound domes fresh water is found to depths of about 187 m and slightly saline water to about 275 m. Fresh and slightly saline water occur over these domes but at lesser depths.

Plumes of anomalously high total dissolved solids in intermediate sands of one aquifer extend outward from some domes (including Long Point) for several miles and are believed related to the domes. Winslow, Doyel, and Wood⁽⁵⁴⁾ and Wesselman believe that the normal movement of groundwater is upward around the domes, but Wesselman notes (written communication, 1972) that changed pressure relationships caused by heavy pumping near some domes may have reversed the movement. Baker⁽⁸¹⁾ reported increases with time in chloride contents of wells near three salt domes in Hardin County, Texas, about 96 km northeast of Houston. The top of the salt in all three domes (Batson, Saratoga, and Sour Lake) is above the regionalized base of the fresh-water-aquifer system. The rates of increase range from 1.25 ppm/year to 77.6 ppm/year. These increases may result from inadequate circulation, inadequate protection of fresh water from oil-field operations, or contamination from underlying salt-water sand or nearby salt domes.

Oil-field brine is widely disposed of by injecting it into deep saline aquifers in the basin. The potential for disposing of mined salt in the form of a brine at any of the six shallow domes is good.

Seismicity

Although the seismic risk of both subprovinces is estimated to be minor or nonexistent, there is a record of slight to moderate damage from earthquakes in both subprovinces between 1890 and 1965.⁽⁴⁴⁾ Only one of the epicenters in the interior subprovince is located in a salt-dome basin (Figure C.25). One earthquake that produced moderate damage and one that produced slight damage are reported east of the northeast Texas salt-dome basin, and a single quake with slight damage is reported from the south margin of the Sabine platform (Figure C.25). Four additional epicenters are reported for the period between December 31, 1966 and January 1, 1972, in the vicinity of the Sam Rayburn Reservoir in east Texas.⁽⁴⁵⁾ However, they may be related to reservoir impoundment and are therefore not plotted on the figure.

In the coastal subprovince, seismic activity is limited to southeast Louisiana and has produced only slight damage (Figure C.25). This seems surprising in view of the reported subsidence that has produced surface fracturing in the Houston area as a result of fluid removal.⁽⁴⁶⁾ Apparently strain release in that area is gradual and at a low level. It is not known whether the reported seismicity (Figure C.25) is related in any way to salt domes. In fact, it is not known whether release of seismic energy is a characteristic feature of salt diapirism. Seismic studies seem to be desirable in areas of subsidence and would probably be very instructive if conducted near salt domes.

Industrial Use

In 1964, production from salt domes amounted to about 74% of the elemental sulfur, about 41% of the salt, and about 12% of the total crude oil produced in the United States. Also, at that time salt domes provided for more than half of the nation's capacity for underground storage of liquefied petroleum gas (LPG).⁽⁴⁷⁾ Industry uses many more salt domes in the coastal subprovince than in the interior subprovince. According to Clark,⁽⁴²⁾ 100 wells have been drilled on the domes of the coastal subprovince to every one drilled in domes of the interior subprovince. Data indicate that about 90% of the shallow domes in the coastal subprovince are being used by the petroleum industry in contrast to about 35% in the interior subprovince. In 1964⁽⁴⁷⁾ only two of the 23 active salt mines were in the interior subprovince. Industry's future use of salt domes in the coastal subprovince is probably reflected in these statistics.

Northeast Texas Salt-dome Basin

The northeast Texas salt-dome basin includes a well defined cluster of domes situated between the Mexia-Talco fault zone on the west and the Sabine platform on the east (Figures C.25 and C.27). Geologically, two of the 20 domes included in the basin on Figure C.27 (Day and Kittrell domes) belong in the south Louisiana salt-dome basin. They are included with the northeast Texas domes in order to conserve on the area covered by Figure C.15 and thereby simplify the illustrations.

Depth to salt in the seven shallowest domes ranges from 34 m for Palestine to 620 m for Whitehouse. Of the seven domes, most is known about Palestine and Keechi.

The northeast Texas salt-dome basin is in an area of gently rolling topography that is crossed from northwest to southeast by three major drainages--the Sabine, Neches, and Trinity Rivers (Figure C.27). Elevations range from about 60 to 215 m and, except for local marshy areas along the major drainages, the area is well drained.

Geology

The northeast Texas salt-dome basin forms the northern part of the east Texas embayment--one of the three large embayments that modify the shape and structure of the Gulf Coast geosyncline. Exposed strata range in age from Cretaceous to Quaternary. A partial stratigraphic section of Tertiary strata in the basin is shown on Figure C.28. Strata of Cretaceous age and those belonging to the Eocene Midway and Wilcox groups are exposed only in the deformed areas around some of the salt domes including Palestine, Keechi, Butler, Brooks, and Steen domes. Most exposures throughout the basin consist of post-Wilcox Eocene strata that have gentle dips into the basin. Quaternary materials consisting of Pleistocene fluvial terrace deposits and Holocene alluvium are restricted to areas along the major and tributary drainage courses.

FIGURE C.27. Unrejected and Rejected Domes in Northeast Texas Salt-dome Basin. Base map, USGS 1:1,000,000; Texas.

FIGURE C.28. Typical Electric and Lithologic Log of Some Tertiary Strata in Smith County, Texas. From Dillard (1963, pl. 3); additional typical electric and lithologic logs within the salt-dome basin are reported by Park (1950, pages 220-221) for the Maydelle field, and by Turner (1950, p. 365) for the South Tyler field; these logs include deeper strata than are shown here.

The upper parts of the seven shallow domes are contained in, and overlain by, strata ranging from Upper Cretaceous to the Sparta Sand of Eocene age. The Upper Cretaceous rocks and the rocks of the Midway Group include limestone, mudstone, shale, clay, marl, and chalk that form complex uplifted blocks on Palestine, Keechi, Brooks, and Steen domes. The top of Whitehouse dome is probably contained in clays of the Midway Group. The tops of Bullard and Mount Sylvan domes are probably contained in the interbedded, poorly cemented quartz sand, silty shale, clay and lignitic clay of the Eocene Wilcox Group. The Wilcox is 310 to 620 m thick and consists of about 40% sand. In ascending order, the important overlying Eocene strata include the Carrizo Sand, the clay-rich Reklaw Formation, the interbedded sand and clay of the Queen City Sand and Weches Formation, and the Sparta Sand.

The Eocene strata form the most important salt-dome geology of the basin. Eocene strata were deposited in alternating marine and continental environments that resulted from repeated transgression and regression of the sea. The marine strata consist mainly of clay and shale with minor sand. The continental strata comprise near-shore sand and lesser amounts of clay, shale, and lignite.

Structurally the northeast Texas salt-dome basin coincides with the Tyler basin, a north-northeast-trending, southward-opening structural depression that is separated from the southern part of the east Texas embayment by the Mt. Enterprise fault zone (Figure C.25).

The top of the Louann Salt is at a depth of about 3250 m in the deeper parts of the salt-dome basin. The Ferry Lake Anhydrite of Early Cretaceous age shows about 775 m of structural relief across the basin.⁽³⁾ The base of the Wilcox Group descends into the basin from a depth of about 130 m below sea level on the flanks to about 465 m along the axis indicating structural relief of about 342 m (Figure C.29).

Eocene strata dip into the Tyler Basin at a rate of about 10 m/km, but the dip increases to as much as 20 m/km in areas within the basin where salt flowage has produced arching and subsidence. Salt flowage has played an important role in localizing subsidiary structures that influenced deposition during the Tertiary age. The subsidiary structures are ridges and troughs that trend northeast. The salt domes occur in the troughs. Sediments thicken and become more sandy in structural depressions where they have collected preferentially. Uplift due to salt diapirism has produced steep dips and faults that have conspicuous surface expression at Palestine, Keechi, Brooks, and Steen domes.

Hydrology

The mean annual precipitation in the northeast Texas salt-dome basin is about 110 cm. About 5% of this water is consumed in recharge to the exposed Eocene strata. Much is lost as runoff.

In general, the base of the Wilcox Group marks the base of the freshwater aquifer system. The base of the Wilcox is at a depth of about 710 m in the

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FIGURE C.29. Diagrammatic Cross Section Trending Northwestward through Smith County, Texas, Showing the Tyler Basin as Expressed in Tertiary Strata; note vertical exaggeration. (From Dillard (1963, pl. 4).

central part of the basin and, as noted above, has about 340 m of structural relief across the basin. The lower beds in the Wilcox are transitional with the clay-rich beds of the underlying Midway Group, and locally they contain saline water similar to the Midway strata.

Almost all data on which this summary is based were acquired from a groundwater report of Smith County, Texas, by Dillard.⁽⁴⁸⁾ There are three main aquifers in the post-Midway Group strata of Eocene age. They are the Carrizo-Wilcox, Queen City, and Sparta aquifers. Within the 1423-km² area of Smith County (believed to typify the rest of the basin) well fields could yield the following quantities of fresh water from the three aquifers, in million liters per day:

Sparta	38
Queen City	12
Carrizo-Wilcox	136

Although far below the potential yields for Mississippi, these data indicate very adequate supplies of fresh groundwater available throughout the basin. The average transmissivities for the three aquifers are as follows, in cubic meters per day:

Sparta	47
Queen City	12
Carrizo-Wilcox	76

In general, groundwater is subjected to water-table conditions in areas of outcrop of the aquifers and to artesian conditions elsewhere. The direction of groundwater movement in the Carrizo-Wilcox aquifer is generally down the dip into the basin, but the greatest recharge area is in the west, and thus the main movement of water is to the east and southeast. Groundwater movement patterns in the other two aquifers are controlled by structural dip, topography, and local discharge to streams that have incised the aquifers. Natural rates of groundwater movement in the basin are slow--probably a few meters per year. Hydraulic gradients are typically about 2 m/km.

Dillard⁽⁴⁸⁾ has noted that the quality of water in the Carrizo-Wilcox aquifer may be adversely affected by salt domes in Smith County. An anomalously high chloride content in City of Tyler Well No. 6 may be a result of moving groundwater coming in contact with the sale of the Tyler dome which, according to available data, penetrates the aquifer to a depth (275 m below the surface) slightly above the pumping level in the well (276 m to 320 m below surface). A systematic study of water quality in other wells that are screened in aquifers penetrated by nearby salt would be very desirable. The published water-quality data for Smith County are inadequate for such a study.

North Louisiana Salt-Dome Basin

The north Louisiana salt-dome basin (Figure C.30) is marked by a northwest-trending cluster of known and suspected salt domes situated between the Sabine platform on the west and the Monroe platform on the east (Figures C.25 and C.30).

FIGURE C.30. Unrejected and Rejected Domes in North Louisiana Salt Dome Basin. Base map, USGS 1:1,000,000; Louisiana and Texas.

There are 15 known domes in which salt has been penetrated by the drill and at least four suspected domes in which salt has not been drilled.

The depth to salt in six of the 15 known domes is in the interval from about 775 to 1995 m. Gas-storage facilities have been developed on four other domes and one dome (Minden) has active petroleum production. Depth to salt is less than 310 m for four (Vacherie, Kings, Rayburns, and Winnfield) and is as little as 30 m on Rayburns. Caprock has been found on the four domes but varies greatly in thickness from dome to dome and over individual domes.

There are at least four suspected salt domes that have surface structural and (or) topographic expression but in which the depth to salt is unknown--Castor Creek, Cedar Creek, Prices, and Prothro. Castor Creek, Cedar Creek, and Prices domes have surface "salines" or salt licks similar to at least four of

the shallow known domes in the basin--Bistineau (465 m), Drakes (275 m), Kings (53 m), and Rayburns (35 m). Also, exposed strata at Prothro dome are uplifted and tilted as a result of salt diapirism similar to at least five other known shallow domes in the basin--Arcadia (435 m), Minden (590 m), Gibsland (270 m), Rayburns, and Drakes. The known deep domes in the north Louisiana salt-dome basin lack "salines" or surface structural expression.

FIGURE C.31. Altitude of Base of Fresh Ground Water and Location of Domes in North Louisiana Salt Dome Basin; chloride of fresh water less than 250 parts per million; contour interval 100 feet, + above sea level, - below sea level. Contour map from Rollo, 1960.

Eight known and suspected domes are distributed throughout Winn and Bienville Parishes, and one dome (Vacherie) is located at the north boundary of Bienville Parish where it adjoins Webster Parish (Figure C.31).

The tops of two deeper salt domes, Coochie Brake and Chestnut, lie within 807 m of the surface. The Coochie Brake and Chestnut domes are located in Winn and Natchitoches Parishes, respectively.

Of the known domes only Winnfield has a mined opening. More is known about the geology and hydrology of the Winnfield dome than any of the others. The mine and shaft are at present in a completely flooded condition as a result of a sudden inflow of water near the shaft on November 17-18, 1965. The mine did not collapse, but minor subsidence and fracturing occurred at the surface, probably because of rapid dissolution by the inflowing water at the upper caprock-salt contact.

Brine production from shallow wells, mostly during the Civil War, is reported for five of the shallow domes in the basin.

The north Louisiana salt-dome basin is part of a broad dissected plain having a gently gulfward slope. It is an area of gently rolling or hilly topography with elevations ranging from 77 to 130 m. The hilly topography is interrupted locally by nearly flat valleys characteristic of the main drainages of northern Louisiana. Four of these southward flowing drainages cross the basin. From west to east they are Bayou Dorcheat, Black Lake Bayou, Saline Bayou, and Little River (Figures C.30 and C.31).

Geology

Comprehensive geologic reports describing the physiography, stratigraphy, and salt-dome geology are available for Webster Parish⁽⁴⁹⁾ and Caldwell and Winn Parishes,⁽⁵⁰⁾ but none is available for Bienville Parish. However, stratigraphy of Webster Parish adequately represents the stratigraphy in the rest of the salt-dome basin. Most of the domes in the salt-dome basin are located in an area where beds of the Claiborne Group of Eocene age are at the surface, but the more westerly domes are located along the western limit of the Claiborne where the underlying Wilcox Group is exposed on the east flank of the Sabine platform. Three of the domes that lie along the western and southern margins of the basin (Bistineau, Coochie Brake, and Castor Creek) are in areas where the Tertiary strata are buried by a thin veneer of terrace deposits of Pleistocene age.

Some salt domes have surface expression. For example, the only Upper Cretaceous strata exposed in Louisiana occur at Rayburns and Prothro domes where they have been uplifted.⁽⁵¹⁾ Although not exposed, Cretaceous strata are reported at or near the surface of Kings and Bistineau domes.⁽⁵²⁾ At Vacherie dome Wilcox strata are exposed in an area where the normal surface strata are the overlying Claiborne Group. Atop the Arcadia, Minden, Gibsland, and Drakes

domes, some exposed rocks of the Claiborne Group are older than the Claiborne rocks surrounding the domes.

The upper parts of the four known shallowest domes probably penetrate highly flexed Upper Cretaceous, Paleocene, and lower Eocene strata. It is common for the older strata to be uplifted about 620 m at the domes. If any or all of the four suspected domes are as shallow as 620 m, their tops would be within the same stratigraphic range as those of the known domes. In general, Claiborne strata are flexed over, rather than penetrated by, the salt cores.

On the basis of subsurface data, rocks of Late Cretaceous age are at least 342 m thick in the basin and consist mainly of marine shale and sandy shale, marl and clay-marl, chalk and impure chalk, and minor sand, limestone, and red clay.⁽⁵²⁾ The Midway Group of Paleocene age is about 170 m thick and consists almost wholly of marine clay. The Eocene and Paleocene Wilcox Group thickens southward from 240 to 775 m and is composed of interstratified sand and clay, probably mostly of continental and deltaic origin. Although generally not in contact with the salt, the Claiborne Group of Eocene age is of special significance because it carries the important supplies of fresh groundwater in the basin.⁽⁵³⁾ The Claiborne ranges from 550 to 800 m in thickness and consists of four formations which are, in ascending order, Cane River Formation, Sparta Sand, Cook Mountain Formation, and Cockfield Formation. These strata record alternating marine and continental deposition. The Cane River and Cook Mountain are mainly clay, and the Sparta and Cockfield are mainly sand and are the principal fresh-water aquifers.

The north Louisiana salt-dome basin is a relatively shallow northwest-trending trough that plunges and opens to the southeast and plays out into the Gulf Coast geosyncline. The trough area subsided more rapidly and received thicker accumulations of Mesozoic and Tertiary sediments than the flanking Sabine and Monroe platforms. The Louann Salt is at a depth of about 3130 m in the north part of the basin and is probably more than 4650 m deep in the southern part. The Ferry Lake Anhydrite of Early Cretaceous age shows about 775 to 930 m of structural relief across the northern part of the basin.⁽³⁾

As noted above, salt diapirism has produced strong local uplift, but little seems known about possible subsidiary deep structures in the basin that may be related to salt movement. Structural contours on the Ferry Lake Anhydrite show some irregularities including two small northeast-trending subsidiary troughs that are possibly related to salt withdrawal during dome formation.⁽³⁾ The main structural irregularities in the basin are those related to salt diapirism.

Hydrology

There is considerable published general information concerning the availability of fresh groundwater in Louisiana.^(53,54) Detailed reports concerning

the surface- and groundwater resources are available for several of the parishes located west and south of the salt-dome basin but not for the parishes within the basin.⁽⁵⁵⁻⁵⁷⁾ Much basic water-supply information, including water quality data, is available for municipal water supplies in the basin.⁽⁵⁸⁾ Additional information is available from the offices of the U.S. Geological Survey in Baton Rouge.

Surface Water. The average annual precipitation in the basin as estimated from published reports on adjoining parishes, ranges from about 120 to 130 cm. About 35% is lost to runoff, and a small percentage is consumed in recharging the aquifers.

The basin area is topographically high relative to the drainage courses of the Red and Ouachita Rivers between which it is situated (Figure C.30). The basin is drained mainly by a dendritic system of streams that flow southward to Red River.

In general, the area is well drained, with small local swamps and marshes restricted to some of the major stream courses and to the "salines" or areas of subsidence that are localized over some of the salt domes. Records for peak runoffs are available from gaging stations near the Cedar Creek and Castor Creek domes,⁽⁵⁹⁾ and those data can be used to establish flood-frequency relationships.

Groundwater. Fresh groundwater is available throughout northwest Louisiana to depths ranging from more than 65 m above sea level on the Sabine platform to 342 m below sea level at Rayburns dome along the axis of the salt-dome basin (Figure C.30). Locally, as at Rayburns dome, the base of the freshwater system extends down into the sands of the Wilcox Group, but throughout most of the basin fresh groundwater is restricted to the strata of the overlying Eocene Claiborne Group. Small amounts of fresh water are found in the valley and the terraced upland deposits of Quaternary age that form veneerlike belts along the major and tributary drainages. Deposits of Holocene age form thin mantles of sand, silt, and clay along stream valleys but do not constitute important aquifers.

Recharge to fresh-water aquifers is principally from rainfall. Discharge from the aquifers is by evapotranspiration, effluent seepage, interaquifer leakage, and pumping from wells. Total natural discharge far exceeds the total discharge from wells.

The depth to the base of fresh water broadly reflects control by regional structural and stratigraphic patterns. For example, the depth increases generally southward toward the Gulf Coast geosyncline and eastward toward the axis of the Mississippi structural trough. On a somewhat more restricted scale the Sabine platform in northwest Louisiana has a pronounced effect on the position of the base of the fresh-water system. A thick clay section belonging to the Midway Group of Paleocene age is at or near the surface over

much of the Sabine platform. The clay prevents circulation of fresh water into and out of the deposits of Cretaceous age. Thus the base of the fresh-water system is shallow over the platform but deepens eastward in the north Louisiana salt-dome basin and becomes abruptly deeper southward where the dip off the platform adds to the southerly dip of the Gulf Coast geosyncline. Within the basin the base of the fresh-water zone (defined as 250 ppm chloride) is quite irregular (Figure C.30), but generally it follows the base of the Sparta Sand of the Claiborne Group.⁽⁵³⁾ On the south side of the approximately east-west alignment formed by Winnfield, Cedar Creek, and Castor Creek domes (Figure C.30), the base of the freshwater zone rises abruptly through about 140 m of section into younger strata of Oligocene and Miocene age. This abrupt step in the base of the fresh-water zone reflects the southern limit of flushing of connate saline waters from the Claiborne strata.⁽⁵³⁾ Where complications due to faulting or withdrawal from wells do not occur, groundwater movement is in the general direction of the regional dip of the strata. However, near the limit of flushing of an aquifer the major component of movement of fresh groundwater will be upward either along faults, through the less permeable confining beds, or, if a salt dome is present, along the structurally disturbed strata that surround the salt mass. The possibility exists that accelerated upward movement of groundwater related to the limit of flushing of the Claiborne strata at Winnfield salt dome contributed to the hydrologic problems encountered at that dome.

Based on a broadly consistent density of control points, the configuration of the basal surface of the fresh-water system is far more irregular in the north Louisiana salt-dome basin than in adjacent areas (Figure C.30). Depressions and rises ranging in area of closure from 5.2 to 260 km² are common throughout the basin. Rollo⁽⁵³⁾ has suggested that salt domes that penetrate to or near the land surface have a pronounced effect on the occurrence of fresh groundwater and may cause some of the erratic patterns in the basin.

It is at least interesting to note that many of the salt domes in the central part of the basin are located where the base of the fresh-water zone is marked by locally oversteepened south to southeasterly slopes. Inasmuch as the normal hydraulic gradient and movement direction in the basin is south and southeast, these relationships seem to preclude the contribution of chloride ions to the aquifers from the salt domes within the limits of detectability of the sampling and analysis system used by Rollo.⁽⁵³⁾ In fact, some of these oversteepened slopes may be related to the flushing mechanism proposed by Rollo.

Seven of the known and suspected domes in the basin have surface "salines" or "salt licks." The depth to salt in the known domes of this group ranges from 35 to 342 m. Because the detailed movement patterns of groundwater around salt domes are not known, the source of salt in the surface "salines" is not known. It could either be leached from uplifted masses of strata that contain saline connate water, or it could be derived from dissolution at the top of

the salt mass and slow upward circulation of the resulting brine through pore spaces as suggested by Goldman.⁽⁴³⁾ However, as Bodenlos⁽⁶⁰⁾ has noted, the second process (upward circulation through pores) does not take into account the relatively low permeability of anhydrite caprock. Reported caprock thicknesses for the seven north Louisiana domes with surface "salines" range from 3.4 m (Kings dome) to 200 m (Drakes dome) indicating that the development of the saline (based on this small sample) is independent of the thickness of caprock. This factor together with the wide range in depth to salt noted above seems to suggest that the salt in surface salines does not derive from dissolution of the top of the salt mass. In either case, the "salines" indicate complex hydrologic patterns at salt domes.

Mississippi Salt-Dome Basin

The Mississippi salt-dome basin extends in an east-southeast direction from northeastern Louisiana across Mississippi to southwestern Alabama--a distance of about 400 km (Figures C.25 and C.32). It averages about 96 km wide and contains 77 known and suspected salt domes. Salt has not been drilled in four of the 14 shallowest domes--Crowville, Sardis Church, Hazlehurst, and Cypress Creek. On the basis of meager drillhole data, the depth to salt at Crowville and Sardis Church is probably greater than 620 m, and at Hazlehurst and Cypress Creek it may be slightly less. The salt tops of four of the 10 known shallow domes are: Bruinsburg, 625 m; Byrd, 638 m; Leedo, 640 m; and County Line, 675 m. The depth to salt is greater than 400 m in 10 known shallow domes.

The Mississippi salt-dome basin covers parts of two natural physiographic regions: 1) the Mississippi alluvial plain in the northwest, and 2) the Gulf Coastal Plain in the southeast. These two regions are separated by a narrow north-northeast-trending physiographic district east of the Mississippi River designated the loess or bluff hills. The Gulf Coastal Plain is characterized by pine-forested hilly uplands traversed by alluviated stream valleys, whereas the Mississippi alluvial plain is nearly flat. Elevations in the Mississippi salt basin range from 77 to 187 m along major divides and from about 15 to 46 m along major streams.

Geology

Surface exposures in the large southeastern part of the salt-dome basin are Tertiary deltaic sediments mantled locally by thin deposits of Pleistocene sand and gravel. A smaller northwestern part is entirely within the broad belt of Quaternary alluvium of the Mississippi Valley. This alluvium forms a thin mantle averaging about 65 m thick above the Tertiary sediments. The tops of the 10 known unrejected salt domes are in Tertiary deltaic strata ranging from the Wilcox Group of Paleocene and Eocene age upward to the Catahoula Sandstone of Miocene age (Figure C.33). These strata consist mainly of poorly lithified sandstones and interbedded shales with minor amounts of marl,

FIGURE C.32. Contour Map Showing Base of Fresh Ground Water, in Feet below Sea Level, in the Mississippi Salt Dome Basin; long-dashed lines show approximate trace of discontinuities in contoured surface, short-dashed lines show location of geohydrologic sections of figures 12 and 13. Modified from Newcome (1965, 1967); Lang (1972); Shaw and others (1966).

FIGURE C.33. Geohydrologic Section across Smith and Scott Counties, Mississippi, Showing the Relationship between the Base of the Fresh Water and the Coastward-dipping Aquifers; line of section shown on figure 11. Modified from Boswell and others, 1970.

limestone, and sandstone. The percentage of interbedded shale and clay in the Miocene and younger strata increases toward the Gulf of Mexico, as does the thickness of those strata. The unrejected domes are located in the southern part of the Mississippi salt-dome basin. It is situated near a transition zone where the stratigraphic section changes from predominantly sand to the north to predominantly shale to the south. Eargle⁽⁶¹⁾ has described the stratigraphy and structure of the Tatum dome area, which includes Lampton, McLaurin, and Richmond domes. Structure and isopach maps of that area show changing patterns of structural development and sedimentation since the end of Early Cretaceous time related to the formation of a middle Tertiary depocenter in the Mississippi embayment.⁽⁶¹⁾ These changing patterns together with the overall position of the transition zone produce stratigraphic complexities in the area of the unrejected salt domes. Additional complexity is introduced by common ancient river channels in which tongues of channel sands were deposited, generally normal to the present strike of the strata. Although complex, the regional structural and stratigraphic framework is well known, and much of the specific geologic data available for Tatum dome can be applied to neighboring domes.

The Mississippi salt-dome basin is a deep, west-northwest trending, asymmetric depression that lies athwart the southwesterly projection of the buried

Appalachian tectonic belt and the south-southeastward projection of the buried Ouachita tectonic belt (Figure C.25). In the east the basin trends westward. An approximately north-south section across it⁽⁶²⁾ shows the top of the Louann Salt at depths ranging from 3165 m, beneath the north flank, to 6420 m, beneath the structural axis. Displacement of the structural axis to the south of the geographic axis (Figure C.25) gives the basin a strongly asymmetric profile. The Louann Salt is inferred to be 1582 m thick beneath the axis in eastern Mississippi.⁽⁶²⁾ Shallow crustal structure interpreted from reversed refraction data along a north-south line located about 64 km west of Williams' cross section clearly shows the asymmetric depression, not only within the sedimentary prism, but also as a basement structure.⁽⁶³⁾ To the west, the axis of the salt-dome basin swings slightly northward and trends across the axis of the Mississippi structural trough into Louisiana, where it apparently terminates against the Monroe platform (Figure C.25). There is no indication of broadening of the basin where it crosses the trough. The salt-dome basin has little or no structural expression in the Tertiary strata in which the tops of the 10 known shallow domes are located. These strata dip from 4 to 20 m/km. The dips are generally gulfward, and the strikes swing in a broad arc, convex to the north, that reflects the combined structural effects of the Mississippi structural trough, the Gulf Coast geosyncline, and, to a small degree, the Monroe platform. Although they are highly exaggerated vertically, the geohydrologic sections shown in Figures C.33 and C.34 illustrate the generalized shallow stratigraphy and structure across the salt-dome basin.

The depth to salt in the domes of the Mississippi salt-dome basin ranges from 130 m to more than 4650 m. The tops of 44 of the 77 domes lie between 620 and 1240 m deep; the tops of only two are less than 310 m deep, and the tops of 12 are more than 3100 m deep. The areal distribution of the depths to salt in the domes is irregular, although the deeper domes occur in the northern part of the basin and the shallower domes occur in the southern part. This explains the fact that there are no potentially suitable domes in the northern part of the basin (Figure C.25). They are all too deep. It is possible that an inadequate salt supply in the Louann Salt or some other factor regulating the diapiric process did not permit continued upward penetration by the several very deep salt diapirs in the northern part of the basin. Some of the variation in depth to salt of the shallower domes may be related to groundwater hydrology.

Hydrology

For availability and quality of surface and groundwater, the Mississippi salt-dome basin is one of the most richly endowed areas in the United States. In most parts of the basin individual wells could yield 3800 to 11,400 liters/min and well fields could produce as much as 95 million liters/day of good to excellent water. Appraisals of the hydrologic potential of multicounty areas

FIGURE C.34. Geohydrologic Section across Marion and Jefferson Davis Counties, Mississippi, Showing Relationship between Base of Fresh Water and Coastward-dipping Aquifers; line of section shown on figure 11. From Taylor and others, 1968.

covering most of the salt basin have been prepared cooperatively by the Water Resources Division of the U.S. Geological Survey and the Mississippi Industrial and Technological Research Commission as informal reports available from the state offices in Jackson. These reports contain much information on surface-- and groundwater hydrology, only a fraction of which is summarized here. Other informative reports include hydrologic appraisals of the Pascagoula,⁽⁶⁷⁾ Pearl,⁽⁶⁴⁾ and Big Black⁽⁶⁵⁾ River basins in Mississippi, all of which cover areas that traverse the Mississippi salt basin. None of these reports, however, contains specific information concerning the hydrology of salt domes.

Surface Water. Annual rainfall varies from about 130 cm in the northwest part of the basin to about 150 cm in the southeast part. About 30% of this water is lost as runoff, a large percentage is lost consumptively through evapotranspiration, and a small part infiltrates to the groundwater reservoirs. Of special interest in this report is the water lost through runoff, because runoff determines, in part, erosion rates, and it produces flood hazards.

Severe floods have occurred in the Mississippi salt basin in recent years. Peak annual discharge data acquired at gaging stations have been used to construct flood-frequency curves for selected recurrence intervals of as much as 50 years for some of the major streams that drain the Mississippi salt basin.⁽⁶⁶⁾ These curves, when manipulated by a shape coefficient for the drainage basin, can be used to estimate the magnitude of future floods for selected recurrence intervals of as much as 50 years at specific sites, according to a method outlined by Wilson and Trotter.⁽⁶⁶⁾ The curves could be projected, with appropriate caution, to estimate for recurrence intervals greater than 50 years if consistent climatic conditions are assumed.

A rough approximation of erosion rates in the Mississippi salt basin can be obtained from data on rates of soil loss for various landusage categories. Soil-loss data from several watersheds in the basin were acquired from Graham Renfro of the U.S. Soil Conservation Service in Fort Worth, Texas (verbal communication, 1972). The acquired data pertain only to soil loss by sheet erosion, and specifically exclude losses due to critical erosion along stream banks, road cuts, road banks, gullies, etc. There are 56 tributary watersheds within the 22,700 km² Pearl River basin, and their average soil loss ranges from 0.6 to 2.5 mm/year. The range of soil loss in the 25,000 km² Pascagoula River drainage basin is similar to that of the Pearl--cultivated land, 1.8 to 10 mm/year; land in pasture, 0.3 to 2.3 mm/year; and forested land, 0.08 to 0.3 mm/year. The data show that forested land has the least soil loss of the three, and that by proper forest management, soil loss on forested land could probably be kept as low as 0.05 mm/year--between 4.6 and 6 m in 100,000 years.

Groundwater. The resource potential of the fresh groundwater system in the Mississippi salt basin has received much study in recent years, and data are available in several reports.^(53,67-73) Although less is known about the saline

aquifer system than the fresh, the saline system has been used locally for brine disposal. At Baxterville oil field near Tatum dome, for example, about 20 million barrels of brine are injected into the system each year, and the cumulative total is over 200 million barrels.

Fresh groundwater is found in the Mississippi salt basin to depths ranging from less than 32 m to more than 930 m below seal level (Figures C.33 and C.34). Strata containing important fresh water include the sand of the Wilcox Group, to a depth of 930 m in a small area in the northeastern part of the basin; the Claiborne Group, especially the Sparta Sand;⁽⁷⁴⁾ lenticular sands in a thick section of undivided Miocene beds; and, in the western part of the basin, blanket alluvial deposits in the Mississippi Valley.

The base of the fresh-water-aquifer system slopes generally southwestward, subparallel to the regional dip of the strata. It is characterized by at least two major discontinuities or steps across which the depth to salt water decreases abruptly to the south or southwest (Figure C.34). These discontinuities reflect the southerly limits of flushing of certain aquifers or groups of aquifers as shown on the geohydrologic cross sections on Figures C.33 and C.34. The northern discontinuity marks the limit of flushing of aquifers of the Wilcox Group, and the southern one the limit for aquifers in the Claiborne Group. All ten shallow salt domes are located south of the southern discontinuity, in the area where fresh groundwater does not exist beneath the Miocene beds. As noted above, the domes tend to be shallower in that area than elsewhere in the salt basin. The possibility exists that the depth to the tops of the salt diapirs is in part determined by movement of fresh groundwater.

The upper parts of the two shallowest domes (Lampton, 405 m, and Tatum, 453.5 m) penetrate upward through the base of the freshwater-aquifer system. Lampton dome appears to extend about 107 m above the regional base of fresh water. Tatum dome extends about 92 m above the regional base of fresh water, and about 240 m above a local depression in the base of the fresh-water system that is related to the rim syncline. The direction and rate of movement varies with adjustments on the buried faults. Sheets⁽⁴⁶⁾ has emphasized the potential hazards of surface fracturing in the Houston area. He recognizes that subsidence related to fluid withdrawal may trigger or accelerate surface movement above some of the buried faults, but he suggests that there is no relation between the location, age, or trend of those faults and the subsidence. He suggests that active surface fracturing in the Houston area is commonly the surface expression of larger buried faults that either belong to the regional system that parallels the coast or to systems related to salt domes.

South Texas Salt-Dome Basin

The south Texas salt-dome basin is a triangular area (Figure C.25) encompassing the lower part of the Rio Grande embayment, a southeastward-plunging syncline tributary to the northeastward-trending Gulf Coast geosyncline. The salt-dome

basin includes only six domes in which the drill has penetrated salt (Figure C.35). However, several additional domal structures in which salt has not yet been penetrated are known in the region. The basin includes Pescadito dome, believed the largest in the United States. It is the farthest west dome of the Gulf region and also one of the deepest known salt domes--over 3230 m. The salt-dome basin also contains, at its southeastern end, the only structure in Texas in which gypsum caprock has been found at the surface--Gyp Hill dome. Domes of the basin, therefore, exhibit some of the qualities of both the interior and the coastal domes, as described in other sections of this report.

The surface of the south Texas salt-dome basin consists of two types of plains, a nearly flat depositional surface on late Tertiary and Quaternary coastal and fluvial deposits, and an inland erosional surface, of somewhat more rolling submature topography developed on Tertiary formations. Between the two is the high west-facing Bordas Escarpment, the western limit and the highest elevations of the depositional surface.

Geology

The outcropping rocks of the south Texas salt-dome basin range from middle Eocene in the western part to Holocene in the east. The stratigraphic section for the east Texas-south Louisiana salt-dome basin (Figure C.36), generally applicable to the south Texas basin, serves as a reference for the following discussion. The larger structural feature that includes the salt-dome basin is the Rio Grande syncline. Although only relatively few salt diapirs have penetrated the overlying rocks, salt is believed to extend beneath a much larger area than that encompassing the area of the domes. Most of the known domal structures produce petroleum.

The Cretaceous and Paleocene formations that crop out in the Rio Grande syncline are marine; they are succeeded upward by alternating marine and non-marine, beds reflecting tectonic movements involving the Gulf Coast geosyncline. The geosyncline appeared to sink periodically, causing transgression of the coastline landward, but continuing sedimentation pushed the shoreline seaward again, and the sediments became brackish to fresh water during the latter part of each cycle.

Sediments of Tertiary and Quaternary age thin and change their facies outward from the center of the Rio Grande Basin, as they do shoreward from the Gulf Coast geosyncline. Wedge-shaped as they are, the units therefore range considerably in thickness and lithologic characteristics.

Formations associated with piercement domes in the south Texas region are, from oldest (Eocene) upward: Laredo Formation, chiefly fossiliferous sands interbedded with chocolate-colored fissile clay, 187 to 215 m thick; the Yegua Formation, chiefly clay with minor beds of sandstone and layers rich in oyster shells, 187 to 248 m thick; the Jackson Group (upper Eocene), about 375 m thick,

FIGURE C.35. Shallow and Deep Domes in South Texas Salt Dome Basin. Base map, USGS 1:1,000,000; Texas.

^{1/}Names used are not necessarily in conformity with U.S.G.S. standards.

FIGURE C.36. Schematic Generalized Stratigraphic Section Pertaining to the Western Part of the East Texas-South Louisiana Salt Dome Basin. From Eargle and others, 1971, figure 2.

of montmorillonitic clay, volcanic ash, and sands grading upward into Frio Clay (Oligocene?) about 92 m thick; the Miocene formations, Catahoula Tuff, Oakville Sand, and Fleming Formation, about 465 m thick; the Pliocene Goliad Sand and Pleistocene coastal terraces, undivided--sands and clays--465 to 620 m thick.

The center of the Rio Grande syncline, the large feature of which the south Texas salt-dome basin is a part, extends southeast from the vicinity of Cretaceous outcrops of southwest Texas to the coast south of Corpus Christi Bay. In this region, therefore, the axis of the basin lies north of the present course of the Rio Grande, and its trend parallels and relates more to that of the Nueces River than to the Rio Grande.

Faulting accompanied basin subsidence as is shown on the tectonic map of the northern Gulf Coast region (Figure C.25). Principal faults nearly parallel the strike of the formations and are generally downthrown to the coast. They are accompanied, however, by compensatory upthrown-to-the-coast faults, forming grabens. The principal faults lie along the landward margin of the salt-dome basin. Movement of salt in the deep subsurface may have contributed to, or perhaps been the main cause of, the faulting.

Two principal fault zones tranverse the south Texas basin and trend much farther northeastward toward the east Texas-south Louisiana coastal basin: the Mirando-Provident City fault zone and the Sam Fordyce-Vanderbilt fault zone (Figure C.25). Movement along the faults was taking place while the sediments were being deposited. Thus, the beds are spectacularly thicker on the downdip (gulfward) side of each fault.

Hydrology

Most of the statements that follow were abstracted from a report on the groundwater resources of Brooks County, Texas,⁽⁸²⁾ and from a manuscript on the groundwater resources of Kleberg, Kenedy, and southern Jim Wells Counties, Texas (G. Shafer and E. T. Baker, written communication, 1972). The statements pertain mostly to the southeast part of the salt-dome basin where the dome, Gyp Hill, is located. The climate is semiarid with average annual precipitation of about 59 cm and average monthly temperatures ranging from 14°C during January to 30°C during July and August. Sandy soils blanket the area and facilitate infiltration of rainwater, but extensive blankets of caliche hold much of the infiltrated water near the surface so that the water evaporates. The area is drained by small intermittent low-gradient streams that flow east and southeast in wide, nearly flat valleys. The area is almost wholly dependent on its groundwater for municipal and agricultural water supplies.

Tropical storms are known to strike the area and deposit in a few days rainfall equal to or exceeding the mean annual precipitation. Torrential rainfall of record-breaking magnitude from hurricane Beulah in October 1967 produced flooding of a 4800-km² area in south Texas.⁽⁸³⁾ The Gyp Hill area received about

50 cm of rainfall, much of which was ponded in numerous shallow depressions. High areas such as Gyp Hill were not inundated, but highways were blocked for several days, and ranching and oil-field operations were hampered for at least 6 months. The hydrologic system has approached but not returned to pre-Beulah conditions 2 years after the hurricane.⁽⁸³⁾

Aquifers are recharged by rainfall in outcrop areas to the west, and groundwater moves eastward under artesian conditions at rates varying from a few meters per year during periods of low rainfall to tens of meters per year during periods of high rainfall. During high rainfall, the piezometric surface rises in the recharge area, resulting in a steepening of the hydraulic gradient and an increase of hydraulic pressure. Usable groundwater is found in several aquifers,⁽⁸²⁾ but the Pliocene Goliad Sand is the main aquifer because it is the most permeable and therefore the most completely flushed of the fresh-water-bearing strata. In some areas saline or slightly saline water is found in aquifers above Goliad beds that bear fresh water of good quality. Adequate supplies of undeveloped fresh groundwater are available from the Goliad Sand in the vicinity of the Gyp Hill salt dome (Figure C.37).

FIGURE C.37. Cross Section C-C' through Part of the South Texas Region in the Vicinity of Gyp Hill Dome Showing Age and Characteristics of Aquifers. Modified from Myers and Dale, 1967, figure 17.

C.2.5. Dome Salt - Paradox Basin

The Paradox Basin is in southwest Colorado and southeast Utah (Figure C.38), an area relatively sparsely populated. The Paradox Basin, which is not a definable physiographic feature, lies well within the much larger Colorado Plateaus province. Basin boundaries are usually determined by the extent of a thick sequence of Pennsylvanian evaporites for which the basin is named.

The spectacular topography of the region is the result of rapid down-cutting by the Colorado River and its tributaries through relatively flat-lying, resistant sandstones. Many of the massive, resistant sandstones form sheer walls in canyons as much as 620 m deep, creating formidable barriers to access. Within the basin three prominent isolated mountain masses rise above the general level of the land. The largest are the La Sal Mountains about 24 km east of Moab; they reach an altitude of 4057 m. The other two masses are the Abajo Mountains near Monticello, Utah, and the Ute Mountains a few miles west of Cortez, Colorado. Southeastward the basin is bordered by the high peaks of the San Miguel, San Juan, and La Plata Mountains. Each of these mountain groups consists of an intrusive igneous complex. Part of the basin is bordered by more gently rising uplifts, including the Uncompahgre, Defiance, Monument and San Rafael. Locally, the famous salt anticlines of the region interrupt the continuity of the general land surface, forming linear flat-floored valleys bounded by steep escarpments. Fracture patterns and joint systems on the flanks of the anticlines have provided planes of weakness leading to the development of natural arches, spires, fins, and other spectacular erosional features.

Most of the area is semiarid, but the precipitation varies with the altitude, and ranges from less than 25 cm in the low deserts bordering the Colorado River to more than 75 cm in the highest parts of the mountain masses. Much of the precipitation occurs as scattered thundershowers during the summer--particularly at the lower altitudes. Winter snows seldom remain on the ground longer than a few days or weeks except on the mountains and the Uncompahgre Plateau. The temperature ranges from about -30°C to about 38°C. Most of the area is covered by pinyon and juniper or by sagebrush, and the streams are bordered by cottonwood, willow, and salt cedar (tamarisk). Ponderosa pine, spruce, fir, and aspen occur at the higher altitudes.

Drainage and Water Supply

The Colorado River is the master stream of the area, cutting diagonally across the northwestern half of the basin (Figure C.38). Major tributary streams include the Green, San Juan, and Dolores Rivers (Figure C.38). Most of the salt-anticline region is drained by the Dolores River, which heads higher in the San Juan and San Miguel Mountains, flows southward to the town of Dolores, then turns and flows northward and northwestward through the area. The fact that the Dolores cuts directly across the middle of Paradox Valley

FIGURE C.38. Paradox Basin

rather than flowing longitudinally through it is responsible for the name Paradox. A small eastern part of the area is drained by the San Miguel River, which also heads in the San Juan and San Miguel Mountains, and which joins the Dolores below Uravan, Colorado.

Most of the small tributaries are ephemeral, as might be expected, but a few streams are perennial, particularly those emanating from the La Sal and Abajo Mountains, and the Uncompahgre Plateau. Except along the flanks of these high areas, springs are relatively few, small, and are mainly in the synclinal valleys. Exceptions are the several large springs that issue from the northeast wall of Paradox Valley and supply water for irrigation. Fred Cater (verbal communication, July 17, 1972) states that these springs flow about $5 \text{ m}^3/\text{min}$ and believes that the water issues from faults that border the northeast wall of the valley. Water may be conducted along these faults and fault zones from the flanks of the La Sal Mountains northwest of the valley.

The principal aquifers and potential aquifers plus Quaternary alluvium supply many stock and domestic wells and a few wells used for irrigation, municipal, or industrial supplies. Some of these wells favorably situated on the flanks or troughs of synclines flow small amounts of water under artesian conditions.

Stratigraphy and Hydrology

In the Paradox Basin a wedge-shaped sequence of sedimentary rocks overlies a basement complex of Precambrian crystalline rocks. Beginning with the Cambrian, every geologic period except the Ordovician and Silurian is represented by these sedimentary strata (Figure C.39). From Cambrian through Mississippian time the area was covered by marine waters. A brief emergence during Early Pennsylvanian time allowed the development of karst topography overlain by a thin regolith. The basin then subsided rapidly at the same time that marginal areas rose, particularly the Uncompahgre uplift to the northeast. Invasion by the sea began a new cycle of marine sedimentation during which the basin was nearly filled with evaporites. By late Pennsylvanian time only a shallow basin remained, and the continued slow retreat of the sea eventually left the entire area above water. The retreat of the sea was accompanied by strong positive movements of the bordering uplifts. Erosion of the exposed crystalline cores of these uplifts then covered the former marine basin with a flood of arkosic debris. Continental sedimentation continued with Late Cretaceous time, when a new invasion by the sea deposited a thick layer of shale. Regional uplift drove out the sea at the close of the Cretaceous, and the area has remained high and dry since. Quaternary erosion has left a basin-filling wedge of sedimentary rock that is about 5580 m thick along the northeastern edge of the basin but thins to about 1860 m southwestward.

FIGURE C.39. Generalized Stratigraphic Chart of Northeastern Paradox Basin.

Rocks older than the Middle Pennsylvanian Paradox Member of the Hermosa Formation are not exposed within the Paradox Basin but have been penetrated by numerous drill holes within the basin and are exposed near Durango along the southeast margin of the basin.

Paradox Rocks

The Paradox Basin's extensive salt deposits of Middle Pennsylvanian age form the Paradox Member of the Hermosa Formation.⁽⁸⁴⁾ Where the evaporite of halite facies is present, the Hermosa Formation can be divided into an upper member above the halite and a lower member below, both members being predominantly carbonate facies (Figure C.39). The Paradox Member exhibits a remarkable cyclic pattern of chemical and clastic sedimentation. The member can be subdivided into 29 evaporite cycles in the deeper part of the basin, and for convenience of correlation each cycle has been given a numerical designation.^(85,86)

The evaporite facies, predominantly halite, extends over an area of about 31,200 km² in southeast Utah and southwest Colorado. Evaporites are thickest in a troughlike depression bordering the ancestral Uncompahgre uplift along the northeast margin of the basin (Figure C.38). The original thickness of evaporites deposited in the trough area is difficult to estimate because of the widespread effects of salt flowage in the numerous diapiric salt anticlines of the region. Where the salt is not so severely disturbed the original thickness of the evaporites is estimated to have been between 1550 and 1860 m. In the salt anticlines the maximum known thickness of evaporites now is nearly 4340 m. Isopach maps of the evaporite facies show thickness variations which are the result of both differential deposition across ancestral structural features and flowage into salt anticlines. The greatest thickness variations occur in the Uncompahgre trough (Figure C.38), where thicknesses may vary from zero to more than 3100 m in a horizontal distance of 4.8 km or less.

A typical Paradox Member evaporite cycle of deposition consists chiefly of chemical precipitates; however, a small amount of detrital and organic material forms an integral part of the cycle. The lithofacies of the cycle include halite with or without potash salts, anhydrite, silty dolomite, and black shale (Figure C.40). The halite and anhydrite units of each cycle are entirely a chemical facies; however, the dolomite and black shale are both chemical and clastic, and contain organic material.

The sequential order of chemical precipitates in a typical Paradox cycle is determined by the degree of solubility of the rock-forming mineral in sea water. If a cycle is defined on the basis of the chemical sequence of rocks deposited, as sea water undergoes a gradual increase and a decrease in salinity, the sequence is (A) silty calcareous dolomitic argillaceous

FIGURE C.40. Typical Evaporite Cycle of Paradox Member. Modified from Hite, 1969.

organic--rich black shale, (B) silty dolomite, (C) nodular and laminated anhydrite, (D) halite with or without potash salts, (C) anhydrite as before, (B) dolomite as before, and (A) black shale as before. In this sense the cycle can be termed a chemical cycle and the lithologic sequence of units shows mirror-image symmetry as A, B, C, D, C, B, A. In detail, however, the Paradox chemical cycle is not symmetrical. At the base of the cycle, units A, B, C, and D are deposited as the result of gradually increasing salinity, and each unit shows a gradual transition with the overlying unit. The halite unit (D), however, neither shows a gradual decrease in salinity nor grades into the overlying anhydrite unit (C). Studies by Raup⁽⁸⁷⁾ of the bromine content of halite in the Paradox Member have shown that salinities gradually increased all the way to the top, or near the top of each salt bed. Further evidence of asymmetry can be seen in those cycles that are potash bearing. The potash deposits, which represent the period of highest salinity in the cycle, usually occur at the top of the halite instead of the middle.⁽⁸⁸⁾ A strongly asymmetrical distribution of detrital quartz through the cycle can also be noted (Figure C.40). The lower dolomite unit (transgressive phase) contains nearly twice as much quartz as the upper dolomite unit (regressive phase).

Boundaries of Paradox cycles were at first defined on the basis of rock units, generally the black shales, but recently it was recognized that a more workable system of cycle boundaries can be outlined if unconformities are used.⁽⁸⁹⁾ In the evaporite cycle the basal contact of each halite unit is transitional with the underlying anhydrite, but the upper contact of the halite unit is sharp. This sharp contact, at the top of the halite unit, is a dissolution surface or unconformity along which several feet of halite have been removed. These unconformities, which interrupt the chemical cycle, result from major incursions of sea water that reduce salinity in the evaporite basin. Using the unconformities as cycle boundaries, the order of units would be C, B, A, B, C, D. The change from transgressive to regressive conditions occurs somewhere near the mid-point of unit A so that the transgressive hemicycle is C, B, A, and the regressive hemicycle is A, B, C, D. Comparison of the transgressive and regressive hemicycles shows considerable condensation of the sedimentary sequence in the transgressive phase. Transgression appears to have been more rapid than regression.

The evaporite cycles of the Paradox Member show a lateral change of facies, similar to the vertical change, from halite through anhydrite to dolomite rock. The lateral change is particularly well demonstrated from northeast (halite) to southwest (dolomite) in the basin.

The anhydrite rock of a Paradox evaporite cycle consists of material that is finely laminated or nodular in appearance. Fine-grained dolomitic mud is present as intervening laminae or as the matrix between anhydrite nodules.

The rock is very dense and impermeable except where fractured. Quantitatively, anhydrite is the least abundant constituent of the cycles.

Dolomite in the Paradox cycle is typically fine grained and may contain as much as 50% by weight of well sorted quartz silt. The rock appears relatively impermeable, but brine may be seen oozing from newly drilled cores in dolomite units. The rock is tough and dense, and where encountered in underground mine workings it generally stands without support. Some flows of brine and/or oil and gas have been encountered in these rocks where they are intensely fractured.

The halite rock of the Paradox Member is essentially the same as the halite of other marine evaporite deposits. The rock is composed of an intergrowth of halite crystals that range from 2 mm to 50 mm in diameter. Average crystal size is about 5 mm. The rock is generally a smoky gray except where trace amounts of potash minerals, which generally contain trace amounts of hematite, color the rock shades of tan or red. Halite rock in the Paradox Basin generally contains no more than 2 or 3% of minerals insoluble in water. Nearly all this insoluble fraction consists of anhydrite, although there may be trace amounts of dolomite, quartz, clay, and talc. The anhydrite occurs in thin, discrete laminae, darker colored bands with indefinite boundaries, and as small euhedra scattered through the halite matrix. More than half of the halite units in the evaporite cycles contain potash deposits. The potash minerals include sylvite (KCl) and carnallite ($\text{KMgCl}_3 \cdot 6\text{H}_2\text{O}$). The economics of these deposits as related to waste disposal sites is treated later in this report. Paradox halite contains some moisture in the form of tiny brine inclusions. A small number of measurements show a range of 0.144 to 0.093% water by weight. In addition to brine, inclusions of hydrocarbons have also been noted. These, however, have been seen only in the coarsely crystalline halite of cycle 10.

The black shale has the most complex lithology of the cycles. About 20 to 30% by weight is carbonate, divided about equally between dolomite and calcite. The remainder of the black shale is clay-size detritus consisting of quartz, feldspar, various clay minerals, and locally as much as 15% organic matter. These shales, like many other black, organic-rich rocks, contain sufficient uranium to create radioactivity anomalies on gamma ray logs. By using these logs, a precise basinwide correlation of these units can be made. The rock is characterized by high porosity and low permeability except where fractured. Fracturing is widespread and can be observed in all cores of these units. In the Cane Creek potash mine, it was necessary to set steel arch supports in all workings where these rocks were encountered. Fractures and joints are nearly always filled with halite and less commonly with potash minerals. The role these fracture systems play in accumulation of hydrocarbons or brine will be detailed in a later section of this report.

Post-Paradox Rocks

Normally the Paradox Member of the Hermosa Formation is overlain by an unnamed upper member of fossiliferous limestone of Middle Pennsylvanian age. But in numerous parts of the Paradox Basin the normal sequence has been changed locally during the development of salt anticlines. In many places, even in the locality of Paradox Valley, outcrops of the Paradox Member include caprock material that may have developed as late as Quaternary time. The distribution and the sequence of post-Paradox rocks vary considerably from one salt anticline to another.

Most of the post-Paradox formations wedge out locally. This results from the fact that many of the salt anticlines were uplifted in a series of pulses, so that some formations either were not deposited over the rising structures or were removed by erosion before deposition of the next younger unit. In southwestern Colorado, the Morrison is the oldest formation to have completely or almost completely blanketed the salt anticlines (Cater, 1970). In eastern Utah, the Morrison again is the oldest formation to have covered the Salt Valley anticline.

Aside from alluvium, the chief post-Paradox strata that overlie the Paradox Member in the salt anticlines are the upper member of the Hermosa Formation, and the Rico and Cutler Formations--all of Paleozoic age. None of these units is an actual or potential aquifer for many miles in any direction.

Structure

The depth of the salt varies considerably throughout the Paradox Basin. Along the southwestern and western edges of the basin the combination of the structural rise over the Monument upwarp and deep dissection by the Colorado and San Juan Rivers exposes gypsum beds of the Paradox Member. Halite is present in the subsurface a few miles downdip from these outcrops. In this same general area drill holes on the axis of the Elk Ridge anticline have penetrated salt at depths of 775 to 930 m. The only other outcrop of the Paradox Member around the edge of this basin is along the southeastern edge about 16 km north of Durango, Colorado. Here again only gypsum beds are exposed, but halite-bearing rocks are present in the subsurface about 24 km southwest of this locality. Halite-bearing rocks may be present downdip a short distance from the outcrop, but less than 3.2 km from the outcrop the Paradox Member is deeply buried beneath towering cliffs of Mesozoic sedimentary and Tertiary volcanic rocks. Along the northeast margin of the basin, bordering the Uncompahgre uplift, the salt-bearing rocks of the Paradox Member are faulted down against the Precambrian core of the uplift and covered by a thick wedge of coarse Permian clastics. The top of the salt in this area is from 4340 to 4650 m below the surface. In the rest of the basin, depths to the salt average about 1860 m, except in the salt anticlines where locally it is brought within 155 m of the surface. Generally speaking, the salt will be

at favorable mining depths only within those areas where the Paradox Member or the upper limestone member of the Hermosa Foundation is exposed.

Salt Anticlines

The dominating structural features in the Paradox Basin are the salt anticlines which occupy a rectangular shaped area about 80 km wide and 192 km long on the northeast side of the basin. This area coincides with the Uncompahgre trough (Figure C.38), which is also the area of thickest salt accumulation.

The salt anticlines can be divided geographically and structurally into five major systems. Each system consists of a long, undulating welt of thickened salt over which the younger rocks are arched in anticlinal form. Surface structural expression is more pronounced along certain segments of each anticlinal system, and these have been given individual names. The anticlines are commonly breached along these segments, forming prominent flat-floored valleys. The major salt anticline systems are 1) Lisbon Valley-Dolores Valley, 2) Moab Valley-Spanish Valley-Pine Ridge, 3) Gypsum Valley, 4) Castle Valley-Paradox Valley, and 5) Salt Valley-Cache Valley-Fisher Valley-Sinbad Valley (Figure C.38). Several smaller anticlines not included in the five major chains lie southwest of Moab along the Colorado River: Cane Creek anticlines, Shafer dome, Lockhart anticlines, Rustler dome, Gibson dome, and Meander anticline.

As the result of geophysical studies, detailed surface mapping, and study of data from numerous drill holes, the gross form of the Paradox salt anticlines is fairly well known. A typical anticline can be considered as three structural components: the floor of nonevaporitic rocks beneath the anticlines, the central core of halite and its associated caprock, and the overlying or flanking beds of nonevaporitic rocks. These components show considerable difference in structural form. These differences are due to rheological contrasts and also to the influence of one component on another. The floor component, influenced by the underlying rigid basement complex, was deformed primarily by block faulting and strike-slip faulting. The central core, which is plastic, was deformed almost independently of the floor or the overlying component. The floor and overlying or flanking strata, although similar rheologically, bear little structural resemblance because they are separated by the thick plastic central core, which absorbs most of the stress transmitted from the crystalline basement.

The floor of a Paradox salt anticline consists of a relatively thin covering (465 to 775 m) of alternating units of limestone, dolomite, shale, sandstone, and quartzite resting on the crystalline Precambrian basement. The sedimentary rocks range in age from early Pennsylvanian to Cambrian. Pre-Paradox strata dip to the northeast under the salt anticline area at about 48 m/km. Through much of the area the beds are relatively undisturbed except

for gentle flexing. Locally, however, deep drilling has shown that pre-Paradox strata are cut by normal faults beneath some of the salt anticlines. The maximum known relief on one of these faults is 1742 m⁽⁹⁰⁾ on the southwest flank of the Paradox Valley anticlines (Figures C.38 and C.41). As diagrammed in Figure C.41, the pre-Paradox strata were thinned by erosion before the deposition of salt. Fault-block structures played an important role in the origin of the salt anticlines. Locally, they may form important traps for petroleum, as they do in the Lisbon Valley anticline (Figures C.38 and C.42).

The central cores of the salt anticlines range in thickness from about 775 to 4340 m. They consist of 70 to 80% halite rock with associated potash minerals; the remainder is made up of marker beds of anhydrite, dolomite, and black shale. These marker beds, which may be up to 65 m thick, impart a confining strength to the halite layers so that the structural behavior of these central cores is different from the behavior expected of an unconfined homogeneous mass of halite. Also, the marker beds permit mapping of the structural detail of the central core, if drill hole records are available.

The structure in the central core of a salt anticline contrasts sharply with the structure of the underlying and overlying nonevaporitic rocks. The core structure is the result of both regional compression and flowage of the Paradox Member into the anticlines. The amount of flowage in an anticline is a function of three factors: 1) thickness of the original salt layer, 2) degree of differential loading on the flanks of the anticline, and 3) whether or not the salt core was confined throughout its history by younger rock layers. Of these three factors the last was probably the most important. Locally there were areas along the axes of anticlines nearest the Uncompahgre uplift where confining younger strata were missing. This allowed large-scale migration of Paradox strata toward and up through these holes in the sediment cover to form diapiric anticlines.

The migration of Paradox strata from flanking synclinal areas to the salt anticlines involved movement of both halite layers and the intervening marker beds as a single mass. This contrasts with the Gulf Coast salt domes in which nonchloride beds are generally absent except for an occasional erratic block.

Contrasting degrees of structural complexity are evident between the diapiric and nondiapiric anticlines. Nondiapiric anticlines, such as Cane Creek (Figures C.38 and C.43), Shafer dome, and Gibson dome, are characterized by relatively simple disharmonic folding and minor faulting. Some of the nondiapiric anticlines such as the Lisbon Valley anticline (Figure C.42) are considerably more complex, but the stratigraphic identity of the evaporite cycles is still preserved. Recumbent folds and low-angle reverse faults are typical structures in the salt core of the Lisbon Valley anticline. In one

FIGURE C.41. Cross Section through Paradox Valley Anticline, Trace of Section Shown on Figure 1. Adapted from Cater, 1970.

FIGURE C.42. Cross Section through Part of Lisbon Valley Anticline, Utah; dark zones in Paradox Member represent clastic-penesaline marker beds; intervening units, several of which are numbered, are salt beds.

FIGURE C.43. Cross Section through Cane Creek Anticline, Utah; dark zones in the Paradox Member represent clastic-penesaline marker beds; intervening units, several of which are numbered, are salt beds. Hite, 1968.

drill hole near the crest of this structure the same bed was repeated seven times by recumbent folding. Detailed studies of folding in the Paradox Member have been made by Evans and Linn⁽⁹¹⁾ in mine exposures in the Cane Creek anticline. These authors classified the orders of folding in this structure on the basis of periodicity (p. 290). The anticline itself was considered a first-order fold. Folds in the marker beds were considered as second order, and the folds involving thin anhydrite laminae in the halite and potash units were given third, fourth, and fifth order ranking. As shown on their mine maps, the second order folds have wave lengths of as much as 130 m, although the average is about 65 m. Amplitudes of up to 23 m were noted on some of these folds. The third, fourth, and fifth order folds have wave lengths and amplitudes of a few meters or centimeters. The second-order folds are most significant to any type of underground mining operation where it is necessary to avoid penetration into the marker beds above or below a particular salt bed.

As might be expected, the structure in the central core is most complex in the diapiric salt anticlines. Although all these anticlines have been drilled, structural interpretation in the central cores of the anticlines is somewhat hypothetical because it is impossible to recognize stratigraphic units of the Paradox Member with any degree of certainty. Identification is difficult because drill penetration of marker beds is at very low angles and these beds thicken rapidly as they approach the ancestral Uncompahgre uplift.

The marker beds in the cores of the anticlines are believed to be deformed into a series of nearly vertical isoclinal folds whose axes parallel the northwest trend of the salt anticlines. Accurate knowledge of the geometry of these folds will come only through extensive drilling, but some reasonable interpretations can be made on the basis of available information. For example, it is known from drilling that complete removal of salt by flowage has occurred in the Dry Creek syncline on the southwest flank of the Paradox Valley salt anticline (Figure C.41). The salt flowed into the Paradox Valley anticline and also into the Gypsum Valley anticline, the next anticlinal system to the southwest. When flowage from the synclines into the anticlines began, it is likely that a nearly equal volume was transferred to each of the adjacent anticlines. If so, then each marker bed might be expected to pull apart somewhere near the midpoint of the syncline. With a complete transfer of salt, the associated marker beds would be subjected to a high degree of crustal shortening. In Figure C.44, a marker bed is shown in the position it might occupy in the Paradox Valley anticline after flowage from the adjacent Dry Creek basin syncline. The original length of the bed, measured from the midpoint of the syncline, would have been about 9.1 km. After flowage the same bed might have been shortened by folding so it would occupy a horizontal distance of only 3.2 km. The depicted bed has a thickness of about 32 km, about average for the area. According to Currie et al.,⁽⁹²⁾ the wave length

FIGURE C.44. Diagrammatic Cross Section through Paradox Valley Anticline Showing the Shortening of Marker Bed in Paradox Member Caused by Flowage from Adjacent Syncline. A to A' represents original length of marker bed, about 5.7 miles; B to B' represents horizontal distance, about 2 miles, occupied by the bed after transfer by flowage from Dry Creek Basin syncline; reduction of original length by 65 percent is accomplished by series of vertical to near vertical folds with amplitudes of 1,000 feet or more.

of folds in a multi-layer sandwich of competent and incompetent layers is dependent on the thickness of the competent layers. Using that concept, a 32-m thick marker bed should fold with a wave length of about 1145 m. This means that a 32-m-thick marker bed should fold eight times in the distance A-A' (Figure C.44). Continued flowage of salt and marker beds into the anticline would eventually shorten the bed to the distance B-B' and decrease the wave length of the folding to about 400 m. With this wave length and degree of shortening, the folds could have amplitudes of 310 m or more. Structures observed in caprock and in drill holes on some of the salt anticlines suggest that the fold geometry described above may closely resemble the actual structures.

Dissolution of halite from the upper surface of the central core has locally developed a caprock of insoluble material along the crestal parts of the salt anticlines. This caprock differs somewhat from the Gulf Coast variety. Both originated as the result of solution of halite; however, the Gulf Coast caprock was formed entirely from anhydrite occurring in the halite as thin laminae and small disseminated crystals. The Paradox caprock includes this type of material and also the marker beds whose lithologies were previously described. All the anhydrite in this residuum has been converted to gypsum. In contrast to the Gulf Coast, there has been no microbial reduction of sulfate and development of a limestone layer.

Caprock in anticlines such as Cane Creek or Lisbon Valley results from the solution of only a few hundred meters of salt. This took place prior to deposition of overlying shales, limestones, and thin anhydrite units of the upper member of the Hermosa Formation. These beds of the upper member are essentially impervious and have prevented further solution of halite over these structures for nearly 250 million years. The protective beds of the upper member were not deposited across such diapiric structures as the Paradox Valley and Gypsum Valley anticlines, and as a result those structures have experienced a long history of salt solution that may have removed as much as several hundred meters of halite. In both types of anticlines the maximum development of caprock occurred at the point of greatest uplift of the central core.

In areas where caprock crops out, and weathering has not been too severe, it is possible to map certain lithologic units, particularly black shales. Continuity of individual units in the caprock may sometimes exceed a kilometer. Structures in the caprock are probably not too dissimilar from the original form when these rocks were still interbedded with halite. A slight modification probably took place as the result of volume change following the conversion of anhydrite to gypsum. These caprocks are porous and locally, in anticlines such as Paradox Valley, they are saturated with brine.

The strata overlying the halite cores of the salt anticlines have a structural expression which differs greatly from the other two components of the anticlines. The difference is due to rheological contrasts and a set of forces not involved in the deformation of the other two components.

The rocks in this uppermost component consist of massive sandstone units, siltstones, shales, and locally, limestones. These units have a combined strength greatly exceeding that of the underlying plastic-like evaporite core. This component ranges in thickness from 1550 to 4650 m. Extreme variations in thickness are due mostly to differential rates of sedimentation in the Cutler Formation of Permian-Pennsylvanian age. The form of folding in this component consists of a gentle arching across the axis of each salt anticline. Dips in beds exposed along the flanks of nondiapiric structures seldom exceed 10 degrees. In the diapiric anticlines, particularly in the Paleozoic rocks, dips may exceed 30 degrees. The arching of these strata over the salt anticlines is the result of an initial horizontal compressional force that later gave way to a vertical force stemming from the upward movement of the central core of evaporites. These structural arches are now broken by high-angle normal faults across which some blocks of Mesozoic strata have been dropped as much as 620 m. This faulting is due to collapse caused by dissolution of the underlying evaporites. Various authors have shown many of these faults extending into the Paradox Member. It is true that some may extend into Paradox caprock, but because they are the result of solution of halite at the caprock-halite interface they cannot extend into the salt itself. The effects of the salt removal and subsequent collapse are particularly evident near the ends of salt anticlines such as Moab Valley and Paradox Valley (Figure C.38). There, Mesozoic strata, although highly faulted and fractured, are continuous across the evaporite core and form a conspicuous downsag. Locally, in some of the diapiric anticlines, isolated blocks of younger rocks appear to have sunk into the Paradox Member. Cater⁽⁹³⁾ suggests that this collapse may be due to isostatic adjustments between the heavier downsagged blocks and the lighter evaporite core.

Geologic Persistence of Paradox Salt Deposits

It has been specified that, for safe waste disposal, an underground repository should remain essentially intact for a long period of time, perhaps about 250,000 years. Despite their highly soluble nature, rock salt deposits show great geologic persistence, as indicated by the survival of extensive salt deposits from every geologic period since the Cambrian. The nature of most salt deposits is such that, unless disturbed by drilling or mining, solution takes place only along the upper surface. Generally, this surface is protected from groundwater by an impervious layer of clay or anhydrite. Good examples are the salt deposits of Cretaceous age in northeastern Thailand. Despite the fact that rainfall in this region commonly exceeds 150 cm/year,

the salt deposits have survived even though they locally occur at depths of less than 65 m.⁽⁹⁴⁾ Even where groundwater is in direct contact with salt, dissolution is effective only where hydrologic conditions permit the escape of saturated brine.

The permanence of the Paradox salt deposits can be estimated by calculating present dissolution rates from the salt load carried by surface drainage in the basin, and by calculating the average dissolution rate since the salt was deposited, based on the volume of present-day caprocks. It is known that the only areas of salt removal, past and present, are along the western and southeastern edges of the salt basin and the crestal parts of the salt anticlines. At most, downdip dissolution of salt along basin edges has probably advanced no more than 3 or 5 km since the close of the Pennsylvanian period (280 million years). Dissolution from the edges of the salt basin should, therefore, pose no threat to sites well within the basin margins.

Some idea of the rate of dissolution of salt from the crests of salt anticlines may be obtained from measurements of the discharge rate and chloride content of the Dolores River, the Colorado River, and several tributaries of the Colorado that drain structures in which salt bodies are exposed. Five of the larger salt anticlines of the Paradox Basin--Gypsum, Lisbon, Paradox, and Sinbad Valleys and the Dolores anticline--are drained by the Dolores River (Figure C.38). At a gaging and sampling station 14.4 km above the mouth, the Dolores carried an average of 480 tonnes per day of sodium chloride during the 16-year period of record--1954 through 1969.

The remaining salt anticlines of the Paradox Basin are drained by the Colorado River, but the determination of the additional salt load picked up by the Colorado below the mouth of the Dolores is hampered by lack of records. For only one year, 1952, is it feasible to determine both the load carried by the Dolores River (measured at Gateway, Colorado), and the net load added to the Colorado below the Dolores (between Dewey Bridge and Hite, Utah).

A total of 620 tonnes per day includes all the common salt for 1952 from all salt anticlines, including the Meander anticline in Cataract Canyon.

The chloride content of the Dolores River is derived in part from dissolution of halite from the Gypsum Valley, Paradox Valley, and Sinbad Valley anticlines. Two other anticlines, the Dolores and Lisbon Valley, are also drained by the river, but the Paradox Member is not exposed in these structures, and there is no evidence of recent salt removal. Solution of halite is probably taking place from the first three anticlines over a combined surface area of about 340 km². This area of active salt removal is obtained by measuring the prominent collapsed parts of each anticline. The average yearly load of nearly 180 thousand tonnes of sodium chloride carried by the Dolores River represents removal of halite from an area of 340 km² at a rate of about 0.28 mm/yr. If

this rate of removal continued over a period of 1 million years, 275 m of halite rock would be stripped from each of the three anticlines. This calculated rate of removal is probably excessive because: 1) it does not consider that a sizable percentage of the sodium chloride carried by the Dolores River may be derived from rocks younger than the Paradox Member and 2) salt is probably being removed from a much wider area than that on which surface collapse has become conspicuous.

The caprock of the salt anticlines provides another means of estimating rates of salt removal. The average thickness of caprock over these structures is about 310 m. Allowing for the increase in volume that results when anhydrite is converted to gypsum, the present 310 m of caprock would represent about 230 m of residual nonchloride material after the dissolution of salt from the anticlinal core. Because the original core of these anticlines averages about 25% nonchloride material, 230 m of residual nonchloride material would remain after the dissolution of salt from about 930 m of halite-bearing rock. At the rate of 275 m per million years, it would take 3.3 million years to remove this much halite, suggesting that the caprock started developing in the early Pliocene. The age of the caprock is unknown. Cater⁽⁹³⁾ has stated "Collapse of the crests of the salt anticlines occurred in two stages apparently widely separated in time. The first stage followed perhaps rather closely the Late Cretaceous folding. The second stage followed epeirogenic uplift of the entire Colorado Plateau in the middle and late Tertiary, and this stage is still continuing." If the caprocks as we see them now began forming at the close of the Cretaceous, about 65 million years ago, then the removal rate of halite would have been about 14 m per million years. If accretion of the caprocks did not begin until the epeirogenic uplift of the Colorado Plateau, which according to Hunt⁽⁹⁵⁾ began during the Miocene, about 20 million years ago, then removal of salt proceeded at the rate of 46 million meters per million years. The question of caprock age is further complicated by data from drill holes on the crests but outside the collapsed parts of the anticlines. One drill hole on the southern end of the Paradox Valley anticline (Petroleum Production Board Government 1 in sec. 19, T. 45 N., R. 15 W., Colorado) penetrated the Cutler Formation underlain by 290 feet of Paradox caprock. Another drill hole on the northern end of the Salt Valley anticline (Defense Plant Corp., Reader 1, in sec. 4, T. 22 S., R. 19 E., Utah) penetrated the Morrison Formation underlain by 200 m of Paradox caprock. At Lisbon Valley anticline a thin caprock is overlain by the upper member of the Hermosa Formation, about 250 million years old. These and other relationships between caprock and overlying strata of Mesozoic and Paleozoic age suggest that at least part of the caprock over the Gypsum Valley, Paradox Valley, and Sinbad Valley anticlines may have developed during Permian time and has probably received continuous additions since that time. Considering then that

a caprock with a present-day thickness of 310 m might be the result of dissolution of halite through a period of 250 million years, the rate of halite removal might be as slow as 0.9 m per million years.

In summary, the exact rate of halite removal from the Paradox salt anticlines is difficult to determine, but a range of 0.9 to 275 m per million years is suggested. Most of the available evidence suggests that the actual rate lies within the lower half of that range.

Oil and Gas in the Paradox Member

Oil and petroleum gases, primarily methane, are found in the Paradox Member by almost every well drilled in the Paradox Basin. These occurrences constitute a hazard to any underground mining operation. They range from trace amounts to high pressure-high volume flows which have on occasion resulted in spectacular blowouts during drilling. Generally these flows dissipate in a matter of a few hours or days. A few wells, however, have been completed for commercial production from these zones. One well on the west flank of the Cane Creek anticline has produced nearly one million barrels of oil over a period of 10 years. The oil and gas is generally limited to marker beds, although small inclusions of liquid hydrocarbons are occasionally noted in drill cores of halite. The marker beds in the Paradox Member have very low permeabilities, and as a result significant accumulation of hydrocarbon is found only where the beds are intensely fractured.

In the Cane Creek potash mine of Texas Gulf, Inc. near Moab, the potash ore lies only a few meters beneath a marker bed. Roof bolts extending up into the bed were commonly sites of hydrocarbon leakage. To safeguard against entry of high-pressure flows into the marker bed, any large accumulation of oil and gas is allowed to bleed off before mining operations proceeded. Unfortunately, the proximity of potash ore and marker bed combined with folding of these beds brought about an occasional intersection of mine workings and marker bed, necessitating all the precautions required in the operation of a gassy mine.

Of all the structures in the Paradox Basin, Salt Valley anticline and Shafer dome are perhaps the best suited for further detailed investigation. Both have favorable, although somewhat different, geologic characteristics and are on or near a railroad. The proximity of both sites to cultural or industrial developments should not be a major disadvantage. The Salt Valley site is bordered on the south by the Arches National Park, but disposal facilities could be located out of view, and at least 8 km from the boundary. The Shafer dome site presents a similar situation. The Dead Horse Point overlook, a Utah State park, is on the canyon rim above the Shafer dome site, but the topography of the area could be used to hide any surface facilities of an emplacement site. The area is already scarred by large evaporation ponds

fully visible from the overlook. The Shafer dome site is not far from the Cane Creek potash property of Texas Gulf Co., Inc. As the potash mine has been flooded and converted to a solution mining operation, there are no men working underground in the area. A site could be located that would be at least 7 km from the nearest underground workings of this mine.

C.2.6. Shales-Rocky Mountains and Interior Plains

The Rocky Mountain system and Interior Plains contain thousands of square kilometers of structurally undeformed Mesozoic rocks in areas of low population density and seismic stability. The regions are partly underlain by thick sequences of Cretaceous shale which have been described in many reports. Except for the possible problems associated with a high content of montmorillonite in the presence of heat, these rocks seem well suited for waste storage or disposal.

As much as 5890 m of alternating marine and nonmarine rocks of Late Cretaceous age were deposited in a $1,500,000\text{-km}^2$ area in the Great Plains and Rocky Mountain states (Figure C.45). Gilluly⁽⁹⁶⁾ estimated that about $4 \times 10^{16} \text{ m}^3$ of sediment were delivered to this area. Of this volume, as much as $2 \times 10^{16} \text{ m}^3$ is estimated to consist of marine clay, silt, and mud. These fine-grained sediments are now represented by the Colorado and Graneros Shales, the Benton Formation, and the Mancos, Carlile, Pierre, Steele, Claggett, Bearpaw, and Lewis Shales (Figure C.46).

The Pierre Shale and its lateral equivalents, the Bearpaw, Claggett, Lewis and Steele Shales, are representative of most of the Cretaceous Shales of the region and are therefore selected for brief discussion. The Pierre (Figure C.47) has a volume of about $8 \times 10^{15} \text{ m}^3$, an outcrop area of about $234,000 \text{ km}^2$, and is concealed beneath younger strata in an area of about $962,000 \text{ km}^2$.⁽⁹⁷⁾ It consists of thick sequences of claystone, shale, bentonitic mudstone, and many thin beds of bentonite. It ranges from less than 155 m thick in the eastern Dakotas to more than 1550 m thick in southeastern Wyoming and central Colorado. The Pierre is largely devoid of aquifers and, although highly porous locally, is essentially impermeable and does not yield water to wells. The moisture content as determined from core samples decreases with depth of burial; it is as much as 35% in near-surface weathered beds and as little as 15% where deeply buried.⁽⁹⁸⁾

The mineralogy of the Pierre Shale is probably typical of the clayey Cretaceous formations in the Great Plains and Rocky Mountain regions. Tourtelot, Schultz, and Gill⁽⁹⁹⁾ report that most samples of the Pierre from the Black Hills and Missouri River areas consist of 65 to 80% clay minerals, 15 to 25% quartz, a few percent feldspar, and small amounts of calcite, dolomite, biotite, pyrite, gypsum, jarosite, clinoptilolite, and organic material. The clay minerals consist of 25 to 45% montmorillonite, 35 to 45% mixed-layer illite-montmorillonite, 15 to 25% illite, and about 5% each of kaolinite and chlorite.

Many areas underlain by the Pierre would invite further study if the montmorillonite in the shale could be tolerated. For example, near the northern end of the Black Hills in Carter County, Montana, the outcropping Pierre is about 650 m thick, generally impermeable, and has a gentle dip (Figure C.48).

— LIMIT OF UPPER CRETACEOUS DEPOSITS	— ACCEPTED THICKNESS, IN FEET
---- LIMIT OF LITHOLOGIC FACIES	---- PARTLY INFERRED THICKNESS, IN FEET
	WHOLLY INFERRED THICKNESS, IN FEET

FIGURE C.45. Thickness of Upper Cretaceous Deposits, Rocky Mountain System and Interior Plains. From Reeside, 1944, Map 2.

FIGURE C.46. Correlation Chart of Upper Cretaceous Formations of Great Plains and Rocky Mountains. Vertical lines indicate depositional hiatus. Slanted lines indicate nonmarine shale and less chalky shale and limestone. Modified from Cobban and Reeside, 1952.

FIGURE C.47. Distribution and Thickness of Pierre Shale and Equivalent Rocks in the Western Interior of the United States. From Tourtelot, 1962.

FIGURE C.48. Generalized Sections Showing Relations of Pierre Shale and Equivalent Rocks in South Dakota, Wyoming, and Montana. Modified from Tourtelot, Schultz, and Gill, 1960, Figure 205.1.

In this area the seismic risk is minor and holes drilled for oil and gas are sparse. The thick shale sequences of the Pierre do not yield water except in the uppermost 6 to 10 m of the formation, where a small amount is obtained from a few wells. The few clayey and silty sandstones in the formation contain minor amounts of water. The calcareous shale of the Niobrara Formation which directly underlies the Pierre may contain a small amount of water at a few localities. The Cretaceous Fox Hills Sandstone and Lance Formation, and the Tertiary Fort Union Formation, which overlie the Pierre, are a multiple aquifer system. Water wells that penetrate this system are common in the area and yield as much as 1900 liters/min. The water in all aquifers of the area is under high artesian pressure.

The Bearpaw Shale is the westward equivalent of the upper part of the Pierre Shale in Montana and southern Canada. The 29 samples from the Bearpaw analyzed by Scott and Brooker⁽¹⁰⁰⁾ contain an average of 5% sand, 41% clay, and 54% silt. The clay fractions contained an average of 77% montmorillonite, 18% illite, and 5% chlorite. From these and other data it seems unlikely that thick sequences of clayey Cretaceous rocks containing less than 10% montmorillonite can be found in this region.

C.2.7. Shales-Eastern U.S.A.

In the eastern half of the United States, three thick sequences of shale and clay appear to be suitable for final underground storage of radioactive wastes: (1) Late Ordovician shales in Ohio and New York, (2) Devonian-Mississippian shales in Ohio and Michigan, and (3) Paleocene clays in Mississippi, Louisiana, and Arkansas (Figures C.49 and C.50).

Physical and chemical properties of shales that make them desirable as host rocks for underground waste emplacement include their low permeability, relatively high plasticity, large ion-exchange capacity, and the distribution of these favorable properties over large areas throughout great thicknesses of some shale sequences. A shale thickness of at least 150 m at a depth of 300 to 900 m below the surface is considered optimum for ensuring the isolation of waste material after burial.

A recent study of caverns mined at depths of 60 to 550 m for storage of petroleum products in the United States⁽¹⁰¹⁾ showed that shales are an almost ideal host rock for mined storage caverns. The locations of mined storage caverns in the eastern United States including the 30 sites with one or more caverns formed in shale units all shown in Figure C.50. Even the two deepest caverns in shale (420 and 550 m) have been mined and operated without water or structural problems.

Shale Deposits of the Appalachian Highlands

The Appalachian provinces contain thick sedimentary deposits of Paleozoic age, but the most favorable deposits are in the Appalachian Plateau. Some shales in the eastern part of the Interior Plains dip into and thicken beneath the Appalachian Plateau, and they are discussed here rather than with "Shale Deposits of the Interior Plains." Illite and chlorite are the dominant clay minerals in most shales of the region, and montmorillonite is rare. Principal thick shales are in the Late Ordovician and Late Devonian-Early Mississippian sequences.

Late Ordovician Shales in Ohio

In central and eastern Ohio, a thick and undeformed Late Ordovician marine-shale sequence dips gently eastward at about one half degree. The unit is generally 270 to 300 m thick and embraces the Eden, Maysville, and Richmond Stages of the Cincinnati Series (Figure C.51). Colton⁽¹⁰²⁾ shows that the unit thickens to 550 to 600 m at the Ohio-Pennsylvanian state line. At and near the outcrop in western Ohio the sequence is gray to black, argillaceous limestone interbedded with calcareous shale. Farther east, away from the Cincinnati and Findlay arches, the limestone layers are thinner and grade into shale, so that in central and eastern Ohio the sequence is largely shale. The eastern boundary of important limestone beds is uncertain and must be

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FIGURE C.49. Physiographic Provinces of Eastern United States and a Map Showing Distribution of Thick Shales Favorable for Storage of Radioactive Waste in Eastern United States; Also Shown are Location of Underground LPG-Storage Caverns Excavated in Shales and Other Rocks. Modified From U.S. Geological Survey national atlas, 1970.

FIGURE C.51. Depth to Top of Late Ordovician Shales in Ohio and New York and Parts of Adjacent States; Thickness of Shale Sequence Ranges from About 300 m Near Outcrop to About 600 m Near 1500 m Depth Isopach Line. Modified from Colton, 1961; Miller, 1975.

established in any subsequent study of the region. In northeastern Ohio, the sequence is similar to that of northwestern Pennsylvania (Figure C.51), and it embraces the Utica black shale, overlain by the Reedsville Shale.

In Ohio, the Late Ordovician shale sequence is underlain by the Trenton Limestone (30 m thick), which is in turn underlain by other carbonates of Ordovician and Cambrian age. The shale sequence is overlain by the Queenston Shale, about 30 m thick, which is overlain by the Clinton sandstones and shales of Silurian age.

The top of the shale sequence is 300 to 900 m below land surface in a long north-south belt about 65 km wide and extends nearly 320 km across central Ohio (Figure C.52). Just to the east is another belt with similar dimensions, where the depth to the top of the shale sequence is 900 to 1500 m; still farther east the shale is more than 1500 m deep.

Principal rivers in the region are the Sciota River in central Ohio and the Muskingum River in east-central Ohio; both rivers flow south and join the Ohio River. Lake Erie is at the north edge of the region of interest.

The Late Ordovician shale sequence has a low permeability, and its great thickness would form a thick impermeable seal above and below storage caverns placed near the middle part of the sequence.

Oil and gas are not produced from the Late Ordovician shale sequence, and the unit is not known to contain any other mineral resources. Most of Ohio's oil and gas production is from younger rocks in eastern and east-central Ohio, and the only production from older (underlying) rocks is from the Trenton Limestone in northwestern Ohio and from Early Ordovician and Cambrian carbonates at a few places in north-central Ohio. Since most production is from overlying strata, not too many boreholes have penetrated the Late Ordovician shales.

Three hydrocarbon-storage caverns have been excavated at depths of 120 to 175 m in the Late Ordovician shale sequence in Ohio.⁽¹⁰¹⁾ They are in Allen, Butler, and Lucas Counties, in the western part of the state. The averages of compressive-strength tests of materials from all three sites ranged from about 200 to 300 atm. It is not certain if these analyses are representative of the in situ compressive strength of the units. All three caverns have operated successfully since 1959-70, without any evidence of water inflow or structural failure. In two of the caverns the shale was subject to air slacking.

Younger rock units have been mined underground at a number of places in the region: coal mines are common in the Pennsylvanian rocks of eastern Ohio; 2 salt mines are in Silurian rocks in Cuyahoga and Lake Counties; and 5 brine-well systems are in Lake, Summit, Meigs, and Wayne Counties.

FIGURE C.52. Map Showing Principal Tectonic Features in Eastern United States. Modified from U.S. Geological Survey National Atlas, 1970.

The region lacks structural deformation, and all of it is in seismic-risk zone 1 (Figure C.52). Several small earthquakes (MM V-VI) have occurred in the Cleveland area, and several small earthquakes and one larger earthquake (MM-VII-VIII) have occurred in southeastern Ohio.

The Late Ordovician shale sequence in central and eastern Ohio appears highly favorable for underground waste storage because:

- The sequence consists of 300 to 600 m of impermeable shale, with some limestone (mainly to the west).
- It is at favorable depths over a large area.
- Much of the area is rural and agricultural lands.
- The sequence contains no known commercial mineral deposits.
- Few boreholes have been drilled through the sequence.
- The sequence has not been greatly deformed or otherwise fractured.
- The region is entirely within seismic-risk zone 1.
- Several successful hydrocarbon-storage cavities have been excavated in the sequence in the western part of Ohio.

The uncertain features include:

- Presence of underground mines and boreholes in younger strata.
- Earthquakes in the Cleveland area and in southeastern Ohio.
- Uncertainty as to the eastern limit of important limestones in the sequence.

Late Ordovician Shales in New York

A thick undeformed Late Ordovician marine-shale sequence dips gently southward at about one degree in central and western New York. The unit is 240 to 365 m thick in central New York, but it apparently thickens to the south and southwest and is up to 600 m thick where it is at moderate depth.⁽¹⁰²⁾ At different places the sequence embraces the Utica (Canajoharie), Frankfort, and Pulaski Shales (Figure C.51). The Utica Shale at the base is dark gray to black, whereas the younger shales are typically medium gray and contain some siltstone.

The shale sequence is directly underlain by several hundred feet of Trenton Limestone, which in turn is underlain by other limestones, dolomites, and sandstones of Ordovician and Cambrian age. In most of New York the shales are overlain by the Oswego Sandstone (Figures C.51 and C.52). The Oswego is also of Late Ordovician age, and it ranges in thickness from 0 to 120 m on the outcrop to as much as 300 m in the deep subsurface of south-central New York.⁽¹⁰²⁾ The Oswego is overlain by the Queenston Shale or is unconformably

overlain by Silurian sandstones and shales. The Oswego Sandstone typically is tightly cemented, and at depth its water is mineralized.

The top of the Late Ordovician shale sequence is 300 to 900 m below ground level in a narrow east-trending belt about 30 km wide and 400 km long (Figure C.52). The unit dips southward and is 900 to 1500 m below the surface in a parallel belt 30 km wide and 480 km long. Both belts are deflected southward in the vicinity of the Finger Lakes in the central part of the region; the bottoms of the lakes are generally 300 m or more below the elevation of nearby hills, and it is the depth of the shale beneath these deep lake bottoms that should be considered in this area. The data are not too precise in the eastern part of these belts, but it appears as though thick shales are also present here at comparable depths; the Canajoharie Shale is reported to be 600 m thick in Montgomery County.

Major drainageways in the area include the Oswego, Seneca, and Genesee Rivers, draining northward into Lake Ontario, and the Mohawk River and tributaries of the Susquehanna River draining east and south to the Atlantic Ocean. In addition, the Finger Lakes are scattered through an area about 80 km by 145 km in the central part of the region, and Lake Ontario and Lake Erie are on the north and west sides of the region.

The shale sequence has a low permeability, and its thickness would result in an impermeable seal above and below storage caverns placed in the middle part of the shales.

Late Ordovician shales do not contain commercial amounts of oil and gas in central and western New York, nor are they known to contain any other mineral resources. Oil and gas production in the region is almost entirely from Silurian and Devonian rocks (younger than the shale sequence), and thus only a few boreholes have penetrated the Late Ordovician shales.

Underground cavities have not been mined in the shale sequence in this region. There are, however, underground mines in younger Silurian gypsums and salts: 3 gypsum mines are located in Erie and Genesee Counties; 3 salt mines are in Tompkins, Livingston, and Yates Counties (the Retsof mine in Livingston County is the largest underground salt mine in the world); and 4 brine-well systems are in Schulyer, Onondaga, and Wyoming Counties.

Ordovician strata in central and western New York have not been deformed by faulting and folding. Much of the region is, however, in seismic-risk zones 2 and 3, and several earthquakes (up to MM VIII) have occurred in the Attica area (Figure C.52). The seismic risk may impose limitations on developing underground-storage sites in the region, particularly toward the west and northwest.

The Late Ordovician shale sequence in the region has many features favorable for underground waste storage:

- It commonly comprises 300 m (or more) of impermeable shale.
- It is at favorable depths over a large area.
- Much of the area is agricultural and forest land.
- The sequence contains no known commercial mineral deposits.
- Few boreholes have been drilled through the sequence.
- The rocks have not been greatly deformed or otherwise fractured.

Unfavorable features include:

- The higher seismic risk to the west and northwest.
- The probable presence of organic matter in the black shales in the lower part of the sequence.
- The presence of underground mines in younger strata.

More data are needed to evaluate the thickness, lithology, and other characteristics of the sequence in a large area of eastern New York.

Devonian-Mississippian Shales in Ohio

A thick sequence of undeformed Devonian and Mississippian marine black and gray shales dips gently to the east and southeast across the eastern half of Ohio into adjacent parts of Pennsylvania, West Virginia, and Kentucky. The angle of dip is less than one half degree. The shale sequence is 180 to 210 m thick in central Ohio, and it thickens eastward to 900 to 1500 m in West Virginia and Pennsylvania near the front of the Appalachian fold belt. The sequence consists of the Olentangy and Ohio Shales of Late Devonian age and the Bedford Shale of Early Mississippian age; the Ohio Shale is commonly subdivided into the Huron, Chagrin, and Cleveland Shale Members (Figure C.53). These shales are generally equivalent to the Chattanooga, New Albany, and Antrim Shales in other parts of the eastern United States.

Hoover⁽¹⁰³⁾ summarized the general characteristics of each named shale in the sequence. The Olentangy Shale contains some limestones and apparently is not more than 30 to 45 m thick. The members of the overlying Ohio Shale are the principal units in the entire sequence: the Huron and Cleveland Shales are dark gray to black and contain carbonaceous matter, and they are separated by the blue-gray Chagrin Shale. All three members of the Ohio Shale contain quartz, illite, and chlorite as principal minerals, but the Chagrin also contains kaolinite as an important constituent. The Bedford Shale at the top is gray and contains some layers of sandstone and siltstone. It also contains quartz, illite, chlorite, and kaolinite as principal minerals. Typically the Bedford Shale is less than 30 to 45 m thick. Thus the Olentangy and the Bedford (at the bottom and top of the sequence, respectively) are

FIGURE C.53. Schematic Diagram Showing Relations of Devonian-Mississippian Shales in Ohio and Adjacent States. From Colton, 1970.

relatively thin and contain non-shale interbeds, and the bulk of the Devonian-Mississippian sequence consists of the Ohio Shale. The black shales predominate in central Ohio, whereas the blue-gray Chagrin Shale makes up most of the unit farther east toward the Appalachian basin. Eastward, into West Virginia, Pennsylvania, and New York, the shale sequence grades into a sequence of shales and oil-producing lenticular sandstones. The area where this gradation occurs is uncertain, but it appears to be near the border of Ohio, West Virginia, and Pennsylvania.

The shale sequence overlies the Onondaga Limestone (Middle Devonian), about 30 m thick, which in turn overlies a Silurian sequence of limestones, dolomites, and sandstones (plus gypsum and salt in northeastern Ohio, northwestern Pennsylvania, and western New York). Above the shale sequence is the Berea Sandstone, an important oil and gas reservoir commonly less than several hundred feet thick, and above the Berea are the sandstones, shales, and coals of the Mississippian and Pennsylvanian Systems.

The top of the Devonian-Mississippian shale sequence is 300 to 900 m below ground level in a large region of southeastern Ohio and adjacent parts of West Virginia, Pennsylvania, and Kentucky (Figure C.53). It is more than 900 m below the surface in a narrow strip farther southeast, but beyond the front of the Appalachian fold belt the sequence is deformed and probably would not be suitable for waste storage. Owing to the great thickness of the shale sequence, storage caverns could be excavated in the lower part (and still be more than 300 m deep) in a large area west of the region where the top of the sequence is 300 m deep; thus another region on the map has been added to show where the lower part of the sequence is more than 300 m deep.

The major river in the region is the Ohio River and its tributaries, the Muskingum, Kanawha, and Allegheny Rivers. At the north end of the region is Lake Erie.

The Devonian-Mississippian shales have low permeability and great thickness (as much as 1500 m), and thus they should form a thick impermeable seal above and below caverns excavated in the middle part of the sequence.

The shale sequence has produced little oil and gas. Some wells have produced 10 to 15 barrels of oil per day for a few days, and natural gas is produced in a number of fields in south-central Ohio and adjacent parts of West Virginia and Kentucky. Oil is produced in large quantities from lenticular sandstones interbedded with the shales just east of Ohio in West Virginia and Pennsylvania. The unit is not known to contain any other mineral resources. The uranium and oil content are typical for black shales: 0.0012% uranium and 4.2 gallons of oil per ton of shale.⁽¹⁰⁴⁾

Boreholes and mines pass through the shale sequence in some areas. Oil and gas are produced from older Silurian strata in central and north-central Ohio, whereas production is mainly from younger Mississippian and Pennsylvanian strata in eastern and southern Ohio and adjacent areas; thus the number of boreholes penetrating the shale sequence is greater in the former area and fewer in the latter area. Salt mines have been dug into underlying Silurian rocks to the northeast in Lake and Cuyahoga Counties, Ohio, and in Tyler County, West Virginia. Brine wells have been completed in Silurian strata in Lake, Summit, and Wayne Counties, Ohio, and in Marshall County, West Virginia. Iron ore was produced from Silurian rocks (Clinton sands) in southern Ohio in the 19th century. Coals are mined underground from Pennsylvanian strata at a number of places in eastern Ohio and adjacent states.

The region lacks structural deformation, and all of it is in seismic-risk zone 1 (Figure C.52). Several small earthquakes (MM V-VI) have occurred in the Cleveland area, and several small earthquakes and one larger (MM VII-VIII) have occurred in southeastern Ohio.

The Devonian-Mississippian shale sequence in the eastern half of Ohio and adjacent areas appears highly favorable for storage of underground waste because:

- The sequence consists of 300 to 1500 m of impermeable shale.
- It is at favorable depths over a large area.
- Much of the area is rural and agricultural land.
- The sequence has a quartz, illite, chlorite, and (in some units) kaolinite mineralogy.
- The sequence is not known for significant mineral deposits, although some oil and gas have been produced locally.
- The rocks have not been greatly deformed or otherwise fractured.
- The region is entirely within seismic-risk zone 1.

The several unfavorable factors include:

- Abundant borehole penetrations for oil and gas locally in central and north-central Ohio.
- Penetration of the shale sequence by salt mines and brine wells in northeastern Ohio.
- Penetration by underground coal mines and petroleum boreholes into younger strata in eastern Ohio and adjacent areas.
- Earthquakes in the Cleveland area and in southeastern Ohio.
- The uncertainty as to where the shales contain important sandstones near the eastern border of Ohio.

- The possibility of producing gaseous or liquid carbon compounds upon heating the black carbonaceous shales.

Less-Favorable Shales

Late Ordovician shales of the thick Martinsburg Formation of northwestern New Jersey and eastern Pennsylvania contain siltstones and sandstones, and the unit has been deformed by complex folding and faulting.

The Hedges and Myers Shales (Mississippian) of western Maryland and northeastern West Virginia are interbedded sandstone and shale sequences and are thick only in the structurally complex Appalachian fold belt. Where other thick Mississippian and Pennsylvanian shales are present in undeformed parts of the Appalachian Highlands they contain interbeds of sandstone, limestone, and coal.

Shale Deposits of the Interior Plains

Sedimentary rocks of the Interior Plains are mainly consolidated limestones, sandstones, and shales of Paleozoic age. In most regions they are flat-lying or dip gently into one of the moderately deep sedimentary basins. It is in these basins that the shales are often thicker. Illite and chlorite are the dominant clay minerals in most shales of the province, and montmorillonite is rarely present. The principal stratigraphic sequences containing thick shales are those of the Late Ordovician and the Late Devonian-Early Mississippian. Thick shales that extend eastward into the Appalachian Plateaus are discussed with "Shale Deposits of the Appalachian Highlands."

Devonian-Mississippian Shales in Michigan

Two thick shale sequences of Late Devonian and Early Mississippian age are present in the Michigan basin. The basin is undeformed except for an anticline in the southeastern part, and the strata dip toward the central part of the basin from all sides at an angle less than one half degree. The two marine-shale sequences together are about 460 m thick in the western half of Michigan (this discussion deals only with Michigan's southern peninsula).

The lower sequence consists mainly of the Antrim and Ellsworth Shales of Devonian and Mississippian age (Figure C.54). The Antrim Shale is characterized as 45 to 90 m of black, hard, thin-bedded, carbonaceous shale containing interbeds of argillaceous limestone in the basal 3-25 m.^(105,106) Overlying the Antrim, and partly equivalent to it, is the Ellsworth Shale, which is greenish gray, silty, and locally dolomitic and is commonly 90 to 180 m thick in western Michigan. The total thickness of the Antrim-Ellsworth Shale sequence in the western half of Michigan ranges from 150 m in the east to 290 m in the northwest. Also included in this sequence are the thin

FIGURE C.54. Cross Sections in Michigan; Upper section shows relationships of Antrim and Ellsworth Shales to Berea Sandstone and Bedford Shale (modified from Cohee and others, 1951); lower schematic section shows depths and shale-sandstone relations in Coldwater and Antrim-Ellsworth Shales.

(20 m) black shales of the Sunbury Formation at the top. In the eastern half of the state the upper part of the Antrim-Ellsworth sequence contains the Berea Sandstone (an oil and gas reservoir) and the Bedford Shale, and the shales in the sequence are only 45 to 150 m thick.

Directly above the Antrim-Ellsworth sequence is the Coldwater Shale of Early Mississippian age. The Coldwater consists of gray and bluish-gray shale with some thin limestone, dolomite, and siltstone interbeds.⁽¹⁰⁵⁾ The thickness ranges from 150 m in the west to 300 m in the central part of the state and from 300 to 335 m in the eastern half of the state. At the base of the Coldwater (directly above the Antrim-Ellsworth) is 3 to 6 m of red limestone and dolomite interbedded with shale. The Coldwater contains lenticular sandstones and silstones in eastern Michigan, and these rock types make up much of the formation in the northeastern one-third of the state. In the southern part of the basin (mainly in Clinton and Ingham Counties) the lower part of the Coldwater contains a sandstone bed 6 to 9 m thick.

The Antrim-Ellsworth Shale sequence overlies the Traverse Group, a series of Middle and Late Devonian limestones 60 to 210 m thick, and this in turn overlies another 3000 m of sedimentary rocks. Above the Coldwater Shale is 120 to 180 m of Marshall Formation of Mississippian age, and this is overlain by about 300 m of younger sedimentary rocks and a cover of glacial drift that ranges from 15 to 240 m thick.

The depth to the top of the Antrim-Ellsworth Shale sequence (where it is more than 150 m thick and where the Berea Sandstone is absent) is 300 to 850 m in a large semicircular area embracing most of the western half of the Michigan basin (Figure C.55). The maximum depth to the base of this sequence is about 1070 m. The top of the Coldwater Shale is 300 to 550 m below the surface in a subcircular area in central and western Michigan. The lower part of this sequence also is at least 300 m deep in a larger area that extends almost to the area where the Coldwater Shale is overlain by glacial deposits. Principal rivers in the western half of Michigan are the Grand, Muskegon, and Manistee, all draining into Lake Michigan to the west.

Shales of both the Antrim-Ellsworth and Coldwater sequences have a low permeability, and because of their total thickness they would form an impermeable seal above and below storage caverns mined in either of the shales. The Marshall Formation, overlying the Coldwater Shale, is reported to be one of the main aquifers of the Michigan basin.⁽¹⁰⁷⁾

Minor amounts of oil and/or gas have been produced from some of the limestones and lenticular sandstones in both shale sequences, but in only a few wells has there been commercial development. Natural gas and oil have been produced from the Berea Sandstone in a number of areas in eastern Michigan. Neither sequence is known to contain any other mineral resources.

Figure 10. Depth to top of Antrim-Ellsworth Shales (Late Devonian-Early Mississippian) in Michigan basin (modified from Cohee and others, 1951; Merewether and others, 1973). Thickness of shale sequence ranges from 500 feet along eastern boundary to 800-900 feet in the west.

FIGURE C.55. Depth to Top of Antrim-Ellsworth Shales (Late Devonian-Early Mississippian) in Michigan Basin (modified from Cohee and others, 1951; Merewether and others, 1973); thickness of shale sequence ranges from 500 feet along eastern boundary to 800-900 feet in the west.

Black shales in the Antrim contain an average of 0.0021% uranium and about 5 to 10 gallons of oil per ton of shale, as determined from samples from each of two boreholes. (104)

Locally, many boreholes have been drilled through both shale sequences to search for and produce oil and gas from the underlying Devonian, Silurian, and Ordovician limestones and dolomites. Several of the Mississippian formations, including the Berea and Marshall Sandstones, have yielded oil and natural gas in the eastern half of the basin. Eight brine-well systems in the western half of the basin penetrate through the shale sequences in Manistee, Mason, Muskegon, Midland, and Gratiot Counties. Two gypsum mines in Kent County have been excavated in Late Mississippian strata and do not penetrate the shale sequence.

The region lacks structural deformation, with the exception of the Howell anticline in the southeast, and all of it is in seismic-risk zone 1 (Figure C.52). Three small earthquakes (MM V-VI) have occurred near the edges of the region under consideration.

Both the Antrim-Ellsworth and Coldwater Shale sequences in the western half of Michigan's southern peninsula appear highly favorable for storage of radioactive waste because:

- The combined thickness of the impermeable shale sequences is about 460 m and each of the units ranges from 150 to 300 m thick.
- The deposits are at favorable depths over a large area.
- Most of the sparsely populated land surface is used for agriculture or is forested.
- The sequence contains no significant mineral deposits.
- The rocks have not been greatly deformed or otherwise fractured in except in the southeast.
- The region is entirely within seismic-risk zone 1.

The several unfavorable factors include:

- Locally abundant boreholes drilled for oil and gas.
- Penetration of the shales by brine wells.
- Underground gypsum mines into younger strata.
- The possibility of producing gaseous or liquid carbon compounds upon heating the black carbonaceous shales.
- The overlying Michigan Formation is an aquifer.

Less-Favorable Shales

The Borden Siltstone (Mississippian) of Illinois is dominantly gray, argillaceous, fine-grained siltstone with lesser amounts of silty shale and fine sandstone. The siltstone is 180 m thick in the southern and eastern parts of central Illinois and is 300 to 900 m deep over a large area. This area is not considered a favorable unit because it is dominantly siltstone; it contains some oil-bearing sandstones; and it is near the New Madrid fault zone and partly within seismic-risk zone 2.

The Kewanee Group (Pennsylvanian) is present throughout most of the Illinois basin and consists of interbedded shales, sandstones, limestones, and coals. It contains almost all of the major coal reserves in the region.

The McLeansboro Group (Pennsylvanian) in the southern half of Illinois consists of 150 to 300 m of alternating shales, sandstones, limestones, and coals.

Pennsylvanian rocks of Kansas, Missouri, Iowa, and northern Oklahoma comprise interbeds of shale, sandstone, limestone, and coal. None of the shales is more than 30 to 60 m thick.

Mississippian and Pennsylvanian rocks of eastern Oklahoma and Arkansas include shales, sandstones, and some limestones, with most of the shales too thin for consideration. The only thick shales are in the Stanley (Mississippian) and Atoka (Pennsylvanian) sequences of the Arkoma basin and Quachita Mountains, but these areas are structurally complex and the shales are deformed.

The Springer and Goddard Formations (Mississippian and Pennsylvanian) of southern Oklahoma are thick units of shale that contain important oil-producing sandstones. They occur in the structurally complex parts of the Ardmore and Anadarko basins and are more than 900 m deep at most places.

The Barnett Formation (Mississippian) consists of shale and some limestone in the Fort Worth basin of north-central Texas. The Barnett is more than 140 m thick in the northeastern part of the basin, but in that area it is more than 900 m below the surface.

The Smithwick shale facies of the Atoka (Pennsylvanian) is locally more than 150 m thick in parts of the Fort Worth basin of north-central Texas. The Smithwick interfingers with the Atoka sand facies to the east, and the boundaries of the thick shale areas are imprecise. Many faults intersect the shale unit to the south, where it is within 600 m of the surface, and farther north in the basin the shale is more than 900 m deep.

The Cisco and Canyon (Pennsylvanian) rocks of north-central Texas comprise shales, limestones, sandstones, and some coals; none of the shales is thick enough to be considered.

Clay Deposits of the Coastal Plain

Coastal Plain sediments consist of interbedded and interfingering sands, clays, and calcareous materials. These sediments dip and thicken toward the sea. Because they are relatively young (Cretaceous, Tertiary, and Quaternary in age), almost all sediments less than 450 m below the surface are still unconsolidated, and they tend to be consolidated only where they are more than 900 m deep.⁽¹⁰⁸⁾ Below the water table, all sediments are saturated with water, and fresh water extends to depths of 30 to 600 m feet in different areas. Clays (unconsolidated shales) are common in the stratigraphic sequence, but thick clays that are widespread are not common. Montmorillonite is the principal clay mineral in almost all units.

Paleocene Clays in Mississippi, Louisiana, and Arkansas

A thick sequence of marine clay is present in the Mississippi embayment area of the Gulf Coastal Plain. The Porters Creek Clay is Paleocene (Early Tertiary) in age and is part of the Midway Group. It consists of 150 to 335 m of dark-gray, montmorillonitic, laminated clay and has the high plasticity that normally is associated with montmorillonitic clays.⁽¹⁰⁷⁾ The thickness increases toward the Mississippi River and toward the Gulf, with the maximum thickness in southwestern Mississippi.⁽¹⁰⁹⁾ The clay sequence is about 120 m thick over the Monroe uplift in northeastern Louisiana and only about 60 m thick over the Jackson dome in west-central Mississippi. Eastward, the Porters Creek Clay thins to about 100 m near the Mississippi-Alabama state line. Precise data are not readily available to the west, but it is possible that the unit is 150 m thick or more for about 160 km into Texas and 300 to 900 m below the surface in a narrow band subparallel to the Gulf Coast.

The Porters Creek Clay is underlain by the Clayton Formation, which is about 15 to 30 m of marl at the base of the Paleocene, and this in turn is underlain by Cretaceous sands and clays (Figure C.56). The clay sequence is overlain by 30 to 60 m of sandy clay, the Naheola Formation, which in turn is overlain by the Wilcox Group of Eocene age. The Wilcox consists of about 60 to 600 m of sands with some clays, and it is one of the most widely used aquifers in the region.

The top of the Porters Creek Clay is 300 to 900 m below ground level in an 80-km wide belt that extends north-northwest across central Mississippi and across smaller areas in southeastern Arkansas and northwestern Louisiana

FIGURE C.56. Stratigraphic Column and Depth to Top of Porters Creek Clay (Paleocene) in Mississippi Embayment (modified from Rainwater, 1964; King, 1969); clay sequence is 500 to 1,100 feet thick.

(Figure C.56). The clay sequence is 900 to 1500 m deep in a narrow belt farther south, which extends mainly through central Mississippi and central Louisiana.

Surface drainage in the region is dominated by the Mississippi River, although other important rivers crossing parts of the region include the Pearl, Big Black, Yazoo, White, Arkansas, Quachita, Red, and Sabine. The combined flow of all these rivers represents the runoff from nearly one-half of the entire United States.

The clays of the Porters Creek have a low permeability, and the sequence acts as an aquiclude between two important fresh-water aquifers. Numerous industrial, municipal, and domestic wells have been completed in the Wilcox aquifer 30 to 60 m above the clay sequence.^(110,111) The sands in the Cretaceous, about 15 to 30 m below the Porters Creek Clay, are also important freshwater aquifers in northern Mississippi and adjacent parts of Arkansas and Tennessee. The Wilcox and the Cretaceous locally produce fresh water from depths as great as 600 m in Mississippi.⁽¹¹¹⁾

Oil and gas are not produced from the Porters Creek Clay in this region, although the overlying Wilcox is productive downdip to the south. The underlying Late Cretaceous strata also produce oil and gas farther south as well as in the northern half of Louisiana. Oil and gas have been produced from still deeper strata at widely scattered sites in the region, particularly in Louisiana, and thus locally many boreholes have been drilled through the clay sequence. The Porters Creek is not known to contain any nonpetroleum minerals, and no mine shafts have penetrated the clay.

Sediments dip gently and lack deformation in most of the region, although several principal structural features are within or near the area of optimum depth (Figure C.56). The Pickens-Gilbertown fault zone extends northwestward across central Mississippi from Clarke County to the Yazoo River, and a few salt domes are present within the region in northwestern Louisiana and just south of the region in Mississippi. Also, the northern part of the region is astride the New Madrid fault zone. Much of this northern area is in seismic-risk zones 2 and 3 (Figures C.52 and C.56), but the remainder is in zone 1.

The Porters Creek Clay has many features favorable for storage of radioactive waste in certain parts of the Mississippi embayment because:

- The clay sequence comprises 150 to 335 m of impermeable clay.
- It is at favorable depths over a large area.
- Much of the region is sparsely populated agricultural and forest land.
- The sequence contains no known significant mineral deposits.
- The rocks have not been deformed or otherwise fractured throughout most of the region.

- Much of the region is in seismic-risk zone 1.

Unfavorable features include:

- A high seismic risk in the north along and near the New Madrid fault zone.
- Local deformation and possible structural movements near fault zones.
- The clay is unconsolidated and montmorillonitic and may present special problems during and after excavation.
- The clay lies stratigraphically between two widely used aquifers that locally yield fresh water from depths of 600 m.
- Runoff from nearly one-half the United States passes through the rivers crossing the region.
- Many petroleum test holes have been drilled locally through the clay sequence.

Less-Favorable Clays

In southeastern Georgia and northeastern Florida, the Hawthorne Formation (Miocene) consists of several hundred feet of clay and is no more than several hundred feet below the surface.

Several units composed predominantly of clay in the Cretaceous and Tertiary of Texas are locally 90 to 150 m thick at and near the outcrop, but data are insufficient to determine whether these units have a suitable thickness and lithology at the optimum depth range of 300 to 900 m. Late Cretaceous units include the Eagle Ford and the Taylor clays and marls. Early Tertiary units include the Wills Point Clay of the Midway Group (Paleocene equivalent to the Porters Creek Clay of Louisiana) and the Cook Mountain-Crockett sequence and Jackson Group, both Eocene. These units are at the proper depths in narrow bands subparallel to the Gulf Coast, extending southwestward from east Texas to the area south of Austin. A more detailed study is needed to determine whether any of these many units are locally suitable for waste storage since these units locally contain layers of silt and sand, and their shallow-subsurface distribution is in the region of high-angle faults.

C.2.8. Shales-Southeastern Triassic Basins

The eastern Triassic basins extend from Nova Scotia to Florida (Figure C.57). They have many features in common but differ in detail.^(101,102) The northern basins have been fairly well studied and most general statements about Triassic basins are based on these studies.

The basins are long, narrow and are filled with from several thousand feet to more than 6100 m of clastic sediments. On the basis of shape they could be considered ideal for the storage of waste products. From the North Carolina-South Carolina boundary north most of the basins are in the Piedmont Provinces and are subaerially exposed. In South Carolina, Georgia and northern Florida the Triassic basins occur in the Coastal Plain and are overlain by varying thicknesses of Cenozoic sediments.

Scattered occurrences of Triassic sediments are reported from the Coastal Plain north of South Carolina but little is known about their lateral extent.⁽¹⁰³⁾

All the Triassic basins which have been explored in detail are bordered on one or both sides by major normal faults. In turn, sediments have been derived from one or both sides of the basin and rest unconformably on Paleozoic and Precambrian deformed rocks similar to those on the flanks of the basin. Precambrian gneiss, schist and granite are more abundant from Virginia south and Paleozoic metamorphosed rocks from Virginia north. Strike faults occur within the basin sediments and are approximately parallel to the border faults. Transverse faults are common and in places the two sets produce a rhombic pattern. Some of the transverse faults extend into the surrounding folded and thrust-faulted Paleozoic rocks.

Border faulting is believed to have been a fairly continuous process during which sediments accumulated as the basins progressively deepened. In conjunction with the development of the border faults, which are believed to be late Triassic, large amounts of basic magma entered the basins. Most entered along the borders and spread into the sediments as sills and dikes. Magma intrude some of the strike and transverse faults in the basins. Most of the basin sills are cut by faults, though it is generally believed that most of the faulting, other than the border faults, occurred after most of the sediments were deposited.

Both the faulting and the dike intrusions spread into the rocks adjacent to the basins. The Triassic sediments were much more extensive than now. In places sediments bury the border faults and spread over the upthrown blocks.

It is believed that the faulting and magma activity is associated with the initial rifting of North America and Africa.^(104,105)

It has been suggested⁽¹⁰⁶⁾ that some of the present isolated basins were originally connected and that post depositional faulting, folding and erosion

FIGURE C.57. Exposed and Buried Triassic Sediments. Modified from Marine and Siple (1974), particularly Georgia-Florida area. If the Georgia-Florida basin is not a complex of basins it is the largest Triassic basin on the east coast. Coastal Plain basins have not been eroded as much as Piedmont basins, so it is possible that a basin of this size exists.

had caused their apparent isolation. Klein⁽¹⁰⁷⁾ presented stratigraphic data, K-Ar data and paleocurrent analyses on basins from North Carolina to Nova Scotia (Figure C.58) that indicated the sedimentary rocks were locally derived from the margins of the basins and concluded that each was a separate basin. There is little doubt that the rocks are locally derived but this does not prove the basins were separated.

The sediments filling the Triassic basins are largely clastic and dominantly red in color. Conglomerates, arkosic sandstones, siltstones, shales and argillites are the most common sediments. Minor limestone and coal are present in some basins. The depositional environment was entirely continental.

Local topographic relief is believed to have been low in the south and moderate to high in the north.⁽¹⁰¹⁾ This is presumably based on the relative abundance of coarse sediments and sediment thickness.

North Carolina

The southern most Triassic basin outcrops are in North Carolina. These will be reviewed briefly in order to obtain some idea of what should be expected in the less thoroughly studied basins lying beneath the Coastal Plain sediments of South Carolina and Florida.

The Deep River Triassic Basin, 58 km long and 8 to 22 km wide, lies in the north central part of the state in the Piedmont Province.⁽¹⁰⁸⁾ The northeast trending basin is a tilted and down faulted trough-shaped block, in which the sediments dip generally southeastward (Figure C.59). The southeast edge is formed by the Jonesboro fault, a northwest-dipping normal fault having a maximum vertical displacement equal to the maximum thickness of Triassic sediments in the basin (2,100 to 3,000 m). The northwest edge is irregular and reflects an erosional contact between the Triassic and Pre-Triassic rocks. Within the basin, longitudinal faults roughly parallel the Jonesboro fault. They are displaced several hundreds or thousands of meters. Cross faults, commonly followed by dikes, have displacements ranging from a few meters to a few hundred meters. Tilting and differential subsidence within the rectangular fault framework has caused many variations from the regional strike and dip.

The basin is bordered by pre-Triassic metamorphic and igneous rocks. The rocks to the northwest are part of the "Carolina Slate Belt" (metamorphosed volcanics and sediments). The rocks on the southeast flank contain some similar rocks but also contain gneiss, schist and granite.

FIGURE C.58. Triassic Regional Dispersal Patterns.
From Klein, 1969.

GEOLOGY OF THE DEEP RIVER COAL FIELD, NORTH CAROLINA

FIGURE C.59. Theoretical Sections Shown in Development of Deep River Triassic Basin, North Carolina. From Reinemund, 1955.

There is a basal conglomerate derived largely from rocks a short distance northwest of the basin; however, most of the sediments were derived from the southeast flank of the basin (Figure C.59). Boulder-bearing aluvial fan sediments were deposited along the southeast edge of the basin by streams flowing northwest. Finer grained sediments were deposited towards the center of the basin. In the center and northwest portion of the basin fine grained sediments were deposited in lake and swamp environments. Thin coal beds were deposited along with these sediments.

Although gross lithologies are similar from outcrop to outcrop, individual beds change rapidly in thickness laterally, and in lithology both laterally and vertically.⁽¹⁰⁹⁾

Most basin sediments contain less than 15% feldspar but some contain more than 25%. Metamorphic rock fragments are abundant. The medium and coarse grain sands are held together by quartz cement, clay paste and by inter-locking of rock particles.

The Triassic sedimentary rocks are more easily eroded than the metamorphic and igneous rocks of the Piedmont. Therefore, the Triassic belt is a trough-like topographic lowland for most of its length.

Bain (personal communication) is presently conducting a detailed study of the Deep River basin using a wide variety of geophysical techniques. He plans to drill a well in the near future. He will use this basin as a model for a later regional study of the Triassic basins of the east coast.

The Triassic Wadesboro basin lies in-line and to the southwest of the Deep River basin. It is bounded on the southeast by a major fault and on the northwest by a number of smaller faults and lies unconformably above rocks of the Carolina Slate Belt.⁽¹¹⁰⁾

A major fault defines the eastern border. It is thought to be an extension of the Jonesboro fault which forms the southeast flank of the Deep River Basin or the continuation of the Governors Creek fault which lies within the Deep River basin. Portions of the fault are covered by Coastal Plain sands.

The fault system on the western edge of the basin are of smaller magnitude and consist of many narrow, elongated horsts and graben filled with fanglomerated. Both longitudinal and cross faults are present in the main basin. The latter cut the western border faults in many places. Dikes accompany many of the cross faults. The basin sediments strike N41°E and dip 20°SE. Sediments are predominantly alluvial.

Fanglomerates are limited to the western edge of the basin. This in contrast to the Deep River basin where the coarse facies occur on the east flank of the basin. Argillite pebbles and cobbles, derived from the Carolinate

Slate, are abundant. Arkosic sands, containing 35-65% moderately fresh angular feldspar, occur in the interior and eastern portion of the basin. The character of both sediments indicate they were derived from source areas close to the basin. Red siltstones and claystones are abundant throughout the basin and occur alone or interbedded with sandstone.

Clastic sediments were derived from both margins but only fine to medium-grained sediments were deposited along the eastern border. Because a major fault lies along this border, they suggest the coarse facies were deposited and had been eroded. It is also possible that the upthrown block was tilted to the southeast and drainage was away from the basin rather than into it.

Of importance in evaluating the distribution of lithologic facies in Triassic basin is the fact that throughout the Deep River and Wadesboro basins have similar structural configurations the coarse facies occur on different sides in the two basins.

The Crowburg basin southwest of the Wadesboro basin includes the southermost outcropping Triassic sedimentary rocks in eastern North America. The wedge-shaped basins are 6.4 km long and as much as 1.6 km wide. Fanglomerates are the only sediments observed in the outcrops.⁽¹¹¹⁾

South Carolina

Triassic sediments have been found in three different localities beneath the Coastal Plain cover in South Carolina. The Florence basin was inferred from magnetic anomalies⁽¹¹²⁾ and seismic data⁽¹¹³⁾ and confirmed from cores.⁽¹¹⁴⁾ McKee et al⁽¹⁰¹⁾ show this basin extending northeast into North Carolina. Top of the Triassic is at 180 m.^(115,116)

Little data are available on this basin but as it is the shallowest known basin in the Coastal Plain it should be seriously considered for further investigation.

Cook⁽¹¹⁷⁾ describes a well near Summerville, 32 km northwest of Charleston which penetrated 300 m of probably Triassic coarse red sandstone and shales with a few pebbles and diabase dike. It is overlain by 480 m of Coastal Plain sediments. Mansfield⁽¹¹⁸⁾ believed this material was Cretaceous. Woollard et al⁽¹¹⁶⁾ found no seismic evidence for a basin in this area.

The northeast trending Dunbarton basin is aligned perpendicular to the South Carolina and extends a short distance into Georgia.

The basin has been outlined on the basis of an aeromagnetic survey, seismic - refraction and seismic - reflection surveys and confirmed by drilling. It is approximately 48 km long and 2 km wide (Figure C.60). The upthrown western flank of the basin consists predominantly of schist, gneiss and some quartzite. There is a 24 m weathered zone between the base of the Coastal Plain sediments and the top of the crystalline rocks. Seismic and

FIGURE C.60. Aeromagnetic Map Showing the Outline of the Dunbarton Triassic Basin and the Location of Wells and Seismic Investigations. From Marine and Siple, 1974.

well data indicate a major normal fault forms the northwest border of the basin though it is not certain whether this is a border fault or a longitudinal fault just outside the basin (it occurs in gneiss). The contact between the basin sediments and the basement gneiss dips 35°SE . The nature of the southeastern boundary is not known but as the sediments dip to the southeast it is assumed to be a normal fault of greater magnitude than the one forming the northwest border (Figure C.61).

Well samples indicate the western portion of the basin contains maroon arkosic siltstone with abundant gneissic pebbles and boulders (fanglomerate). Near the center of the basin the sediments consist of red to maroon mudstone, gray-brown fine-to-medium-grained sandstone and pink to buff medium-to-coarse-grained arkosic sandstone. Clay is abundant in the two types of sandstone. Calcareous cement is common.

The calcareous cement is thought to have been formed from water which moved downward through the fanglomerate along the western margin and upward in the basin center where evaporation caused carbonate and other salts to concentrate.

The sediments on the southeastern flank consist half of maroon clay stone and siltstone containing gray calcareous nodules and half grayish-brown very fine-to-fine-grained arkosic sandstone with calcareous nodules. Thus, there is a systematic decrease in grain size from northwest to southeast. This resembles a half-graben distribution.

A thin (1 to 2.5 m) weathered zone occurs between the base of the basin sediments and the top of the underlying crystalline metamorphic rocks. The Triassic sediments are overlain by 250 m to 300 m of Coastal Plain sediments. There is a 3 m to 14 m mildly weathered zone between the top of the Triassic and base of the Coastal Plain sediments. This zone is overlain by 12 m to 40 m (thickening from northwest to southeast) of gray sandy clay which should form an effective caprock.

Georgia

No unequivocal Triassic basins have been identified in the Georgia Coastal Plain; however, numerous wells have penetrated diabase dikes and sills and red arkosic sediments beneath the Cretaceous and it is reasonable that some of these are Triassic in age. K-Ar dates on some of the dikes indicate a Triassic age. In some wells these rocks are underlain by Paleozoic sediments.

Milton and Hurst⁽¹⁰³⁾ have indicated the counties in which true crystalline metamorphic rocks have been encountered immediately under Coastal Plain sediments. These occur along the northwestern updip portion of the Coastal Plain, bordering the Fall Line. They have also outlined the counties in which probably Triassic rocks (basalt, diabase, volcanic ash, arkose?, sandstone in

FIGURE C.61. Postulated Sequence of Geologic Events for the Dunbarton Triassic Basin. From Marine and Siple, 1974.

siltstone) were found. Relatively little section has been penetrated in most wells, as once diabase, basalt or granite (arkose) is encountered drilling usually stops.

In Echols County (south central Georgia) Upper Cretaceous sediments rest on Paleozoic shales and sandstones. Basement rock to the east and northeast of Echols County are largely altered volcanic ash and are similar to and considered to be equivalent to the slate belt rocks of the Piedmont. Two recently drilled wells in Camden County and one in Charlton County are reported to have penetrated Paleozoic black shales, siltstone and quartzite at depths of 1340 to 1400 m.

Northwest of this area sandstones and shales along with diabase dikes are the most common rocks. Five wells drilled in 1973-74 in Lowndes County (immediately west of Echols County) penetrated red shales, siltstones, sandstones and diabase dikes. These sediments are believed to be Triassic and have been called the Eagle Mills formation. The top of the Triassic occurs between 1220 and 1340 m. The deepest penetrated 1140 m of possible Triassic and bottomed in Paleozoic.

X-ray analyses of shales from the thick Triassic section in Lowndes County indicated they have a clay suite similar to those reported for the North Carolina Triassic. The clay suite is complex containing varying amounts of illite, montmorillonite, chlorite and kaolinite. The first two clays are predominant. Montmorillonite is the major clay in many of the samples.

To the north and northwest of Lowndes County most of the Triassic sediments described by Milton and Hurst⁽¹⁰³⁾ are arkoses or granite wash. A well in Mitchell County encountered Triassic at 2000 m and penetrated 325 m of sand and shale. A well in Thomas County bottomed in Triassic at approximately 1800 m. A recent well in Dodge County drilled through 365 m of "granite wash" or weathered granite. Top of Triassic in this area is 760 to 900 m.

The limited well data suggest there are major deposits of Triassic sediments in the central and southwestern Coastal Plain of Georgia. The sediments in the northwestern part of the area are arkosic sands, whereas, shales and silts are apparently more abundant to the southeast (Lowndes County). This may be one large basin.

Recently published cross sections⁽¹¹⁹⁾ also indicate a Triassic basin in this area. The interpretation is presumably based on more complete data obtained by the oil companies. A north-south cross section (Figure C.62) shows a thin (300 to 600 m) deposit of Triassic extending from Crisp County to Lowndes County. Drilling suggests there is nearly 1200 m in Lowndes County so the thickness may be underestimated.

FIGURE C.62. Cross Sections Showing Location of Triassic Basins in Georgia Coastal Plane. From AAPG, 1975.

An east by south-east cross section shows steep normal faults in Pulaski and Dodge Counties with a maximum of 1280 m of Triassic. The Triassic is in the shape of a wedge which thins out near the coast in Wayne County (Figure C.62).

Earlier McKee et al.⁽¹⁰¹⁾ indicated a Triassic basin existed in the southwestern half of the Coastal Plain. They indicate the axis of the basin swung west into Alabama and southwest into Florida and thickened in both Alabama and Florida. The shape and location are similar to that of the Apalachicola Embayment which persisted in this general area from the Paleozoic to the Miocene.

In Georgia the Triassic sediments are overlain from 600 to 2400 m of Coastal Plain sediments. They would seem to be too deep to be of value for shallow waste disposal. Most of the area with less than 600 m of Coastal Plain sediments, is underlain by crystalline metamorphic rocks, though, small isolated basins could well occur in this area.

North Florida

Barnett⁽¹²⁰⁾ has completed a comprehensive evaluation of the basement rocks in Florida and southern Georgia. He shows the presence of a deep Triassic graben underlying the region of the Apalachicola or Southwest Georgia Embayment (Figure C.63). It is nearly 96 km wide and filled with coarse red feldspathic and arkosic clastics containing numerous diabase sills. Maximum thickness is unknown but should exceed 1200 m. The upthrown blocks consist of Paleozoic sediments which have suffered low-grade metamorphism.

The Triassic rocks are overlain by from 1200 m to over 3000 m of sediments which thin towards the northeast into Georgia.

Arden⁽¹²¹⁾ explored the smaller Triassic basin to the west of the Apalachicola Embayment. A north by northeast trending seismic cross section throughout the central portion of the Florida Panhandle showed the presence of a faulted Triassic basin approximately 24 km wide with a maximum thickness of 1800 m. It is overlain by 1800 m of Gulf Coastal Plain sediments. Gravity and magnetic data confirm the interpretation.

Hydrology

Hydrologic data on the Triassic sediments are relatively scarce. Data have been published on two groups of wells drilled in the Dunbarton basin.⁽¹²²⁾

The first three wells provided the following results. Laboratory analysis of core samples show that all rocks have low hydraulic conductivity (10^{-5} to 10^{-8} m per day). This is caused by the poor sorting and high clay content of all samples. Effective porosity of the sandstone cores was ~8% and mudstone cores ~0.5%.

FIGURE C.63. Pre-Jurassic Erosional Surface Showing Major Triassic Graben in Northern Florida and Southwest Georgia. A second basin occurs in the central Panhandle. From Barnett, 1974.

Eight year field tests of the well in the fine-grained sediments on the southeast flank of the basin indicated the hydraulic conductivity was 2.4×10^{-7} m per day; the transmissivity of the region surrounding the well was 4.6×10^{-6} m per day; the average permeability was 2.4×10^{-7} m per day. The permeability values of the fanglomerates on the northwest flank were comparable to those in the fine-grained sediments. This indicates the ever-present clay matrix is the factor limiting and controlling permeability and transmissivity and grain size are of little consequence.

This materially increases the volume of sediments that could be considered suitable for the storage of radioactive material. Basin faults were not detected.

Additional geophysical data and analysis suggested the presence of several longitudinal faults in the basin. The faults do not extend into the Cretaceous indicating they are more than 100 m.y. old. Two wells were drilled on either side of one of the faults.

Cores from the weathered zone in the upper 30 to 45 m of Triassic rocks swelled and cracked when exposed to wetting and drying. Cores of the Triassic from both holes were predominantly maroon, fine-grained, dense mudstone. One well was slanted to intercept the geophysical fault but none was encountered.

Of the five wells; two have heads above land surface and 68 m and 110 m above the head in the overlying Coastal Plain aquifer. These two wells are located in the northwestern portion of the basin. Because of the very low permeability no well has yet achieved a water level that is in equilibrium with that of the formation, though in two months time one well was near equilibrium. The geopressures are real and pose a serious problem both from the standpoint of contamination of shallow aquifers and stability of storage facilities.

In order to obtain some additional data on the hydrologic characteristics of the Triassic, several North Carolina ground water reports were examined.

In the Deep River basin near Durham, North Carolina the rocks dip 10° to 20° southeast. Though the porosity and permeability is generally low, water percolating downdip through the more permeable sandstones develops sufficient head to produce flowing wells in some localities. Wells in the Triassic sediments yield from 6 to 1.2 liters per minute per meter of well. Wells commonly have from 15 to 30 m of uncased depth and produce 26 to 42 liters per minute. The yield per meter of well decreased with depth. In part this is related to the development of secondary openings formed by weathering near the surface.

Bain (personal communication) says there is some porosity at depth but virtually no permeability.

Where diabase dikes are present downgradient from the direction of groundwater movement, impounding can occur. This makes the best location for water wells. Some such mechanism might explain the geopressures in the Dunbarton basin.

To the east in the Raleigh area⁽¹²³⁾ Triassic sediments yield from 0.12 to 1.4 liters per minute per meter of well or from 11 to 24 liters per minute per well. Yield decreases with depth. Rocks are largely arkosic sandstones, argillaceous sandstones, shales and fanglomerates. Rocks are largely impermeable and water is stored in and moves along joints and fracture planes.

The ground water studies indicate that minor water movement does occur in the Triassic sediments but generally is related to fractures. Water production decreases with depth (maximum depth usually less than 90 m) suggesting that at depths of several hundred meters there should be little ground water movement.

Studies have been made^(124,125,126,127) of the hydrologic characteristics of the basement crystalline-metamorphic rocks near the Savannah River Plant. These rocks are presumably typical of the rocks that flank and underlie the Triassic basins.

The upper thousand meters of rock are pervaded by small fractures and hydraulically connected. Transmissivity is low. Response to withdrawal is slow. Also present are fractures open wide enough to permit transmission of waters at a faster rate. These fractures occur in vertically restricted but laterally extensive zones.

Field tests indicated there were numerous intervals that produced from 4 to 70 liters per minute of water. One third of the core samples with the highest values, had permeabilities that averaged $0.02 \text{ } \mu\text{pd/m}^2$.

Injection tests into most sections were so low as to be unmeasurable. Swabbing tests indicated that the average permeability is $0.012 \text{ } \mu\text{pd/m}^2$; however, some fracture zones have a coefficient of transmissibility of 2000 $\mu\text{pd/m}$. The apparent permeability of hydraulically transmissive zones is $41 \text{ } \mu\text{pd/m}^2$; some zones exceed this value.

Tracer tests were conducted and calculations indicated the coefficient of permeability was $14 \text{ } \mu\text{pd/m}^2$; porosity, due to fractures, was 0.10%; flow velocity was 1.2 meter per day. The measured velocity was 5.5 times higher than the calculated velocities based on gradient, permeability and porosity.

The crystalline rocks are apparently more permeable and transmit water more rapidly than the Triassic sediments.

C.2.9 Carbonates

A limited number of carbonate rocks appear to have the basic physical properties essential to the site selection criteria for a nuclear waste repository. Several examples of dry excavations in carbonate rocks are described in the following section.

The Barberton Site

The Barberton mine is 64 km due south of Cleveland, Ohio and about 10 km southwest of Akron, Ohio (Figure C.64). The room and pillar mine is 700 m below ground level or 390 m below sea level. Nearly 140 km of mine tunnels have been developed through the upper 14 m of the Columbus Limestone. In ascending order, the strata above the mined interval consist of 1.5 m of limestone, nearly 520 m of shale, some 125 m of alternating shale and sandstone, 20 m of conglomeratic sandstone, and 6 m of glacial drift. (Figure C.65).

The location of mine relative to regional structure and stratigraphy may be the salient reason for the dryness of the mine. The important contributory factors for this dryness are:

- The mine is beneath approximately 520 m of dense impermeable shale that apparently has little or no secondary permeability induced by fracturing. Although strain directions are nearly the same at depth as at the surface, open joints have only developed near the surface (manifest in outcrops). Also, major faulting is absent in Ohio.
- The mine is on the flank of a structural high - the Cambridge arch. Percolating meteoric water above the shale aquiclude would tend to migrate down-dip away from the crest. The regional water table in the Barberton area should be inclined westward toward the trough of an adjacent structural low - Parkersburg-Lorraine syncline.
- The mantle of glacial drift acts in two capacities. First, it serves somewhat in the fashion of a sponge retaining a very large percentage of the meteoric water in the area, and secondly, the till fraction of the drift serves as a "first level" aquiclude above the thick shale sequence below.
- The depth of the mine itself is an important factor. Due to lithostatic confining pressure, rocks at this depth ordinarily have little or no porosity.

Geomorphology. The environ of the Barberton mine is a glacially modified stream-dissected plateau known as the Allegheny Plateau. Based upon depth to bedrock contours the plateau prior to Pleistocene glaciation exhibited a mature stage of dissection in the erosional cycle. The glaciers subdued the

FIGURE C.64. Location of Barberton Mine

FIGURE C.65. Generalized Stratigraphic Section, Barberton Mine, Ohio. After C. R. Stauffer, 1944.

relatively rugged topography and invoked profound drainage changes. Relief on the bedrock surface ranges up to several hundred feet. The present glaciated surface could be classed as typical kettlemoraine replete with poor drainage conditions.

Bedrock Geology. The bedrock of northeastern Ohio consists of a thick wedge of Paleozoic strata deposited on a Precambrian basement that slopes southeastward from an elevation of about 600 m below sea level near Findlay, Ohio to 9100 to 12,000 m below sea level in Virginia and eastern Pennsylvania.⁽¹³⁹⁾ The basement high is along the Findlay Arch, a northeastward extending branch of the Cincinnati Arch, and the basement low to the southeast is within the trend of the Appalachian mobile belt. The depositional basin between the two tectonic elements mentioned above was essentially free from major subsidiary positive (uplift) or negative (downwarp) tectonic elements during Paleozoic deposition, hence the rocks, in general, represent sediment that accumulated in a slow, uniformly subsiding basin where periodic uplifts in distant source areas provided deltaic wedges of clastics. The shape of the basin and the character of the sediments within it are shown in Figure C.66.⁽¹⁴⁰⁾

The units that appear to play the most important part in maintaining dry conditions in the Barberton mine are the shales that were deposited during late Devonian and early Mississippian time. This overlying shale sequence is approximately 520 m thick and forms an aquifuge.

The slow subsidence in the depositional basin and the absence of subsidiary tectonic elements permitted the accumulation of this thickness of shale. Figure C.67⁽¹⁴¹⁾ illustrates a cross-section of the Paleozoic bedrock of northeastern Ohio.

Tectonics and Seismicity. No major faults have been mapped in Ohio (Figure C.68). Those that have been described are of small displacement and limited geographic extent. The nearest faulting of any significance to the Barberton area is associated with the graben-like features found in the Cleveland mine of the International Salt Company. Most of the faulting at this site is very small scale, but one fault in the mine has a displacement of 14 m.

Jointing in the rocks of northeastern Ohio comprise two sets that have orientations of N 25° - 80° E and N 25° - 80° W.⁽¹⁴²⁾ The joint system at the surface correlates well with the strain directions ascertained for the rocks within the Barberton mine.

FIGURE C.66. Sections of Allegheny Synclinorium and Findlay Arch, Along a Line Extending from Eastern West Virginia to Northwestern Ohio; (A) Structure Section; (B) Stratigraphic Diagram Along Same Line as Structure Section, With Effects of Deformation Eliminated; Section (B) Demonstrates that the Apparent Thinning of the Formations in Southeast Part of Section (A) is Due to Exaggeration of Vertical Scale, and that the Rocks of the Synclinorium were not Laid Down in an Independent Depositional Basin. Compiled From Various Sources.

FIGURE C.67. Cross Section of the Paleozoic Rocks of Northeastern Ohio, Distinguished by Age; the Line of Section is Determined by the Well Locations Shown on the Upper Map; Note that the Vertical Scale of the Cross Section is Enormously Exaggerated. From Banks, 1970.

FIGURE C.68. Map of Ohio Showing Structural Features; the Heavy Broken Lines Indicate the Major Structures; the Minor Folds are Indicated by Broken Lines Representing the Crests of the Anticlines and Solid Lines for the Crests of the Synclines; the Approximate Position of the Faults, with the Exception of the Bowling Green Fault, is Indicated by the Letter F; the Joints in the Rock Formations are Indicated by Solid Straight Lines Tipped by Barbs; the Joints in the Coal Beds are Straight, Light Dotted Lines. From VerSteeg, 1944.

This portion of Ohio falls within USGS seismic risk zone number one: minor earthquake damage. Primer⁽¹⁴³⁾ reports five earthquake epicenters with intensities of 5-6 (MM) in the northeastern portion of Ohio. He attributes the cause of these earthquakes to salt mining operations in that region.

Hydrology. Glaciation has had a major influence on the development of the present hydrologic regime in the Barberton area. Surface drainage has been modified by both the erosional and the depositional effects of the ice sheets. Pre-glacial valleys were scoured and later filled with drift prior to the development of the present drainage which has evolved over the last 11,000 years. A major divide near Barberton separates northward drainage to Lake Erie from southward drainage to the Ohio River via the Tuscarawas River.

Subsurface water is also influenced by the glacial deposits. In general, the widespread lacustrine deposits associated with former glacial lakes and the ground moraine have low permeability and hence contain little or no available water. On the other hand glacial outwash (stratified drift) deposits are considered the best aquifers in the region, especially in the form of valley drains filling buried valleys now occupied by present surface drainage.

Yields of water from wells in bedrock are generally much lower than from glacial outwash deposits, but are considerably better than from the till deposits. Pennsylvanian sandstones and conglomerates are the leading aquifers. Well logs of deep wells throughout much of Ohio indicate that no water is encountered in the Devonian-Mississippian shale sequence. All waters reported in these logs, and supposedly these are brines, occur within the driller's "Big Lime" - the Silurian-Devonian carbonate sequence. The so-called "1st water" occurs at the ORISKANY horizon, and the "2nd water" is reported below the salt-bearing Silurian strata at the "Newburg" horizon.

Halite stalactites forming around drilled pressure-release holes in the mine roof at Barberton indicate that connate water is present in the upper Columbus-Delaware Limestone or in the overlying Huron Shale.

The Kansas City Limestone Mines

The lower 3.6 to 4.3 m of the Bethany Falls limestone is commonly mined leaving approximately 25% behind as pillars. Features of the overlying strata that contribute to the development of underground space are illustrated in Figure C.69. Usually the upper 1.5 to 2.1 m of the Bethany Falls must be left intact to assure roof stability. Also, the Galesburg Clay, the Stark Shale, and a full thickness of the Winterset Limestone are considered important in the mine's stability. The clay and shale serve as an aquiclude. The "peanut zone" shown in Figure C.69 is apparently very unstable and must be left intact.

FIGURE C.69. Graphic Column Showing the Bethany Falls Limestone and Associated Strata that Figure Significantly in the Development of Underground Space.
From Gentile, 1975.

The Hushpuckney Shale at the base of the Bethany Falls normally forms the floor in the excavations. It has created some problems by arching or heaving between pillars - in some cases several inches over a period of a few years. Explanations offered for this phenomenon include: (1) expansion due to pressure release when the limestone is removed, (2) volume increase due to sulfate minerals forming from the weathering of sulfide minerals, and (3) expansion of the shale due to water absorption.

Geomorphology. Kansas City is located at the confluence of the Kansas and Missouri Rivers athwart the physiographic boundary that separates the Dissected Till Plains to the north and the Osage Plains to the south. The Kansas glacial terminus north of the Missouri River marks the edge of a maturely dissected area of till and loess. The Osage Plains south of the Missouri River is underlain by some 760 m of gently dipping Paleozoic strata that have been differentially eroded to form rather abrupt east-facing scarps. The Bethany Falls Limestone forms one of the most prominent scarps in the northwestern portion of Missouri.

Glaciation has had an impact on the drainage and on the present topography over at least the northern portion of this region. Glacial drift ranging up to 18 m thick covers a subdued pre-glacial topography that is not unlike the present surface of the Osage Plains. Loess mantles much of the land surface - averaging 90 to 120 m in thickness, but locally up to 30 m. Loess deposits thin to the northeast decreasing to less than 1.5 m thick just south of the Missouri and Kaw Rivers.

Bedrock Geology. The rocks of major concern in the Kansas City investigation comprise the Kansas City Group of the Pennsylvanian Missourian Series (see Figure C.70). Strata of this series as well as the subjacent and superjacent Pennsylvanian series represent marine and non-marine sediments deposited in cyclic sequences. The tectonic elements of the Midcontinent during this span of time were active enough to provide highland sediment sources and subsidence for marine waters. Paleo-environments represented by these rocks are: marine shelf, fluvial, deltaic, and swamp. Irregular subsidence during deposition created cyclic patterns of rock types vertically as well as an array of complex lateral facies. A map (see Figure C.71) by Wanless, et al⁽¹⁴⁴⁾ illustrates the complexity of lateral variations of rock types rather typical of midcontinental Pennsylvanian strata. His map portrays only one cyclothem out of many that were superposed during Pennsylvanian time.

The Kansas City Group at Kansas City is about 75 m thick of which about half consists of limestone units ranging from 0.9 to 9 m in thickness. The Kansas City Group thins in northeastern Kansas to about 50 m and increases in thickness to 180 m in southern Kansas. Three subgroups are recognized in the Kansas City Group: Bronson, Linn, and Zarah.

FIGURE C.70. Columnar Section of Kansas City Group in Kansas City, Identifying Subgroups. From Parizek and Gentile, 1965, After Howe, 1961.

FIGURE C.71. Lithofacies Map of Brercton Cyclothem. From Wanless, et al., (1963).

In the Bronson subgroup the Hertha, Swope, and Dennis formations are dominantly limestone. The Sniabar Limestone Member of the Hertha Formation occurs directly above the Swope Formation and is important in the roof stability of the underground mines in the Bethany Falls. The total thickness of the Bronson in the Kansas City area is 20 to 25 m, but it ranges up to nearly 60 m thick in the subsurface.

The Linn subgroup overlying the Bronson is mostly shale with relatively thin, but areally persistent limestone interbeds. Total thickness for the subgroup ranges from 20 to 30 m. Several of the lithic units of the subgroup change in thickness in short lateral distances in the Kansas City area. Westward in Kansas the Noxie Sandstone Member of the Chanute Shale fills erosional channels cut as deep as the upper part of the Winterset Limestone.

Above the Linn subgroup is the Zarah which in the Kansas City vicinity is dominantly shale; however, the Argentine Limestone Member of the Wyandotte Formation is frequently exploited for aggregate in Kansas. The limestones of the Wyandotte Formation (Argentine and Farley) are frequently not present in outcrops in northern Missouri and southeastern Kansas. Sandstone is abundant in parts of the uppermost section of the Zarah formation. The Swope Group consists of the basal Middle Creek Limestone, overlain by the Hushpuckney Shale, and the top Bethany Falls.⁽¹⁴⁵⁾

The Middle Creek Limestone is bluish gray, hard, fine-grained and fossiliferous. Surface outcrops break along vertical joints into rectangular shaped blocks. The Middle Creek, which extends from Iowa to southeastern Kansas, comprises an 20 to 30 cm thick single bed or two beds separated by shale that total 75 cm thick.

The Hushpuckney Shale is 1.2 to 1.7 m thick at Kansas City. The lower part of the member is dark and fissile and contains sporadic gray phosphatic nodules between the shale laminae. The upper phase is micaceous and calcareous, and weathers into small chips. Commonly, when a greater than normal ceiling height for underground openings is required, the Hushpuckney Shale is excavated down to the Middle Creek.

Two distinct limestone phases of the Bethany Falls in the Kansas City region have a combined range in thickness from 6 to 7.5 m. Over the outcrop area from northwest Missouri to southern Kansas, the Bethany Falls ranges from 3.6 to 9 m thick. The lower phase of the Bethany Falls is fine-to-coarse-grained gray to chalk-white limestone. Discontinuous shale partings separate the limestone beds; stylolites and dendrites are commonly present. This lower phase is more fossiliferous than the upper phase. The upper phase occurs above a thin calcareous shale usually about midway in the section. The upper Bethany Falls limestone phase ranges from 3 to 3.6 m in thickness

and is thick-to-massive bedded, mottled or chalk-white. The upper phase commonly has scattered oolites; some calcite veinlets, vugs, and dendrites also occur.

At Kansas City, the Galesburg Formation, next unit above the Bethany Falls, is made up of 0.9 to 1.2 m of gray or dark gray, blocky shale with poorly defined bedding. Upper layers of this shale are carbonaceous.

The Dennis Formation consists of the Canville Limestone, Stark Shale and Winterset Limestone. The Canville at Kansas City is about 5 cm of dark blue, fossiliferous limestone near Pleasant Hill and some thin, hard, coarse-grained fossiliferous beds near Tarsney in eastern Jackson County. The Canville is generally absent in Missouri and the Stark Shale rests directly on the Bethany Falls. The Stark comprises 0.6 to 1.2 m of shale in Kansas City; the lower part is dark, platy shale containing sporadic phosphatic nodules along the bedding planes. The upper shale phase is black, brown, or gray clay shale. The clay shale content of the Stark Shale creates a condition of low permeability and thereby plays an important role in preserving the dry status of the excavations in the underlying Bethany Falls limestone.

The Winterset Limestone Member of the Dennis Formation is 6 to 10 m thick in Kansas City and nearly 15 m thick in eastern Kansas. The Winterset has a variable lithology, texture and bedding. The thickness of the beds may be from a few inches to several feet. The Winterset is usually bluish-gray, with algae vugs lined with calcite; it is commonly oolitic near the top. Nodular and discontinuous bands of blue or white chert occur in layers of hard, dark blue and siliceous limestone in the upper half of the member. Locally thin shale partings separate individual limestone beds; some shale beds in the Winterset are 0.6 m thick. The primary importance of the Winterset to the occurrence of dry and stable excavations in the Bethany Falls is that, where full thickness of the Winterset exists, it serves both to preserve the stability of subjacent excavations and to anchor roof bolts penetrating from the ceiling rocks of the Bethany Falls.

Tectonics and Seismicity. The Kansas City area is part of the "Prairie Plains monocline." Exposed Pennsylvanian strata dip 1.1 to 3.8 m per kilometers westward and northwestward. This regional structure developed in post-Pennsylvanian-Permian time and pre-Cretaceous time.

A limited amount of faulting can be seen in outcrops surrounding the Kansas City area, and numerous faults have been postulated on the basis of subsurface maps constructed on several horizons above and below the Bethany Falls limestone. Faults with displacements up to 9 m have been mapped in Johnson County, Kansas. Minor folding has been postulated on the basis of subsurface contour maps. Most workers attribute the mild structural deformation to differential compaction and possible recurrent movement in the basement rocks throughout Paleozoic time.

Jointing is a prominent feature in Bethany Falls outcrops. Large blocks (3 to 6 m in width and breadth) occur along valley slopes. Only one joint set exists throughout the region - N 30° - 60°E. Near mine portals joints are open and show signs of solution. Water often drips through these open joints. Within the mines the joint set persists, and although some excavations reportedly have no history of seepage, irregular surfaces of joints and the presence of a white precipitate suggests solution.

Goebel, et al⁽¹⁴⁵⁾ report that region has been free of destructive earthquakes, and that the earthquakes that have occurred have been infrequent and of low intensity. The Kansas City area lies within seismic risk Zone 2 - minor damage.

Hydrology. In many respects the ground water regime in the Kansas City area is similar to the Barberton, Ohio region. Valleys filled with glacial outwash hold large quantities of potable water. Yields of 1900 to 3800 lpm have been reported from some wells.

The major water sources in the region are the alluvial deposits along the Kansas-Missouri River systems. Yields from such deposits are commonly 1100 to 3000 lpm, but range from less than 380 lpm to 7600 lpm. Much of the alluvium is glacial, and it was deposited by meltwater during the waning of the Wisconsin ice.

Superjacent to Pennsylvanian bedrock on the uplands and slopes south of the glaciated terrain is a regolith that consists of colluvium, slope wash, residuum, and thin loess which serves as a source for shallow domestic and stock water supplies. Water moving through this regolith recharges ground water in the bedrock.

Yields of water from bedrock wells in the region are small and usually of poor quality. Bedrock wells around Kansas City generally yield less than 190 lpm, with 90% yielding less than 38 lpm, and 50% having sustained yields of less than 11 lpm. Shales are generally considered aquicludes; however, the Hushpuckney Shale subjacent to the Bethany Falls is frequently an aquifer. Fractures believed to be caused by shrinkage of colloidal matter in the black sediment and the regional joint set create a permeability suitable for yielding from 0.5 gpm to 5 gpm.

Limestone aquifers locally provide sources for small domestic supplies of water. Wells penetrating limestone aquifers are usually less than 30 m deep, but some limestones are known to produce water from as deep as 70 m.

The Bethany Falls Limestone supplies water from a depth of 60 m in northeastern Johnson County, Kansas.

Criteria for Dry Limestone Excavations

Although the two sites reviewed in this section exhibit certain similarities - geomorphological status, presence of impermeable shales, general structural setting, and low seismic risk - the geology of the Barberton site appears to offer the greater potential for dry storage over a longer duration of time.

Both mine sites have impermeable shale above the mine, but the greater stratigraphic continuity and greater thickness of the shales at Barberton offer a greater margin of safety. The general stratigraphy of Devonian-Mississippian rocks exhibits more lateral and vertical consistency than that of Pennsylvanian rocks of the Kansas City mines. For example, Goebel, et al⁽¹⁴⁵⁾ pointed out that a maximum thickness of the Winterset Limestone was necessary above the mine for maximum stability. The superjacent Noxie Sandstone Member can occur in scour and fill channels as low in the section as the upper Winterset. Also, the Bethany Falls interval grades southward into clastic rocks. These conditions would not only jeopardize the structural strength of the rocks, but also possibly create unfavorable hydrologic conditions.

The tectonic elements of the western and central Midcontinent region (Indiana to Nebraska) are more complex than those of the eastern Midcontinent (E. Indiana and Ohio) (Figure C.72). Recurrent movement in some of the tectonic elements may have occurred (Figure C.73).

A tentative set of criteria for locating and developing potentially dry subsurface space in carbonate rocks can be established based upon knowledge derived from the two investigations cited above. These criteria are:

- Low-angle regional dips; perhaps 5.7 to 7.6 meters per kilometer, but not in excess of 17 meters per kilometer
- Thick sequences of impermeable shale or comparable rocks that can serve as an impervious barrier above the excavations. The optimum thickness may depend on the composition of the shale as well as the presence of jointing.
- Glacial cover may enhance the potential for dry excavations. If the glacial drift is sufficiently high above the mined zone, it can serve in two general capacities: the ground moraine serves as a first-level aquiclude; and the stratified drift serves as a "sponge" retaining much of the meteoric water.
- Plateau-like regions are most desirable because they are generally related to the proper underlying conditions, i.e., ridges and valleys or mountainous terrains are usually related to deformed rocks.

FIGURE C.72. Major Structural Features of the Midcontinent.
From Snyder, 1968.

FIGURE C.73. A Schematic Map Showing the Intersection of the Pre-Pennsylvanian, Illinois-Wisconsin Arch with the Pre-Cretaceous and Post-Jurassic Siouxia Arch. From, Leversen, 1960.

- The area should be free from complex folds, and faults that are presently active or which have been active in the geologic past.
- The excavated carbonate rock should be massive, dense limestone or dolostone free from deleterious insoluble residues. The rock should have a compressive strength capable of withstanding the bearing-load of the overburden.
- Jointing should be absent. The depth of the strata to be excavated is an important factor as to whether or not open joints exist in the rocks.

C.2.10 Talc-Serpentine--Eastern U.S.A.

Major talc deposits in the Appalachian Mountain region are principally found in two contrasting geologic associations. In one type, the talc is associated with serpentinites, ultramafic and mafic igneous rocks, whereas in the other, the talc occurs with carbonate-rich sedimentary rocks.⁽¹³⁵⁾ Both kinds are confined to regions that have been folded and regionally metamorphosed. In order to gain a better perspective, certain general features of these two geologic associations are reviewed first.

The talc-serpentine-ultramafic rock association occurs within folded and metamorphosed terrane of the Piedmont, Blue Ridge, and New England Provinces. Excellent discussion about the general geological features of these rocks is given by Phillips and Hess,⁽¹³⁶⁾ Hess,⁽¹³⁷⁾ Chidester,⁽¹³⁸⁾ Cady et al.,⁽¹³⁹⁾ Chidester et al.,⁽¹³⁵⁾ and Miller et al.⁽¹⁴⁰⁾

The ultramafic bodies in this first case occur as hundreds of relatively small pods to larger masses several tens of miles in extent, and define a long linear belt extending along the core of the Appalachian Mountains from Newfoundland to Alabama. Within the United States, however, these ultramafic complexes exhibit significant differences in this trend.

Throughout Vermont and part of Massachusetts, numerous ultramafic bodies are present and consist of both relatively unserpentinized peridotite as well as completely serpentinitized and steatitized masses. In southern Pennsylvania and Maryland where large ultramafic bodies are found, parts of the largest bodies are relatively unserpentinized even though most consist dominantly of the mineral serpentine. Apparently steatite is not very common. By contrast, the ultramafic bodies in southern Virginia contain much serpentine and soapstone. The ultramafic units in northwest North Carolina are relatively unserpentinized peridotite and dunite; the bodies in southwest North Carolina and northern Georgia remain chiefly dunite. Toward the southwest, the ultramafic bodies are largely composed of soapstone.

Virtually all of the ultramafic complexes are more or less concordant with the foliation and layering of the enclosing rocks. In general, the age of the ultramafic rocks is uncertain but most are probably early Paleozoic in age (Miller et al., 1968). The bodies are typically intruded into metamorphic rocks of Precambrian and possibly Cambrian age, although some have been cut by pegmatites dated as late Paleozoic in age.

Talc and serpentine form by the metamorphic alteration of various ultramafic rock types such as dunites, peridotites, and pyroxenites. (135,136,138,139) Within this association, however, talc deposits can form as a consequence of two different geologic processes. In the first case, talc may be formed in relation to contact metamorphism produced by intrusion of granitic pegmatites into serpentinites; the talc occurs at the contact between the pegmatites and the serpentinites.

Most major talc deposits, however, are formed during regional metamorphism of ultramafic and mafic igneous rocks. In this environment, lenticular-shaped ultramafic bodies eventually become serpentinitized cores typically surrounded by successive shells of talc-carbonate and talc (steatite). The talc-carbonate zone is of variable thickness and may commonly completely encompass most of the ultramafic body. Although this rock is typically massive it may be cross cut by shears that extend into the adjacent serpentinite zone. The talc typically consist of shredded particles, whereas the carbonate forms irregular grains and masses and is dispersed throughout the matrix. The steatite zone commonly varies in thickness from 1 cm to 12 cm and consists of nearly pure talc. Although it is generally massive, this zone may locally be flaky or schistose. Because of its low resistance to weathering, the steatite zone is rarely seen in outcrop except where quarried. The adjacent country rock is usually characterized by concentric shells of chlorite and biotite rock (termed the "blackwall zone") at the contact with the steatite.

The general thought is that the serpentinite forms as a direct alteration of the primary ultramafic minerals and subsequently it is altered to talc and talc-carbonate through metasomatic reaction with the country rock during regional metamorphism. (141) According to Phillips and Hess, (136) the grade of metamorphism in general controls the nature and distribution of the talc as well as the composition of the outer "blackwall."

Talc deposits associated solely with mafic igneous intrusions also occur in a few localities. A notable occurrence of this type is a soapstone deposit at Schuyler, Virginia. Although steatitization has replaced much or all of the mafic igneous rock mass during either regional metamorphism or hydrothermal activity, there is generally no accompanying episode of serpentinitization. (135)

The talc-carbonate association is found within the Appalachian Mountain Region at these three significant localities: (1) Gouverneur District of New York; (2) Chatsworth District, Georgia; and (3) Murphy Marble Belt District in western North Carolina. The Gouverneur deposits occur in the Adirondack physiographic province whereas the latter two are found in the Blue Ridge Province. All three are important commercial deposits of talc.

Other well known examples of the talc-carbonate association occur elsewhere in the United States and commonly constitute valuable economic deposits. Among the latter are those commercially mined in the Ennis District of southwest Montana and in the Inyo-Panamint District of eastern California.

The talc associated with sedimentary rocks is usually contained within dolomitic marble, although some deposits also occur in silicate rocks. In this sedimentary setting, the talc bodies are stratigraphically controlled, although structural controls can aid in localizing the talc within favorable stratigraphic intervals. Either low-grade regional or contact metamorphism from adjacent intrusives can be responsible for the introduction of silica (SiO_2), and in some cases magnesium oxide (MgO), into the carbonates to form the talc.

Commercial deposits of talc in this association vary widely in mineralogy and fabric. Some bodies contain relatively pure talc, whereas others have substantial quantities of unaltered host rock and/or such alteration minerals as chlorite, carbonate, quartz, and tremolite. As a result, the talcose rock can vary from fine to coarse-grained and from dense-massive to schistose.

Structural Setting

Inasmuch as the linear talc-serpentinite-ultramafic rock association is confined to the Piedmont, Blue Ridge, and New England Provinces, the Chatsworth and Murphy Marble Belt districts occur in the Blue Ridge, and the Gouverneur deposits lie exclusively within the Adirondack Province, only these areas' structural aspects are discussed in detail.

The general structural features of the Blue Ridge, Piedmont, and New England Provinces in the Appalachian Mountain Region are extremely complex. An excellent summary of the structural geology of the first two regions is given by Miller and others,⁽¹⁴⁰⁾ while most structural features within the New England Province are discussed by Zen⁽¹⁴²⁾ and Cady.⁽¹⁴³⁾

Major folds and low-angle faults are prominent along the northwestern border of the Blue Ridge Province. Along these faults, Precambrian and Lower Cambrian rocks have been thrust northwestward over Paleozoic strata of the Valley and Ridge Province. In part, many of these faults cut larger folds, but in many places the faults are themselves folded. Smaller faults and folds

are abundant in the sedimentary and metamorphic rocks throughout this province as well as the adjacent Piedmont. Large anticlines, synclines, and domes are more prominent throughout the Blue Ridge Province. Little is known, however, of the largest structural features. Because of a lack of good exposures, there is less knowledge about the Piedmont along these lines than for the Blue Ridge.

We have more complete knowledge on one major structural feature, the Brevard Fault Zone. Although the exact amount of displacement is not known with certainty, this major lineament represents a feature of regional proportions in separating the Blue Ridge from the Piedmont. The fault is presently inactive and has been for the past 200 million years as shown by the deformed Triassic diabase dikes which cut rocks in the fault zone. Many other faults in the Appalachian Mountains, especially in the Blue Ridge and southern Valley and Ridge Province, are also inactive at this time. Fractures and incompetent gouge zones that are commonly associated with these faults provide zones of weakness which can potentially result in certain engineering problems.

The Green Mountain section of the New England Province structurally consists of a central anticlinorium flanked by a major, thrust-faulted synclinorium to the west. Along the eastern flank, homoclinal structure persists.⁽¹⁴⁴⁾ The basement rocks that form the core of the mountains are Precambrian metamorphics intruded by granites and mafic rocks. Overlying these Precambrian units is a broad zone of intensely folded and regionally metamorphosed Lower Paleozoic eugeosynclinal volcanic and sedimentary rocks. Ultramafic rocks were intruded into these rocks prior to metamorphism, and in turn these rocks were intruded by various igneous plutons of Middle to Late Paleozoic age. To the west equivalent-age sedimentary units are thrust-faulted and folded. The metamorphism and principal folding here represent a Middle Devonian event.

The latest structurally-related events in the Appalachian Mountain Region involve localized deposition of Triassic non-marine sediments in graben-fault basins within the Piedmont and associated intrusive and extrusive igneous activity. Throughout the Cenozoic, regional uplift and possible tilting undoubtedly modified this mountain trend as it experienced continued erosion.

The Adirondack Province regionally shows a general structure consisting of anticlinal and synclinal folding. Numerous high-angle normal faults are particularly common on the eastern and southern peripheries of this province. These faults are thought to be of Precambrian age although there is an indication that some have been reactivated in Paleozoic or possibly Mesozoic times.⁽¹⁴⁵⁾

The existence of earthquakes within the Appalachian Mountains indicates that some fault movement and deformation may still be occurring. In general, these seismic events are not directly associated with any identifiable structural features⁽¹⁴⁶⁾ although minor faulting appears to have some seismic association.⁽¹⁴⁷⁾

Major Talc-Serpentine Regions and Districts

Vermont Region. The ultramafic rocks in Vermont define a linear belt, elongated in a north-south direction, that extends through the central part of the state (Figure C.73). Most bodies lie east of the axis of the Green Mountain Anticlinorium, and dip steeply eastward.

The ultramafic complexes are in general pod-like to lenticular bodies that range from 9 m to 10 kilometers in length and vary in thickness from a few meters to 1500 m. Most are, however, not longer than 300 m nor thicker than several hundred feet.

Faults in the region are generally small and represent subordinate features. As Chidester and others⁽¹⁴⁸⁾ have however noted, minor faults are found in many ultramafic bodies exposed in talc mines. These faults typically strike parallel to the elongation direction of the ultramafic bodies, are essentially vertical and do not appear to affect the country rock appreciably. Zones of gouge and fault breccia as much as 1.2 m wide can be found along some faults although total displacement in most cases remains indeterminate. The suggestion has been made that these faults probably reflect a tendency of the steatitized ultramafic bodies, in contrast to the country rock, to deform by slippage rather than by flexure-rupture.

The ultramafic complexes exhibit a wide variation in their degree of alteration and much of the information in this section has come from the extensive work on these rocks by Cady and others.⁽¹³⁹⁾ These bodies consist either of largely unserpentinized dunite or peridotite, or a completely serpentinized mass with talc-carbonate and steatite rims. Several large bodies consisting of only talc-carbonate rock also are known. In contrast, smaller completely serpentinized and extensively steatitized bodies occur in central Vermont, but since erosion here is not as deep, complete relations are not observable. Many very large, partially serpentinized ultramafic bodies are characterized by a reddish-brown weathering surface called "buckskin". These bodies typically have central cores of massive, dark-green antigorite with minor accessory and relic minerals. The usually poorly exposed margins consist of schistose serpentinite, composed of antigorite and chrysotile with minor brucite and talc. Hard, dense and white calc-silicate rocks called

rodingites may exist in contact with the country rock. Within these bodies there can be appreciable slip and cross-fiber chrysotile asbestos. The veinlets, up to 2.5 cm in width, cut across other minerals. This occurrence of asbestos does not persist in completely altered bodies.

The completely serpentinitized ultramafic bodies typically contain schistose serpentinite cores, with zones of talc-carbonate and steatite developed along the outer margins.

The sheared and schistose serpentinite and steatite are rarely exposed, although the talc-carbonate rock can be resistant to weathering.

The largest bodies of both types occur in the northern part of the state near the Missisquoi Valley area. Although most of the ultramafic complexes are only partially serpentinitized, several large talc-carbonate and serpentinite bodies are present here. One large north-south trending talc-carbonate body lies just east of Troy Village and has a surface exposure four miles long and three-tenths of a mile wide. This particular deposit, however, has not proved satisfactory for commercial development and only one small prospect pit has been driven into it. Two other talc-carbonate bodies occur adjacent to Belvidere Mountain, but each is too small for further consideration.

Several large asbestos quarries are located in this general area along the southeastern slopes of Belvidere Mountain near Lowell and Eden. About 30% of the asbestos produced in the United States comes from this locality which is currently being mined by the Vermont Asbestos Mines Division of the Ruberoid Company. The principal source rock for this commercial asbestos is schistose serpentinite, located along the borders of the ultramafic bodies.

The largest ultramafic body in Vermont is located in the southern part of the state near East Dover. This complex is about four and one-half miles long and approximately one mile wide. According to the Vermont Geological Survey,⁽¹⁴⁴⁾ much of the southern part of this body is only partially altered while the northern part consists of serpentinite and talc-carbonate. The serpentinite is also cut by numerous talc-carbonate veins which northward make up most of the periphery.⁽¹⁴⁹⁾

The numerous mines, prospects and potential deposits in Vermont are discussed by Chidester and others.^(148,135) Several of the larger mine operations are worthy of some comment here.

The Johnson Mine, operated by the Eastern Magnesia Talc Corporation, is the largest in the state with an annual production approximating 45,500 metric tons per year. The deposit being worked is a steeply dipping serpentinite and talc-carbonate lens enclosed in schistose country rock. Measuring some 300 m

in length and 180 m in width, this body is about 25 m thick. Current mining is nearly 150 m below the surface and is concentrating on the recovery from the talc-carbonate zone.

The Waterbury Talc Mine at Moretown is operated by the same firm, but averages only about one-half the annual production of the Johnson Mine. This mine is located within a vertically pinching lens some 1370 m long and more than 150 m wide. Mining is currently more than 30 m below the surface in a body which extends nearly 220 m in depth. Talc-carbonate and steatite zones, both of variable thickness, are present. The country rock consists of schists and greenstones.

Near Chester, the Vermont Talc Company produces some 4500 metric tons of talc annually from the Windham Mine. Talc is currently being mined from two 300 meter long lenses which are 60 and 100 m wide respectively. At this locality, talc-carbonate forms outer shells approximately 45 m thick around each of the serpentinized bodies.

Although not currently being exploited commercially, the Barnes Hill prospect has been well-studied by Chidester.⁽¹³⁸⁾ The geological relationships, as shown on Figure C.74, illustrate well the nature of a completely serpentinized and partially steatized ultramafic body in Vermont.

Southeast Pennsylvania-Maryland Region. Many large serpentinite bodies and several talc deposits are located within the Piedmont Province of southeast Pennsylvania and Maryland. These bodies occur within metamorphosed terraine consisting dominantly of schists and metavolcanic units. As indicated by the Maryland Geological Survey⁽¹⁵⁰⁾ and the Pennsylvania Topographic and Geologic Survey,⁽¹⁵¹⁾ several distinct, large serpentine bodies with associated talc occur within this district.

The first is a serpentinite body oriented in an east-west direction along the Pennsylvania border with northeast-striking extensions into Maryland. The talc deposits associated with this body are called the State Line District north of the Susquehanna River, and the Jarrettsville-Dublin District south of this river in Maryland.⁽¹⁵²⁾ This body appears as nearly continuous for approximately 32 km in length. Width varies up to 2.4 km. A smaller body approximately 6.4 km long by 1.1 km of a mile wide is also present in the State Line District.

This major ultramafic body is surrounded by country rock consisting of greenschist-grade metagabbro and schist and has been described by Lapham and McKague⁽¹⁵³⁾ and Lapham⁽¹⁵⁴⁾ as consisting of massive to schistose, highly serpentinized peridotite within much of its exposure in Pennsylvania, especially along its northern boundary. Talc and vermiculite also occur in this area, and some chromite was economically recovered in the past from this

FIGURE C.74. Index Map of Vermont Showing Occurrences of Ultramafic Bodies and Location of the Barnes Hill Talc District.

district. Along its southern contact, the ultramafic body is only partially serpentized. In Maryland, the western portion is completely serpentized and locally altered to talc.

Within the southern district, several additional ultramafic bodies, occur north of this principal serpentine occurrence. The largest is up to 3.2 kilometers long and 1.6 kilometers wide, and most are completely serpentized. One relatively large talc deposit that has been recently mined occurs near the town of Dublin, and exhibits a peripheral schistose talc zone up to 180 m in width.⁽¹³⁵⁾ This deposit is associated with a late-stage pegmatite which has intruded the serpentinite.

Several distinct ultramafic bodies in the Baltimore County region have been studied.⁽¹⁵²⁾ Along the western edge of Baltimore County and southward through Howard County, a 3.2 kilometer wide serpentine body extends for nearly 30 kilometers. This body is aligned in a northwest orientation and dips 30 to 40 degrees to the northwest. The serpentized ultramafic rock displays borders of talc schist and chlorite blackwall, and has been commercially mined near Marriottsville where a talc body extends for nearly 180 m.⁽¹³⁵⁾

In the general City of Baltimore region, there exist several ultramafic bodies described by Pearre and Heyl.⁽¹⁵²⁾ One deposit is part of the Baltimore gabbroic complex and is only partially altered. The other, referred to as the Bare Hills deposit, is well-jointed and has a center altered to soapstone. Numerous chromite prospects occur in this occurrence.

Several elongated ultramafic bodies intruded into schist also occur in Montgomery County.⁽¹⁵⁵⁾ These bodies, the largest of which is 6.5 by 1.6 kilometers in principal direction, are described as containing massive serpentinite, thin zones of schistose antigorite, and peripheral talc-carbonate rock. Some small soapstone lenses are also abundant from this apparent alteration of original gabbroic intrusions.

Schuyler District, Virginia. Although not well described in detail, a belt of ultramafic rocks exists within the Piedmont Province from the North Carolina border to the Potomac River near Washington, D.C. Within this trend, important deposits of talc and serpentine occur near Schuyler, Virginia.

At Schuyler, described in detail by Hess⁽¹⁵⁶⁾ and Chidester and others,⁽¹³⁵⁾ there are five separate parallel units of metagabbro intruded into schistose and gneissic bedrock. These intrusions, which are parallel to the regional structure, have subsequently been folded into a northwest-plunging anticlinal-synclinal structure on whose southeast limb are exposed the talc deposits. These talc deposits range in thickness up to 55 m and reach about 450 m in length. The dip is eastward at an angle of 45 degrees.

Hess⁽¹⁵⁶⁾ suggests that the original igneous body was a differentiated hornblende gabbro that varied from ultramafic material near the bottom to areas of felsic segregations near the top. As these bodies were affected by regional metamorphism, the ultramafic section altered to talc-chlorite and talc-carbonate material, whereas the more gabbroic portion was altered to nontalc bearing rock. The soapstone contains a broad mixture of materials including actinolite, chlorite, talc and carbonate. Some large soapstone occurrences have been mapped in areas adjacent to the North Carolina state line.⁽¹⁵⁷⁾ Although these bodies range up to 13 kilometers in length, none have ever been described in detail.

Total production of talc from three of these metamorphosed intrusions near Schuyler approximates two million metric tons. At present, a division of the Georgia Marble Company quarries some soapstone in the area for dimension stone.⁽¹⁵⁸⁾

Although the principal talc deposits of North Carolina occur in the Murphy Marble Belt District, the geologic map of North Carolina (North Carolina Division of Mineral Resources, 1958) shows numerous, relatively large ultramafic bodies within a belt approximately 55 kilometers wide throughout the Piedmont and Blue Ridge Provinces of western North Carolina and northeast Georgia. These occurrences have linear dimensions up to 3.2 kilometers and are generally elongate.

Most are relatively unserpentinized, although many of the smaller bodies have been completely altered to soapstone.⁽¹⁵⁹⁾ For the most part, the original igneous material was intruded into hornblende gneisses and schists, being elongated in a direction coincident with the lineation of the gneiss.⁽¹⁶⁰⁾

According to the detailed descriptions of these ultramafic bodies as given by Hunter,⁽¹⁶¹⁾ many of the larger ones appear similar. All are characterized by nearly pure dunite cores with serpentinized margins; the degree of serpentinization probably averages close to 20 percent. The outer portions of the partially serpentinized zones are typically schistose, and commonly composed of talc and vermiculite at the country-rock contact. On the average, these occurrences typically contain approximately 40% dunite, 55% serpentinized dunite and five percent talc-vermiculite.

In virtually all the larger ultramafic bodies, the talc-vermiculite zone is characteristically thin, being no thicker than 30 m in any case, but averaging close to 15 m.

Although talc is not mined from these rocks in western North Carolina, appreciable corundum was mined there many years ago. More recently, some olivine for refractory use was mined, and disseminated chromite occurs in subeconomic concentrations in several dunitic bodies.

Major Talc-Carbonate Districts

Murphy Marble Belt, North Carolina-Georgia. This district forms part of a narrow linear belt, approximately 135 kilometers long, that extends northeast from Canton, Georgia to near Wasser, North Carolina.^(159,162) Talc deposits occur within the Murphy Marble Unit, which in part represents the overturned, northwest limb of a large anticline.

Figure C.75 presents a geologic map of this district together with several regional cross sections that show the local structure. Commercial talc deposits are confined to a zone of white, fine-grained dolomitic marble near the center of the Murphy Marble Unit. These talc deposits, which range in size from small pods to lenses as large as 210 m in length and 43 m in thickness, have their long axes parallel to the plane of bedding. Most of the talc is relatively pure and exhibits a flaky and slightly fibrous habit. Some is also fine-grained. Minor impurities include quartz, carbonate and tremolite.

Van Horn⁽¹⁶²⁾ believed that these talc deposits originated as the result of two distinct episodes. The first involved the regional metamorphism of a slightly sandy carbonate horizon into a dolomitic marble. Later, hydrothermal solutions metasomatically replaced the dolomite along shear zones to give rise to the talc lenses.

Production from this district has yielded several hundred thousand metric tons of talc. Most of this production has come from several mines near the town of Murphy. Since several deposits contain very pure talc, crayon talc is also currently recovered.

Chatsworth District, Georgia. Talc deposits occur near the city of Chatsworth throughout an area of less than 25 square kilometers. The country rock in which these talc deposits occur consist of moderately folded Precambrian schists and gneisses of the Blue Ridge Province that are overlain by Precambrian slates, sandstones and dolomites. A major post-metamorphic thrust fault, which lies immediately to the west of this district, separates these Precambrian rocks from Cambrian shales of the Valley and Ridge Province. The talc forms lenticular bodies in a metasedimentary schist in close association with a coarse-grained gneiss. The host Cohutta schist appears as remnants in the gneiss leading Furcron and Teague⁽¹⁶³⁾ to consider this as evidence for the existence of a thrust fault in the vicinity.

In these remnants, the talc deposits irregularly pinch and swell in tabular bodies and lenses that range from a few to tens of meters in thickness and as much as several hundred meters in length. Many extend several hundred meters down dip. These deposits consist of a highly foliated and massive rock composed dominantly of talc, dolomite and chlorite and referred to

FIGURE C.75. Geologic Map of the Murphy Marble Belt District, North Carolina, Showing Principal Talc Deposits; Relationships on Two Enlarged Structural Cross Sections are also Shown.

commercially as grinding talc. Distinct lenticular masses ranging from one to tens of meters thick and a few tens of meters long of nearly pure talc (crayon talc) also occur within the deposits. Needham and Hurst⁽¹⁶⁴⁾ note that these bodies of crayon talc are generally located where the dip changes along flexures.

The origin of these bodies is somewhat debatable. Needham and Hurst⁽¹⁶⁴⁾ have assumed that the talc deposits originally formed from metasedimentary bodies of dolomite within the enclosing schist, although Chidester and others⁽¹³⁵⁾ suggest the possibility that these talc bodies may also have originally been derived from peridotites. In either case, it is fairly certain that the original rock was replaced by serpentine and/or chlorite that was in turn later replaced by talc.

Needham and Hurst⁽¹⁶⁴⁾ have mapped 26 remnant bodies although only three are currently being mined, and these have been responsible for the district's principal production. Aggregate production to date exceeds one million metric tons.

Gouverneur District, New York. The talc deposits in this district are described in detail by Engle⁽¹⁶⁵⁾ and Chidester and others.⁽¹³⁵⁾ These deposits occur in elongated zones within a northeast trending belt of impure marbles of the Precambrian Grenville series. This marble is highly deformed and represents a part of the metamorphosed strata that occur along the south-east flank of a northeast-trending anticlinorium. Within this marble unit, commercial zones of talc pinch and swell in a sinuous folded pattern that is roughly conformable with the adjacent marble layers.

Zones of talc generally have composite lengths greater than five miles, widths as great as 120 m, and may extend down dip as much as 760 m. The dip is variable, but averages around 45 degrees to the northwest. These zones appear to be localized with regard to secondary structures, such as shear zones, that probably developed during metamorphism.

A highly variable mixture of talc, tremolite, anthophyllite, serpentine, carbonate and quartz is present in the commercially worked deposits. In most the talc averages around 25%, but is subordinate to tremolite. The thought is that initially the tremolite and talc developed as the result of a metamorphic reaction between quartz and dolomite. These and other minerals, however, also appear to have formed from the introduction of hydrothermal fluids along shear zones.

In this district, which is the largest and has the most productive deposits of this type in the Western Hemisphere, there are a number of mines operating within two distinct talc-bearing units.⁽¹³⁵⁾ One deposit near Talcville is nearly 90 m thick and supports two active mines. In another talc zone near Balmat Corners, four mines are currently operating throughout

an interval which averages about 38 m in thickness. The Gouverneur District has cumulatively produced more than seven million metric tons of talc.

Hydrology

Surface Water. Although there are wide variations locally, the average precipitation throughout the Appalachian Mountain Region is 1.2 m annually. In general, rainfall is lower along the east side of the mountains and in the northern section where most talc and serpentinite bodies are known. That water which is not lost by virtue of evapotranspiration or percolation-infiltration into the subsurface becomes surface runoff and moves seaward along well-developed stream systems. In most areas discussed in this report, surface drainage generally flows eastward toward the Atlantic Ocean.⁽¹⁶⁶⁾ In northwestern Vermont, however, the surface drainage moves westward into Lake Champlain, and then northward into the St. Lawrence River. Runoff from the western Adirondack Province in New York flows directly into the St. Lawrence River for eventual transit into the North Atlantic Ocean.

Because the Appalachian Mountains are generally well vegetated, soils are commonly well-developed and thick except for very steep or rock-strewn slopes, and forest litter and root networks are effective absorbing media, runoff from natural surfaces is not as pronounced as in some other areas. Disregarding the effects of urbanization and stream modifications for the moment, more water has an opportunity to be intercepted and absorbed into soil or other surficial debris. Appreciable stream flow may thus be ascribed to soil and ground water discharges in addition to conventional overland flow. These conditions lead to less drastic erosion and downcutting than experienced in less humid geomorphic settings.

Because of a wide range of variables, such as resistance to weathering-erosion, topographic position, geomorphic history in relation to stream-network development and the nature of enclosing and overburden rock units, no concrete statement can be made about the possible effects of surface water with regard to any specific talc-serpentine body. Data revealed no pronounced favorable or unfavorable examples among the bodies studied with reference to relative downcutting by streams, exhuming through stream erosion or diversion of streams due to especially resistant conditions. For this sort of information to be meaningful, each individual talc-serpentine body requires a more thorough study of the local geomorphic-stream cycle.

Ground Water. Within the Piedmont, Blue Ridge and New England Provinces, there are few highly productive ground water aquifers.⁽¹⁶⁷⁾ Several of the districts discussed in this report are however located in proximity to reasonably productive aquifers. These include: (1) the Gouverneur District

of New York, which lies astride a carbonate aquifer that extends along the northwest boundary of New York; (2) mafic-ultramafic bodies which are adjacent to important crystalline-rock aquifers that cover a relatively broad area in western North Carolina; and (3) the Murphy Marble District of North Carolina, which is associated with an important carbonate aquifer. In the latter locality, this aquifer, which is capable of producing upwards of 7500 liters per minute in wells and springs, is considered a favorable site for development of large ground water supplies.⁽¹⁶⁸⁾

Other areas within the Appalachian region also have the capability of producing moderate amounts of ground water. Within the Piedmont and Blue Ridge Provinces, the most important factors controlling ground-water yields are topography, the thickness of soil, and the amount of weathered residuum overlying the bedrock. In general, higher yields are normally attained in valleys and lowlands as compared to hills and upland slopes. These provinces are considered to be capable of yielding up to 95 liters per minute from individual wells.

The least productive ground water region is the Blue Ridge Province, which is underlain by dense, sparsely fractured crystalline rocks of low permeability. Soils are generally thin and the rocks weathered only to shallow depths. Meyer and others⁽¹⁶⁸⁾ have noted that little usable ground water is found below a few hundred feet. Wells in this region typically yield less than 55 liters per minute, although springs may be numerous after relatively heavy rainfall.

The Piedmont Province provides a somewhat more favorable site for ground water because rocks here have experienced less erosion due to the more moderate topography and elevations. Hence, weathering has produced a thicker mantle of soil which overlies additionally fractured bedrock. Ground water thus can be more readily found in the lower part of the soil zone, in weathered saprolite or regolith and in the bedrock-fracture system. Because of depth limitations with regard to both weathering and effective fractures, water usually is not recovered in large quantities below about 90 m.

Ground Water in Talc Deposits. On the basis of the hydrologic properties of talc itself, talc mines are relatively dry. The water found in these mines comes predominately from either mine structures or dillholes exposed to the surface or from higher-level mine workings which transmit water from near-surface sources to deeper mine openings. Where the adjacent country rock is fractured or jointed and the talc has been mined directly up to the wall-rock contact, the influx of percolation into subsurface mine workings can be pronounced. Flows of up to several hundred liters per minute were observed at the Earnest Mine at its upper levels in the Chatsworth District.

At the Vermont Corporation mines, mining is currently underway on two 24-meter levels, the deepest being 50 m, within a talc-magnesite body 300 m long and 105 m wide. Under these conditions only a very small amount of water was issuing from joints and fractures in the adjacent chlorite-schist country rock. Similar circumstances are observed at the Eastern Magnesium Talc Corporation mine at Johnson, Vermont and the Windsor Minerals, Inc., mine near Windsor, Vermont.

Ground-water influx is a minor problem in the mines around Murphy, North Carolina. Any water which flows into the workings issues from joints and minor solution cavities within the adjacent marble. The talc body itself was noted as being quite dry.

Although the majority of water observed in the Chatsworth mines came from higher-level openings or rock fractures in the breached contact zone, some very slight seepage was observed from fractures and faults in the host talcose schist. These occurrences, some 60 plus meters down in the Earnest Mine, were rare and yielded only droplets of water. The talc bodies themselves were essentially dry.

Most observations thus suggest that talc itself is for the most part extremely dry. Even if the country rock can transmit appreciable water, ground-water seepage into the talc appears minimal if proper mine design is practiced and any surrounding contact zone is not severely breached. The low permeability of talc may be related to its very weak physical strength which appears to result in appreciable self-sealing through slippage and possible flowage. The pureness of the talc is also important.

Ground Water in Serpentinite Deposits. The permeabilities and porosities of serpentinite are largely unknown inasmuch as few underground structures or mines exist in this rock mass. Commercial operations in this rock are mainly open pit. In addition, no data concerning laboratory measurements of permeability and porosity in either serpentinite or talc were found in the course of this study.

Based on field observations, however, the suggestion is that partially serpentinitized ultramafic rocks and serpentinites composed mainly of massive antigorite are structurally sound and exhibit physical strengths similar to many other competent crystalline rocks. On the other hand, schistose serpentinite appears to be structurally much weaker, having physical properties more akin to those of schistose talc. More investigation regarding the permeability of schistose serpentinites and altered igneous bodies in general appears in order to ascertain the response of such rocks to ground water inflow.

C.2.11 Metamorphic Rocks-Canadian Shield

Each of the continental masses of the earth is characterized by a nucleus of very old Precambrian rocks which are exposed at the surface and known as shields. The name Precambrian simply indicates that these rocks are older than the fossiliferous Cambrian strata. Radiometric age measurements place the time boundary between the Cambrian and older series at approximately 570 m. y. (10^6 yrs.) ago. The oldest dated Precambrian rocks are 3600 to 3800 m. y. old, and in Greenland, Labrador, and Minnesota considerable work has been done to establish their age and to decipher the complicated geologic history.

The principal areal exposure of Precambrian rocks in North America is in Canada and measures many thousands of square kilometers ($>1.8 \times 10^6$ square miles); hence the shield in North America is referred to as the Canadian Shield. Figure C.76 is an old but useful illustration showing how some of the larger concepts relating to the development of the Canadian Shield evolved. The principal geologic province and the approximate age or range in age of the rocks are shown. The Canadian Shield is roughly an annular exposure around Hudson Bay. As shown in Figure C.76 in the vicinity of Hudson Bay and on the outer perimeter the Precambrian rocks of the Shield are covered with relatively thin layers of Paleozoic and younger sedimentary rocks. Thus drill holes through the younger rocks encounter the Precambrian rocks at depths roughly proportional to the distance away from the Shield. The greater part of the North American continent is underlain by Precambrian rocks, but only in areas adjacent to the Shield is it common to refer to the Precambrian basement as the subsurface extension of the Shield.

The Canadian Shield (Figure C.76) extends into the northern parts of Minnesota, Wisconsin, and Michigan, and in New York in the Adirondack Mountains. Because of similarities in lithology, structure, and the common factor introduced by the large Precambrian cherty iron formations and the derived iron ores the region including Minnesota, Wisconsin, and Michigan has commonly been referred to as the Lake Superior district.

Physiographic

The international boundary commonly becomes a geographic limit in discussions pertaining to the Canadian Shield. The physiographic regions of Canada are shown in Figure C.77. Similarly, physiographic regions are recognized in the United States, and there are significant differences in the physiographic development of Canada and the United States; however the discussion here is concerned largely with the gross features. Most of Canada and a large part of the northern United States were glaciated during the Pleistocene (Figure C.78).

FIGURE C.76. The Canadian Shield Showing the Geologic Provinces After Collins, Farquhar, and Russell (1954); the Indicated Radiometric Ages are the Approximate Time of a Major Orogeny in Each Province and are Based on Data Prior to 1961; since then a Considerable Volume of New Data has Become Available and Many Changes Have Been Made. From Goldrich and Other, 1961.

FIGURE C.77. Physiographic Regions of Canada; the Relationships Between the Physiographic Subdivisions and the Geologic Provinces. From Douglas, 1968.

FIGURE C.78. Extent of Continental Pleistocene Glaciation in Northern North America. From Geological Survey of Canada, 1957.

The glaciers did considerable work in eroding and transporting materials and in levelling the region so that today it is characterized by relatively low relief, for the most part rarely exceeding 100 m. Elevations rising 200-300 m above the general level are prominent features, locally named mountains. The topography of the Canadian Shield, however, was largely the result of stream erosion and was developed in late Precambrian or early Paleozoic time. Glaciation has modified the ancient surface, reducing some of the highlands and filling in some of the lowlands.

Structural

Although the structure in the Precambrian rocks is complex, it has been intensively studied because of its close relationship and control of mineralization. Figure C.79 is a generalized map showing the structural trends in the Canadian Shield. On the basis of physiographic, structural, and also lithologic characters the Canadian Shield was divided into structural provinces and subprovinces (Figure C.80).

Geologic Provinces

Radiometric age measurements on mineral and rock samples from the Canadian Shield developed age patterns that further characterize the structural provinces, and on the basis of the various features mentioned above the Canadian Shield has been divided into geologic provinces which are essentially the same as the structural provinces (Figure C.80). For example the large Superior Province that extends south into the United States is characterized by east-west structural trends of greenstone-granitic gneiss belts. These rocks were first formed 2600 m. y. or more ago. They were involved in the Kenoran (Algoman) orogeny and locally show the effects of metamorphism resulting from younger igneous activity, folding, shearing, and so forth.

Stratigraphy and Crustal Thickness

With the introduction of radiometric age determinations, correlation of formations in the Precambrian received great attention, and very considerable progress has been made in defining the Precambrian stratigraphic succession. This subject for all of the Canadian Shield is too large for review, and some attention is given to the Precambrian stratigraphy in the discussion of the Lake Superior region in a later section of this report. Canadian geologists have long used a two-fold classification of Precambrian rocks. The oldest rocks are the Archean and were first formed 2600 m. y. ago or prior to that time. These are the rocks of the greenstone-granitic gneiss belts mentioned above. The younger Proterozoic rocks are more platform-type of sediments deposited on the more or less stabilized Archean craton.

FIGURE C.79. Generalized Structural Trends in The Canadian Shield. From Douglas, 1968.

FIGURE C.80. Structural Provinces and Subprovinces of the Canadian Shield. From Douglas, 1968.

The east-west structural trend of the Archaen rocks is shown in Figure C.81. Estimates of the thickness are shown, but because of the complexity of the structure these figures should not be taken too literally.

Very little is known about the crustal thickness of the Shield, and our information at present is based on rather limited seismic studies. In general the crustal thickness is of the order of 40 km. (range 35-40 km). The thickness apparently increases beneath Lake Superior (50 km) and decreases beneath the Hudson platform at Hudson Bay (25 km).

Lake Superior Region

The Lake Superior region as used here includes adjoining Ontario as well as Minnesota, Wisconsin and Michigan. It is convenient to divide the Precambrian rocks into three major divisions used by the Minnesota Geological: Early, Middle and Late Precambrian with time boundaries of 2600 and 1800 m. y. ago. A similar three-fold division with somewhat different time boundaries was used by the Ontario Department of Mines in compiling the most recent geological map of the Province. The Geological Survey of Canada uses a two-fold division of Archean and Proterozoic with a time boundary of 2560 m. y. The Proterozoic is divided, from oldest to youngest, into the Aphebian, Helikian, and Hadrynian. The time boundary between the Aphebian and the Helikian is 1800 m. y., corresponding to the time boundary between the Middle and Late Precambrian. In the usage of the Geological Survey of Canada the terms Kenoran and Hudsonian orogenies correspond to the Algoman and Penokean orogenies.

Early Precambrian or Archean Rocks. In general the same stratigraphic elements listed in Table C.1 are found throughout the Lake Superior region, but the formal terminology differs in different areas. The Archean sequence is largely composed of basaltic and andesitic lavas; pyroclastic material ranging from rhyodacitic to rhyolitic in composition; graywacke-shale sequences altered to metagraywacke-slate; banded iron formation with minor carbonate and related sedimentary rocks. These rocks were folded and intruded by numerous dikes and large masses or plutons of tonalitic to granodioritic magma in an early phase of the Algoman (Kenoran) orogeny. The section in northeastern Minnesota, for example from oldest to youngest, includes the Ely Greenstone and Soudan iron-formation that were folded and intruded by numerous dikes of rhyodacitic composition, the Northern Light of Gneiss of trondhjemitic composition, and the Saganaga Tonalite. In the Rainy Lake district the Keewatin lavas that correspond to the Ely Greenstone are underlain by an older metasedimentary group, the Coutchiching Series of Lawson (1913). In the Rice Bay area of the Rainy Lake area the Coutchiching is 200 m or more in thickness.

FIGURE C.81. East-West Trend of Archean Structural Belts and Estimated Thickness of Archean Sequences.

In northeastern Minnesota and also in the Rainy Lake district in Ontario the early Archean volcanic and pyroclastic rocks were folded and intruded by numerous dikes and plutons of tonalitic composition. These masses started their ascent as magma but continued to rise as diapirs so that they were uncovered and provided coarse detritus to adjacent basins forming conglomeratic wedges in the Knife Lake sediments of northeastern Minnesota and in the Seine sediments in Ontario. The erosion surface or unconformity was recognized by Lawson in Ontario and by early workers of the Minnesota Geological Survey, and on the basis of geological relationships, the Laurentian orogeny was defined. The unconformity is well exposed in Cache Bay of western Saganaga Lake, and in the country to the south, and also on Shoal Lake and along Highway 11 in the Rainy Lake district. The conglomerates are wedges in the thick graywacke-shale sequences of the Knife Lake group in Minnesota and the Seine Series of Ontario.

Middle Precambrian or Aphebian Rocks. The general nature of the Middle Precambrian rocks in the Lake Superior district are best known for the extensively developed banded iron-formations. In Ontario, the type locality of the Huronian rocks, the sequence is dominantly elastic sediments with little or no iron formations. This fundamental difference in lithology is largely the basis for objection to the use of Huronian as a time-stratigraphic term. Radiometric dating of the Nipissing diabase,⁽¹⁶⁹⁾ in the Blind River-Bruce mines area, Ontario, places a minimum age of 2100 m. y. on the Huronian Group. Radiometric age determinations on argillites and rocks related to the iron-formations in the Labrador trough and in the Lake Superior district are considerably younger (1600-1850 m. y.), but these ages may simply reflect folding and low-grade metamorphism. Thus the Huronian Group of the type locality in Ontario may be older than the Animikie Group of the Mesabi and Gunflint districts, or it may simply be a coarse clastic facies and of similar age. For the present this problem remains unresolved.

Late Precambrian or Helikian and Hadrynian Rocks. The Late Precambrian rocks in the Lake Superior region are referred to the Keweenaw. The best known rocks are the Middle Keweenaw volcanic rocks, the North Shore Volcanic Group and the Duluth Gabbro or the Duluth igneous complex. A correlation chart for the Late Precambrian rocks in the Lake Superior region is given in Table C.3.

Igneous and Metamorphic Rocks. Three major periods of folding and igneous activity have been recognized in the Lake Superior region as follows:

- Keweenaw igneous activity (ca. 1100 m. y.)
- Penokean orogeny (ca. 1850 m. y.)
- Algoman orogeny (ca. 2700 m. y.)

During these periods the rocks described above were folded and intruded by magmas with the result that their primary features were changed in varying degree ranging from low-grade (greenschist facies) to high-grade (granulite facies) metamorphism.

The crystalline metamorphic rocks and associated igneous rocks are naturally extremely well developed in the Canadian Shield, and these present excellent prospects for excavations which should be well suited for the long-term storage of radioactive waste materials.

Structure. The structure of the Lake Superior region is complicated because of the three major periods of orogeny and igneous activity, but the major structural elements are fairly well known. Mountain building or orogeny accompanied by metamorphism and igneous activity develops crystalline metamorphic and igneous rocks that are more or less amenable to radiometric age measurement; therefore, it appeared logical to attempt to apply some of the principles developed in the study of geologically young mountains to the Precambrian.⁽¹⁷⁰⁾ Since this early work the concepts of plate tectonics have undergone rapid development. Present-day tectonic concepts are closely related to the movement of large plates with the major earth structures developing at plate junctions. It is not uncommon to see a statement to the effect that the plate-tectonics concept has revolutionized geology and completely displaced the geosynclinal hypothesis of mountain building.

The plate tectonic concept, however, is not easily applied to the ancient rocks of the Precambrian, and some sort of compromise is useful. Most geologists will accept the hypothesis of mobile zones common to both the geosynclinal and the plate-tectonic concepts. Thick wedges of immature and rapidly accumulated sediments and volcanic rocks typical of the eugeosyncline or of the active plate margins are an essential feature. The piles accumulate to thicknesses of the order of 30,000 to 50,000 feet. Possibly in early Precambrian time, because of greater heat flow, the crust may have been thin and more flexible or mobile; nevertheless, the cycle worked out for more recent mountains applies to Precambrian belts, and the wedges or piles of conglomerate, graywacke-shale, and volcanic rocks were strongly deformed and depressed to depths in the mantle where partial melting generates tonalitic to granodioritic magma. These then moved upward invading the supracrustal cover rocks.

One feature of the folded mountain chains that is accepted by structural geologists is uplift and subsequent erosion exposing plutonic igneous rocks and metamorphic rock of amphibolite to granulite facies. Uplift or vertical translation was by blocks uplifted along faults. There is much evidence in the Precambrian rocks to support this model.

The development and application of radiometric dating techniques to Precambrian rocks have an interesting history. Once it was thought that concordant mineral ages by two or more techniques were definitely the age of crystallization or origin. Later it was discovered that both K-Ar and Rb-Sr mineral ages commonly date a metamorphic event subsequent to initial formation. Now it is widely accepted that K-Ar and Rb-Sr mica ages may date the time of uncovering, that is, the time of uplift and erosion that uncovered the presently exposed rocks to a depth at which the radioactive decay systems were stabilized in the minerals.

Early Precambrian. A large volume of radiometric age determinations on minerals and rocks in the Lake Superior region indicate that the interval of time for uplift and erosion required to stabilize the decay systems in minerals ranges from 50 to 150 million years. In the Early Precambrian rocks of the Lake Superior region, for example, whole-rock Rb-Sr isochron ages and U-Pb zircon concordia ages indicate that the complete cycle from sedimentation to collapse of the depositional trough with subsequent folding, intrusion, and metamorphism probably took 20-50 m. y. The major orogeny is dated at approximately 2700 m. y. ago. Mineral ages, as well as disturbed Rb-Sr isochrons, indicate that the area was uplifted and eroded to depth at which stabilization of the decay systems was achieved 2550 m. y. ago. During this period faulting was an important process. Horizontal displacements of considerable magnitude are known in the Archean rocks, but of particular interest to us are the large vertical translations that were achieved. These vertical movements were accomplished along major fault zones that permitted block movements. The fault pattern in northeastern Minnesota is shown in Figure C.82.

Middle Precambrian. In east-central Minnesota the Penokean granites have been dated by a number of techniques. The post-kinematic granites in the Saint Cloud area were emplaced approximately 1820 m. y. ago. Mineral ages indicate that uplift with erosion stabilized the isotopic systems in biotite approximately 1770 m. y. ago. During the 50-million year interval uplift was accomplished by movements of blocks and shearing in the granite is an obvious feature. In eastern Minnesota in the Denham area large fault zones have been found in the field following extensive age measurements in the laboratory. A well-recrystallized cataclastic zone in the McGrath Gneiss (2700 m. y. or older) has been dated at 1770 m. y. ago.

Late Precambrian. The structure of Lake Superior is a basin or syncline formed by downwarping. The Middle Precambrian North Shore Volcanic Group (Table C.2) was intruded by the gabbroic magma (Duluth Gabbro), and it first was postulated that with release of the pressure responsible for the intrusion, the whole region sagged forming a lopolith.⁽¹⁷¹⁾ Taylor,⁽¹⁷²⁾ however, ascribed the origin of the Duluth syncline to structural evolution independent

FIGURE C.82. Fault Pattern in Lower Precambrian Rocks in the Region of Ely, Minnesota. From Sims and Morey, 1972, p.47.

TABLE C.2. Correlation Chart of Late Precambrian Rocks in the Lake Superior Region. From Sims and Morey, 1972, p. 282.

			Southern Minnesota	N. Wisconsin- W Michigan	Isle Royale, Michigan	Northeastern Minnesota	Thunder Bay, Ontario	Eastern End, Lake Superior		
UPPER PRECAMBRIAN	KEWEENAWAN	Upper		Bayfield Group
			Hinckley Sandstone	Chequamegon Sandstone					Jacobsville Sst	Hinckley Sandstone
			Fond du Lac Formation	Devils Island Sandstone					Fond du Lac Formation	
		Middle	Solar Church Formation	Ontonagon Group	Freda Sandstone
				Nonesuch Formation						
				Copper Harbor Conglomerate	Copper Harbor Conglomerate					
		Lower		Chengwatana volcanic group	Portage Lake lava series
				South Range volcanic rocks	Portage Lake lava series					North Shore Volcanic Group
				Bessemer Quartzite
				Sioux Quartzite	Barron Quartzite
MIDDLE PRECAMBRIAN		Igneous and metamorphic rocks	Igneous and metamorphic rocks			Metasedimentary rocks	Metasedimentary rocks	Metasedimentary rocks

FIGURE C.83. Generalized Map of the Lake Superior Region Showing the Known and Inferal Distribution of Kewunawan Volcanic and Sedimentary Rock and Some of the Major Faults; also Shown is the General Distribiton of the Late Precambrian Sioux Quartzite in Southwestern Minnesota and Adjacent Parts of South Dakota and Iowa. From Craddock, 1972.

FIGURE C.84. Rates of Relative Crustal Movement in Meters per Century.

of the Duluth Gabbro. More recently Chase and Gilmer⁽¹⁷³⁾ have attempted to relate the Keweenaw igneous activity to a plate tectonic model in which the Midcontinent gravity high marks a continental rift system.

Figure C.83 is a generalized geologic map of the Lake Superior region in which Craddock⁽¹⁷⁴⁾ has shown the known and inferred distribution of Upper Precambrian rocks. Craddock describes the rocks briefly and discusses the probable thickness in some detail. The cumulative thickness of the Middle Precambrian plateau basalts is about 12,000 m. Major faults in Craddock's compilation (Figure C.83) are the Douglas fault and the Hastings-Lake Owen fault. The area between the faults is now structurally high, and Craddock named it the St. Croix horst. A major cross fault, the Belle Plaine fault, is shown in Figure C.83. Actually, there are numerous strike faults and also cross faults beyond those shown in Figure C.83.⁽¹⁷⁵⁾ Some of the faults displace Cambrian and Ordovician strata, and some of the faults, first formed in Late Precambrian time were activated in Paleozoic time.

Glacial Rebound

A noteworthy feature of the Canadian Shield as well as some of the platform-type sedimentary sequences which overlap it, is that a heavy load of ice covered most of it in recent geologic time. Melting of the extensive ice sheets has unloaded these rocks and various amounts of rebound have taken place and are still continuing, although at a lessening rate. For example Frost and Lilly.⁽¹⁷⁶⁾ Measured vertical uplift in feet per century in an area north of Quebec City P.Q. where relative movements between 0.18 m and 0.15 m per century were reported. The area of negative movements overlapped a seismic zone northeast of Quebec City.

Crustal movement in the Great Lakes region has been known for more than a century. Clark and Persoage⁽¹⁷⁷⁾ present a table of relative crustal movement on the Great Lakes. Thus, Thunder Bay Ont. (formerly Port Arthur) is rising about 0.3 m per century relative to Duluth Minn. even though Duluth itself is rising at a rate of about 0.15 m per century. (Figure C.84).

Recent measurements⁽¹⁷⁸⁾ at Manitoulin Island, Ontario suggest a constant uplift of 2.2 mm. per year over the past 5000 years, relative to the Lake Huron outlet area at Sarnia Ont.

Average rates of post glacial uplift for northern and eastern North America are shown on an attached map by Andrews.⁽¹⁷⁹⁾

Walcott⁽¹⁸⁰⁾ infers from the amplitude of gravity anomalies, that a substantial amount of further uplift will occur and that the relaxation time is of the order of 10,000 to 20,000 years. Rebound uplifts of several hundred feet have been reported from Sweden since deglaciation there.

This type of rebound is regional in nature and should not have a significant effect on large stable structural blocks. Its significance lies more in its bearing on lake drainage, ship canals, and things of similar engineering interest.

Seismicity

The Canadian Shield is generally considered to be a stable area but this is not true in the strict sense. There is a history of earthquakes that resulted in major damage in a zone that extends northeasterly from Lake Ontario along the St. Lawrence river. This zone is shown on various seismic risk maps.

The important question that arises is - how does one evaluate the probability that an earthquake will cause damage in a certain area or at a particular place? Various forms of seismic risk maps have been developed in attempts to answer this question. Such maps are based upon the seismic history of the region mapped. Usually this consists of some record of epicenter location, intensities and damage that resulted. Their use as risk maps is based on the assumption that future earthquakes will occur where they have occurred in the past. A problem that arises is that records in one area may be more complete, and encompass a longer time, than in another. There is a scarcity of recorded earthquakes in or near the shield rocks of the Lake Superior region. A recent earthquake, occurred on July 9, 1975 in west-central Minnesota near the "Goldich" line which marks the southern boundary of the Lake Superior province. This recent earthquake had a magnitude of 4.6 and an intensity between V and VI. Its location was 96° west longitude, 45° 35' north latitude. The center was 65 kilometers southwest of the last previous event in 1950, and about 100 kilometers southwest of a documented earthquake that occurred over 100 years ago.

An earthquake of some what greater magnitude is recorded on the various seismic maps, in the upper peninsula of Michigan, prior to 1938. The only earthquakes shown for Wisconsin were in the most southerly part of the state near the Illinois border. None are shown for the Wisconsin Precambrian areas. No earthquakes have been recorded in the areas north and west of Lake Superior nor in any part of Minnesota except along the important discontinuity referred to as the "Goldich" line.

All the seismic risk maps that have come to our rattention classify the Lake Superior region among the seismically most stable of the continent. It seems clear that selection of an in-rock storage site in these regions, well away from major fault zones should provide minimum risk of seismic instability.

Hydrogeology

The major part of the shield region is underlain by crystalline rocks. Only minor areas are covered by relatively unmetamorphosed and by undisturbed sedimentary rocks. Most of these rocks are covered by a thin discontinuous mantle of surficial material which is commonly thicker in the valley areas than on the uplands. Although the topography is locally rugged, relief is usually no more than a few hundred meters. Annual precipitation varies from near 0.56 m at the western edges of the shield to 1.1 m near the St. Lawrence River, one third to one half of it as snow.

In general, groundwater in the crystalline rocks occurs only in fractures so that the potential amounts depend directly on the number, size and permeability of such fractures.

At depths exceeding 900 m no water was observed, and none was reported by the mine staffs. Water has been reported at one mine at Elliot Lake Ontario at that depth, in a thrust fault zone. At an adjoining mine where the working area is a thousand meters or so below the thrust fault no water was found. The mines visited included workings in, or partly in, gneisses, iron formation, quartz-pebble conglomerate, basic volcanics, shale, noritic and siderite.

It is worth noting that in the active mining areas there are numerous primary fractures such as shear zones, faults and joints but these are dry, despite the fact that extensive mining has disturbed the rock equilibrium and created secondary permeability in the wall rocks. The implication is that in areas where such extensive disturbance has not occurred, the fractured rock should be impermeable. At depths between 600 and 900 m, some water occurs in some mines, at depths less than 600 m water commonly occurs in fractures and should be expected.

The Creighton mine near Sudbury Ontario at 2100 m, was the deepest mine visited. It is dry at that depth and is dry to above 900 m from surface. It seems clear that at depths exceeding 900 m there is very little probability of encountering waterbearing fractures. At depths beyond 1500 or 1800 m the rock pressures are such that the probability of water entering an underground opening through the rock is exceedingly small. At such depths the particular rock type probably will have little bearing on permeability. Experience in deep mines confirms this general statement.

Whether the above would be true for major crustal flows is not known. However a storage site should not be selected along or very near a linear, feature of that kind.

There seems to be little data readily available on the chemistry of groundwaters in Precambrian rocks.

In general groundwater in the Canadian Shield is similar to the acid surface waters and has a similar low pH.

The groundwater flow systems are comparatively short and shallow. In some mines the groundwater is affected by the minerals of the ore deposits and may be very acidic and corrosive. This is however a rather local condition that would be unlikely except in or near sulfide mineralization.

It should be pointed out that the chemical nature of the water in near-surface fractures should bear some relation to the overlying surficial deposits. For example, if a high carbonate glacial drift overlies an area, the pH of the cleft water ought to be higher than in areas where the drift is derived from crystalline rocks of non-carbonate type.

At depths, any waters encountered which are stagnant should be expected to reflect the chemistry of the enclosing rocks rather than the nature of the surficial cover. There is also the possibility of minor amounts of original water trapped within the rocks and not introduced later through fractures.

C.2.12 Flood Basalt--Northwestern U.S.A.

Flood basalt covers approximately 75,00 square miles (200,000 square kilometers) in the Pacific Northwest--chiefly in southeastern Washington, northeastern Oregon, and adjacent parts of Idaho--where its average thickness has been estimated to be approximately half a mile or one kilometer. Flood basalt occurs in no other physiographic province of the United States. The central part of the Pasco basin has the thickest known section of flood basalt, exceeding 3000 meters. Therefore, this basin has been selected as a comprehensive example of the occurrence of flood basalt for the purposes of this report.

Present evidence indicates that the flood basalts of the Pacific Northwest were extruded over a time period of at least eight million years during the Miocene and early Pliocene epochs.^(181,182,183,184,185) A principal immediate source presumably was the Grande Ronde-Cornucopia dike swarm (known also as the Chief Joseph dike swarm) along and near the eastern margin of the flood-basalt province. This swarm comprises thousands of individual dikes distributed over an area about 264 Km long and 120 Km wide, from Huntington, Oregon, on the south to Almota, Washington, on the north and from Riggins, Idaho, on east to Dayton, Washington, on the west. Most dikes of the swarm tend N. 10° W. ± 10°. ⁽¹⁸⁶⁾

Certain of the youngest and best exposed flow layers of the flood basalts have been traced several tens of miles on the land surface. Also, during about the last 15 years a definitive stratigraphic succession has been demonstrated from petrographic and geochemical differences among the several layers. Thus, the whole flood-basalt sequence presently is designated as the Columbia River Basalt Group; its lower part, of Miocene age, as the Picture George

Basalt Formation; and its upper part, of upper Miocene-lower Pliocene age, as the Yakima Basalt Formation. (187,188,189,182,190,183,184,191,185)

Figure C.85 is a columnar diagram of the upper, and probably the greater, part of the Yakima Basalt Formation in the central part of the flood-basalt province.

The cited maximum known thickness of the flood basalt occurs in a 3250 m exploratory well, Rattlesnake Hills No. 1, drilled in 1957-1958 by the Standard Oil Company of California about 3.2 km west of the Hanford Reservation and on the north flank of the Rattlesnake Hills, in sec. 15, T. 11 N, R. 24 E. By report the well bottomed in the basalt. The well was re-entered in 1967 by Battelle Pacific Northwest Laboratories for geophysical logging and other tests. (192,193)

Stratigraphy

In succession downward from the land surface, the rock units of the Hanford Reservation and its vicinity are: (1) alluvium, the deposits of present streams, cobble gravel to silt; (2) glaciofluvial and fluvial deposits laid down by an ancestral Columbia River, largely unconsolidated granule or pebble gravel and sand, locally interlayered silt or cobble-to-boulder gravel, maximum thickness about 60 m, forming and underlying the terraces that extend over most of the Reservation; (3) the Ringold Formation, somewhat consolidated and stratified silt and fine sand plus a conglomerate train, maximum known thickness about 180 m, conspicuously exposed in the White Bluffs east of the Columbia River and extensive beneath the glaciofluvial and fluvial deposits of the Reservation; and (4) layered basalt of the Columbia River Group, more than 3000 m thick as has been noted, described at greater length in following paragraphs.

Newcomb⁽¹⁹⁴⁾ describes the character and stratigraphic relations of these several rock units. Figure C.86 shows the general extent, in the Pasco Basin and its vicinity, of the Columbia River Basalt Group and of the overlying sedimentary deposits.

Columbia River Basalt Group

Layered Extrusives. In and near the Hanford Reservation, the extrusives of the Columbia River Basalt Group (largely if not exclusively of the Yakima Basalt Formation) form a succession of flow layers which evidently were very nearly horizontal as emplaced. As exposed at the land surface, individual layers are about 3 to 45 m thick and each is notably uniform in thickness throughout its traceable length. Cored bore holes have penetrated flow units as much as 115 m thick that show no clear evidence of a time break. Certain prominent layers have been "walked out" over distances of tens of miles. As has been noted, in other layers distinctive petrographic or, even more

FIGURE C.85. Columnar Diagram of the Yakima Basalt Formation in the Central Part of the Flood-Basalt Province. After Ledgerwood and other, 1973; and Atlantic Richfield Hanford C., 1975.

FIGURE C.86. Sketch Map of the Pasco Basin and Vicinity, Washington, Showing Selected Geologic Features. Modified after Newcomb, 1970.

diagnostic, geochemical characteristics have been shown to persist over similar distances. The geochemical characteristics found to be most useful in the Pasco Basin are concentrations of, and ratios among, CaO, MgO, K₂O, TiO₂, Ba, and Cr.

Each individual layer is generally delineated at the top of a two-tiered pattern of joints, somewhat platy above and columnar below;⁽¹⁹⁵⁾ the columnar parting may transect virtually all of a relatively thin flow layer. If the layer is thick, its basal part may display a second colonnade, while its central part commonly has few and widely spaced transverse joints. Also, the uppermost part, generally to a depth of not more than 3 m below the top of a flow layer, may be slightly to moderately vesicular. Few of the vesicles appear to be interconnected or more than about an inch in greatest dimension. Aside from these joint partings and vesicles the typical extrusive rock is generally without void space; cavernous openings such as lava tubes are virtually unknown.

Although conspicuous and commonly open in outcrops and shallow excavations, joints such as have been mentioned appear to be generally snug at depth. Thus, from holes DH-4 and DH-5, bored in 1971-1972, 98.4% was recovered from a total of 2900 m cored, generally in pieces at least 0.3 m long. Segments of 100% core recovery ranged from 3 m to 500 m long, and averaged 80 m long; these suggest rock of good engineering quality in several zones hundreds of meters thick. The drilling record noted one 0.3 m and one 1.8 m apparent void, 2.4 m of "fractured" or "badly broken" rock, and 17.5 m of "poker chip" partings.⁽¹⁹⁶⁾ In holes DB-1, -2, -4, -5, and -7, bored in 1973-1974, 96.7% was recovered from a total of 1080 m cored.⁽¹⁹⁷⁾

Interbeds. Several sedimentary beds intervene in the upper 460 m of the Yakima Basalt Formation, but notably are lacking in the lower part of the extrusive sequence. Those interbeds which have been discriminated and named from geologic mapping on the land surface in the vicinity of the Hanford Reservation are, in ascending order, the Vantage sandstone member, Squaw Creek diatomite, Quincy diatomite, Mabton bed, Selah member, and Rattlesnake Ridge member (see Figure C.85). These are described variously as sandy, tuffaceous, or diatomaceous.^(198,188,199,182,190) Cored bore holes within the Reservation penetrate as many as seven interbeds from about 6 to 30 m thick. Newcomb⁽¹⁹⁴⁾ notes tuffaceous interbeds cropping out in the basalt sequence or penetrated by wells at numerous places in and near the Reservation.

Tectonic and Seismic Features of the Basalt

Figure C.86 shows contours drawn on the top of the Yakima Basalt (Columbia River Basalt Group) where it is overlain by younger deposits, within and adjacent to the Hanford Reservation. Because this contoured surface, of

pre-Ringold age, initially was of low relief, the contours essentially show structural forms. In sequence from the north the concealed tectonic segments of the basalt are: (1) the gentle Pasco syncline, whose axis underlies the generalized course of the Columbia River and swerves from an eastward to a southward trend; (2) an asymmetric, eastward-plunging anticline which prolongs Umtanum Ridge and whose crest, within the Reservation, protrudes through the younger deposits to form Gable Butte and Gable Mountain; (3) the Cold Creek syncline, whose sinuous axis trends eastward beneath the high terraces in the central part of the Reservation, then swerves southward to merge with the axis of the Pasco syncline about 16 kilometers north of Richland; and (4) a moderately sharp and probably asymmetric anticline which prolongs Yakima Ridge southeastward.

More conspicuous are the bold, emergent anticlines which enclose the Pasco Basin: (1) on the north, boldly asymmetric Saddle Mountain; (2) on the west, eastward-plunging and irregularly asymmetric Umtanum and Yakima Ridges, mentioned previously; and (3) on the southwest, comparatively gentle but likewise asymmetric Rattlesnake Hills and the prolongation across the Yakima River, Red and Badger Mountains.

Structural relief of the Pasco Basin, from the deepest parts of the Pasco and Cold Creek synclines to the crest of the Rattlesnake Hills, is somewhat more than 1,050 m. Brown⁽¹⁹³⁾ implies that this relief has developed at the mean rate of 10,000 years per m, and so implies that deformation began about contemporaneously with emplacement of the youngest Yakima Basalt members. He suggests also that the deformation continues today, possibly at the same mean rate. Tillson⁽²⁰⁰⁾ postulates, from apparent progressive subsidence of first-order bench marks in the area, that "a gradual basining * * is proceeding at an average rate of 1 mm/yr." This postulated rate is equivalent to 1000 years per meter.

Faults in and near the Hanford Reservation have been noted or mapped by several investigators; of these faults, the principal three are: (1) a high-angle fault which extends about 65 kilometers along the north flank of Saddle Mountain (see Figure C.86), and which is probably more than 100,000 years old; (2) the Rattlesnake Hills-Wallula fault, which prolongs the northeastern flank of the Hills, and which extends southeastward about 50 kilometers from the Yakima River to and across the Columbai River; and (3) the Wallula Gap fault, which is sub-parallel to and about 2.5 kilometers southwest of the Rattlesnake Hills-Wallula fault, and which extends southeastward from the Columbia River about 20 kilometers. The latter two faults may merge southeastward and, with other faults in the southeastern segment of the Horse Heaven Hills anticline, may well be segments of a linear, deep-seated, regional structural feature that extends from the Rattlesnake Hills at least to the vicinity of Milton-Freewater, Oregon, a distance of about 100 kilometers. Deformation and

displacement along this regional feature probably was largely complete by the end of Ringold time, say 300,000 years ago. However, reported rupture of the land surface during the Milton-Freewater earthquake of 1936 suggests that at least some elements of the regional feature still may be tectonically active. (201,202,194)

Within the Hanford Reservation, by detailed mapping and trenching, Bingham⁽²⁰²⁾ identified two low-angle faults striking northeast across Gable Mountain near its center and western end, respectively. Along the central fault, vertical displacement was about 20 m at maximum but apparently died out into sharply folded rocks. Evidently, both folding and faulting were products of the tectonic forces that shaped the Gable Mountain anticline. Bingham⁽²⁰²⁾ cites evidence that "the latest structural movements occurred more than 100,000 years ago."

Other reported faults within the Reservation are subordinate to the folds which have been described. Recent trenching by bulldozer along several of the mapped fault traces failed to demonstrate stratigraphic displacement and led to alternative explanations: landslide and local erosion into the crest of a sharply-asymmetric fold (Isaacson, R. E., oral communication). In general, past identifications of minor faults in the basalt probably should be considered debatable in the absence of stratigraphic correlation by intensive petrographic and geochemical examination.

Seismicity of the Hanford Reservation has been investigated for more than a decade; the common finding has been expressed aptly by one investigator: "Hanford does not lie in * * a persistently active seismic zone." The writer is informed of only three notable earthquakes during the last century and within a range of 160 kilometers: (1) March 6, 1893, location of epicenter unknown; (2) the Corfu earthquake of November 1, 1918, epicenter placed generally along the Saddle Mountain fault about 30 kilometers north of the center of the Reservation, magnitude about IV (modified Mercalli scale) at epicenter; and (3) the Milton-Freewater earthquake of July 16, 1936, epicenter about 110 kilometers to the southeast, magnitude VII at epicenter but IV at Richland and Pasco. (203,201,204)

Microseismic events in and near the Hanford Reservation have been recorded by a network of seismographs established by the U.S. Geological Survey in 1968; preliminary findings summarize as follows: (1) 166 events from March 23 to December 31, 1969, about one fourth probably due to explosions 80 to 200 km distant from the center of the network; (2) 105 events and a few explosions within 75 km; (3) all microearthquakes of Richter magnitudes less than 2.0--that is, generally less than the threshold of non-instrumental human preception; (4) some 230 events in a small area near the Columbia River

in the southeastern part of the Reservation and a few events in the western part; (5) a zone of microearthquakes trending northwest in the vicinity of Saddle Mountain north of the Reservation. In general, microearthquake activity is notably less than in regions of Nevada and California which generally are considered to be seismically active.⁽²⁰²⁾

Hydrologic Characteristics of the Flood-Basalt Sequence

The rocks of the Columbia River Basalt Group constitute a moderately productive aquifer throughout the flood-basalt province, wherever those rocks are in the zone of saturation. Water-bearing openings have been presumed to be joints, contact zones between flow layers, and interbeds between the flow layers. The typical joint pattern and the lithologic character of interbeds have been described briefly.

Although in part coarse-grained and potentially productive of water locally, the interbeds in the upper part of the flood-basalt sequence are largely fine-grained and generally act more to constrain than to facilitate movement of water. This is summarized by Newcomb⁽¹⁹⁴⁾ in the statement that "these sedimentary layers * * have yielded less water to wells than the more permeable stratigraphic zones of the basalt."

Two generalizations from the experience of drilling into the flood basalts for supplies of water lead to a critical conclusion in regard to hydraulic conductivity and transmissivity of that rock type: (1) Water in the flood basalts is extensively unconfined, so that wells of diverse depths show a common piezometric level. Thus, natural vertical hydraulic gradients of substantial magnitude within the basalt appear to be the exception rather than the rule. (2) Wells drilled for minimal water supplies have penetrated widely different depths into the zone of saturation, leading to a common interpretation that water-yielding capacity of flood basalt ranges greatly and erratically. On the other hand, wells drilled for the larger water supplies commonly penetrate hundreds of feet into the zone of saturation; among these deeper wells, specific yields appear to approach a proportion to the depth of penetration.

From the preceding generalizations it is concluded tentatively that, at any coarse scale, the flood basalts tend toward isotropy in hydraulic conductivity. In other words, hydraulic gradients aside, the potential for water to move across stratification (presumably via columnar joints in great part) tends to equal that for water to move parallel to the stratification (presumably via flow-layer contacts and platy joints).

Exceptions to the first generalization do occur, most notably in local bodies of confined water whose piezometric level is substantially above the water table, and locally above the land surface. Newcomb⁽¹⁹⁴⁾ explains this

as a common effect of tectonic deformation, in these words: "The forces of deformation produced excessive strain on some of the basalt and developed planes of shear. Whether the deformation is in the form of tight folding with interflow slippage or solely in the form of shear rupture and fault displacement, gouging and grinding of the rock generally reduced its permeability at the shear planes. These zones of crushed rock can be barriers to groundwater movement. Such barriers cause the damming of groundwater and, in some areas, may confine groundwater so that it has pressure levels relatively close to the surface."⁽²⁰⁵⁾

In and adjacent to the Hanford Reservation, the piezometric surface for water in the flood-basalt sequence appears to be sharply trough-shaped about an axis which, in geographic position, about coincides with the generalized course of the Columbia River. Also, in vertical position the axis coincides approximately with, or locally is above, mean stage of the river. Such coincidence suggests strongly that water in the basalt is tributary to the river.

GLOSSARY OF GEOHYDROLOGIC TERMS (217)

Technical terms, particularly those with somewhat obscure meanings, are defined in this glossary. For terms not defined in this glossary the reader is referred to the "Glossary of Geology and Related Sciences," compiled by the American Geological Institute, and Webster's New World Dictionary, College Edition.

In some instances, definitions have been enlarged beyond the standard textbook definitions. No attempt has been made to cite the source of definitions, and those definitions that have been enlarged may deviate slightly from the standard definitions.

Alluvial Fan. A sloping, fan-shaped mass of loose rock material deposited by a stream at the place where it emerges from an upland into a broad valley or a plain.

Alluvium. All detrital material deposited permanently or in transit by streams.

Amphibole. A group of dark, rock-forming, ferromagnesian silicate minerals which are closely related in crystal form and composition and which have abundant and wide distribution in igneous and metamorphic rocks.

Anorthosite. A granular plutonic igneous rock composed almost exclusively of soda-lime feldspar.

Anion. An ion that is negatively charged.

Anticline. A fold, the core of which contains stratigraphically older rocks, and which in simplest form, is elongate and convex upward with the two limbs dipping away from each other.

Aquifer. A formation, group of formations, or part of a formation that contains sufficient saturated permeable material to yield significant quantities of water to wells and springs.

Aquitard. A natural rock or fine-grained, unconsolidated unit of low permeability which is stratigraphically adjacent to one or more aquifers and through which water movement is markedly retarded or impeded.

Argillite. A compact rock, derived from claystone, siltstone, or shale, which has undergone a somewhat higher degree of induration than is present in these sedimentary rock types but which is less clearly laminated than shale, does not have the fissibility of shale, and lacks the cleavage distinctive of slate.

Artesian. Pertaining to an aquifer is one that is confined so that its hydraulic head rises above the top of the aquifer unit; thus an artesian water body is one confined and under hydraulic pressure.

Basement. A complex of undifferentiated rocks that underlies the oldest identifiable rocks in the area.

Basin. A depressed area generally having no surface outlet; a segment of the earth's crust that has been downwarped by the accumulation of sediments on it; a synclinal tract or area in which the rocks dip toward a central point, and in which folding occurred subsequent to deposition; an elongate, fault-bordered intermountain area.

Batholith. An igneous intrusion greater than 40 square miles in surface exposure, composed predominantly of granitic rocks.

Bauxite. A rock composed of an impure mixture of earthy hydrous aluminum oxides and hydroxides and is the principal commercial source of aluminum.

Bentonitic. Pertaining to a rock containing bentonite, a clay formed from the decomposition of volcanic ash.

Biotite. A complex silicate of aluminum, potassium, magnesium, and iron with hydroxyl that is a widely distributed and important rock-forming mineral of the mica group.

Block-faulting. A type of vertical faulting in which the crust is divided into structural or fault blocks of different elevations and orientations.

Brackish. A somewhat general term applied to mineralized water of concentrations intermediate between those of brine and fresh water.

Breccia. A coarse-grained clastic rock composed of large, angular, and broken rock fragments cemented together in a fine-grained matrix.

Caliche. Calcareous material that forms on or near the surface of stony soils of arid and semiarid regions; thought to be genetically associated locally with capillary fringe of the water table; and is inferred to form when evaporation exceeds precipitation.

Cap Rock. A low-permeability bed of anhydrite and gypsum which overlies a salt body, or plug in a salt dome.

Cation. An ion that is positive charged.

Chimneying. The process of progressive collapse of rock overlying an explosion-produced cavity resulting in a tall underground cylinder (chimney) of broken rock.

Chlorite. A group of green hydrous silicate minerals containing magnesium and iron, with or without aluminum; that these minerals are widely distributed, especially in metamorphic rocks.

Clastic. Pertaining to or being a rock or sediment composed principally of broken fragments derived from preexisting rocks or minerals.

Cleavage. The property or tendency of a rock to split along secondary, aligned fractures or other closely spaced, planar surfaces, produced by deformation or metamorphism.

Coal Measures. A succession of sedimentary rocks (or measures) consisting of clays or shales, sandstones, limestones, and conglomerates with interstratified beds of coal; a group of coal seams interbedded with the above strata.

Conglomerate. A coarse-grained clastic sedimentary rock composed of rounded to subangular fragments larger than 2 mm in diameter, such as granules, pebbles, cobbles, and boulders, set in a fine-grained matrix of sand, silt, or cementing materials, such as calcium carbonate, iron oxide, and silica. The consolidated equivalent of gravel.

Consolidated (Material). In geology, natural materials that have been made firm, cohesive, and hard.

Crystalline Rock. An inexact but convenient term designating an igneous or metamorphic rock, as opposed to a sedimentary rock.

Depositional Environment (Sedimentary Environment). A geographically restricted environment where sediment accumulates under similar physical, chemical, and biological conditions.

Diagenesis. Process involving physical and chemical changes in sediment after deposition that converts the sediment to consolidated rock.

Diamond Pipe. Term used for an occurrence of kimberlite (periodotite that has been converted to a hydrous magnesian silicate) in volcanic pipes large enough and sufficiently diamond bearing to be minable.

Diapirism. The piercing of overlying rocks by a mobile core, such as a salt body or an igneous intrusion.

Discharge. In groundwater hydrology, water that issues naturally or is withdrawn from an aquifer.

Dome. A dome-shaped landform or rock mass; a large igneous intrusion whose surface is convex upward with sides sloping away at low but gradually increasing angles; an uplift or an anticlinal type structure, either circular or elliptical in outline, in which the rock dips gently away in all directions, for example, a salt dome.

Dunite. A coarse-grained plutonic igneous rock composed almost entirely of olivine.

Ephemeral Stream. A water course carrying surface water part of the time but dry or having only underflow in the streambed materials part of the time.

Epidote. A basic orthosilicate mineral containing calcium, aluminum, and varying amounts of iron, and commonly occurring in metamorphic rocks.

Fault. A fracture or fracture zone along which there has been displacement of the sides relative to one another parallel to the fracture.

Fault Block. A crustal unit bounded by faults, either completely or partly.

Fault System. A system of parallel faults that are related to a particular deformational episode.

Feldspar. Any of an important group of rock-forming minerals that are essentially silicates of alumina and some other base, such as potash, soda, or lime.

Flood-Plain Deposit. Sandy and clayey sediment deposited by river water that spread out over a flood plain.

Foliation. A general term for a planar arrangement of textural or structural features in any rock type, but most commonly applied to metamorphic rocks, such as cleavage in slate or schistosity in a metamorphic rock.

Garnet. An important group of silicate minerals rich in alumina, iron, lime, and magnesia that occur chiefly as accessory constituents of metamorphic rocks and less commonly of igneous rocks.

Geohydrologic. Of or pertaining to the geology and hydrology. An abbreviation combining the adjectives geologic and hydrologic.

Loess. An unconsolidated or weakly consolidated sedimentary deposit which is composed dominantly of silt-sized rock and mineral particles and which is deposited by wind.

Low-Angle Fault. A fault, the dip of which is no more than 45° (or less).

Mafic. Pertaining to or composed dominantly of magnesium rock-forming silicates.

Magnetism. The development, movement, and solidification to igneous rock, or magma, a naturally occurring mobile rock material, generated within the earth and capable of intrusion and extrusion.

Matrix-Hole Concept. A type of radioactive waste repository consisting of several drill holes spaced moderately close together.

Mesa. An isolated, nearly level landmass standing distinctly above the surrounding country, bounded by steeply sloping erosion scarps on all sides, and capped by layers of resistant, nearly horizontal rocks.

Metallic. Of or belonging to metals, containing metals, more particularly, valuable metals that are the object of mining ore.

Mica. A group of silicate minerals of aluminum and other bases, especially potassium, magnesium, and iron, and characterized by a great perfection of cleavage in one direction that produces thin, though, elastic plates or laminae.

Monocline. A unit of strata that dips from the horizontal in one direction only, and is not part of an anticline or syncline.

Nepheline Syenite. A plutonic rock composed almost entirely of alkali-feldspar and nepheline which is a silicate mineral containing appreciable amounts of potash, soda, and alumina.

Olivine. An olive-green, common rock-forming ferromagnesian silicate mineral of mafic, ultramafic, and low-silica igneous rocks.

Outcrop. A part of a body of rock that appears, bare and exposed, at the surface of the ground.

Permeability (Permeable). The relative ease with which a porous medium can transmit a liquid under a hydraulic gradient.

Plagioclase. The group of common rock-forming feldspar minerals that contain varying mixtures of sodium and calcium.

Playa. A flat-floored area composed of thin, evenly stratified sheets of fine clay, silt, or sand, and representing the lowermost or central part of a shallow completely closed or undrained, desert lake basin in which water accumulates after a rain and is evaporated, usually leaving deposits of soluble salts.

Peridotite. A coarse-grained plutonic igneous rock composed chiefly of the mineral olivine but also containing considerable amounts of other ferromagnesian minerals.

Pluton. A body of intrusive igneous rock of any shape or size.

Pluvial (Period). Pertaining to a period of time or a climate in which rainfall or precipitation is abundant.

Porosity. That property of a rock or soil which enables the rock or soil to contain water in voids or interstices, usually expressed in percentage or a decimal fraction of void volume as compared to total volume.

Pyroxene. A group of dark, rock-forming silicate minerals closely related in crystal form and analogous in chemical composition to the amphiboles, and found chiefly in igneous rocks.

Pyroxenite. An ultramafic plutonic igneous rock chiefly composed of pyroxene.

Quartz Monzonite. A coarse-grained igneous rock, intermediate in composition between granite and granodiorite, which contains quartz and about equal amounts of the alkali and soda-lime feldspars.

Recharge. In hydrology, a source or means for replenishment of water withdrawn or discharged from an aquifer. An aquifer in hydraulic equilibrium will discharge a quantity of water about equal to the amount of recharge.

Rift. A long fairly relative narrow depression formed along lines of multiple fracture.

Rift Valley. An elongate narrow trough or valley formed by the sinking of a strip of the earth's crust between two more or less parallel nearly vertical faults.

Rift Zone. A system of crustal fractures.

Rise (Marine). A broad, elongate, and smooth elevation of the ocean floor.

Sedimentary Basin. Geologically depressed area with thick sediments in the interior and thinner sediments at the edges.

Seismicity. The phenomenon of earth movements as manifested by earthquakes.

Shield. A continental segment of the earth's crust which has been relatively stable over a long period of time and which has exposed crystalline rocks mostly of Precambrian age; in general, representing the oldest rocks of the continent.

Sill. A tabular igneous intrusion that parallels the planar structure of the surrounding rock.

Sink Hole. A funnel-shaped depression in the land surface generally in a limestone region often connected to a subterranean passage developed by solution.

Static Head. The height above a standard datum of the surface of a column of water (or other liquid) that can be supported by the static pressure at a given point.

Stock. An igneous intrusion less than 40 square miles in surface exposure.

Strain. Deformation resulting from applied stress; proportional to stress.

Stratum. A single layer of homogeneous or gradational lithology, deposited parallel to the original dip of the formation.

Strike-Slip Fault. A fault, the actual movement of which is parallel to the strike of the fault.

Syncline. A fold, the core of which contains stratigraphically younger rocks, and which in simplest form, is elongate and concave upward with the two limbs dipping toward each other.

Tectonic. Pertaining to the forces involved in, or the rock structures and external forms resulting from the deformation of the earth's crust.

Tectonism (Diastrophism). Crustal movement produced by earth forces, such as the formation of plateaus and mountain ranges; the structural behavior of an element of the earth's crust during, or between, major cycles of sedimentation.

Terrace. A long, narrow, relatively level or gently inclined surface, generally less broad than a plain, bounded along one edge by a steeper descending slope and along the other by a steeper ascending slope, and containing unconsolidated material.

Thrust Fault. A fault with a dip of 45° or less in which the upper rock mass appears to have moved upward relative to the lower rock mass.

Transmissivity. Volume of water flowing through a 1-ft width of aquifer of given thickness under a unit gradient (1 ft vertically for each foot laterally) and at the viscosity prevailing in the field. Mathematically, it is the product of permeability and aquifer thickness.

Ultramafic. Pertaining to igneous rocks composed chiefly of ferromagnesian, dark minerals.

Underthrust. A type of thrust fault in which the lower rock mass has been actively moved under the upper, passive rock mass.

Uplift. A structurally high area in the crust, produced by positive movements that raise or upthrust the rocks, as in a dome or arch.

Upwarping. Uplift of a regional area of the earth's crust.

Water Table. That surface of an unconfined water body at which the pressure is atmospheric.

Zeolite. Any of a large group of minerals that are hydrous silicates of aluminum with sodium and calcium, or rarely, with barium and strontium.

Zeolitization. The process by which feldspars and other aluminosilicates of a rock are altered to zeolites.

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APPENDIX D

THE OCEAN FLOOR

THE OCEAN FLOOR*THE OCEAN ENVIRONMENT

The oceans are divided into three principal physiographic provinces, each occupying about a third of the world ocean area:

1. Continental margins (continental shelf, inland seas, marginal plateaus, continental slope, continental rises)
2. Ocean basin floor (abyssal plains, abyssal hills, oceanic rises, deep sea trenches)
3. Mid-oceanic ridge (ridge flank and crest, rift valley and rift mountains, fracture zones).

Any of these major provinces may be interrupted by volcanic seamounts or islands and in the following discussion each province is treated in categorical order. It is helpful to remember that, with few exceptions, geologic processes on the continents are primarily erosional whereas in the oceans they are principally depositional.

The Continental Margin

Between the beach and the landward edge of oceanic crustal basalt lies the "wetted perimeter" of the continents-the continental margin. This is the most dynamic environment of the ocean, where seasonal temperature changes in the water are high, chemical and biological processes are most variable, and geology is most complex and unpredictable. Here lie most of the remaining pools of hydrocarbons, as well as most of the world's great fishing grounds. Surface material of the continental margin changes radically over short distances, ranging from hard rock to gravel to clay within a few miles to tens of miles.

Most of the continental shelves (200 to 300 m deep) and marginal plateaus (to 1000 m deep) are covered with silt, sand, and gravel (e.g., the Blake Plateau); most inland seas (e.g., the Japan Sea) are floored with finer material. The continental slope (200 to 2500 m deep) and rise (2500 to 4500 m) are generally covered with silty clay. The thickest deposits of this organic fine-grained sediment, often interbedded with sand layers, are found beneath the deltas, aprons, and cones off the mouths of major rivers. A combination of factors, including the high organic content (potential oil sources) and interbedded sand layers (potential reservoirs) makes these areas particularly

* Extracted from W. P. Bishop and C. D. Hollister, "Seabed Disposal-Where To Look," Nuclear Technology, vol. 24, pp. 425-443, 1974.

suitable for hydrocarbon production. In these regions the rate of sediment accumulation is relatively rapid, and this, combined with relatively steep slopes, provides optimum conditions for sediment failure (slumping, turbidity currents, etc.). Evidence that these processes occur is abundant.

Bottom currents on the continental shelves are strong but variable, with velocities often exceeding 50 cm/sec, and strongly influenced by wind direction, storms, and tides. Currents on the deeper marginal plateaus are influenced more by steady-state wind-driven circulation. Velocities of bottom currents here are of the order of 10 to 25 cm/sec. Current velocities on the slope are generally of the same order and are principally driven by differences in salinity or temperature (thermohaline) or are associated with internal waves and tides.

Abyssal bottom currents on continental rises on the western sides of ocean basins generally vary between 10 and 35 cm/sec, are also thermohaline, and tend to be constant in direction. Abyssal currents on eastern sides of the major oceans are generally weak and variable.

Also of importance to a discussion of the ocean environment is the position of the continental margins relative to areas of crustal growth or destruction. Because most continents are moving, it is useful to think of their surrounding margins either as being on the "leading" (destructive) or "trailing" (constructive) sides. Leading edges are characterized by very narrow seismically active continental shelves (e.g., the west coast of the Americas), and precipitous, rocky, and faulted continental slopes. Shelves on trailing edges are generally broad, flat, and seismically inactive (e.g., east coast of the Americas). Clearly it would be meaningless to average shelf characteristics of these two very different margin types. These fundamental differences must be kept in mind during the following discussion of the deep ocean floors.

Ocean Basin Floor

This province, occupying another one-third of the ocean area, is the deepest (4-1/2 to 11 km) of the three provinces and includes the flat abyssal plains, the gently rolling regions of abyssal hills, and the deepest portion of the oceans--the deep-sea trenches.

The flat abyssal plains (gradients $<1:1000$) are built through deposition by the catastrophic and periodic turbidity currents that sweep vast quantities of coarse continental debris from the continental margins to this final resting place by rapid (hundreds of cm/sec) underwater avalanches. The destructive power and speed of these currents have been determined through the failure sequences of submerging communication cables. Sediments recovered

from abyssal plains are typically silty clays interbedded with thick coarsely graded layers of sand and gravel. Holes drilled in these regions by the *Glomar Challenger* often collapsed through failure of these incohesive coarse interbedded deposits. The temperature of bottom water on these plains does not vary seasonally, and bottom currents are extremely weak and variable, with tidal velocities of 2 to 5 cm/sec predominating.

The abyssal hills that lie seaward of the abyssal plains are a few kilometers in diameter and a few tens of meters high and are thought to be old ocean crust originally formed as extrusions of pillow basalt from the MOR spreading center. As the crust spreads away from the MOR it cools and sinks, reaching a depth of 5 km in about 50 million years. The vast abyssal-hill province (e.g., most of the North Pacific) is generally covered with 50 to 100 m of brown zeolitic clay overlying a few meters of limestone that was deposited when the hills were part of the much shallower MOR. The mid-plate abyssal hills are seismically passive. These regions are also located in central portions of the wind-driven surface-flowing current gyres that are typically quite stable and relatively unproductive biologically. Bottom currents, although topographically influenced, are weak and variable and their effects on the bottom have not appreciably altered in 50 million years. From the viewpoint of stability, the vast (10^8 km^2) regions of the mid-plate/mid-gyre abyssal hills are unparalleled.

The dynamic deep-sea trenches (7 to 11 km deep) often form the landward boundaries of abyssal hills, especially in the Pacific and Indian Oceans. Here the ocean crust, colliding with and presumably being overridden by lighter crustal rock, is being destroyed at a rate of 2 to 4 cm per year. The arcuate circum-Pacific trenches represent some of the most dynamic regions on the planet. Many high-intensity earthquakes occur in or near trenches. This activity triggers massive submarine slides and is associated with extensive volcanism.

The technology does not exist for drilling into these great ocean deeps, so our data come primarily from piston cores tens of meters in length, and from seismic profiles. These data show wide variation in the sedimentary column. Some trenches are nearly filled by flat abyssal plains, some are almost empty, and some show very highly deformed landward edges where sediment has been plastered against the leading front of the advancing continent. Those near major sediment sources are nearly filled with 2 to 3 km of debris; many have at least 1/2 km of sand, silt, and clay, often containing turbidite sand and gravel beds. The material is deposited at rates from 1 to 100 cm per 1000 years.

Mid-oceanic rises (e.g., the Bermuda Rise) are broad low swells of abyssal hills generally 4 to 5 km below the surface. They may be associated with concentrations of seamounts or islands. Sediment on oceanic rises tends to be thicker than is found on abyssal hills, often exceeding 200 to 500 m. It is a mixture of pelagic ooze and clay and is generally silty clay in texture. The notable increase in sediment thickness results primarily from lack of dissolution of the remains of pelagic plants and animals in water depths shallower than the carbonate compensation depth (the depth at which carbonates falling from the surface are dissolved), which is generally between 4 and 5 km.

Mid-Oceanic Ridge (MOR)

This 40,000-km-long, globe-circling sea-floor spreading center forms the "constructive" plate-boundary. In the center of this symmetrical two-way expanding ridge lies the hot, seismically active rift valley (2 to 3 km deep) where new crust is being continually extruded. Sediment thickness in this shallowest and youngest part of the ocean is generally too small to be detected except as a thin, current-winnowed veneer of pelagic carbonate seen in bottom photographs. Pelagic carbonate ooze may attain 5- to 10-m thickness in low-lying pockets of the rift mountains on either side of the rift valley, and from there, in either direction, sediment thickness increases linearly with distance, age, and depth, from the ridge where, at the carbonate compensation depth, the cold bottom water is able to dissolve the lime faster than it can be deposited, yielding a sedimentation rate of <1 cm/1000 yr.

Water temperatures and currents on the shallowest portions of the rift mountains vary markedly in response to wind-driven currents flowing over and around rugged mountainous relief. Bottom-water velocities may at times exceed 50 cm/sec. The temperature of the central portion of the rift valley may approximate that of molten basalt (1200°C), yet water circulation dissipates this heat so quickly that no water temperature rise can be measured at the sea surface directly above this tremendous ($20,000 \text{ km}^2$) heat source.

These three provinces comprise the ocean environment, about which a significant amount of data has been acquired (Table D-1).

From the data at hand, certain very good first approximations can be made about the rates and kinds of processes that are active in the three provinces. We will now examine these provinces by the criteria of geologic stability and predictability of processes, as we know them today.

THE ENVIRONMENTAL STABILITY AND PREDICTABILITY OF OCEANIC PROVINCES

Assuming that one of the most important criteria for waste disposal is geologic stability and thus predictability over periods of up to a million

years, we will now compare the ocean provinces with regard to their frequency of catastrophic events and the rates of natural processes.

Most of our data necessarily come from interpretation of the past record, which is contained in the sediments. The bulk of this information is derived from piston cores 5 to 10 m in length and 6 to 8 cm in diameter, which contain a record that goes back millions of years. The Deep Sea Drilling Project of the National Science Foundation (NSF) has obtained sediment data from over 400 drill holes that in some instances penetrate the entire sediment column and up to 0.5 km (leg 37) into oceanic basalt. This record encompasses tens to hundreds of millions of years, and often the record indicates change. For example, each change in color, texture, and composition observed in recovered sediment has been caused by some change in depositional conditions. A completely uniform sample, on the other hand, indicates relative uniformity of conditions and material.

Another much shorter time-set of data concerning natural catastrophes comes from seismographs that have been recording earthquake activity for tens of years. Current meters measure velocity and direction for periods of weeks to months, and bottom photographs give us an instantaneous view of the modern sea-floor environment. Occasionally, sedimentary features seen in these photographs indicate "long-term" current effects that may be averaged for periods of months or longer. Seismic profiles show the reflecting horizons of the upper kilometers of unconsolidated sediment. Interpretation of these horizons (whether disturbed, faulted, etc.) yields significant qualitative insight into geologic stability over periods of millions of years.

Samples and seismic profiles taken on the continental margin often reveal conspicuous layering of sedimentary horizons. Cores contain sand interbedded with silt or clay, indicating rapid changes in depositional processes and/or sediment sources. Profiles reveal abundant layering that is often severely contorted. Abrupt color changes suggest variations in source accumulation rates, or degree of oxidation. Erosional unconformities or hiatuses in the record are also commonly found in margin cores. Thus, we infer that the continental margin has been relatively unstable over geologic time, and it appears likely that it will continue to be a relatively unstable and unpredictable region.

Areas of high sedimentation rates have been cited as desirable disposal sites because of natural burial processes. However, at a sedimentation (burial) rate as high as 1 m per thousand years, which is high for even the most active fans and cones, only 250 to 500 m of sediment will accumulate over a waste container during the time its contents remain dangerous. The technology already exists to penetrate and thus emplace objects 1 to 2 km beneath the sea floor in these regions.

TABLE D-1. Characteristics of the Oceanic Provinces

Physiographic Provinces of the Oceans	Mid-Oceanic Ridge				
	Fracture Zones	Flanks	Crest	Rift Valley	
ENVIRONMENT					
Water depth (km)	4 to 6	3 to 5	0.5 to 2	1.5 to 3	
Local relief (m)	1000's	100's	1000's	1000's	
Regional slope (deg)	10's	2 to 5	10's	< 5	
Bottom temp (°C)	< 3	2 to 4	3 to 5	5 to 15	
Texture of bottom sediment	pebbles to sand	sand silt, clay	boulders to sand	boulders to sand	
Sediment thickness above igneous rock (km)	0 to 2	$\frac{1}{2}$ to 2	0 to $\frac{1}{4}$	0	
DYNAMIC PROCESSES					
Rate of sediment accumulation (cm/1000 yr)	0 to 10	2 to 4	0 to 5	0 to 1	
Currents {	Non-tidal currents (cm/sec)	10 to 20	3 to 5	10 to 50	10 to 20
	No. of near-bottom measurements	< 10	< 10	< 10	< 10
Erosion or deposition	E	D	E	E	
Earthquake frequency	low to high	very low	moderate	very high	
Biological activity	low	moderate	moderate	low	
Frequency of sediment failure	moderate	low	high	very low	
Volcanic activity	moderate	low	high	very high	
CHARACTERISTICS GERMANE TO SUB-SEA FLOOR DISPOSAL					
Geologic stability (predictability)	low	moderate	low	very low	
Areal extent (km ² × 10 ⁶)	←————— 120 —————→				
Accessibility by man	low	low	high	moderate	
Present Value of Natural Resources, × 10 ⁶ \$ per Annum	Biologic	←————— 1 × 10 ³ —————→			
	Sand, gravel	←————— 0 —————→			
	Mineral	←————— 0 —————→			
	Hydrocarbon	←————— 0 —————→			

The continental margins off leading edges of moving continents, like the western United States, are seismically very active. Earthquakes on trailing edges like the east coast of the United States are much less frequent. On leading edges the combination of high seismic activity, steep slopes, and high rates of soft sediment accumulation results in optimum conditions for sediment failure, e.g., slumps, slides, turbidity currents, etc.

TABLE D-1 (Continued)

Ocean Basin Floor				Continental Margin			
Seamounts Islands	Trenches	Abyssal Hills (Mid-Plate)	Abyssal Plains	Deltas/Fans	Rise	Slope	Shelf
0 to 3 1000's 10's 2 to 20 gravel and sand 0 to $\frac{1}{2}$	7 to 11 10 to 100 2 to 20 1 to 3 sand, silt, clay 1 to 5	5 to 6 10 to 100 <1 <2 clay $\frac{1}{4}$ to $\frac{1}{2}$	4 to 6 <5 <1 <2 clay, gravel $\frac{1}{2}$ to 2	0.1 to 5 10 to 100 $\frac{1}{2}$ to 5 3 to 5 sand, silt, clay 5 to 10	3 to 5 5 to 10 <1 1 to 3 silt, clay 5 to 10	0.2 to 3 100's 4 to 10 3 to 5 silt, clay 2 to 10	0.1 to 0.5 5 to 10 <1 2 to 20 gravel and sand 2 to 20
< 0 to 5 5 to 25 10's E moderate very high moderate high	1 to 10 2 to 10 < 5 D very high moderate high high	< 1 2 to 10 < 10 D very low very low very low very low	2 to 5 < 5 10's D very low low low very low	10 to 500 5 to 30 10's D very high to low, depending on position of plate boundary high very high low	5 to 50 5 to 30 100's D moderate moderate low	< 0 to 3 5 to 20 100's E high very high low low	< 0 to 50 10 to 50 1000's D/E very high low low
very low 10 very high	very low 6 very low	high 130 very low	moderate 10 to 20 very low	very low 5 to 10 moderate	moderate 20 low	very low moderate	very low 50 very high
1×10^3				9×10^1			
0				3×10^2			
0				6			
0				3×10^4			

At the seaward edge of the continental margin are the flat abyssal plains generally best developed off trailing edges, and the deep sea trenches usually found off leading edges. The plains represent the last residing place of seaward-moving continental detritus which becomes entrained in turbidity currents that flow at speeds of tens of meters per second across the continental

CATACLYSMIC/SPORADIC EVENTS

Earthquakes
 Volcanism
 Slumping of sediments
 Turbidity currents
 Rapid currents causing erosion
 Ice rafting of boulders
 Biological windfalls
 Refocusing of tidal circulation
 Magnetic reversals
 Intrusion by man

slope and rise. The frequency of these flows depends on the rate of continental erosion and sediment transport to the top of the continental slope, and the highest frequencies have been recorded from cable ruptures off the Congo Canyon, where approximately 50 occur each century.

Deep-sea trenches are found at the base of most leading edges of continental margins, and here high seismic activity, combined with proximal sediment supply and steep slopes, provides maximum instability.

Trenches are thought to be the result of processes of underthrusting or subduction of oceanic crust beneath continental crust. This process cannot be observed directly, but is inferred from the location and magnitude of seismic events, crustal layering, and a tremendous negative gravity anomaly that suggests a significant mass deficiency in the trench region. Here the crust and mantle are being deformed faster than the earth can adjust itself. The bottoms of the trenches are of the order of 10 km across, and if we assume that subduction is occurring at the maximum credible rate, the present floor of the trench will be subducted beneath the continental plate in approximately 250 000 years—a time comparable to that during which nuclear wastes

CONTINENTAL MARGINS

(Shelf, Slope, Rise including Deltas, Cones, Fans)

High resources including

food	recreation
minerals	sand and gravel
hydrocarbons	habitation?

Shallower (<4000 m)

Unstable

Strongest currents

Variable conditions (temporally and geographically)

High sedimentation or erosion

Biologically active

TRENCHES

Seismically active
 Volcanic activity
 Unstable sediments
 slumping
 sliding
 turbidity currents
 Adjacent to continents
 detrital sediments
 upwelling and biological productivity
 rich fishing grounds
 Deepest areas
 Subduction might provide permanent disposal
 Severest technological problems

will cease to be dangerous. Little or nothing is known about the piling up of sediments on the leading edge of the continental plate, or about crustal deformations in the trench area. These uncertainties place in doubt the actual subduction of any particular piece of trench floor, and hence whether such a location would be a practical waste depository.

The dynamic processes of the MOR are better understood than are those of the trenches. Here new molten lava flows onto the sea floor, causing crustal growth, and shallow but severe earthquakes occur at rates of up to 10 per day. Currents, strongest on the highest mountains near the ridge crest, are generally weak and variable on the deeper (>3 km) and more subdued flanks of the ridge. Cores from this region show a relatively high degree of variability, with changes in texture, color, and composition, indicating varying conditions of deposition.

MID-OCEANIC RIDGE CREST

Unstable
 seismic (almost constant earthquakes)
 volcanism
 No sediments
 Rough topography
 Shallower (2 to 4 km)
 Hot (molten basalt near surface)
 Strong and variable currents (>50 cm/sec in some areas)

As we progress further from the rift valley to depths of over 4-1/2 km on the lowest flanks of the MOR, we again reach the Abyssal Hills or Swales where stability and uniformity become a dominant characteristic, where biologic activity is at its lowest, and where the very cold currents move slowly. Sedimentation is in the form of a rain of fine-grained material from the surface.

Cores from abyssal-hill regions tend to be very homogeneous, consisting of abyssal brown clays with an occasional manganese nodule or micro-nodule. Bottom currents are weak, slopes are gentle (<5 deg local, <1/2 deg regional), sediment is uniform, and major earthquakes are unknown or, as yet unrecorded. All indications from the cores, seismic profiles, and our emerging understanding of global deep circulation attest to the stability of these regions over tens of millions of years.

MID-PLATE/GYRE (abyssal hills)	
Stable	
aseismic	
sediments (no slumping or turbidity currents)	
>10 ⁵⁻⁶ yr since last major changes	
Invariant conditions (temporally and geographically)	
Slow currents (2 to 5 cm/sec)	
Low bio-productivity (low on surface, very low on bottom)	
Low relief (40 ft/mile)	
Manganese nodules are the only identified resource	

The abyssal hills occupy over a third of the total deep ocean floor, or about a quarter of the surface of the planet. These vast areas (10⁸ km²) probably represent the least "useful" part of our planet and will be among its very last habitable regions.

By far the largest expanse of abyssal hills-about the size of the United States-lies in the North Pacific, south of the Aleutian Trench, north of Hawaii, west of the Aleutian Abyssal Plain off Canada, and east of the Emperor seamounts. This area is being studied as a "type-region" that best exemplifies the stable mid-plate/mid-gyre resource-limited real estate.

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CONCLUSIONS

If mankind must deal with radioactive wastes, it is incumbent upon us to search all of the earth's surface for areas that might be useful in this

regard. About 70% of the earth's surface is covered by the oceans, and from a brief survey of the 70%, the authors reached several conclusions:

1. The mid-plate/mid-gyre regions of the major oceanic basins are the most stable and perhaps most otherwise useless regions on earth.
2. Emplacement of waste canisters well below the surface of the sediments, likely deep into the basalt, seems possible and appears to provide a set of barriers to the release of radionuclides.
3. While there are not sufficient data now to assess rates and kinds of sea-floor processes, or the technical feasibility of seabed disposal of radioactive wastes, enough is known to define the questions and the systematic interdisciplinary effort required to answer them.
4. The central North Pacific-the largest mid-plate/mid-gyre region in the world-is one suitable and convenient area for the initiation of such studies.

In conclusion, we have identified a type of area on the planet that deserves careful study as a possible repository, and we cannot presently define any insurmountable technical barriers to such a disposal concept.

APPENDIX E

GOVERNMENT REGULATIONS PERTAINING TO THE
MANAGEMENT OF RADIOACTIVE MATERIALS

APPENDIX E

GOVERNMENT REGULATIONS PERTAINING TO THE
MANAGEMENT OF RADIOACTIVE MATERIALS

The following is a list⁽¹⁾ of the major areas of waste management covered by regulations in the Code of Federal Regulations:

- 10 CFR Part 20 - Radiation Protection Standards
- 10 CFR Part 73 - Physical Protection Standards (Security) for Fixed Facilities and Transportation
- 10 CFR Part 30 - Possession and Use of Byproduct Materials (Licensing)
- 10 CFR Part 40 - Possession and Use of Source Materials (Licensing)
- 10 CFR Part 70 - Possession and Use of Special Nuclear Materials (Licensing)
- 10 CFR Part 50 - Utilization and Production Facilities (Licensing)
- 10 CFR Part 71 - Containers for Transportation of Nuclear Materials (Licensing)
- 10 CFR Part 51 - Implementation of NEPA
- 10 CFR Part 150 - Agreement State Program
- 40 CFR Part 220-227 - Sea Disposal Permit Program

Table E.1 indicates the status of waste disposal/storage activities as reported by the U.S. Nuclear Regulatory Commission in November 1975.

FEDERAL REGULATIONS REGARDING TRANSPORTATION

Regulations concerning transportation of radioactive materials have historically been a matter of widespread practical interest because of the amount of transport in this area. These regulations and their implementation are discussed below.

Regulations governing the transport of radioactive materials have been established to prevent the loss or dispersal of material during shipment and to insure the safety of the public and the transportation workers. There is overlapping responsibility for regulating the safe transport of radioactive materials. Primary responsibility at the Federal level lies with the Department of Transportation (DOT) Materials Transportation Bureau and the Nuclear Regulatory Commission (NRC).

A "Memorandum of Understanding" between the two agencies was signed in 1966 and revised in 1973.⁽²⁾ This memorandum calls for cooperation between DOT and NRC and delineates the responsibilities of each agency. DOT is responsible

TABLE E.1 Status of Waste Disposal/Storage Activities⁽¹⁾

MAJOR REGULATIONS ESTABLISHED BY NRC				
<u>Waste Type</u> <u>- Source</u>	<u>Industry</u> <u>(or ERDA)</u> <u>Strategy Etab.</u>	<u>Waste Producing</u> <u>Facilities</u>	<u>Offsite</u> <u>Storage/Disp.</u>	<u>Det'd Reg.</u> <u>Criteria</u> <u>Etab. by NRC</u>
HLW				
- Commercial	YES	YES	YES	PART
- ERDA	PART	--(d)	(YES)	NO
Hi-γ TRU				
- Commercial	NO (in PREP.)	YES	YES ^(a)	NO ^(a)
- ERDA	PART	---	(YES)	NO
Lo-γ TRU				
- Commercial	PART	YES	YES ^(a)	PART ^{(a)(e)}
- ERDA	PART	---	(YES)	NO
LLW (non-TRU)				
- Commercial	YES	YES	YES ^(b)	PART ^(e)
- ERDA	YES	---	(YES)	NO
Gases				
- Commercial	NO	YES	YES	NO
Mill Tailings	NO (or PART)	YES	NO ^(c)	PART
Decom Facilities				
- Commercial	NO (or PART)	YES	YES	PART
- ERDA	NO (or PART)	---	---	---

a. Proposed regulation on TRU gives ERDA possession; licensing by NRC was not made specific in the Energy Reorganization Act.

b. Shared with Agreement States

c. Based on previous legal interpretations of definitions of source and byproduct materials.

d. Except specific licensed ERDA facilities.

e. Shared with EPA.

for promulgating and enforcing safety standards governing packaging and shipping containers and for the labeling, classification, and marking of all packages. DOT also implements safety standards for the mechanical condition of carrier equipment and qualifications of carrier personnel. NRC develops performance standards for package designs and reviews package designs for Type B, fissile and large quantity packages (defined below). DOT requires NRC approval to use these packages.⁽³⁾ The Federal Aviation Administration (FAA), the Interstate Commerce Commission (ICC), and the U.S. Coast Guard also exercise some regulatory authority over the shipment of radioactive materials.

The transportation or packaging for transport of radioactive material is subject to issuance of the appropriate licenses. Applicants for a license to

package or to transport radioactive material must show by a combination of analysis and experiment that the proposed package or transport vehicle satisfies all the requirements set forth in the Code of Federal Regulations. The application must describe proposed controls or precautions to be used in the loading, unloading, handling and transport of radioactive material, and the procedures to be followed in the event of an accident or delay in shipment. Inspection and accountability procedures must also be described.

The following Federal Regulations are applicable to the transport of radioactive materials:

- Title 49 Code of Federal Regulations Parts 170-199 (49 CFR 170-199) - Department of Transportation regulations governing the transport of hazardous materials.
- 10 CFR 71 - Nuclear Regulatory Commission regulations governing the packaging and shipment of radioactive materials.
- 14 CFR 103 - Federal Aviation Administration regulations for shipment of radioactive materials by air.
- 46 CFR 146 and 149 - U.S. Coast Guard regulations governing the shipment of radioactive materials by water.
- 10 CFR 73 - Nuclear Regulatory Commission regulations for the protection of special nuclear material in transit.

DOT and NRC regulations are the most important for shipments in the Nuclear Fuel Cycle. These regulations will be reviewed here.

Transport Group

Radioactive materials are classified for transportation purposes into one of seven transport groups according to their radiotoxicity and potential hazard if released to the environment.* Transport Group I is the most restrictive. Materials such as plutonium are placed in this group. Materials such as tritium gas with a relatively low hazard potential are placed in Transport Group VII.

Quantity Designations

A shipment of radioactive material is classified as Type A, Type B, or Large Quantity of Radioactive Material depending on the amount of radioactivity it contains. The dividing line between the quantity types is dependent on the Transport Group of the material being shipped. (See Table E.2).

* It is anticipated that the U.S. Regulations will be changed in the near future to agree in substance with 1973 IAEA Regulations. The transport group designation has been replaced with a different system in these regulations. (See References 4 and 5).

TABLE E.2 Maximum Activities for Normal Form* Type A and B Quantities of Radioactive Material by Transport Group

<u>Transport Group</u>	<u>Type A Quantity, Ci</u>	<u>Type B Quantity, Ci</u>
I	0.001	20
II	0.05	20
III	3	200
IV	20	200
V	20	5,000
VI and VII	1,000	50,000

NOTE: Any quantity greater than Type B is defined as a "Large Quantity" of radioactive material.

* Normal form material is everything that does not meet the "Special Form" criteria outlined below.

If a shipment contains materials from more than one Transport Group, the most restrictive group will generally apply for the determination of quantity type. (See 10 CFR 71.4(p)).

Low Specific Activity (LSA) Materials

Some shipments of material with a relatively low hazard potential are exempt from certain of the requirements in the regulations even though they may contain Type A or B quantities of radioactive materials. LSA materials include uranium or thorium ores, unirradiated depleted or natural uranium, unirradiated thorium; aqueous solutions of tritium oxide containing less than 5 mCi/ml; and materials having uniformly distributed activity in which the concentration per gram does not exceed the following:

- 0.1 μ Ci of Group I materials
- 5.0 μ Ci of Group II radionuclides
- 300 μ Ci of Group II and IV materials.

Externally contaminated nonradioactive materials may be considered LSA provided the contamination is not readily dispersible and the average surface contamination per square centimeter does not exceed the following:

- 0.1 μ Ci for Group I radionuclides
- 1.0 μ Ci for other materials.

Special Form Material

Radioactive material in special form may be transported as if it were Transport Group V regardless of the radioactive species present. Special form material meets one of the following criteria:

1. Radioactive material in solid form having no dimension less than 0.5 mm (or one dimension greater than 5 mm). Will not sublime, melt or ignite in air at 538°C. Remains nondispersible after immersion for one week in water at 20°C or air at 30°C.
2. Radioactive material contained in a capsule that will satisfy the above requirements.

Special form material must remain intact after a test that includes a 9.1-m drop, impact of 1.4 kg steel rod dropped from a height of 1 m, heating to 800°C for 10 min and immersion in water for 24 hr.

Fissile Classification

A shipment containing fissile materials (^{233}U , ^{235}U , ^{238}Pu , ^{239}Pu , ^{241}Pu) is classified according to the amount of control which must be exercised to prevent nuclear criticality in shipment.

- Fissile Class I packages may be shipped in unlimited numbers in any arrangement, requiring no nuclear criticality considerations.
- Fissile Class II packages may be shipped together in any arrangement but their numbers are limited to an aggregate Transport Index (see definition below) of 50. For purposes of nuclear criticality, safety, individual packages must have a transport index between 0.1 and 10.
- Fissile Class III packages must be controlled in transport by special arrangements between the shipper and the carrier to provide nuclear criticality safety.

Certain shipments of fissile materials are exempt from the normal requirements for packages containing fissile material. These packages meet one of the following criteria (among others - see 10 CFR 71.9):

1. contain less than 15 g fissile material
2. contain Th or U with less than 0.72% fissile material
3. contain less than 350 g fissile material with a maximum of 5 g in any 0.028 m^3 (1 ft^3) of the package.

Transport Index

The transport index is a number placed on a package of radioactive material to designate the degree of control to be exercised by the carrier during transportation. The transport index is the larger of the following numbers:

1. the highest radiation dose rate in millirem/hr at 91 cm from any accessible external surface of the package; or
2. for Fissile Class II packages the number 50 divided by the maximum number of such packages which may be transported together without criticality safety considerations.

Except in a sole use vehicle, the aggregate transport index of all packages in any storage area during transportation must not exceed 50. (Shipments of Fissile Class II packages can never exceed an aggregate transport index of 50.) The transport index also determines the labeling requirements for the package.

Special Nuclear Material (SNM)

A shipment of radioactive material is designated special nuclear material if it contains 5000 g or more of ^{235}U (contained in uranium enriched to 20% or more), or 2000 g or more of ^{233}U or plutonium, or any combination of the three computed by the formula.

$$\text{grams} - \text{grams } ^{235}\text{U} + 2.5 (\text{grams } ^{233}\text{U} + \text{grams Pu}).$$

10 CFR 73 requires that shipments of special nuclear material be afforded additional physical protection in transit to prevent theft or sabotage. These protective measures include:

- NRC approved pre-planned carrier procedures
- hand-to-hand receipts
- tamper-indicating seals on all containers
- minimum container weights of 227 kg for shipments in open vehicles
- no cargo transfers enroute
- periodic radio communication between transport vehicle and carrier dispatcher
- armed escorts or specially designed transport vehicles with disabling features.

Radiation Dose Rate Requirements

Radiation dose rates from shipments of radioactive materials are also limited by the regulations. Larger maximum radiation dose rates are allowed for vehicles being used exclusively to transport radioactive materials than for nonexclusive use vehicles. For vehicles not in exclusive use the following limits apply:

- 200 mrem/hr at any point on the external surface of the package
- 10 mrem/hr at 91 cm from the package

For vehicles being used exclusively for the transport of radioactive materials the limits are as follows:

- 1000 mrem/hr at 91 cm from the external surface of the package (closed transport vehicle only)
- 200 mrem/hr at any point on the external surface of the vehicle (closed transport vehicle only)

- 10 mrem/hr at 1.8 m from the external surface of the vehicle
- 2 mrem/hr in any normally occupied position in the vehicle.

These dose limits are illustrated in Figure E.1 for truck transport.

Packaging

The type of packaging required for a shipment of radioactive material is dependent on the type, quantity, form and fissile classification of the material to be shipped. 49 CFR allows shipment of radioactive materials in DOT specification containers, in containers licensed by NRC or in containers authorized by special permit from DOT.* The specification containers are described in the regulations. All radioactive material shipping containers must meet the general package requirements for radioactive material shipping containers 49 CFR 173.393 and the general requirements for all packaging used in interstate commerce contained in 49 CFR 173.24.

Low specific activity material shipped in exclusive use vehicles is exempted from most of the packaging requirements of 49 CFR. Basically only strong, tight packaging is required. The radiation dose limits discussed previously are still applicable, however.

Type A quantities of radioactive material must be shipped in DOT specification 6M packaging, approved Type B packaging or packaging that meets the requirements of DOT specification 7A (See 49 CFR 178.350). In addition to meeting the general packaging requirements of 49 CFR 173.24 and 49 CFR 173.393, specification 7A packaging must be capable of maintaining its shielding integrity and preventing the dispersal of its contents during a test simulating conditions normally incident to transportation. These test conditions are outlined below. The shipper must maintain on file for at least one year after the latest shipment a complete certification and supporting safety analysis report demonstrating that the packaging is in compliance with Specification 7A.

Test Representing Conditions Normally Incident to Transportation

- Heat - direct sunlight at an ambient temperature of 54°C in still air.
- Cold - an ambient temperature of -40°C in still air and shade.
- Pressure - one half standard atmospheric pressure.
- Vibration - normally incident to transport.

* Special permits are no longer issued for radioactive material shipping containers. The special permits now in effect will be allowed to expire without renewal. In the future all radioactive material will be shipped in DOT specification or NRC licensed containers.

LIMITS FOR NON-EXCLUSIVE USE VEHICLE

LIMITS FOR EXCLUSIVE USE VEHICLE

FIGURE E.1. Radiation Dose Limits (Reference 5)

- Water spray - sufficient to keep the exposed surfaces of the package except the bottom wet for 30 min.
- Free drop - 1-1/2 to 2-1/2 hr after the water spray test at a distance of 1.3 to 1.2 m (depending on package weight) onto an unyielding surface.
- Corner drop - onto each corner in succession (each quarter of each rim for cylinders) from a height of 30 cm onto an unyielding surface (applies only to wood or fiberboard containers of less than 45 kg and to all Fissile Class II Packages).
- Penetration - impact of the hemispherical end of a 5.9 kg steel cylinder 3.2 cm in diameter dropped from a height of 1 m onto the most vulnerable surface of the package.
- Compression - a compressive load against the top and bottom of the package in its normally transported condition. The load is the greater of five times the weight of the package or $1.38 \text{ neutrons/cm}^2$ times the minimum horizontal cross section of the package. (This test applied only to packages weighing less than 4535 kg.)

Type B quantities of radioactive material may be shipped in DOT specification 6M containers, in packaging authorized by NRC or in packaging meeting the 1973 IAEA requirements (provided DOT has validated the approval of the foreign competent authority). Type B packaging submitted to NRC for approval must be able to maintain integrity through the test conditions outlined below simulating transport accident environments as well as the test simulating conditions normally incident to transportation outlined above.

Hypothetical Transportation Accident Test Conditions

(Tests applied sequentially to determine cumulative effect on the package.)

- Free drop - through a distance of 9.1 m onto an unyielding surface, striking the surface in the position which would produce maximum damage.
- Puncture - a free drop through a distance of 1 m onto a cylindrical steel bar 15.2 cm in diameter with the edge rounded to a radius of 6.4 mm. The length of the bar is selected for maximum damage (minimum length is 20.3 cm). The bar is mounted on an unyielding horizontal surface and the package is dropped in a position expected to produce maximum damage.
- Thermal - exposure to the equivalent of an 800°C fire for 30 minutes with no artificial cooling for 3 hrs after the exposure.
- Water immersion - (fissile material packages only) at a depth of 91 cm for 8 hours.

Large quantities of radioactive materials are shipped in Type B packaging. The packaging used must provide adequate shielding and heat removal capabilities for the radioactive material to be transported.

Fissile material may be shipped in DOT specification 6M or 6L (Type A quantities only) packaging or in packaging approved by NRC.

Packaging approved by NRC for fissile materials must meet some requirements not posed on containers for other radioactive materials (See 10 CFR 71.38 - 71.40). These conditions insure a wide margin of safety on the number of packages which can be shipped at one time. They also insure that a critical mass could not be assembled in the event of a transportation accident.

Special Regulations Applying to Plutonium

Because of its hazardous nature, there are some special regulations in effect for the shipment of plutonium. Plutonium may presently be shipped as a liquid (in nitrate solution) or as a solid (plutonium oxide). After June 1978, plutonium in excess of 20 Ci per package must be shipped as a solid (10 CFR 71.42). The plutonium must be double packaged. The inner container must maintain its containment integrity as the outer container undergoes the tests for Type B packaging. The separate outer container must meet the requirements for packaging of material in normal form.

Federal regulations prohibit shipments of plutonium in passenger aircraft in quantities exceeding 20 g or 20 Ci, whichever is less (10 CFR 71.42). However, a recently enacted Federal statute has placed an embargo on shipments of plutonium by air until it is shown that the shipping containers will not release their contents in an airplane crash environment.

STATE REGULATIONS

Although Federal agencies dominate the regulatory process for the transport of radioactive materials, state governments also exercise some control over these shipments. State highway departments regulate gross vehicle weights, vehicular dimensions and other parameters for radioactive shipments just as they do for other kinds of shipments. About one-half of the states have adopted the U.S. DOT Hazardous Materials Regulations to cover intra-state shipments. Several states have adopted or proposed additional regulations concerning radioactive materials.^(2,7) These include:

- special routing of radioactive shipments
- advance notification for shipments of large quantities of materials
- state inspections of some types of radioactive shipments
- prohibition of certain types of shipments within the states
- prior approval for radioactive shipments
- requirement of exclusive vehicle use for radioactive shipments

- use of pilot vehicles
- speed restrictions for radioactive shipments
- specific hours of movement
- accompaniment of all shipments by radiation monitoring personnel.

The variation of regulations between adjacent states can often require special considerations for interstate shipments.

There is a potential conflict between some of the proposed state laws and the provisions of the National Transportation Act of 1974. (Public Law 93-633 signed in 1975) This law prohibits the states from adopting laws or regulations more stringent than Federal regulations unless the state regulations improve transportation safety. Even in this case, such rules can be adopted only if they do not unreasonably burden commerce.

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