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Geothermal Resources of the Northern Gulf of Mexico Basin

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ABSTRACT

Published geothermal gradient maps for the northern Gulf of Mexico basin indicate little or no potential for the development of geothermal resources. Results of deep drilling, from 4000 to 7000 meters or more, during the past decade however, define very sharp increases in geothermal gradient which are associated with the occurrence of abnormally high interstitial fluid pressure (geopressure). Bounded by regional growth faults along the landward margin of the Gulf Basin, the geopressured zone extends some 1300 km from the Rio Grande (at the boundary between the United States and Mexico) to the mouth of the Mississippi river. Gulfward, it extends to an unknown distance across the Continental Shelf.

Within geopressured deposits, geothermal gradients range upwards to 100 °C/km, being greatest within and immediately below the depth interval in which the maximum pressure gradient change occurs. The 120 °C isogeotherm ranges from about 2500 to 5000 m below sea level, and conforms in a general way with depth of occurrence of the top of the geopressured zone. Measured geostatic ratios range upward to 0.97; the maximum observed temperature is 273 °C, at a depth of 5859 m.

Dehydration of montmorillonite, which comprises 60 to 80 percent of clay deposited in the northern Gulf Basin during the Neogene, occurs at depths where temperature exceeds about 80 °C, and is generally complete at depths where temperature exceeds 120 °C. This process converts intracrystalline and bound water to free pore water, the volume produced being roughly equivalent to half the volume of montmorillonite so altered. Produced water is fresh, and has low viscosity and density. Sand-bed aquifers of deltaic, longshore, or marine origin form excellent avenues for drainage of geopressured deposits by wells, each of which may yield 10,000 m^3 or more of superheated water per day from reservoirs having pressures up to 1000 bars at depths greater than 5000 m.

Introduction

Observed geothermal gradients and measurements of the thermal conductivities of the rocks of the outer crust of the earth suggest that the geothermal flux is relatively constant worldwide; in tectonically « quiet » regions it does not vary appreciably, even from the continental land mass to the deep sea floor (JACOBS, RUSSELL, WILSON 1959, p. 102). Tectonically « active » regions, in contrast, are characterized by wide variations in the geothermal flux; it may be either abnormally high, or abnormally low. Some authorities believe that geothermal energy released along preferred belts is the primary cause of crustal deformation (VENING MEINESZ 1952, p. 528-553). Heat generated more than a few hundred kilometers beneath the earth's surface could not, under existing geothermal gradients, reach the surface by conduction alone in more than a billion years. It therefore seems likely that the development of higher than « normal » (WHITE 1965, p. 2) geothermal gradients and heat flow rates at or near the earth's surface must involve mass transfer through a considerable depth range, 5 to 20 km or more, in geologically brief time intervals. Igneous intrusion and volcanism provide classic evidence of mass transfer upwards from deep in the crust or upper mantle.

The sites of igneous intrusion and volcanism in the present and recent geologic past fall within rather well-defined belts. These same belts define major earthquake epicenter alignments (RICHTER 1969, p. 173; KNOPOFF 1969, p. 1280; RICHTER 1968, pp. 404-408). Tensile rupture of the crust has occurred along some belts (e.g., the mid-Atlantic ridge) whereas elsewhere compression has caused superficial crust to be thrust into the mantle, forming trenches (e.g., the Unique Bolero in the Spanish Mediterranean).

Although not defined by alignment of earthquake epicenters, modern geosynclines are the sites of large-scale subsidence, in which enormous masses of sediments have been « transferred » downward into the crust 5 to 10 km or more in geologically-short time intervals. Faults with more than 1 km of throw are not uncommon in the Gulf of Mexico basin, and normal faulting is clearly its most important structural feature (MEYERHOFF 1968, p. 377). Such geosynclinal downwarp must produce large changes in the physical and chemical state of the underlying upper mantle, as well as of the sedimentary fill (LEHNER 1969, p. 2473). Poorly-consolidated sediments with an appreciable water content are subjected to intense heat and enormous pressure. In effect, geosynclinal subsidence with rapid sedimentary fill may be likened to igneous intrusion, except that the motion is downward, the intrusive mass is cold, and the intruded rocks are hot.

Because the water content of young geosynclinal sediments is appreciable, even at depths of 6 to 8 km in abnormally-pressured basins, the geothermal regimes are closely related to the hydrology. If the sediments are mainly clay of a swelling variety, as in the northern

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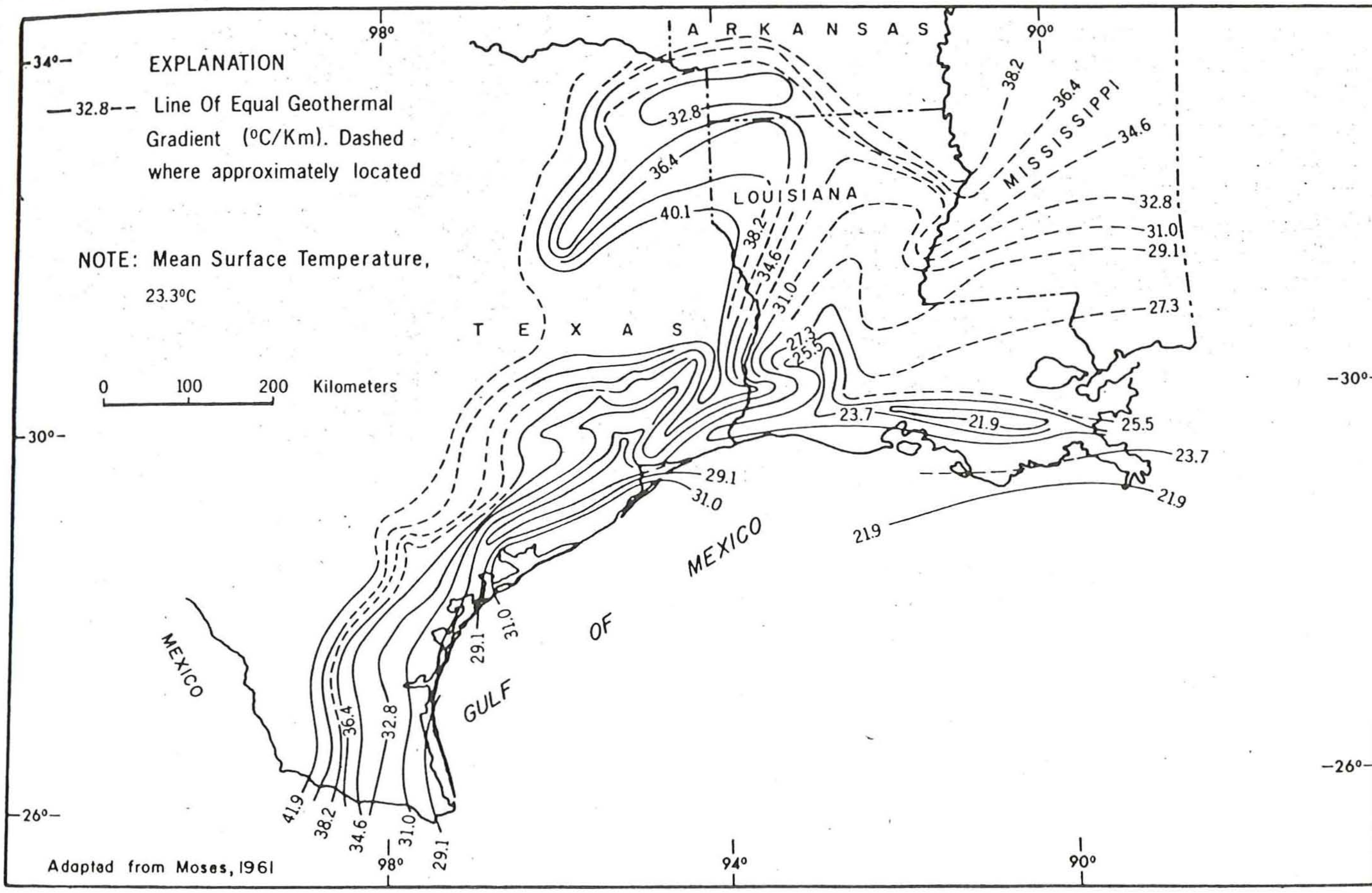

FIG. 1. — *Geothermal gradients in the northern Gulf of Mexico basin.*

Gulf of Mexico basin (MILNE, EARLEY 1958, p. 328; KERR, BARRINGTON 1961, p. 1697), then endothermic diagenesis of clay minerals may alter greatly the geothermal flux, reducing heat flow to the land surface (BURST 1969, pp. 73-93; MEYERHOFF 1968, fig. 4, p. 382). Published geothermal gradient maps of southwestern United States (NICHOLS 1947; MOSES 1961) show the northern Gulf of Mexico basin as a negative geothermal area (Figure 1). These maps are based upon temperatures measured at depths generally less than 2 km; they show decreasing geothermal gradients in the direction of the geosynclinal axis. They cannot be used to predict temperature conditions in the underlying zone of abnormally-high fluid pressure; and with the very best data on the thermal conductivity of the sediments, they would not indicate the true magnitude of the regional geothermal flux in the northern Gulf of Mexico basin.

Geosynclines are the forerunners of mountain chains (e.g., DE SITTER 1964, p. 390) consisting of folded and uplifted sediments, in many places intruded by igneous rocks. The Gulf Coast geosyncline now has dimensions comparable to those of the ancient geologic past at the time of their maximum development. Already the deeper sediments in the geosyncline must be undergoing metamorphism; and because of the appreciable residual water content of sediments in the zone of incipient metamorphism, there must be a continuing upward discharge of hydrothermal fluids from it.

The geothermal resources of the northern Gulf of Mexico basin differ from those of known geothermal regions in the source of the waters and in the hydrology of the systems. Thermal resources of deep geosynclinal basins are not dependent upon recharge and deep circulation of meteoric water; their thermal waters are derived from the sediments themselves, and fluid depletion occurs with energy release. However, the very large amount of water in storage makes possible a continuing large-scale rate of energy production.

Geologic setting

The northern Gulf of Mexico basin occupies the Gulf Coast geosyncline (BARTON, RITZ, HICKEY 1933, p. 1446) (Figure 2). This great structural trough extends some 1500 km from northeastern Mexico to western Florida.

According to MEYERHOFF ET AL. (1968, p. 377), « Five basic geological factors have affected the development of the Gulf Coast geosyncline since its beginning. The first of these is the structural grain of the Paleozoic Ouachita orogenic belt which borders the north and northwest sides of the Gulf coastal plain. The lines of structural weakness inherited from this tectonic belt almost certainly controlled the geometric form of the

Gulf Coast geosyncline. Second, a depression (the Gulf of Mexico) already existed and, therefore, was accessible as a potential site of geosynclinal formation. Third, subsidence kept pace with deposition in the geosyncline. Fourth, a thick salt sequence of Late Triassic to Middle Jurassic age imparted an important element of structural mobility to the geosyncline. Finally, beginning in Paleocene time, the rising Rocky Mountains supplied a high volume of sediments to the geosyncline ».

Subsurface conditions are well known between the Rio Grande Embayment and southern Alabama, where the deposits have been explored by some 300,000 test wells and innumerable intensive geophysical surveys. In two centers of deposition, one adjacent to the coast of southern Texas and the other adjacent to the southern coast of Louisiana, sediments of Mesozoic and Cenozoic age are known to exceed 15 km in thickness (Table 1). In southern Louisiana alone, their volume exceeds 1,600,000 km³, of which more than half have been deposited in the last 26 million years (MEYERHOFF 1968, p. 376, 400).

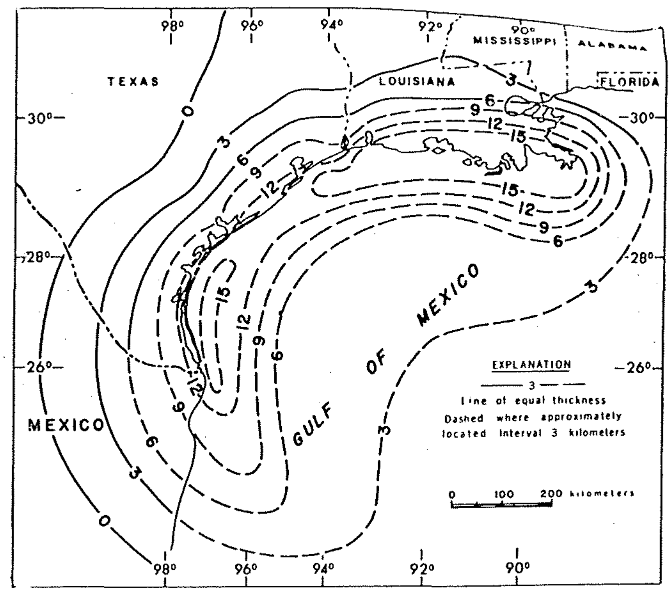

FIG. 2. — *Thickness of Cenozoic deposits in the Gulf Coast geosyncline.*

TABLE 1. — *Deposition rates in areas of thickest sediments, Texas and Louisiana Gulf Coast (modified from* MEYERHOFF ET AL. *1968).*

Age	Duration (million years)	Thickness of Sediments: Coastal Louisiana (feet)	Coastal Louisiana (meters)	Coastal Texas (feet)	Coastal Texas (meters)	Sedimentation Rate (cm/1,000 years) Louisiana	Texas
Holocene and Pleistocene	2	8,000	2,400	2,000	600	120	30
Pliocene	12	8,000	2,400	3,000	900	20	25
Miocene	12	25,000	7,500	10,000	3,000	62.5	7.5
Anahuac, Frio, and Vicksburg	11	17,700	5,300	16,000	4,800	48.5	43.5

The deposits are mainly noncarbonate clastic rocks, sand and clay, or sandstone and shale. Facies distribution reflects deposition from the north and west (Figure 3) resulting in a progressive Gulfward shift of the deltaic and prodelta sediments as basin filling continued. The Gulf Coast geosyncline developed true geosynclinal proportions only during the last 25 to 30 million years (BORNHAUSER 1958, p. 341). And, although rates of deposition in the Texas and Louisiana depocenters have varied widely during this time interval, there appears to have been a progressive acceleration. Although not indicated in Table 1, it is known that the Pleistocene and Holocene deposits off the Louisiana coast exceed 3600 m in thickness (FAILS, SACHS 1967, p. 62).

As suggested in Figure 3, more than half the volume of these deposits is clay; and according to MILNE and EARLEY (1958, p. 328), « montmorillonite, the predominant clay mineral in the Mississippi River and delta sediments, is apparently the stable product of soil development and rock weathering in the drainage basin of the Mississippi River ». Extensive studies of cores from oil test wells to depths of 6000 m and more confirm that, before thermal diagenesis, montmorillonite commonly forms 60 to 80 percent of the clay fraction.

BURST (1969, p. 81) shows that, after recent burial, 72 percent of the clay is of the « swelling » variety. He states (p. 73) that what he terms « second stage dehydration » results from thermal diagenesis of montmorillonite, yielding an amount of water equal to « 10 to 15 percent of the compacted bulk volume » of the rock; that this semisolid interlayer water begins to be liberated at depths where formation temperature exceeds 80 °C; and that release is virtually complete at depths where formation temperature exceeds 120 °C. However, about 20 percent of the montmorillonite remains unaltered, probably because there is insufficient potassium present in the sediments to enable complete conversion to illite.

Diapiric salt, in the form of domes and massifs (ATWATER, FORMAN 1959, p. 2592) derived from the deeply-buried Jurassic Louann evaporite, forms the principal structures from which the production of petroleum in the Gulf of Mexico basin in obtained. HALBOUTY (167, p. 5) reports 409 domes in the basin where salt has been penetrated (Figure 4). The density of rock salt (halite) is 2.165 at standard temperature, whereas the average density of unconsolidated sediments in the basin ranges from about 2.2 to 2.5. Salt behaves as an ideal plastic at temperatures above 300 °C (HANDIN,

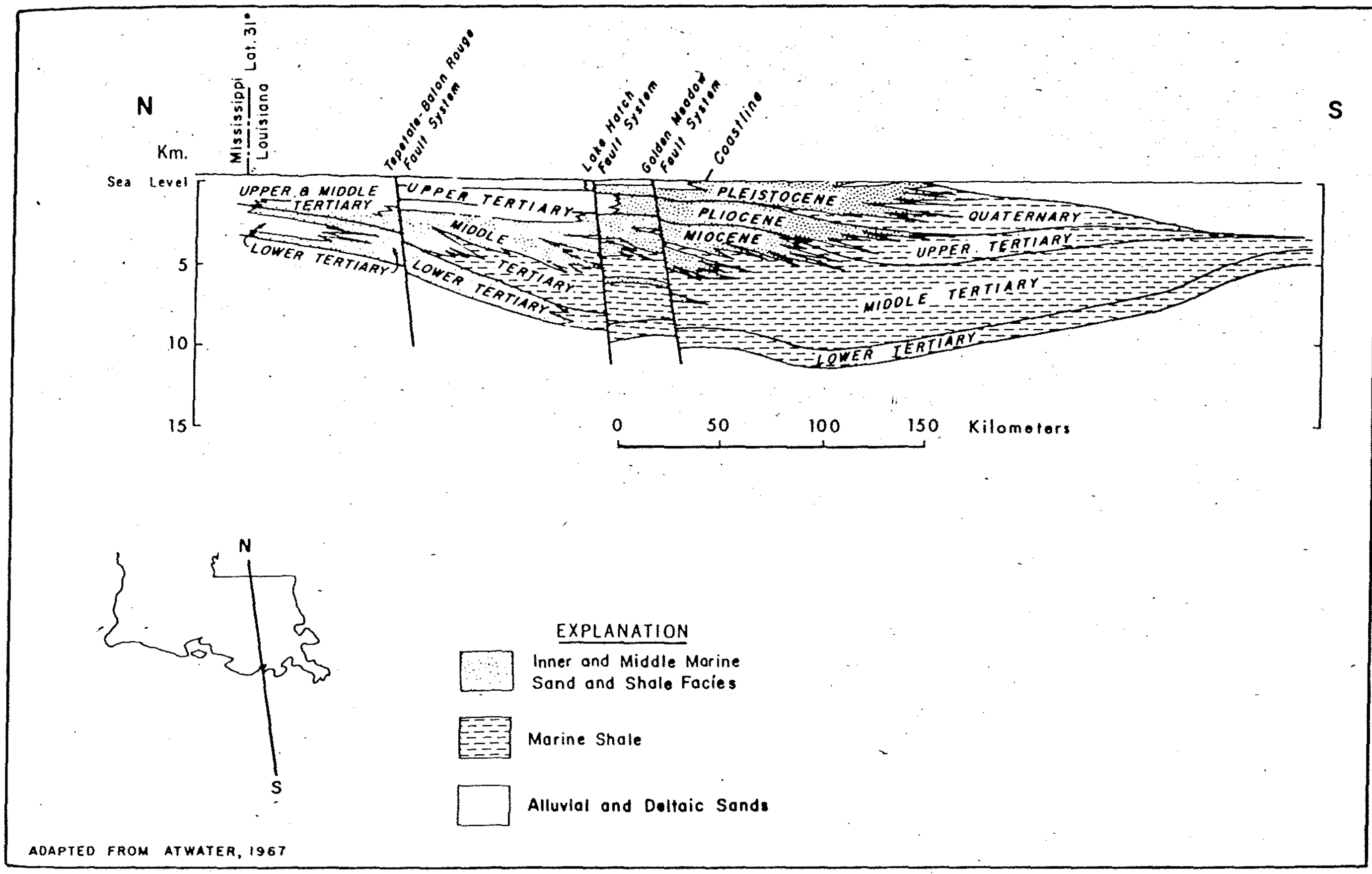

FIG. 3. — *Geologic cross section through southeastern Louisiana and adjacent Continental Shelf.*

HAGER 1958, p. 2901) and its strength at 300 °C is less than 15 percent of its strength at 25 °C. Observed temperatures at depths greater than 5 km (HOUGH, COUVILLION 1966, pp. 142-158) indicate that ideal plastic flow of salt may be the rule at depths greater than about 6 km in the northern Gulf Basin.

Where the salt diapir rises through abnormally-pressured marine or prodelta clay, it is enveloped by a highly fluid mud sheath. The temperature of the deposits pierced is raised locally because the thermal conductivity of the salt plug is several times greater than that of the adjacent sediments. Where the rising temperature exceeds 80 °C, diagenesis of montmorillonite in the enveloping beds occurs, producing a clay slurry or slush. The density of this slurry is comparable to that of salt, whereas its fluidity is much greater; it commonly escapes upward along the flanks of the salt plug and intrudes the sediments above the salt. Such clay diapirs associated with salt domes are not uncommon in the northern Gulf Basin (ATWATER, FORMAN 1959, p. 2594; ATWATER 1967, p. 9), where they are known to attain thicknesses of several thousands of feet (KERR, BARRINGTON 1961, pp. 1697-1712). While fluid, these diapirs are excellent thermal insulators.

Where the top of a diapir enters the main sand series of a deltaic sequence pressure-connected to the atmosphere, its salt is rapidly dissolved and the insoluble residues accumulate, forming « caprock ». Here, the upward movement of the diapir is much reduced, partly because of the increased « drag » of sandy sediments, but mainly because of the cooling effect caused by solution of salt and the acceleration of upward heat loss as convective flow occurs in the sand sequence above the diapir. Salt loses its plasticity rapidly as it cools below 300 °C.

Thus, the areal and vertical distribution of salt in the depositional mass is related to facies distribution, not only because sediment density varies with sediment type, but also because salt diapirism through sandy sediments of the normally-pressured zone produces effects that differ greatly from those associated with diapirism through shaly sediments (ATWATER, FORMAN 1959, pp. 2592-2622).

The most characteristic structural feature of the northern Gulf of Mexico basin is the growth fault. Defined by OCAMB (1961, p. 139), growth faults are normal faults « which have a substantial increase in throw with depth and across which, from the upthrown to the downthrown block, there is a great thickening of correlative section ». Regional growth faults, generally parallel to the axis of the geosyncline, are formed by intersecting arcuate faults whose throw may be constant for many miles; individual fault zones, with net downward displacement gulfward, can be traced for hundreds

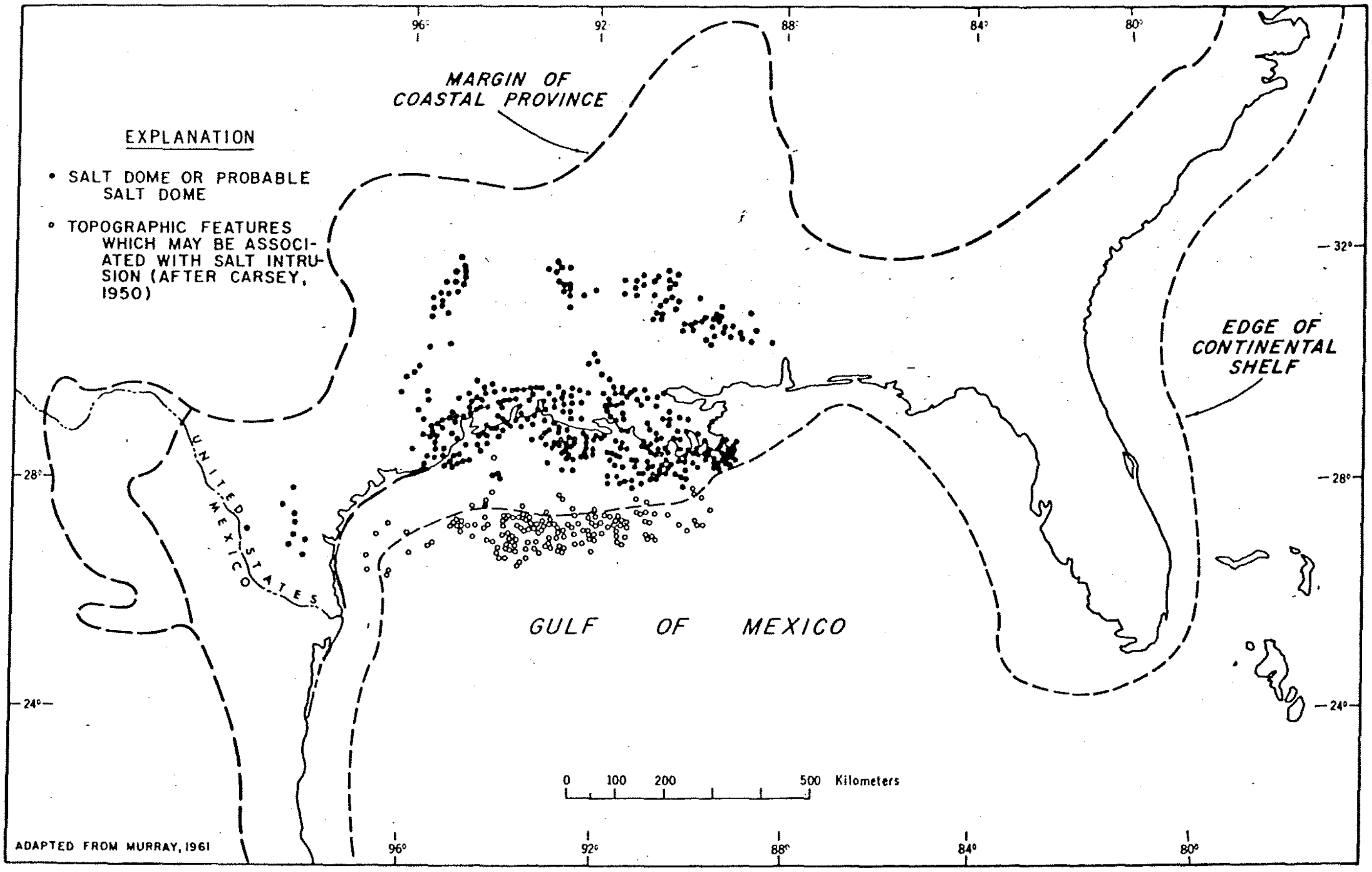

FIG. 4. — *Salt domes and inferred salt domes in the northern Gulf of Mexico basin (after* MURRAY 1961*).*

of miles. The throw may be 100 m, or it may be 1 km or more; and the dip of the fault plane, averaging about 50° at shallow depth, decreases progressively until the fault plane parallels the bedding plane at depths greater than about 7 km.

Parallel belts of regional growth faults reflect the distribution of sediment facies, each belt marking the transition, at depth, from more competent sandy deltaic sediments to less competent clayey prodelta and marine sediments. Growth faulting results in regional reversal of the dip of beds; but its most important effect is the restriction of fluid movement, through compartmentalization of aquifer systems.

Hydrology of the deposits

If deeply-buried sediments of the Gulf Basin drained freely in response to overburden pressure, the clay-compaction rate and loss of water with deepening burial would tend to follow the curves of WELLER (1959, pp. 273-310) and DICKINSON (1953, p. 428) (Figure 5). Above depths ranging from about 2000 to 3000 m in most young sedimentary basins free drainage does occur, although at different rates in deposits of different geologic age (DALLMUS 1958, pp. 883-931); but below that depth there is generally no systematic relation between the porosity of clay (mudstone) and depth of burial (Figure 6). Changes in the clay porosity gradient with depth in the Gulf Basin are a direct function of effective stress, but changes in the porosity of sand with depth appear to be independent of effective stress (ATWATER, COWAN, CARTER, MILLER, written commun., 1966). Data for 17,367 cores of sand from south Louisiana wells show a consistent loss of porosity with depth at a rate of 4 percent for each km over a range of 6.25

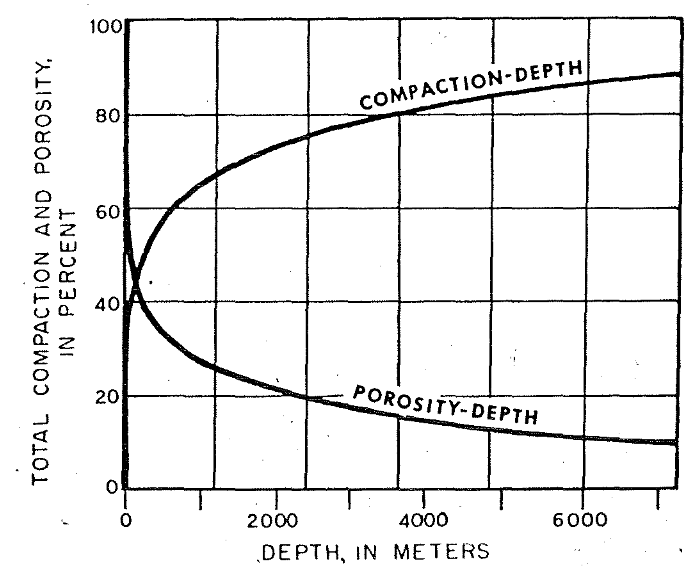

FIG. 5. — *Relation of compaction and porosity of clay to depth of burial, with free drainage (from* DICKINSON 1953*).*

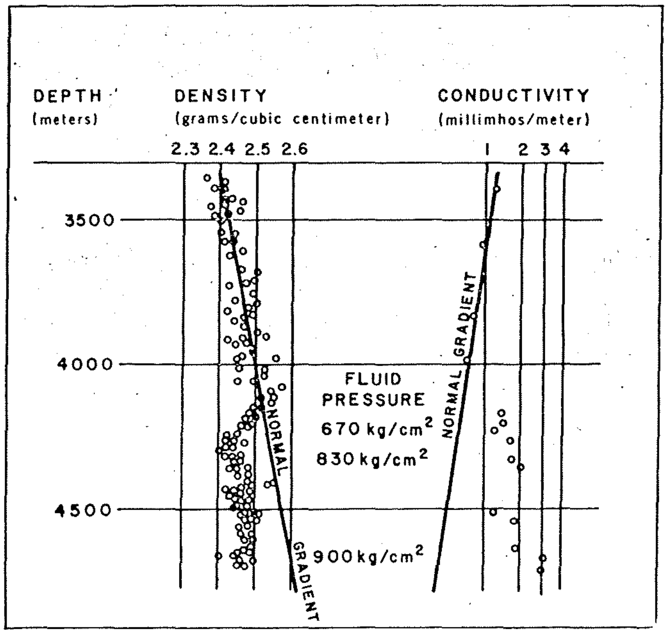

FIG. 6. — *Reversal of the clay density gradient and the electrical conductivity gradient in zones of abnormally high fluid pressure.*

km. The average porosity at a depth of 6.25 km was 14 percent. ATWATER's data agree well with those presented by MEYERHOFF ET AL. (1968, p. 530) for 6906 cores of Miocene sandstone from south Louisiana wells, at depths ranging from 1.5 to about 5.5 km.

Whether such statistical data are meaningful for an area in which the depth to the top of geopressured deposits ranges from less than 1.5 km to more than 5.5 km is questionable. Results of recent deep drilling indicate porosity of sands in the geopressured zone at depths below 6 km is considerably greater than extrapolation of ATWATER's depth-porosity curve would lead one to expect (VIDRINE 1969, p. 747).

To accomplish the observed general decrease of porosity of sediments with increasing depth of burial, there must be a continuing upward flow of escaping fluids, mainly water. Estimates of the rate of water loss from the sand fraction of the sedimentary mass can be made, using the average rate of porosity decrease with depth described above; but estimates of the rate of water loss from the clay fraction, initially comprising 50 to 70 percent or more of the deposits, cannot. According to BURST (1969, p. 80), dehydration of marine clay (shale) after deposition occurs in three stages (Figure 7). After the initial dehydration (commonly referred to as compaction), the water content is about 30 percent; after the second dehydration, about 16 percent; and after the third dehydration, about 5 percent. Remaining water is free pore water. As indicated above, the second dehydration is thermally dependent, and the depth zone at which this process occurs can be defined only if the geothermal conditions at depth are known. The fluid redistribution model for the northern Gulf of

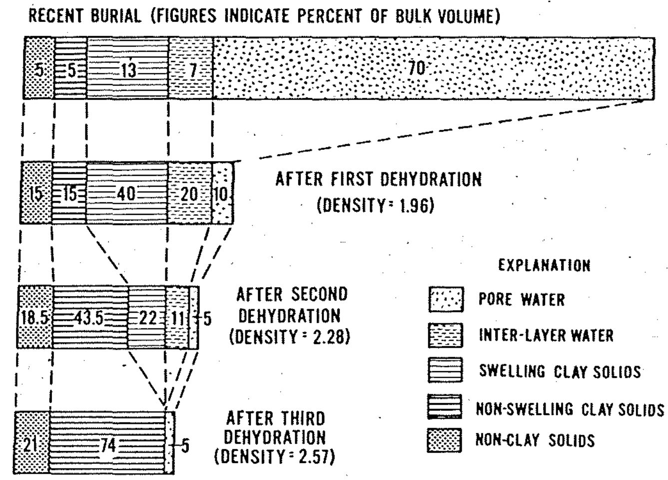

FIG. 7. — *Dehydration of marine shale with increasing depth of burial (from* BURST 1969*).*

Mexico basin (Figure 8), based on geothermal gradient data mapped by MOSES (1961), indicates that second dehydration may begin at depths as shallow as about 2 km, or as deep as about 4.5 km, depending upon the local geothermal gradient. It suggests that flushing of generated free pore water would be completed at depths ranging from 3 km to about 6 km, if drainage of the sediments were unimpeded. Actual conditions do not fit this theoretical model for two principal reasons: (1) the geothermal gradient increases with depth, and its plot is not a straight line, as shown; and (2) geopressuring of the deposits reflects fluid confinement; the system is much less dynamic than the hydropressure system that is assumed to exist.

The immediate effect of the thermal diagenesis of montmorillonite is to reduce drastically the load-bearing strength of clay beds. Generated free pore water supports part of the overburden load; the marked increase in interstitial fluid pressure in clay beds causes expulsion of water into adjacent sand bed aquifers, unless

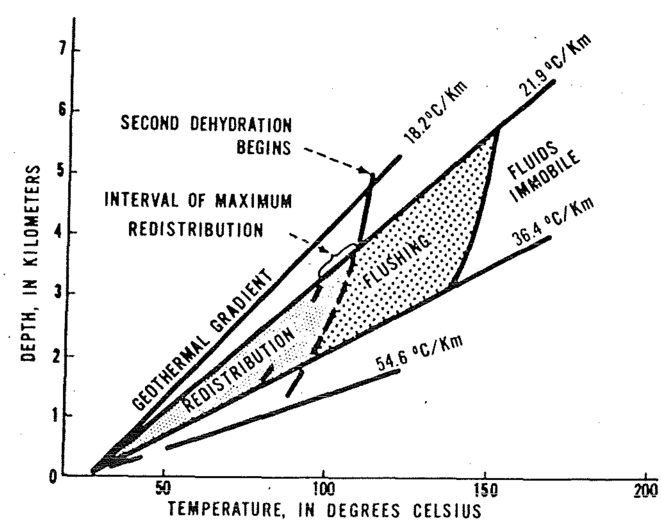

FIG. 8. — *Fluid redistribution model for the northern Gulf of Mexico basin.*

or until the fluid pressure in the aquifers is the same, as a consequence of their compartmentalization. This loss of load-bearing strength with thermal diagenesis may be the factor most responsible for the common occurrence of the top of the geopressured zone in the depth interval between the 80 °C and 120 °C isogeothermal surfaces.

No change in the density of bound and intracrystalline water in clay, and in the free pore water produced by diagenesis must, therefore, occur to produce geopressure. The loss of load-bearing strength could be entirely responsible. A marked decrease in the density of clay beds immediately below the top of the geopressured zone is always noted, however, (FOSTER, WHALEN 1966; HOTTMAN, JOHNSON 1965; WALLACE 1966) and this may be the result of hydraulic inflation of the beds as they become slushy. DICKEY (1968, fig. 2, p. 610) has shown that the top of the zone of undercompacted shale marks the top of the geopressured zone; and PENNEBAKER (1968, pp. 1-5) has demonstrated a method of mapping the top of the geopressured zone and estimating interstitial fluid pressure within it, by mapping formation density using conventional seismic data. PENNEBAKER's method is being rapidly applied to define the occurrence of abnormal fluid pressure beneath the Continental Shelf in the Gulf of Mexico.

Geopressure. The structural and stratigraphic settings of abnormally high fluid pressure in the northern Gulf of Mexico basin are described by DICKINSON (1953, pp. 410-432); and the hydrodynamics have been analyzed by the writer (JONES, 1969, pp. 803-810). The conditions of occurrence were defined by DICKINSON and applied by CHARLES STUART of the Shell Oil Company, under the term « geopressure », to « any pressure which exceeds the hydrostatic pressure of a column of water (extending from the stratum tapped by the well to the land surface) containing 80,000 milligrams per liter total solids ». A geopressure occurs where escape of fluids from a sedimentary sequence is impeded, especially by faulting, to the extent that the interstial fluid pressure reflects a part of the weight of the overlying deposits. The location and depth of occurrence of geopressured zones in the Gulf Basin are shown in Figure 9.

Water salinity. The release of bound and intracrystalline water from prodelta and marine clay during thermal diagenesis, and the upward movement of free pore water as compaction progresses with deepening burial, have resulted in a regional water salinity distribution characteristic of the Basin (JONES 1969, pp. 48-60, and appendix A, pp. 96-98). The dissolved solids range from less than 1000 to 300,000 mg/l and more; there is a general progressive freshening of formation water with depth, and with the age of the deposits (TIMM, MARICELLI 1953, pp. 403-407). The depth-salinity diagram, Figure 10, is based upon analysis of

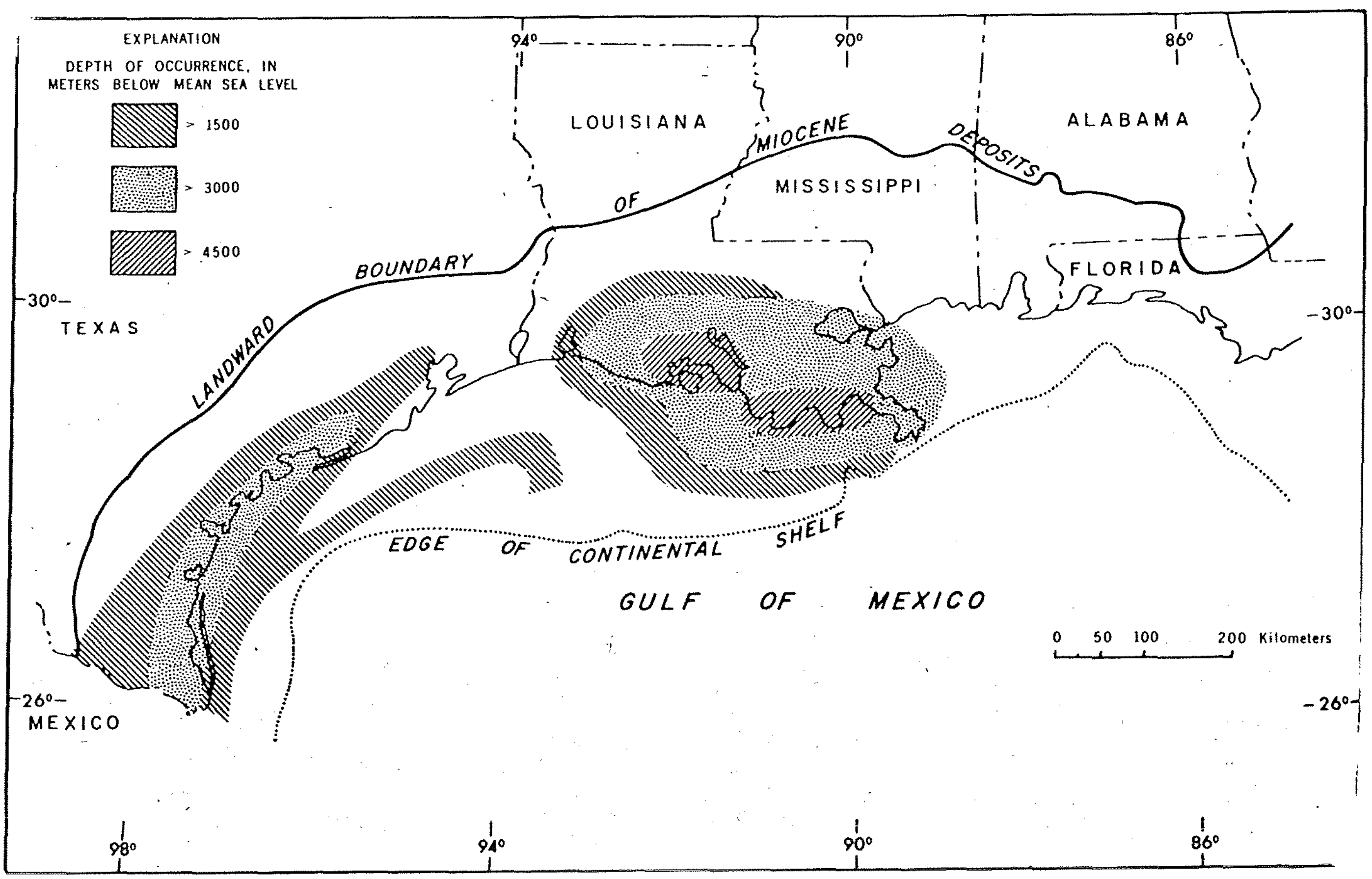

FIG. 9. — *Location and depth of occurrence of the geopressured zone in the northern Gulf of Mexico basin (from* JONES *1969).*

748 samples from wells in 19 counties in the South Texas Coastal Plain. The average salinity of water from successively greater depth intervals of about 0.3 km shows that both the Frio and the Wilcox Formations have maximum salinity at a depth of about 2.5 km, and that there is a decrease in salinity with increasing depth. Dissolved solids are as low as 5000 mg/l at depths greater than 3 km in the Wilcox Formation; and at depths greater than 4 km in the Frio Formation (JONES 1969, pp. 50-53).

Geotemperature regime

Sources of temperature data. Bottom-hole measurements by maximum-reading thermometers attached to the sondes of borehole geophysical logging devices provide information on temperature at successively greater depths in some 300,000 oil-test wells in the northern Gulf Basin. Because these measurements are generally made before thermal equilibrium has been reached, they have long been regarded with skepticism by most investigators.

Recent studies by RAMEY (1962, pp. 427-433) and SCHOEPPEL, GILARRANZ (1966, pp. 667-673) indicate that, when properly recorded, bottom-hole temperature data can be used effectively in studies of regional geothermal gradients. Observed temperatures are « likely to be within 5 percent of the formation temperature » (SCHOEPPEL, GILARRANZ 1966, p. 672). In April 1968, after a thorough review of logging procedures and methods of bottom-hole temperature correction, the Research Committee of the American Association of Petroleum Geologists (Chairman, J. M. PARKER, of Kirby Petroleum Co., Houston, Texas) initiated a geothermal survey of North America. The project was assigned to the Oklahoma State University, and arrangements were made for input of bottom-hole temperature data through the Association membership.

Studies of the geotemperature regime leading to the conclusions and interpretations presented in this paper were based mainly on bottom-hole temperature data recorded on geophysical log headings. Reservoir temperature data derived from bottom-hole tests and temperatures of produced fluids (at depth) were used to confirm log-heading data, and for analysis of the relation of temperature to interstitial fluid pressure (JONES 1969, p. 804).

Temperature distribution. The average geothermal gradient in normally-pressured sediments of the northern Gulf of Mexico basin ranges from about 20 °C/km to 40 °C/km over depth intervals of 1 to 2 km, and are rather uniform and predictable in a given locality. Data supporting this conclusion led to the preparation of the maps of NICHOLS (1947) and MOSES (1961). However, as deeper drilling became possible (with the development of effective methods for controlling abnormally high interstitial fluid pressure) it became apparent that the « normal » depth-temperature relationship no longer applied. Predicted temperatures were, in general, considerably less than observed temperatures. Data for many wells that penetrate the geopressured zone show that, at a depth of a few tens of meters below the top of geopressure, there is a sharp increase in the geothermal gradient (Figure 11).

To explore this relationship in an area for which many data are available, detailed studies were made in the Rio Grande Embayment of the South Texas Coastal Plain. Bottom-hole temperature records for about 600 wells were analyzed, and maps were prepared of the 120 °C and 150 °C isogeothermal surfaces (Figure 12). The thermal diagenesis of clay minerals is well advanced in deposits having a temperature of 120 °C, where the clay has reduced density and thermal conductivity. Records of the interstitial fluid pressure in the deposits confirm that the 120 °C isogeotherm is generally deep within the zone of geopressure, and thus in the zone of higher than « normal » geothermal gradient.

In the Rio Grande embayment area, the relief on the 120 °C isogeothermal surface is more than 1.2 km, and on the 150 °C isogeothermal surface, it is more than 1.8 km. Most areas in which the isogeotherms are at minimum depth are known as structural highs. Regional structure is reflected in the greater depth of the isogeotherms Gulfward. The top of the geopressured zone is also at greater depth Gulfward in this part of the Basin.

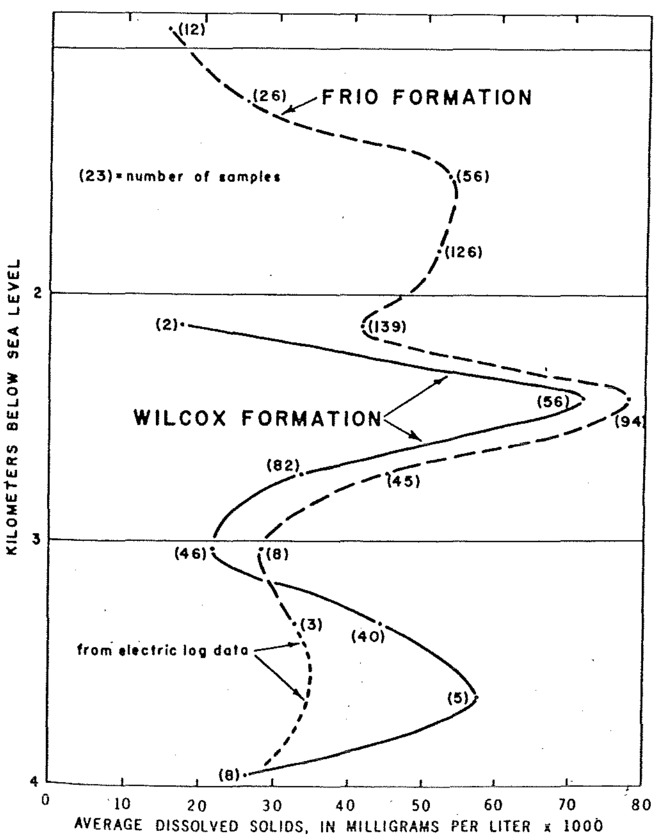

FIG. 10. — *Salinity variation with depth and geologic formation in the south Texas Coastal Plain.*

The geothermal gradient between the 120 °C and 150 °C isogeothermal surfaces (Figure 13) ranges from less than 25°C/km to more than 100°C/km. Sharply defined areas of either high or low geothermal gradients are structurally controlled. Mapped gradients are within the geopressured zone, and they bear little resemblance to the gradients shown on the maps of NICHOLS (1947) and MOSES (Figure 1). It is apparent that the distribution of temperature, areally and with depth, cannot be predicted within the geopressured zone.

Geothermal gradients in wells drilled deeply into the geopressured zone show no sign of decrease downward. The deposits may be geopressured to the zone of metamorphism.

Geotemperature and geopressure relationships. The geothermal significance of geopressure has been discussed by the writer (1969, p. 804) as follows: « That the geotemperature regime is related to the occurrence of geopressure is not surprising, because the movement of water is the most important factor in terrestrial heat flow in sedimentary basins. Checking the upward flow of water greatly reduces the rate of upward flow of heat, and geopressured reservoirs become overheated ». He further states (p. 808): « The thermal conductivity of the mineral grains of sediments is generally four to five times greater than that of water. For this reason, the thermal conductivity of clay varies inversely with its water content ».

LEWIS and ROSE (1969, pp. 1-5) recognize the relation of high temperatures to overpressures, and attribute it to the insulating effect of undercompacted sediments (Figure 6 above) in geopressured zones. They state (p. 1) that « ...wherever there is an insulating layer in the earth's crust, there is a buildup of heat beneath the insulation ». They add that « an overpressured zone does constitute a thermal barrier because it is undercompacted; the greater the water content, the greater the insulating value of the zone ». They offer analytical proof (p. 2) for their theory that high pressure zones are heat insulators; and they present field data to demonstrate that geothermal gradients below, within, and above geopressured zones do, in fact, conform with their theory.

Variations in the thickness, water content, and areal extent of undercompacted sediments alter the heat flux in accord with two independent but simultaneous effects. The « end effect » is due to boundary conditions where rocks of different thermal conductivity are in contact; refraction of heat flow lines as they pass through such boundaries is a function of the geometry of rock bodies and their relative thermal conductivity (GUYOD 1946, pp. 20-27). The « porosity effect » is due to the change in the thermal conductivity of a bed or formation as a consequence of variation in its water content. Because geotemperature conditions in nature tend to approach equilibrium, the temperature beneath an insulating bed or zone must be high enough to produce a geothermal gradient through the bed sufficient to cause heat flow through it, *at a rate equal to the universal geothermal flux.* The temperature observed at any depth in the earth, therefore, is a direct function of the thermal conductivity of the rocks between it and the land surface.

The heat of geopressured zones is held in storage by virtue of the insulating properties of their low-density clay beds, and by the low mass transport of fluid upward from the system. The maximum temperature reached in such a setting depends upon the thickness and porosity of the geopressured section, the rate of water loss from the section, and the heat requirements of diagenetic processes in the section. It also depends upon the areal extent of the geopressured section, lithologic variations within it, and time.

Results of deep drilling in the northern Gulf of Mexico basin indicate that the geopressured zone extends to very great depth, perhaps to the ultimate depth at which fluids are free to move. LEWIS and ROSE, in considering the thickness of the geopressured « insulating » zone, comment (p. 3) that « If the center of the insulating zone were ever passed, we should see the gradient break back toward lower values, but apparent-

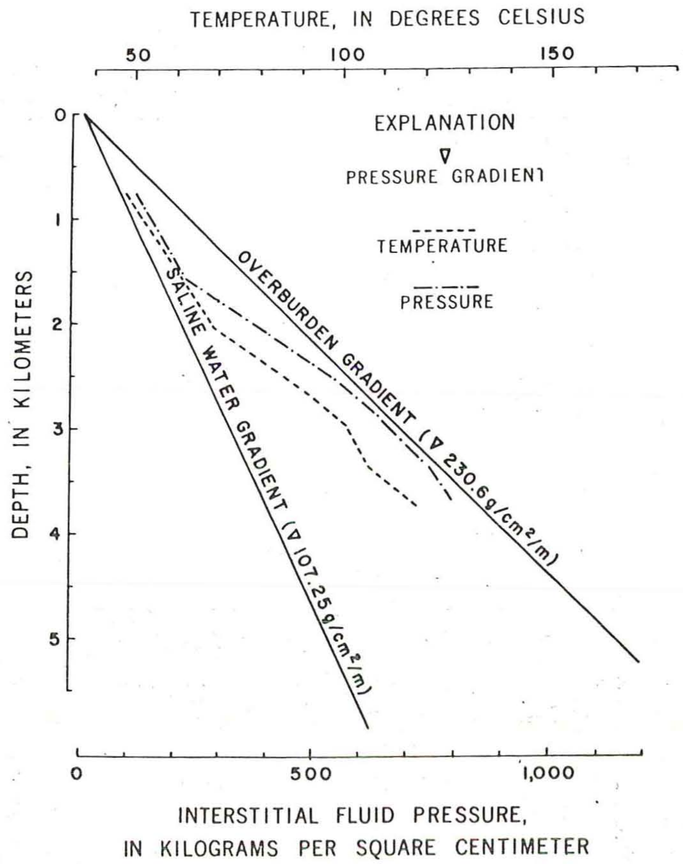

FIG. 11. — *Changes in the pressure and temperature gradients with depth, in deposits penetrated by a well in Cameron County, Texas.*

ly the center has never been reached. It is not certain that the compaction ever does return to normal in this area ».

Role of salt diapirs. If indeed, the insulating geopressured zone does extend to great depth, the importance of salt diapirs in establishing and maintaining the geothermal flux may be greater than generally believed. The known and suspected domes (Figure 4) are mostly deep seated; they usually do not penetrate the main sand series, but stop in the massive geopressured shale below. The domes range from 3 to 8 km in diameter and form an appreciable part of the depositional mass where they are abundant. Although the specific heat of halite is only 0.204 cal/g°C, the immense volume intruded into the overlying sediments from depths of 15 km or more, must have a profound effect upon the geothermal regime of the basin. Coupled with its high thermal conductivity of 17×10^{-3} cgs (compared to that of the sediments intruded, from 1 to 8×10^{-3}cgs) its mass distribution favors widespread heat transfer. The salt diapirs actually resemble a nest of heating rods thrust upwards into the basin deposits.

The maximum subsurface temperature observed in Gulf Basin deposits is 273°C at a depth of 5859 m

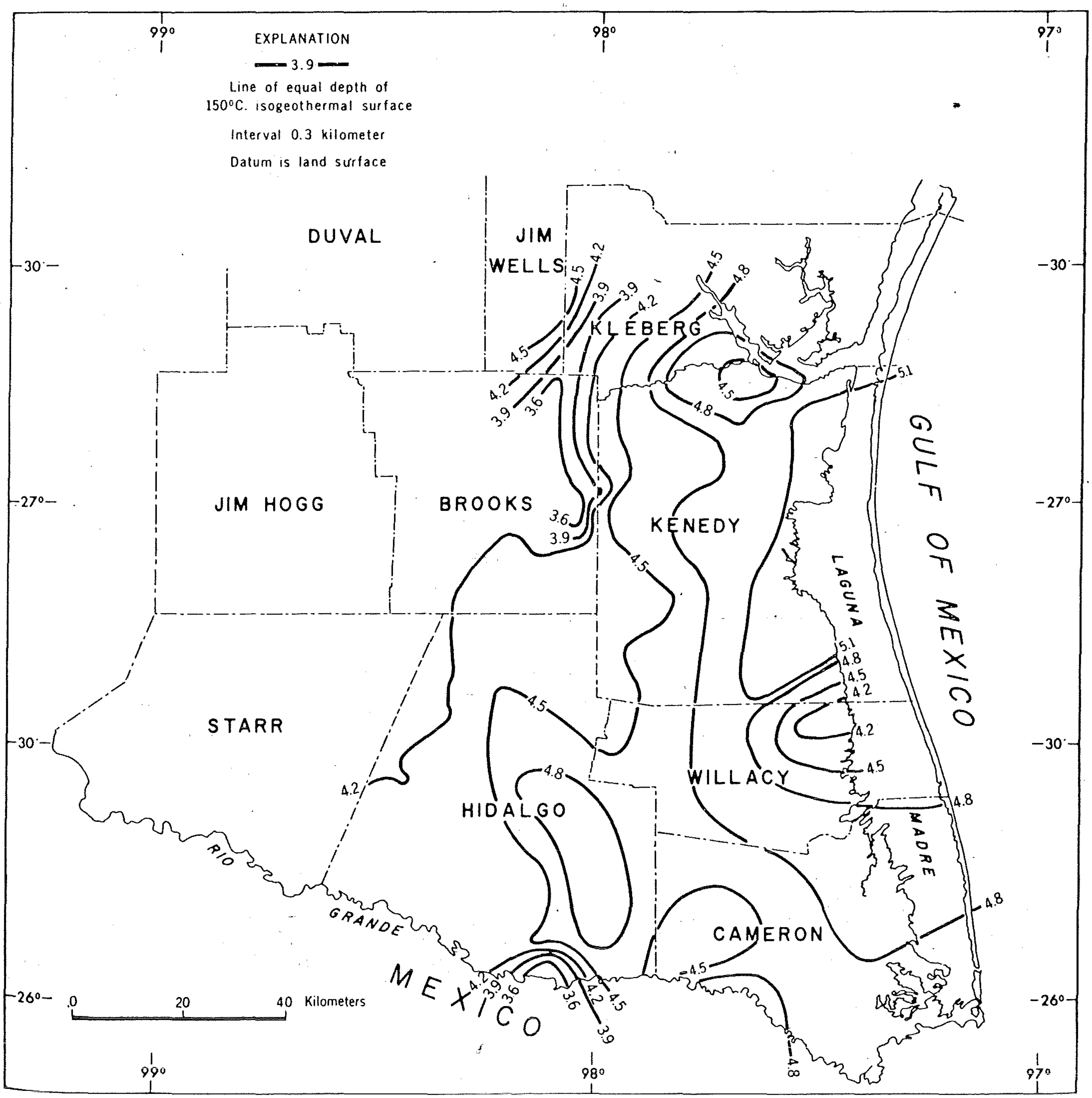

FIG. 12. — *Depth of occurrence of the 150°C isogeothermal surface in the south Texas Coastal Plain.*

in Matagorda County, Texas. This reflects an average geothermal gradient of 42.7 °C/km. At this gradient, a temperature of 300 °C may be expected at a depth of about 6.5 km; at 300 °C salt behaves as a perfect plastic. There can be little doubt that the low thermal conductivity of geopressured clayey sediments is an important factor in salt diapirism, and that salt diapirism plays an important role in the geothermal regime.

Geothermal resources development

Many thousands of wells in the northern Gulf of Mexico basin now produce fluid hydrocarbons and water at temperatures far above the boiling point of water at atmospheric pressure (Figure 14). Before being plugged and abandoned, many test wells produce superheated water or steam; many completed as gas or distillate wells gradually become superheated water producers and are plugged. Some have produced more than a million barrels of salty water hot enough to be self-distilling, while recompletion efforts were in progress. Where the water was produced from geopressured reservoirs, casing-head pressure ranged from a few hundred to 500 atmospheres or more. Reservoir pressure depletion with production was reduced by influx of

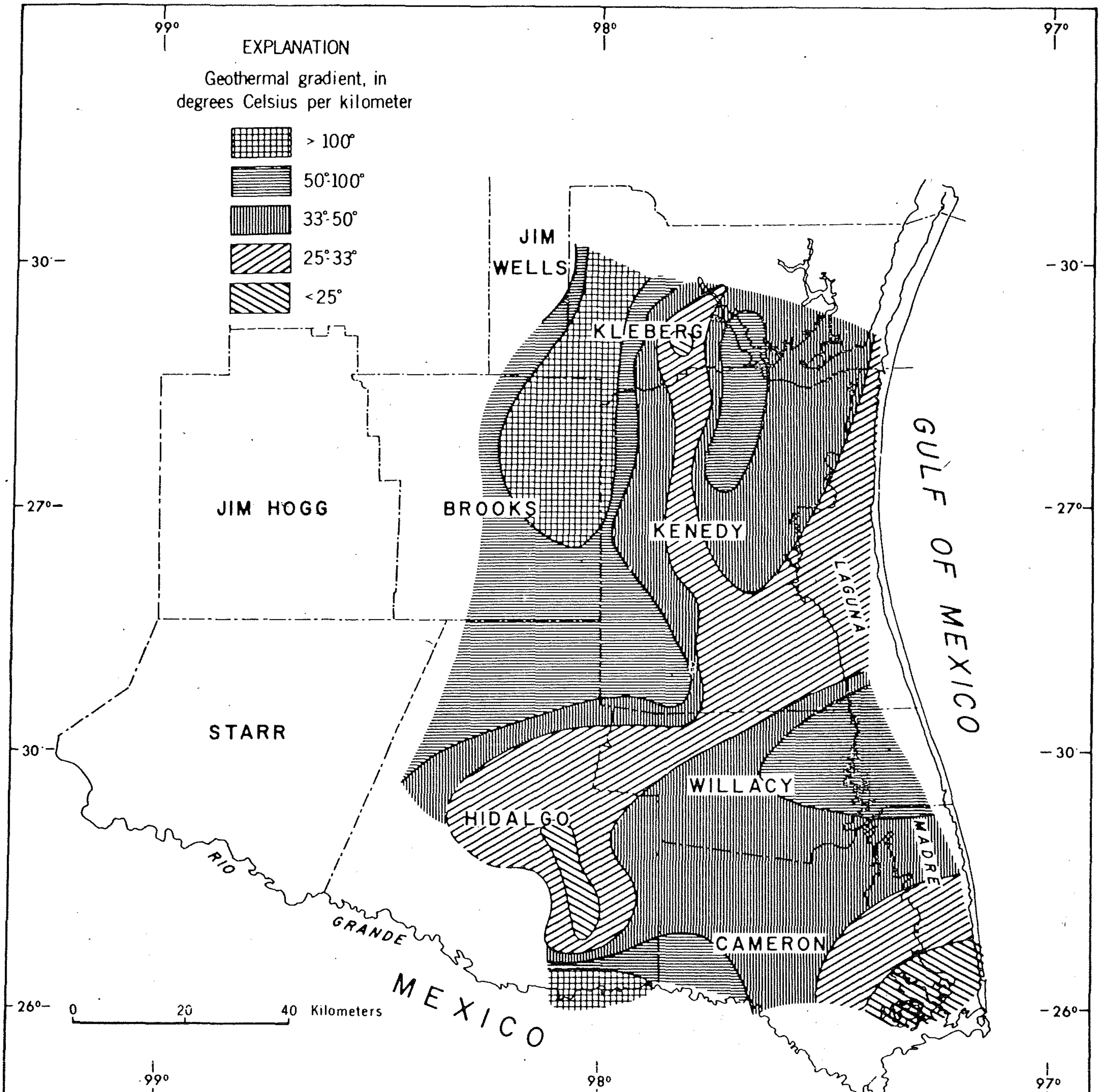

FIG. 13. — *Geothermal gradient between the 120°C and 150°C isogeothermal surfaces in the south Texas Coastal Plain.*

water, probably from undercompacted clay beds adjacent to the sandbed reservoirs (ATWATER 1967, pp. 11-17).

Careful selection of geothermal reservoirs to be developed, coupled with properly engineered well fields and production schedules, should enable appreciable development of the resource throughout the Gulf of Mexico basin wherever temperature, pressure, and water salinity are favorable. Production of geothermal fluids from these reservoirs is comparable to production of hydrocarbons: it is a depletion process. However, the volume of geothermal fluid in storage in the basin is large; depletable volumes, without adverse effects, mainly land subsidence, depend upon the size and thickness of the reservoir tapped, and upon the rate at which undercompacted sediments bounding it will yield their water. Production from well-defined reservoirs known to underlie areas where land subsidence would not pose serious problems, or from beneath the Gulf of Mexico, would be desirable.

Produced thermal waters could now be used advantageously in parts of the Basin as a source of fresh water supply, by self-distillation; as a source of steam or hot water for secondary recovery of fluid hydrocarbons, in pressure-depleted oil fields; or for electric power generation. Methods of developing the geothermal water and power resources are described in Shell Oil Company Patents Nos. 300,258,069 and 300,330,356. However, geothermal power is not likely to be attractive in the northern Gulf of Mexico basin during the immediate future, because the power generation technology is based on oil and gas, which provide a relatively inexpensive source of energy.

Conditions comparable to those in the Gulf Basin may occur in other geosynclinal basins where hydrocarbons are not produced, or where self-distilled fresh water may be an economic alternative in the face of critically short supply. Or the deep basin geothermal waters may be valuable for the chemicals they contain. Disposal of saline water concentrate will be a necessary part of any development plan.

Summary and conclusions

Geosynclinal basins that appear to be negative geothermal regions may, in fact, have appreciable geothermal resources. The northern Gulf of Mexico basin was rapidly filled with noncarbonate clastic sediments during the Cenozoic era; an appreciable part of the de-

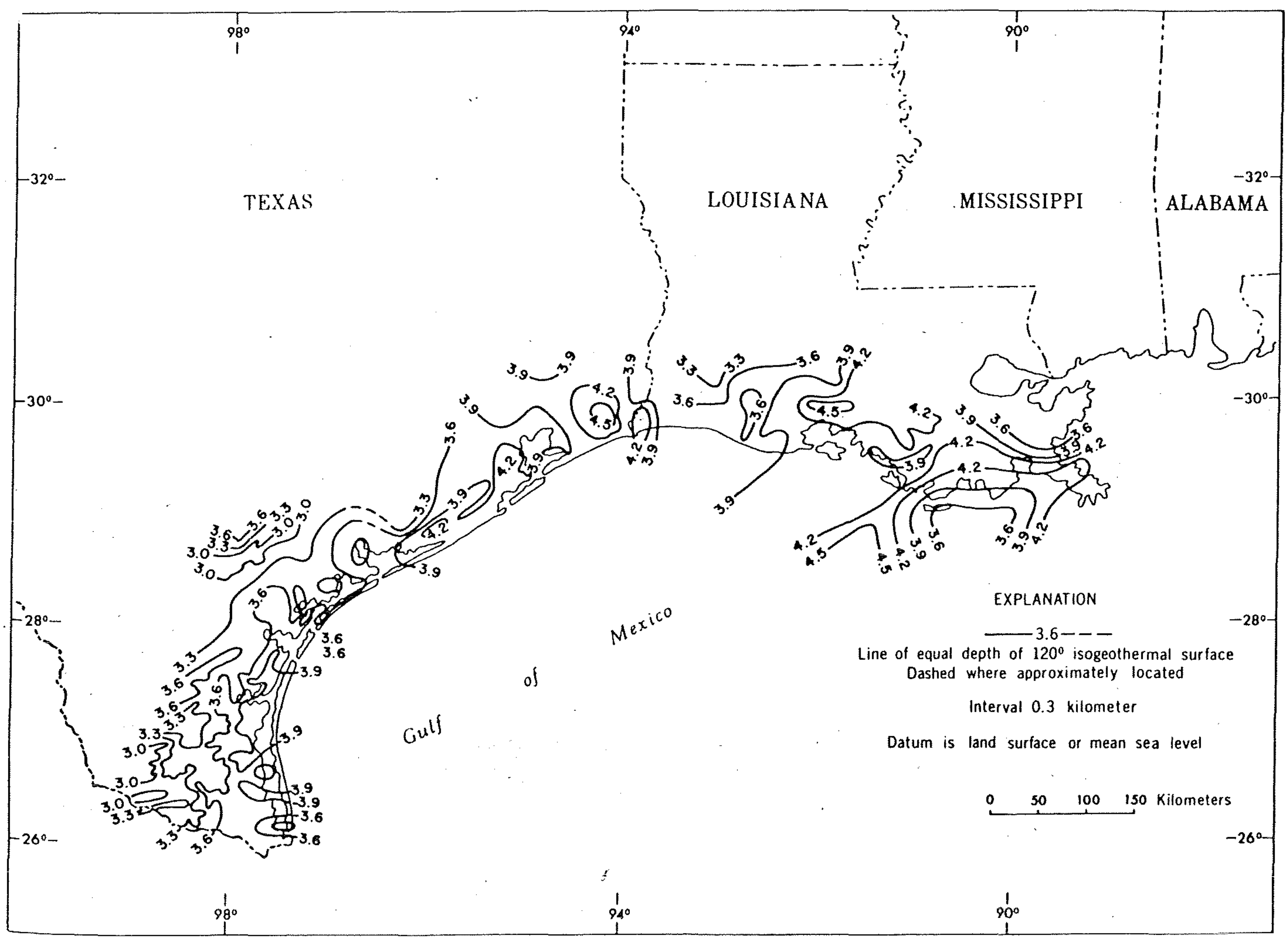

FIG. 14. — *Depth of occurrence of the 120°C isogeothermal surface in the northern Gulf of Mexico.*

posits was a swelling variety of clay susceptible to thermal diagenesis. Beneath these clastic sediments was a thick and widespread Mesozoic evaporite deposit, mainly halite. Subsidence of the geosyncline kept pace with deposition, and structural features developed that cut off and compartmentalized the sand-bed aquifers. The very thick sedimentary mass was intruded from below by hundreds of salt diapirs that probably had, at the time of intrusion, temperatures in excess of 300 °C.

Below a depth of a few kilometers the sediments were not able to drain freely; their interstitial fluid pressures became abnormally high beneath broad areas and to great depth. Coupled with thermal diagenesis of clay, during which up to 15 percent of the bulk volume of the rock became pore water, hydrodynamic sealing produced a very thick, basinwide sheet of undercompacted sediments. These « water-logged » sediments have very low thermal conductivity and form an effective insulator, restricting the upward flow of geothermal heat. Thus, a very large reservoir of residual heat is formed by the geopressured zone; geothermal gradients have been steepened sufficiently to produce the universal geothermal flux. It is this residual heat that is available for development in the northern Gulf of Mexico basin; it can be produced in the form of superheated water, as is common in geothermal energy systems.

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