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FACTORS AFFECTING PETROLEUM ACCUMULATION BENEATH THE EASTERN UNITED STATES CONTINENTAL MARGIN

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ABSTRACT

Evidence published about the geology of the continental margin of the eastern United States indicates that it is attractive for petroleum exploration. Thicknesses favorable for oil and gas generation exist beneath Georges Bank, the Baltimore Canyon area, the Blake Plateau, and the continental rise. Mesozoic clastics and carbonates to the north and carbonates to the south are favorable reservoir and source rocks. Salt domes, basement highs, faults, and other structures could provide suitable traps. Continental fragmentation allowed evaporite formation and accumulation of organic matter in basins during Mesozoic subsidence. Reefs and deltas also formed during this time interval.

INTRODUCTION

One of the largest areas on earth with geological conditions favorable for petroleum accumulation but not yet tested by drilling is the continental margin of the eastern United States. In contrast to the situation in the United States eastern offshore, drilling has been proceeding in offshore Canadian waters since 1966, and a few wells have been drilled in the Bahamas. Almost all the continental shelf of southeastern Canada has been under exploration permit since 1965, and through 1974 at least 85 wells had been drilled. Areas are now under permit there up to 650 km offshore.

Several factors in addition to size make the eastern United States continental margin attractive for oil exploration: it is underlain by a large volume of marine sedimentary rocks; its Mesozoic history is similar to that of the Gulf of Mexico northern continental margin, one of the major oil producing provinces of the world; petroleum has already been found on the eastern Canadian continental margin, an area which is a northern extension of the region under discussion; and nearby industrial centers simplify logistics.

Acknowledgments

I thank Van Price and David Snipes for helpful comments and suggestions.

TOPOGRAPHY

The northern part of the continental margin, from North Carolina to Maine, comprises the continental shelf, the continental slope, and the continental rise (Figure 1). The shelf has an average slope of less than 1° toward the ocean and is up to 420 km wide off Maine. At the shelf edge, at depths from about 80 to 160 m, the inclination of the sea floor increases to about 7°, forming the continental slope. The depth of the outer edge of the slope varies from about 1400 m off Nova Scotia to about 2500 m off North Carolina (Figure 2). The continental rise forms a ramp from the base of the slope out to the abyssal plains, with an inclination of less than 1°. The rise is up to 500 km wide, and its outer edge reaches depths of more than 5000 m (Emery, 1966; Maher and Applin, 1971; Emery and Uchupi, 1972).

The southern part of the continental margin is more complex and consists of the continental shelf, the Florida-Hatteras Slope, the Blake Plateau, the Blake Escarpment, the Blake-Bahamas Abyssal Plain, and the Blake Outer Ridge. The Blake Plateau has depths from about 500 to 1100 m and averages about 900 m. The Blake Escarpment attains depths of 5000 m. The entire continental margin is about 2500 km long off the eastern United States and covers a total area of about 1,570,000 km^2. That part of the margin shallow enough to be accessible to the present operating procedures of the oil industry, the shelf, occupies about 420,000 km^2, and the Blake Plateau underlies about 180,000 km^2 (Emery, 1966; Maher and Applin, 1971; Emery and Uchupi, 1972). The target is obviously a large one.

SEDIMENT THICKNESS

The North Atlantic Ocean basin appears to have begun developing during the latter part of the Triassic Period and the early part of the Jurassic, approximately 190 million years ago. Since that time sediments have accumulated on the subsiding edge of the continent as North America separated from Africa and Europe. It is possible that pre-Jurassic rocks referred to as "basement", contain oil, but this discussion will be limited to the younger rocks.

Estimates of post-Triassic sediment thickness in several areas of the continental margin are encouraging: 8 km in the Georges Bank area (Schultz and Grover, 1974), 12 km in the Baltimore Canyon area (Mattick and others, 1974), and 6 km in the Blake Plateau area (Sheridan

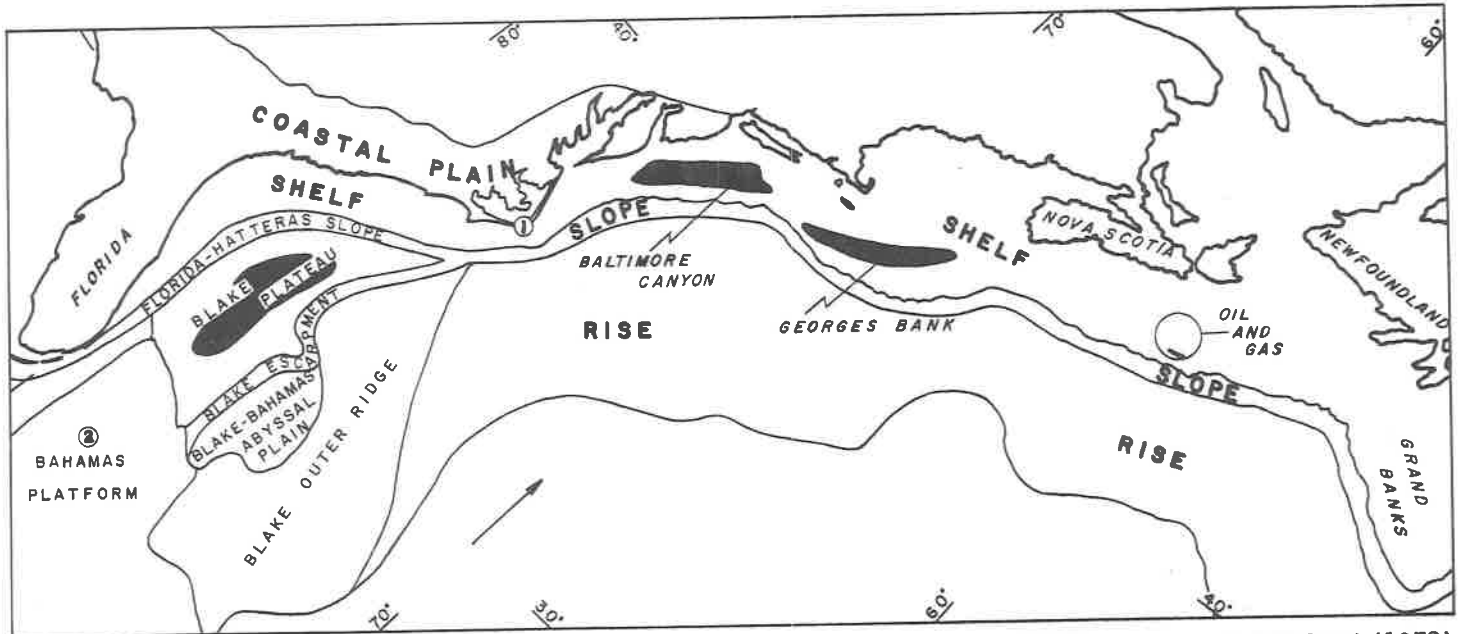

Figure 1. Physiographic units of the continental margin, after Emery (1966), Emery and Uchupi (1972), and Maher and Applin (1971). Dark areas are underlain by unusually thick sedimentary sequences, after Mayer and Applin (1971). "1" marks location of the Hatteras well, and "2" is location of Bahamas well mentioned in text.

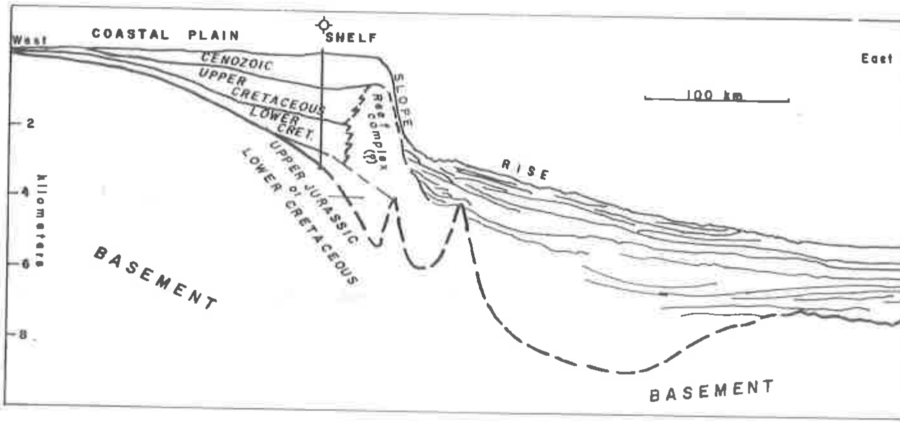

Figure 2. East-west cross-section through the Hatteras well, after Emery and Uchupi (1972). Vertical exaggeration is 25X.

and others, 1966) (Figure 1). These areas are underlain by basins, possibly fault controlled, where unusual thicknesses of sediments have accumulated because of great subsidence. Sedimentary thicknesses beneath the slope and rise are probably more than 10 km in places (Drake and others, 1968). The total volume of Mesozoic and Cenozoic sediments derived from the continent in the area of the continental margin has been estimated as 10 million km^3, almost as much as the 12 million km^3 estimated for the northern Gulf of Mexico province (Gilluly and others, 1970).

SEDIMENTARY ROCKS

The composition and texture of the sediments beneath the continental margin cannot be known with certainty until drilling begins, but there is enough information available from the Atlantic Coastal Plain, offshore Canada, the Bahamas, and geophysical data from the margin to make some fairly confident statements about the rocks offshore. The rocks of the Coastal Plain may be observed in outcrop, and many wildcat wells and water wells provide information about subsurface stratigraphy. The strata of the northern Coastal Plain are mostly sandstones and shales, generally dipping at low angles toward the ocean and extending beneath the shelf. South of North Carolina carbonates are common, the proportion of limestone gradually increasing southward to the mostly carbonate platform of Florida.

The wells on the shelf off Nova Scotia (Figure 1) penetrated Mesozoic and Cenozoic rocks more than 4600 m thick, encountering salt overlying redbeds at the bottoms of some wells. The salt is thought to be of Jurassic age. One well penetrated more than 760 m of salt (McIver, 1972), and evaporites are thought to be much thicker in the subsurface basins. Jurassic carbonates and evaporites overlie the salt, and are in turn overlain by Jurassic sandstones, shales, and limestones. Cretaceous sandstones and shales of a regressive sequence lie above the Jurassic rocks. Cenozoic shales and sands complete the section. The entire sequence appears to grade into coarser clastics updip and become almost entirely shale toward the ocean basin (McIver, 1972). About 40 percent of the section is Jurassic, about 50 percent Cretaceous, and about 10 percent Cenozoic.

Oil and gas occur in Cretaceous and Jurassic strata in Nova Scotia offshore wells, and good porosities and permeabilities are indicated. One well tested at the rate of 36 million ft^3 of gas per day from three zones and 300 barrels of oil per day from a fourth (Austin, 1973). Tests between 700 m and 1900 m in another well yielded hydrocarbons from 17 to 18 zones (Howie and Hill, 1972). Five wells were good enough to be considered new field discoveries (Crosby, 1974; Bryant and others, 1975). Shows have also been found in wells on the Grand Banks shelf.

A well drilled on Cape Hatteras, North Carolina (Figure 1) provides the nearest thing to a look at continental shelf strata in the central part of the continental margin. The well penetrated about 870 m of Cenozoic sediments, mostly sandstones and carbonates, overlying about 1920 m of Cretaceous sediments (with some possible Jurassic in the lower part) consisting of shales and sandstones with some carbonate beds. Below the Cretaceous is a section of about 215 m of probably Jurassic sands and shales (Brown and others, 1972). Porosities of up to 41 percent and permeabilities of more than 2100 millidarcys were recorded in the Cretaceous rocks (Spangler, 1950). Permeabilities on the order of 10,000 millidarcys, calculated from hydrologic transmissivity tests, have been reported from water wells in the Cretaceous of Delaware (Kraft and others, 1971).

It is likely that shelf strata are more carbonate rich than equivalent Coastal Plain deposits, because clastic source areas are farther from the sites of deposition. South of North Carolina the Coastal Plain sediments are more carbonate rich, and the Blake Plateau is probably almost entirely carbonate. A well drilled on the island of Andros in the Bahamas (Figure 1) penetrated about 2670 m of Cenozoic carbonates and bottomed in Cretaceous carbonates at 4445 m (Spencer, 1967). High porosity, in some cases cavernous, was encountered in the Cretaceous section. Porous and permeable carbonates are excellent reservoir rocks, the greatest oil producing region on earth, the Persian Gulf area, being largely a carbonate region. Fine-grained limestones and black shales, which are good petroleum source rocks, are known to be present along the southern continental margin (Lancelot and others, 1972).

Cretaceous reefs are apparently present along the outer edge of the Blake Escarpment (Heezen and Sheridan, 1966; Ewing and others, 1966; Meyerhoff and Hatten, 1974) and could be reservoirs. They could also have provided appropriate environments for the accumulation of organic matter in the back-reef environment by restricting circulation. Reefs of about the same age are highly productive in Mexico. The possibility of clastic reservoirs in the area also exists (Meyerhoff and Hatten, 1974). Wells in the south Florida area have penetrated bedded anhydrite and shales in the Jurassic and Cretaceous section (Meyerhoff and Hatten, 1974) which could serve as reservoir seals. Several small fields produce from Cretaceous limestone in southern Florida, and shows have been reported from other parts of the southern carbonate area. Oily odors and asphaltic specks have been detected in Tertiary samples from the shelf off northern Florida (Emery and Zarudski, 1967).

The thick sedimentary section beneath the continental rise apparently consists of seaward-thinning beds of fine-grained pelagic sediments interbedded with relatively coarse-grained sands deposited by turbidity currents and slumping (Emery and others, 1970). The turbidites might serve as reservoirs, and the lutites should be good source beds. The environment of the slope is particularly favorable for organic accumulation (Rogers and others, 1973), and turbidity currents could bury the material on the rise before oxidation occurs. Harrington (1969) has suggested that the presence of thick turbidite sequences beneath the rise and abyssal plains implies that large quantities of hydrocarbons have been generated along the continental margin. Methane and ethane have been detected in Tertiary sediments on the rise (Lancelot and Ewing, 1972). Although the slope lies at depths beyond the normal range of petroleum operations, it could become an important region when deep water drilling techniques are more extensively used.

STRUCTURES

The Atlantic margin has not been considered in the past to be a province of structural complexity. The prevailing concept, that of a homocline with dip toward the ocean, is generally true but now has to be modified because of subsurface information which has become available within the last few years. Recent studies by Brown and others (1972), Gibson (1970), and Sheridan (1974) have presented geologic interpretations emphasizing the structural complexity of the region.

Basement Highs

Traps associated with basement highs include anticlines formed by basement movement and by differential compaction of the overlying sediments, and terminations of porous sand bodies against the highs. The rifting which apparently affected much of the continental margin during the early Mesozoic (and during the Carboniferous along the northern continental margin) would tend to produce a horst and graben pattern with traps developing above the upthrown blocks. Pre-Mesozoic erosion of the basement surface could also have formed highs.

Basement highs and anticlines possibly associated with basement irregularity have been reported from several areas in the coastal plain (Brown and others, 1972; Dietrich, 1960), including Delaware (Kraft and others, 1971), Virginia (Sabet, 1973), South Carolina (Colquhoun and Comer, 1973), and Georgia (Cramer, 1969). On the shelf, basement irregularities associated with fault basins have been found off Nova Scotia and Newfoundland (Figure 3), and in the Gulf of Maine and near Georges Bank area (Ballard and Uchupi, 1972). Highs have also been found on the shelf off New England (Garrison, 1970) and in the Baltimore Canyon area (Drake and others, 1959; Minard and others, 1974). Gentle, large amplitude folds, possibly related to basement structure, also occur on the shelf and in the Blake Plateau off the Carolinas, Georgia, and Florida (Sheridan and others, 1966; Bunce and others, 1965; Olson, 1974; Uchupi, 1970). Extensions beneath the shelf of large positive structures such as the Cape Fear arch and the Norfolk high could also provide traps.

A series of arches and ridges near the edge of the shelf and rising to within 1500 m of the sea floor has been inferred by geophysicists (Figure 2). It extends from Nova Scotia to the Baltimore Canyon area, and perhaps connects with a ridge system underlying the outer edge of the Blake Plateau. Another ridge system appears to underlie the rise from the Grand Banks to the Blake Plateau (Emery and others, 1970; Drake and others, 1959; Mayhew, 1974). The origin of the ridges is unknown, but they may be fault controlled, produced by rifting when the Atlantic opened in early Mesozoic time. They could also be reef structures similar to the Blake Escarpment reefs or a combination of reefs and basement highs. Diapirs have also been suggested as a possibility. Whatever their origin, the structures are favorable for petroleum accumulation. The ridges could also have inhibited oceanic circulation, thus allowing organic material to be preserved and retained for petroleum generation.

Although not properly falling into the category of basement highs, igneous intrusions into the continental margin sediments, such as suggested by Mattick and others (1974) and Sheridan (1975) in the Baltimore Canyon area and by Ballard and Uchupi (1972) in the New England shelf, might possibly form anticlinal traps. Fractures and weathered zones in the basement rocks might make the basement itself a potential reservoir rock.

Salt Domes

Salt domes of probably Jurassic age have been drilled off Nova Scotia and on the Grand Banks (Figure 3). Seismic studies have detected diapiric structures between the Grand Banks and the Scotian shelf

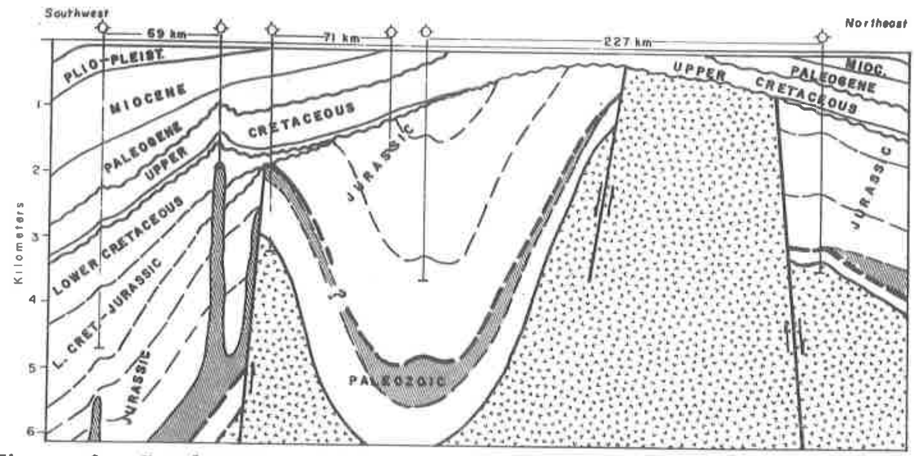

Figure 3. Southwest-northeast cross-section through the Grand Banks, after Upshaw and others (1974), showing basement highs, faults, salt domes, and unconformities. The salt (shaded) in the central syncline is thought to be Paleozoic. Other salt bodies are considered Jurassic. Vertical exaggeration is 38X.

(Keen, 1970) and on the rise and abyssal plain southeast of the Grand Banks (Watson and Johnson, 1970). Diapiric structures possibly formed by salt movement have been detected seismically in the Bahamas (Ball and others, 1968; Lidz, 1973), and a possible diapir has been found near the shelf edge off Delaware (Sheridan, 1975). Bedded salt has been drilled in several places in Florida.

On the other side of the Atlantic, opposite the United States continental margin, features detected seismically have been interpreted as salt diapirs at several places offshore from Spain and down the western bulge of Africa at least as far south as Cape Verde (Pautot and others, 1970). This suggests that the Atlantic Ocean with its restricted circulation of early Mesozoic time might have been bounded by a series of evaporite basins, and it is possible that the gap in the distribution of known evaporites (Figure 4) along the western border of the Atlantic between Florida and Nova Scotia will be filled when drilling begins off the United States (Gibson, 1970; Rona, 1970). Anhydrite fragments were found in cuttings from the Hatteras well, and wells on the Coastal Plain in the Chesapeake Bay area, North Carolina, Georgia, and Florida have chloride concentrations greater than 100,000 ppm in their deeper sections (Manheim and Horn, 1968).

The Triassic fault troughs of the Newark type, exposed on land from South Carolina to Nova Scotia and known from several places in the subsurface of the Coastal Plain, also exist beneath the shelf in the northern parts of the margin and probably extend the entire length of the shelf (Ballard and Uchupi, 1972; Olson, 1974). These basins would

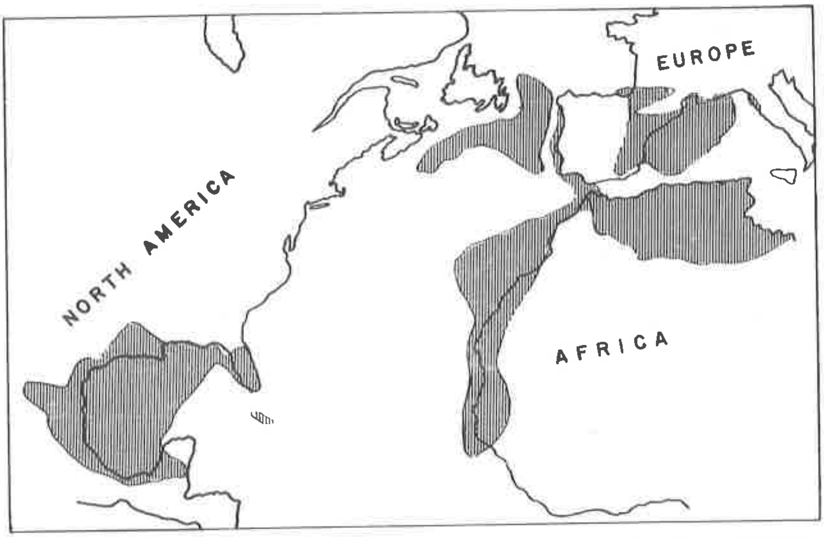

Figure 4. Position of continents around the North Atlantic during the Jurassic, with areas underlain by evaporites shaded, after Schneider and Johnson (1970).

form favorable areas for restricted circulation and so become excellent places for evaporite deposition under the proper climatic conditions.

Faults

Faults have been found in the Coastal Plain, and geophysical data suggest that faulting could be extensive offshore. Evidence of faulting in the Coastal Plain (Brown and others, 1972) has been reported from Delaware (Jordan and others, 1974), Maryland (Jacobeen, 1974), Virginia (Sabet, 1973), South Carolina (Colquhoun and Comer, 1973; Tarr, 1974), Georgia (Cramer, 1969), and Florida (Pirkle, 1971). Lidz (1973) has described normal faults in the Bahamas.

In the Grand Banks area, well and seismic control indicate that normal faults cut both the basement and the Mesozoic section (Amoco and Imperial, 1974) (Figure 3). Faulting has been reported from the New England shelf (McMaster, 1971). Large-scale faulting has also been inferred for the Blake Plateau area (Sheridan and others, 1969), and slickensides have been found on the Blake Escarpment (Sheridan and others, 1971).

The presence of a major structure discontinuity along the shelf south of New England, the 40th parallel anomaly, has been suggested (Drake and others, 1968). The New England seamount chain to the east is thought to be an extension of this zone, as is the bend in the Appalachian system to the west. Recent hypothesis of extensive strike-slip deformation in the continental margin area has suggested other possible areas of wrench faulting (Sheridan, 1974; Ballard and Uchupi, 1972), and these areas offer possibilities for petroleum accumulation of the kinds described by Hardin (1974) and Moody (1973).

Seismic profiles have revealed slump structures at the base of the slope, apparently formed by material sliding off the slope down onto the rise (Emery and others, 1970; Hoskins, 1967). These might serve as hydrocarbon traps, especially those slumps which have an anticlinal form.

Unconformities

Unconformities have long been recognized in the Coastal Plain. At least some of these might be expected to diminish toward the ocean basin, but shelf unconformities have been identified. In the Grand Banks, a distinct angular unconformity separates homoclinal Cretaceous and Tertiary rocks from underlying deformed strata of Jurassic age (Figure 3). Other unconformities within the Cretaceous, at the Cretaceous-Tertiary boundary, and within the Tertiary have been recognized here, but these are more disconformable than angular (Upshaw and others, 1974). Seismic and well data from the Scotian shelf have been interpreted as indicating angular unconformities between Jurassic and Cretaceous strata, within the upper Cretaceous section, and in the Cenozoic sediments (King and others, 1974).

Well data and seismic profiles farther south have revealed angular unconformities near the Cretaceous-Tertiary contact in the Norfolk area (Minard and others, 1974) and in the Tertiary section near Charleston, South Carolina (Colquhoun and Comer, 1973). Geophysical control and shallow cores have revealed numerous unconformities with the Tertiary section of the shelf (Meisburger and Duane, 1969; Uchupi and Emery, 1967).

Lateral Permeability Changes

Pinchouts appear to be common in the Coastal Plain subsurface and are likely to be associated with basement highs and salt domes. Channel sands and barrier bar sands should be present in those parts of the sedimentary section which formed in shallow environments (Maher and Applin, 1971). Pinchouts of turbidite bodies at the base of the shelf are also prbbably present.

Cretaceous reefs occur in the Bahamas and along the Blake Escarpment. They may also extend farther north beneath the edge of the shelf (Mattick and others, 1974; Kraft and others, 1971). Zones of dolomitization with high porosity might also form traps in the carbonates.

COASTAL PLAIN

Many wildcat wells have been drilled in the Atlantic Coastal Plain and all have been dry holes. Some shows have been found, however, in several areas, including New Jersey, North Carolina, and Georgia (Maher and Applin, 1971). Discouraging results in the Coastal Plain are probably caused largely by the relatively thin sedimentary section which averages about 600 m. The thinness is unfavorable because source beds must be buried deeply enough to encounter temperatures and pressures sufficient for generation and migration of hydrocarbons. LaPlante (1974) found that depths required for significant hydrocarbon generation in rocks of late Tertiary age in the Louisiana Gulf coastal plain are 3000 to 4000 m. It is possible, of course, that migration has brought petroleum into the shallower sediments in some areas.

GEOLOGIC HISTORY

Petroleum accumulation is influenced not only by thickness, rock types, and structures, but also by the sequence of geologic events which occurred during the development of the geologic province and the environment in which these events occurred. Continental margins appear to be especially favorable for hydrocarbon accumulation (Hedberg, 1970; Pratt, 1947). Organic productivity is high because of abundant nutrients being brought from the continents, and the relatively shallow water on the shelf allows nutrients to be recycled from the sea floor. The nearby landmass can also provide a large supply of clastics to bury the organic material, and unstable margins allow thick accumulations of sediments to build up in areas of rapid subsidence. Instability of a margin can also generate structures such as faults, unconformities, and slump structures. Shelf-edge ridges, conducive to preservation of organic material and evaporite formation, and to development of traps, seem to be common on margins (Hedberg, 1970).

The history of the east coast continental margin begins with the formation of the Atlantic Ocean in early Mesozoic time. Tensional stresses associated with the widening of the Atlantic caused a horst and graben system to form in the basement. Triassic redbeds and Jurasic evaporites, carbonates, and clastics were deposited in the basins between the upthrown blocks. This sequence of events provided basement highs, evaporites (found in most of the productive areas of the world) for salt dome formation, source rocks deposited under reducing conditions, and reservoir rocks.

In late Jurassic time, normal marine conditions began to prevail over most of the area, and carbonates and clastics succeeded the evaporites in the basinal area. A long, narrow seaway apparently extended from the Gulf of Mexico, along the eastern United States continental margin, and into the Persian Gulf area, forming a continuous marine area with similar environmental conditions, and bounded by North America to the northwest, Europe to the north, Asia to the northeast, and Africa to the south (Irving and others, 1974). The probability of high heat flow at the continental margins during the Mesozoic, due to the nearness of the Mid-Atlantic ridge, is favorable, high heat flow being an aid to migration (Tarling, 1973).

The sequence of Triassic redbeds, overlain by Jurassic salt, in turn overlain by shallow water carbonates, found beneath the Scotian shelf, if very similar to that of the Gulf Coast province. It is also interesting that in the Persian Gulf area, Triassic redbeds underlie the prolific Jurassic carbonate reservoirs, many of which occur above basement uplifts. There are also similarities between the Atlantic margin and the Tampico province in the Gulf of Mexico, production there being from reservoirs above uplifted basement blocks and platforms (Olson, 1974).

Thick sediments accumulated on the Atlantic margin during Mesozoic time. Timing of subsidence makes the area attractive, because more than 70 percent of the world's reserves occur in rocks of late Mesozoic age (Irving and others, 1974). It is also intriguing that the time of maximum rate of subsidence, the Cretaceous, was also a time of widespread reducing conditions in the western North Atlantic (Ewing and Hollister, 1972). Streams flowing down from the Appalachian area to the west built up thick sequences of deltaic clastics during Cretaceous time in the north. The oil-bearing Jurassic and Cretaceous section beneath the Scotian shelf contains deltaic deposits (McIver, 1973). A major Cretaceous deltaic sequence probably extends all the way across the shelf in the Baltimore Canyon area and may attain a thickness of 5000 m, and Miocene deltaic deposits also appear to be present there (Minard and others, 1974). Deltaic sediments have been reported in the Cretaceous and Tertiary of the New England shelf (Garrison, 1970; Minard and others, 1974).

During the Cretaceous, reefs grew around the perimeter of the Gulf of Mexico and are prolific oil producers in Mexico. The Cretaceous reefs in the Bahamas, the Blake Plateau, and perhaps farther north along the shelf edge, may be an extension of this trend. During Mesozoic time, the Atlantic margin was in tropical latitudes where most of the world's oil has been generated. The geographic situation with respect to latitude and ocean currents was remarkably similar, apparently, to that of the Gulf of Mexico (Irving and others, 1974). The petroleum generated in the area has had a good chance of being preserved by being on the trailing edge of a continent rather than on a plate margin.

Yarborough (1971) listed four environments most favorable for hydrocarbon accumulation: deltaic, rapidly subsiding carbonate shelf or lagoon, reef complex associated with a carbonate shelf, and turbidite environments. All four occur on the east coast continental margin, the first in the northern part, and the second and third in the southern part and perhaps in the deeper part of the section in the north, and the fourth on the continental rise.

Of the factors emphasized by Hedberg (1964) as being important in petroleum accumulation, the Atlantic continental margin has marine sediments of favorable age (Mesozoic), suitable thickness, a history of subsidence, with a barrier to open circulation seaward during part of its history, high organic productivity, evaporites, a nearby source of clastics, unconformities, deltas, reducing environments, fine-grained sediments, and reservoir bodies.

Estimating petroleum resources in undrilled areas is a hazardous business. The United States Geological Survey has recently estimated that the outer continental shelf, in depths up to 200 m, contains 2 to 4 billion undiscovered recoverable barrels of oil and 5 to 14 trillion ft^3 of gas at 75 percent and 25 percent probability levels (Miller and others, 1975). The Atlantic margin probably will not be as productive as the Gulf Coast, but could be a major producing area. Factors which may prevent the Atlantic margin from being as productive as the Gulf margin are:

1) The Atlantic lacks a thick Tertiary section. Most of the subsidence here was during the Mesozoic, and most of the subsidence in the northern Gulf margin occurred during the Cenozoic (Gibson, 1970). Tertiary reservoirs are particularly good because the sediments have not had their porosities and permeabilities reduced by compaction and cementation as much as older rocks.

2) The smoothness of the outer shelf and slope as revealed by seismic profiles suggests that salt domes are not as numerous as in the Gulf Coast.

3) There may be a lack of sealing evaporites in some areas of the southern carbonate province (Meyerhoff and Hatten, 1974).

4) The smaller drainage area per foot of shoreline along the Atlantic has probably caused the numbers and sizes of deltas to be small as compared to the Gulf of Mexico province.

CONCLUSIONS

Geophysical data from the continental margin indicate that thicknesses favorable for petroleum accumulation exist beneath Georges Bank, the Baltimore Canyon area, the Blake Plateau, and the continental rise. Jurassic and Cretaceous clastics and carbonates to the north and carbonates in the south, as indicated by wells on the Grand Banks and the Scotian shelf, the Atlantic Coastal Plain, and in the Bahamas, are favorable reservoir and source rocks. Structures for trapping petroleum are present and include salt domes, traps associated with basement highs, normal faults, wrench faults, unconformities, pinch-outs, reefs, and slump structures.

The geologic history of the margin makes the area attractive.

Continental fragmentation provided basins for evaporites and organic accumulation and caused basement highs to develop. Maximum subsidence occurred during the Mesozoic under conditions suitable for petroleum generation. Reef growth and deltaic sedimentation also occurred during this time.

The east coast continental margin will probably not be as prolific as that of the Gulf of Mexico margin, but known geologic factors make it an attractive area for exploration.

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POST-CRETACEOUS FAULTING IN THE HEAD OF THE MISSISSIPPI EMBAYMENT

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ABSTRACT

Mapping to date has shown less than 25 faults that offset Cretaceous and younger strata in the northern part of the Mississippi Embayment (north of 36° latitude). However, in this seismically active area other methods in addition to strict field data can be used to delineate possible faults. In this study about 70 hypothetical faults were interpreted from topographic lineations and well data. This number may be even larger if known faults that cut the Cretaceous along the northern perimeter of the Embayment persist southward.

A pattern of dominant northeast-trending and subordinate north- and northwest-trending faults is presented. The validity of this pattern can be tested under the hypothesis that seismically active faults extend from focal depth to land surface. Twenty-two earthquakes in the study area were analyzed by Street, Hermann, and Nuttli (1974). Fourteen of the seismic solutions nearly coincide in location and strike with geologically inferred faults. The other eight agree in strike with the nearest geologically inferred fault.

INTRODUCTION

General Location

The Mississippi Embayment is the portion of the Gulf Coastal Plain that extends northward up the Mississippi Valley to southern Illinois (Figure 1). This area, in which few faults have been mapped, has been the site of violent (up to MM X-XII) earthquakes. This paper

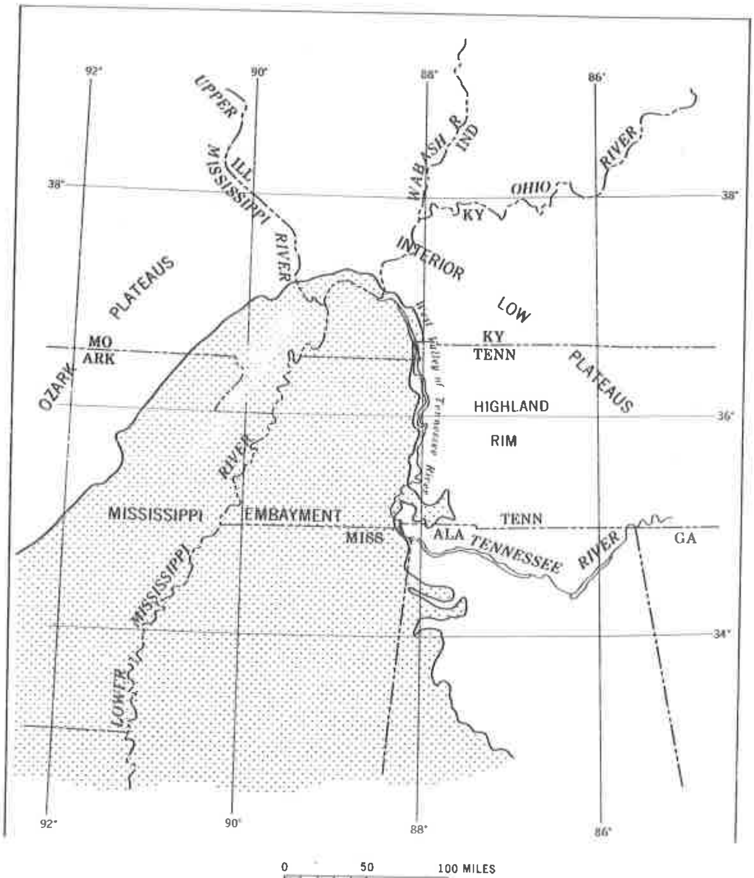

Figure 1. Location of the area of this report in relation to physiographic provinces.

is concerned with the structure of the area and particularly the fault pattern.

As used in this paper, the "head" of the Mississippi Embayment is the region north of 36° latitude. In this area the trough of the embayment syncline changes from "V-shaped" to "spoon-shaped".

This study has both theoretical and practical interest--it can increase knowledge of the general geology and tectonic history of the region, indicate possible faults that may be related to oil or ore deposits, and provide criteria for evaluating earthquake hazard.

Statement of the Problem

Geologists are trained to present conservative interpretations for limited field data. When geologists compile data in preparing a geologic map, faults are commonly shown only if they have been demonstrated by direct observation, repetition or omission of strata or fault gouge or breccia. Many faults are not recorded on geologic maps prepared from limited field data; these may be clearly demonstrated later by more detailed studies involving excavation, drilling or geophysics. In view of the problem of earthquake hazard, however, the conservative concept demands that any suspected fault be shown, because earthquakes are commonly considered to be associated with faults. This is particularly applicable to siting critical structures such as dams, bridges and power plants. Even faults that are inferred from a minimum of evidence should be shown on a "hazard" map, although they may not actually be present.

The northern part of the Mississippi Embayment is an earthquake-prone area. However, because outcrops of the unconsolidated Cenozoic and Mesozoic formations are relatively rare and temporary and generally masked by a blanket of alluvium and loess, the Embayment is an area where faults are difficult to discover and to trace by conventional field methods. Even so, some faults are known in the Embayment, and in adjacent Paleozoic rocks to the north and east, where exposures are better, faults are known to be abundant. Figure 2 shows the surface faults that have been delineated on published maps of the area. The purpose of this paper is to suggest a pattern of additional faults in the Embayment, and then to attempt to test the validity of these inferred faults.

Acknowledgments

This paper is a result of investigation of the Mississippi Embayment, supported in Tennessee and adjacent areas by the Tennessee Division of Geology and the Tennessee Valley Authority. Data used in Arkansas and Missouri were obtained as a part of studies conducted for the Arkansas Power and Light Company and Dames and Moore Company.

Information on drill holes and faulting was freely contributed by other geologists in the various states: for Tennessee, William S. Parks, Geologist, U. S. Geological Survey, and Anthony Statler, Geologist, Tennessee Division of Geology, were helpful; for Kentucky, R. W. Davis, T. W. Lambert and W. W. Olive and R. O. Plebuck of the U. S. Geological Survey; and for Arkansas, William S. Caplan, Geologist, of the Arkansas Geological Commission.

John M. Wilson, Geologist, Tennessee Planning Commission, drew many of the lineations used in this study.

Information and advice from numerous colleagues has been invaluable. For the region in general, thanks are due to C. W. Wilson, Jr., Professor of Geology Emeritus, Vanderbilt University, E. E. Russell, Professor of Geology, Mississippi State University; David Leeds and Steve Ryland, Seismologists, Dames and Moore Co.; J. [illegible]

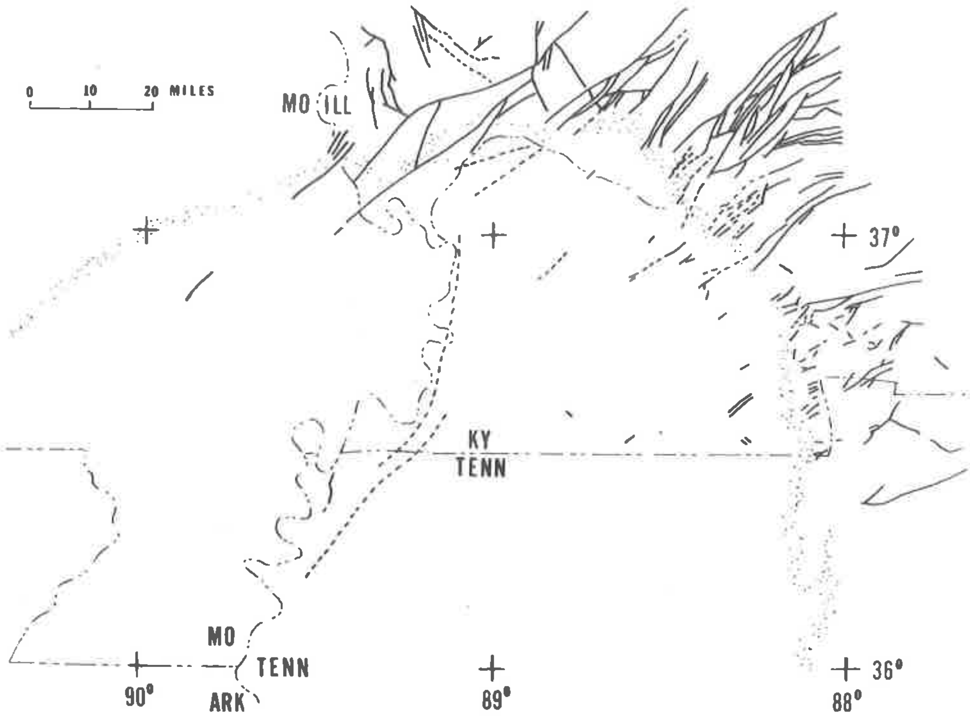

Figure 2. Outcropping faults as drawn by previous workers. To qualify for inclusion on this map a fault must have been drawn as a surface fault on the most recent geologic map with a scale of 1/500,000 or larger. Note that faults are abundant at the edge of the Embayment to the north and east and rare within the Embayment. The present study suggests many more faults within the Embayment.

Kellberg, Chief Geologist, and J. M. Fagan, Geologist, of the Tennessee Valley Authority, and Howard Schwalb of the Kentucky Geological Survey.

Review by J. M. Kellberg, R. C. Milici, S. W. Maher, and R. J. Floyd has resulted in many improvements in the manuscript.

Basic Data For This Study

Basic data for the study consist primarily of previously mapped faults (Figure 2), elevations of contacts in drill holes and on geologic maps, and lineation from topography and streams on quadrangle maps. Most of the data have been obtained from drill hole records of State and Federal geological organizations, and from their published and unpublished detailed geologic maps. Organizations that provided information were the United States Geological Survey, State geological surveys of Arkansas, Illinois, Kentucky, Missouri, and Tennessee, the United States Army Corps of Engineers, and the Tennessee Valley Authority. Most lineations were derived from alignment of contours and stream patterns on topographic quadrangle maps, principally the 1:24,000 series. Other lineations were observed on one ERTS image of an area including western Tennessee and Southeast Missouri.

Previous Investigations of Faulting

Since 1944 three studies have been made of the relationship of faulting to lineations in this region. H. N. Fisk (1944) drew a fault pattern for the entire lower Mississippi Valley (Figure 3). His work was primarily based on lineations visible on air photos, together with trends of river valleys. Krinitzsky (1950) continued this study using photo lineations and tests of particular areas by drilling. Krinitzsky considered the faulting to be related to earthquakes. Later, Stearns and Wilson (1972) drew fault patterns based on lineations in Tennessee and Kentucky near the Mississippi River where maximum bending in the Embayment trough has occurred. The bending zone had been suggested in 1962 by Stearns and Marcher (Figure 4), and Stearns and Wilson's map is shown in Figure 5. Many of the "faults" inferred from lineations by these workers likely are joints rather than faults, because many lineations in adjacent areas of excellent outcrops are joints or joint swarms. The maps of Fisk and of Wilson and Stearns are reproduced here because they were presented in limited edition reports of the Corps of Engineers and Tennessee Valley Authority and therefore may be difficult for some readers to obtain.

Several detailed studies in particular areas have been based on drilling and outcrop data rather than lineations. For example, studies were made by Caplan (1954) in Arkansas; Ross (1964) for part of southern Illinois; Schwalb (1969) in Kentucky; Grokskopf (1955) and McCracken (1971) in Missouri. A regional map of faults was made by Heyl and others (1965) for the faulted fluorspar district partly in, but mainly outside, the northern part of the Embayment; Schwalb and Stearns (1971) drew regional maps of the northern part of the Embayment showing faulting.

Detailed information on faulting in specific areas is published on numerous geologic quadrangle maps in Tennessee, Kentucky and

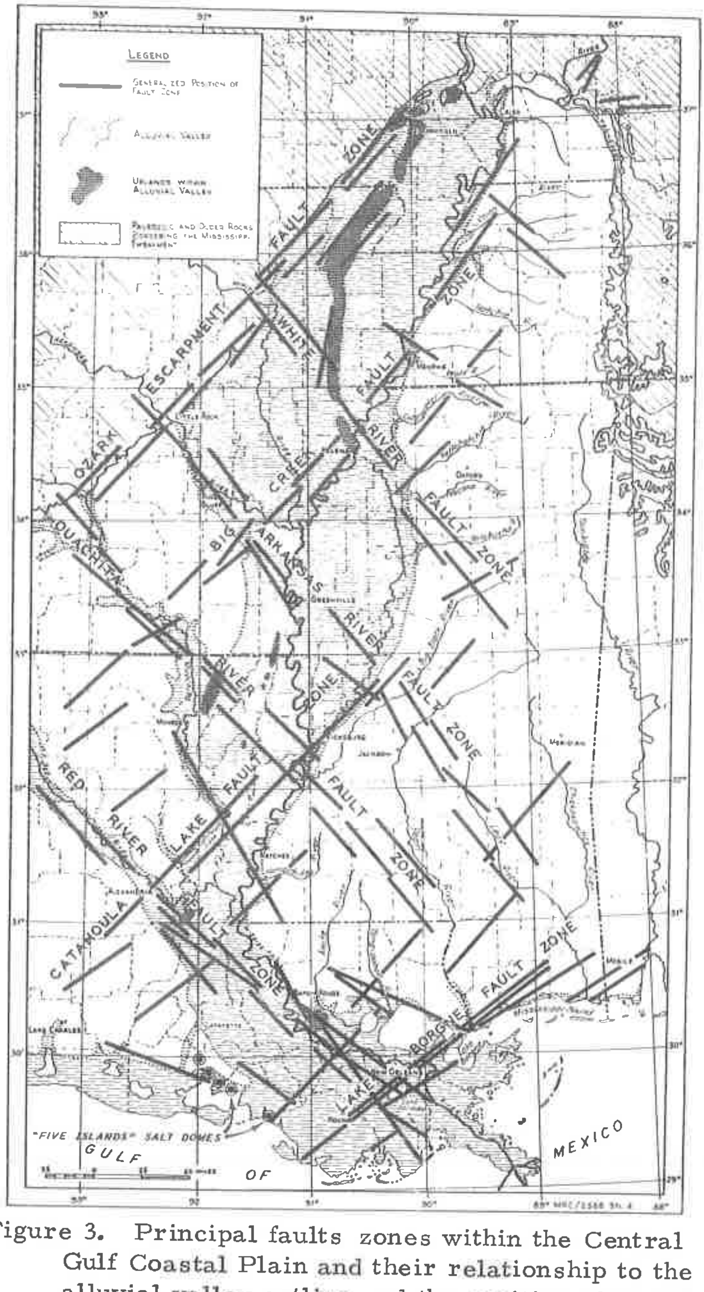

Figure 3. Principal faults zones within the Central Gulf Coastal Plain and their relationship to the alluvial valley outline and the position of coastal plain streams (from Fisk, 1944, Fig. 6). The utility of this map is demonstrated by the Big Creek Fault Zone, which has been verified by drilling and surface mapping in Tennessee and Kentucky.

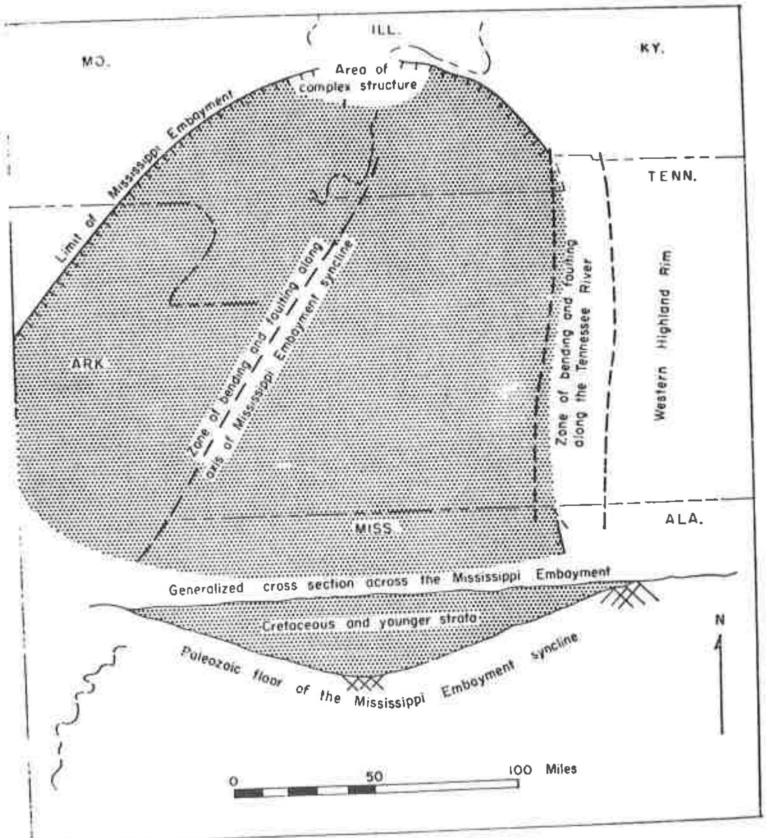

Figure 4. Embayment structure after Stearns and Marcher (1962, Fig. 4). Note that they divide the embayment into two blocks separated by the faulted bending zone at the embayment syncline axis.

Illinois. Examples include the Hickman quadrangle within the embayment in Kentucky (Finch, 1971), the Rockport quadrangle along the eastern margin of the embayment in Tennessee (Wilson and Russell, 1969), the Calvert City quadrangle at the edge of the embayment in Kentucky (Amos and Finch, 1968), and the Illinois portion of the Smithland and Paducah quadrangles at the head of the Embayment (Ross, 1964). Indeed, detailed geologic mapping of the eastern and northern edges of the Embayment has been completed and most of these maps have been published. Cretaceous, Paleocene and Lower Eocene are best exposed in these areas but are overlapped by younger Eocene sediments and covered by alluvium to the west in Missouri and Arkansas. Here structure is known only from drilling records.

Around the head of the embayment from the Missouri-Illinois

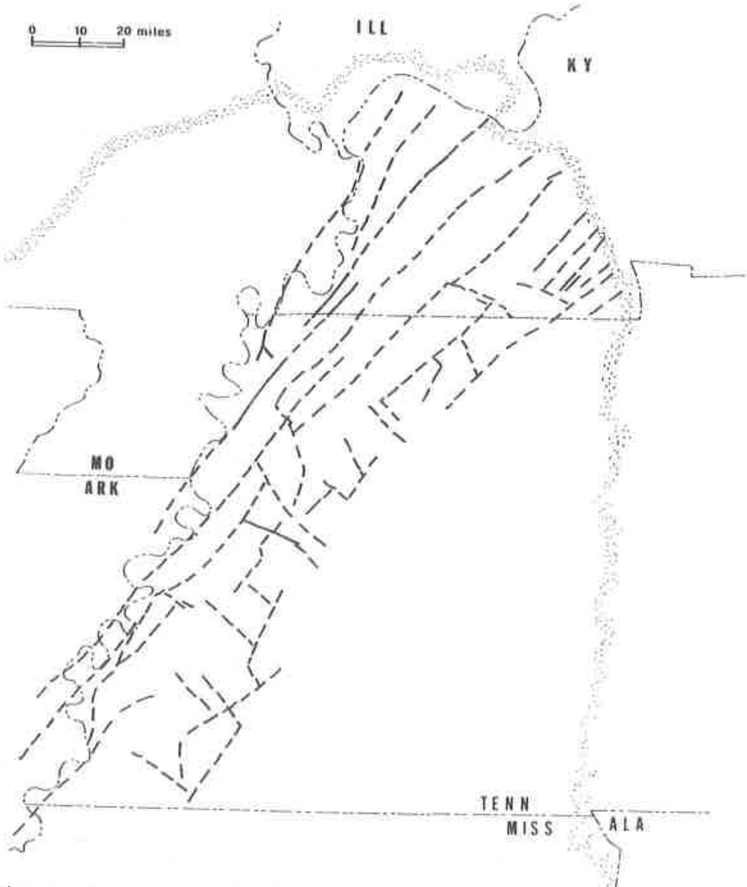

Figure 5. Faults in the Northern Mississippi Embayment and vicinity as mapped by Stearns and Wilson (1972, modified from Fig. 45). Most faults in the embayment (dashed lines) are based only on lineations. Stearns and Wilson only drew faults east of the Mississippi River.

border, across Illinois and Kentucky, faults are mapped from Paleozoic rocks into Cretaceous and Paleocene where they have been drawn as terminating (Figure 2). We will show that by using lineations, together with data from geologic maps and drill holes, that faults can be interpreted as extending across the embayment, cutting through Cretaceous, Paleocene and younger strata, and many may still be active.

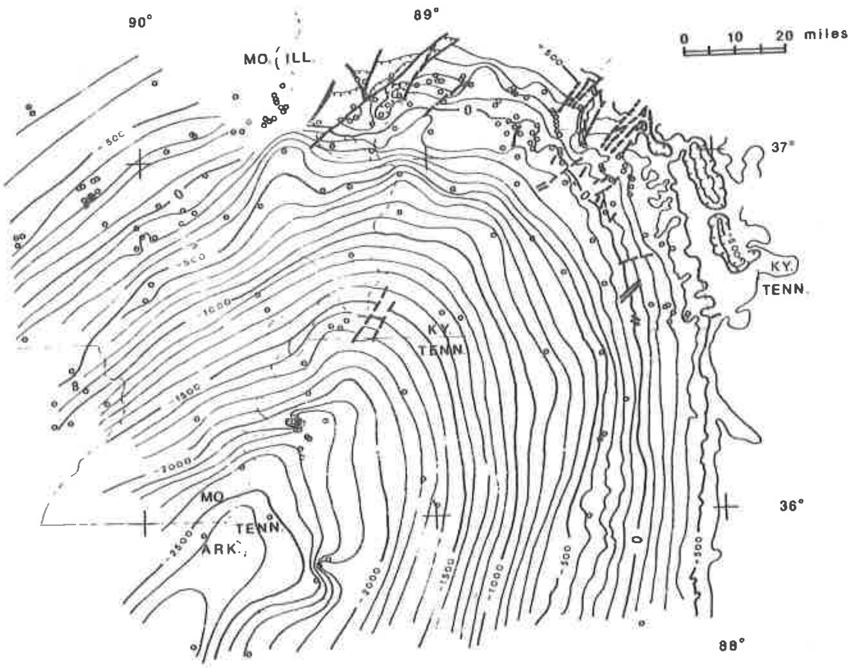

Figure 6. Configuration of the top of Paleozoic-base of Cretaceous unconformity; contoured, smoothly so as to ignore all faults except those mapped at the surface.

SIMPLE STRUCTURE CONTOUR MAPS

Figure 6 is a contour map of the contact between the base of the Cretaceous and the top of the Paleozoic, based upon all available surface and subsurface data. It is contoured as smoothly as the data premit. Faults are shown only where they are mapped at the surface (shown on Figure 2). Figure 7 is a similar map for younger datum (top of the Paleocene Porters Creek Clay). This pair of smoothly contoured structure maps together with the faults shown on Figure 2 could be considered as a complete and correct presentation of the structural geology of the area in a conventional conservative manner. However, the writers here advance the hypothesis that such maps greatly underestimate faulting, and believe that the likely fault pattern in the northern Embayment should be explored, particularly because it is a region of frequent earthquakes, and has experienced faulting in geologically recent and even historical time (Fuller, 1912; Krinitzsky, 1950).

INTERPRETIVE STRUCTURE CONTOUR MAPS

The simple structure contour maps can be redrawn to show additional faults. The same data are used along with lineations from

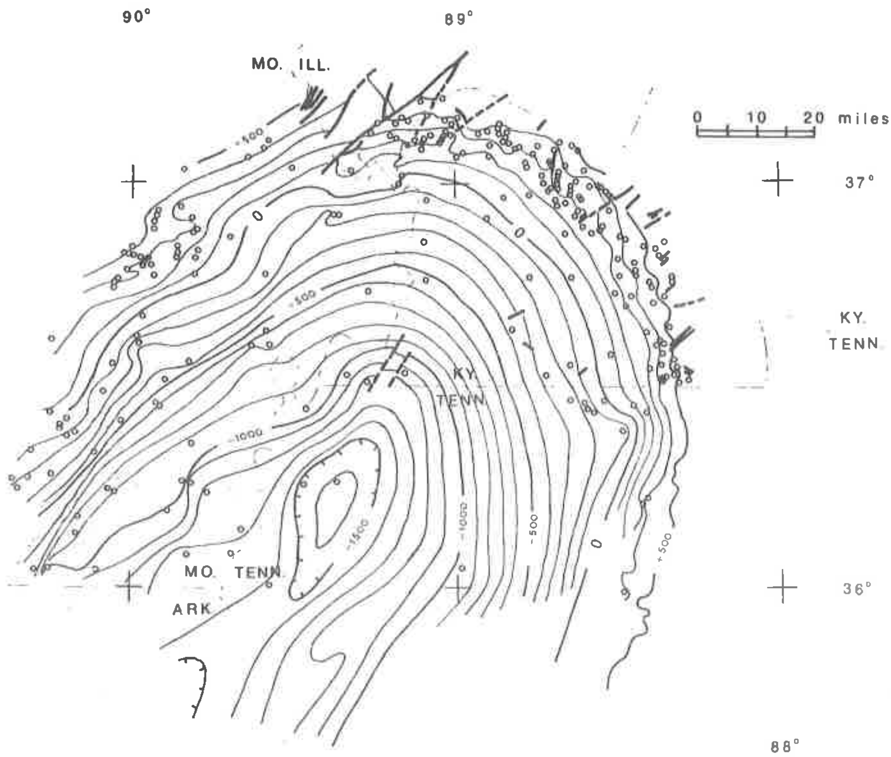

Figure 7. Configuration of the top of the Porters Creek Clay (Paleocene); contoured smoothly so as to ignore all faults except those mapped at the surface.

topographic maps (Figure 8).

Drill hole data can be contoured either simply without faults (as on Figure 6 and 7) or interpretively with faults (Low, 1951). Low's method can be refined by use of lineations. A summary statement of this method follows. A convolution or irregular spacing of structure contour lines may be either a fold or a concealed fault between control points. The working hypothesis is that a fault is present, but the precise location and trend of the fault cannot be deciphered from drill holes alone. Topographic lineations may suggest a precise trend (azimuth) but not the actual location; also closely spaced lineations suggest an actual location, as a fault may be at, or very close to, such a swarm of lineations. Once the fault is drawn, contours are reshaped to conform with more evenly curved and more evenly spaced configuration.

Figure 9 shows the structure of the base Cretaceous-top Paleozoic unconformity with interpretive faults. Figure 10 is a similar map of the top of the Porters Creek. These maps were derived by the process outlined above, and the writers believe that they present a more

Figure 8. Lineations in the northern Mississippi Embayment. Nearly all are from topographic maps (7 1/2-minute quadrangles (scale 1:24,000) or 1 x 2 degree A. M. S. maps (scale 1:250,000)). Longer ERTS lineations are shown in the Missouri "bootheel". These lineations are one of the main bases for location and trend of the hypothetical faults that are drawn on Figures 9 and 10.

realistic general portrayal than Figures 2, 6 and 7.

The fault pattern shown on these maps is generally consistent with the southward projection of the northeast-trending faults of the fluorspar district, which are the northernmost faults on Figure 2.

The dominant northeast trend is also parallel to that of the "Big Creek Fault Zone" of Fisk (1944). The northwest set is also parallel to faulting suggested by Fisk. This is not surprising since the writers and Fisk both used lineations. Though consistent with the work of others and resulting from a rational process, these faults are hypothetical and remain to be proven.

CROSS SECTIONS

Another way to extend faulting from the known, at the edge of the embayment, to the unknown, in the interior, is to "view the structure

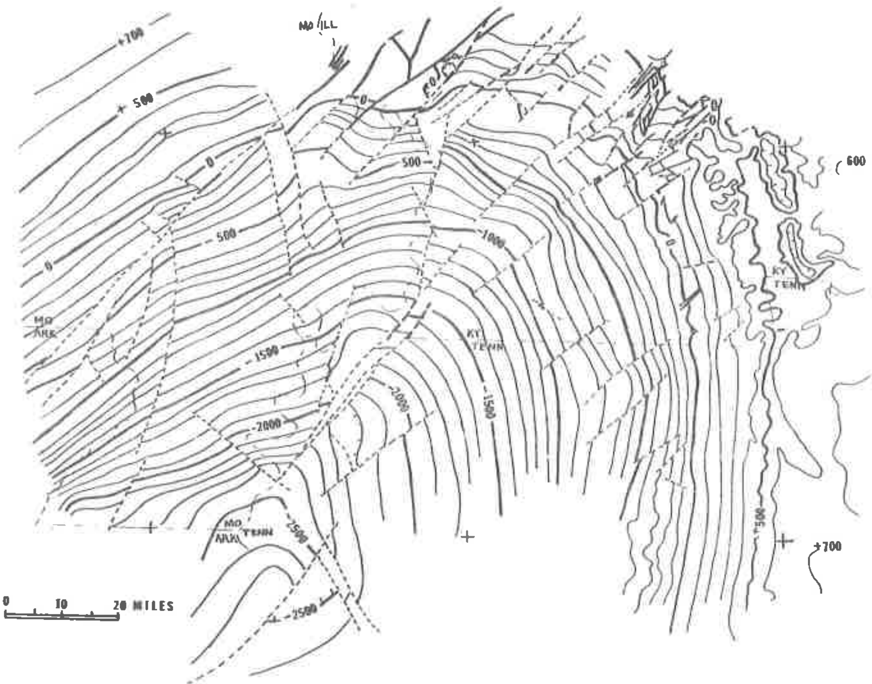

Figure 9. Structure of the Top Paleozoic-Base Cretaceous unconformity showing fault interpretation. Faults on Figure 2 which cut Cretaceous and younger formations are assumed to also cut this datum. Down dip where Porters Creek Clay is exposed, this datum is buried, and few drill holes exist; the detailed trends of Porters Creek Clay contours were used to shape contours and interpret faults on this datum. Well control points are shown on Figure 6.

down the plunge" just as one can project a cross section in a plunging fold belt by viewing a geologic map in a down plunge direction. Figure 11 shows the location of four cross section lines. Two of these extend around the northeast perimeter of the Embayment, where detailed surface mapping and shallow drilling have provided closely spaced control. The other two cross the Embayment where less faulting is known because control is widely spaced. Control for these sections is the interpretive maps, Figures 9 and 10; they will be described first.

The trough section (Figure 12) is drawn to show the characteristic shape and likely fault relationships across most of Tennessee, and northern Arkansas and Mississippi. Here the embayment has a distinct "V-shaped" cross section, and faulting appears to be concentrated in the zone of sharp bending at the bottom of the trough. From here

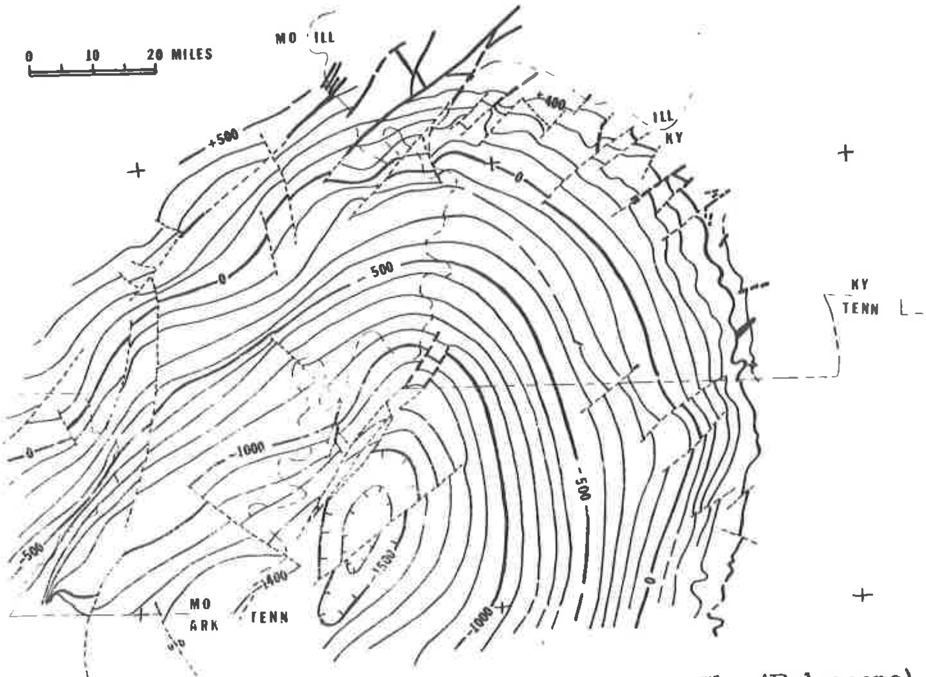

Figure 10. Structure of the Top of Porters Creek Clay (Paleocene) showing fault interpretation. Fewer faults are shown for this datum than the older base Cretaceous. It is likely that some faults shopped moving before the Paleocene. Also, fewer faults are shown because there are fewer well control points, and oil test well records less frequently indicate the top of Porters Creek Clay than the deeper top of Paleozoic rock surface.

northward the embayment has a "spoon" shape and the syncline is more rounded in cross section, and faults appear to be more widely distributed.

The section in Figure 13 crosses the Embayment through southernmost Kentucky. This section will be referred to again, and for convenience it is named the "transverse" section. The "transverse" section has a U-shape and faults occur all the way across the Embayment. Most of these faults are the hypothetical faults drawn on Figures 9 and 10. The actual number of faults across this section may be greater or less than those shown.

It is necessary to draw two sections around the perimeter, because the base of the Cretaceous and top of the Porters Creek Clay do not crop out in the same place. Figures 14a and 14b are structure cross sections of the Cretaceous base and the top of Porters Creek Clay, respectively. They are drawn around the northeast perimeter of

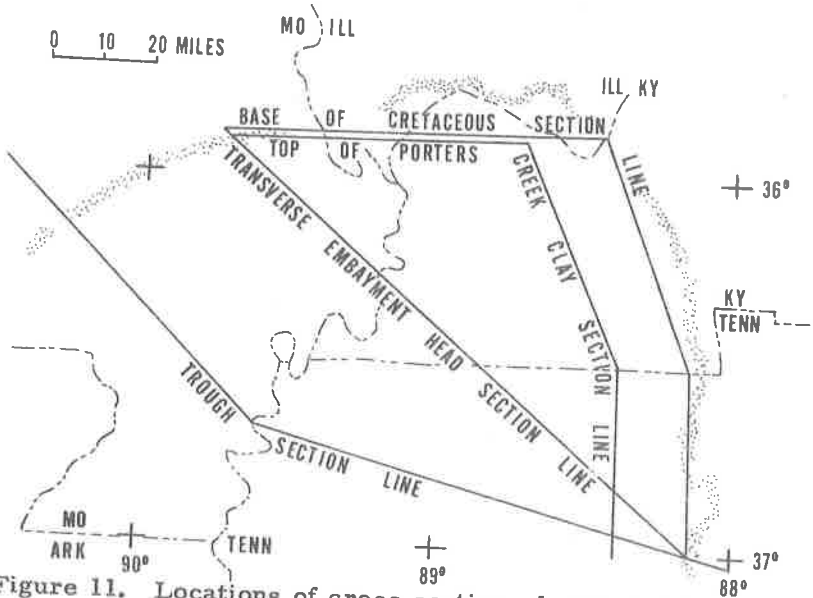

Figure 11. Locations of cross sections drawn on Figures 12 to 18.

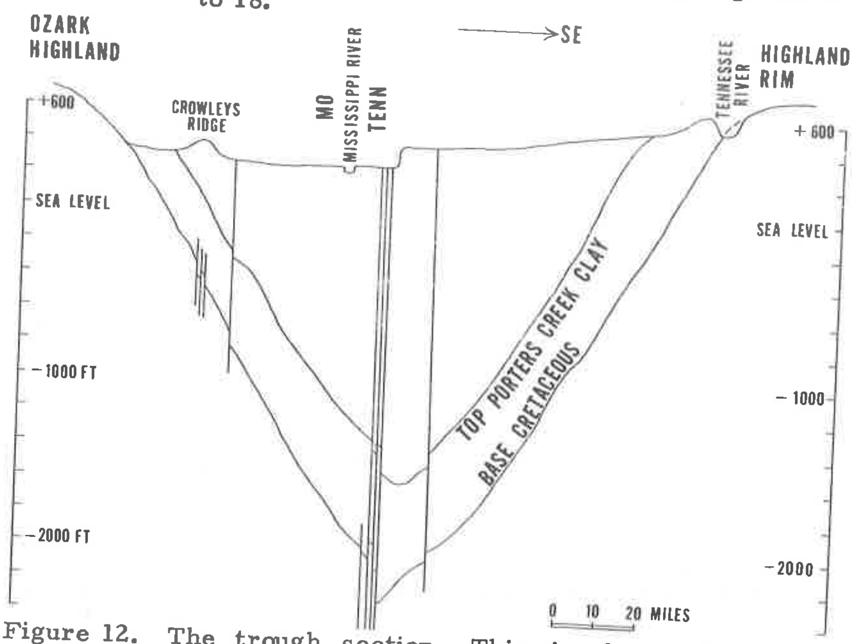

Figure 12. The trough section. This is the southernmost cross section line. It is drawn near the northern limit of the area where the embayment has a distinct "V-shaped" trough, and faulting is concentrated in the bending zone at the bottom of the trough. Faults are vertical on this and subsequent sections because of the large vertical exaggeration (about 300).

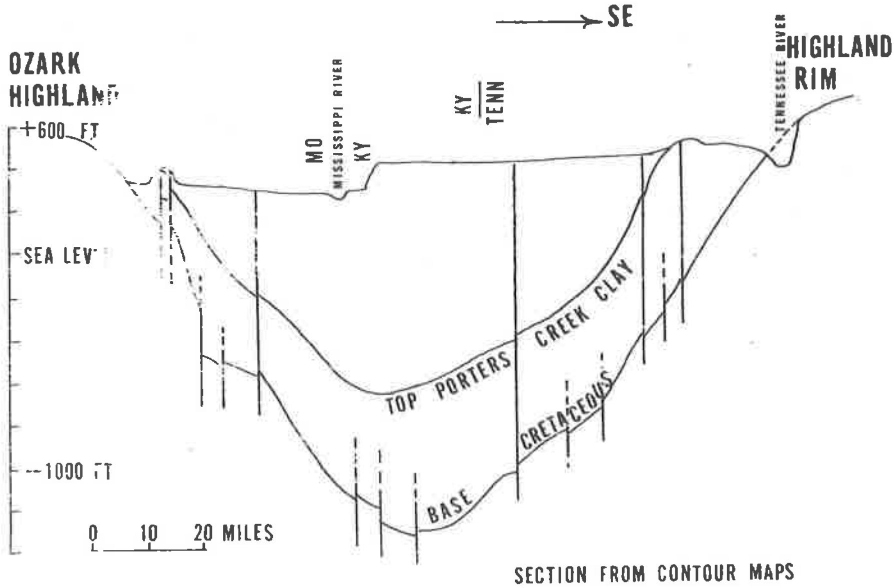

Figure 13. The "Transverse Section". This section extends a-cross the Mississippi Embayment from the Ozark High-land, through Kentucky and Tennessee. This section is drawn from the contours and faults on Figures 9 and 10. This section will be compared with structure at the perimeter of the embayment by projection. Faults are vertical because of the large vertical exaggeration.

the Embayment where many faults are known from surface mapping and closely spaced drilling. These cross sections are much more detailed than Figure 13 because of the large amount of surface control, but they are distorted because they curve around the perimeter rather than extend across the Embayment symmetrically. They will be referred to as "perimeter sections."

In order to compare the sections, the "perimeter sections" are projected into the line of the "transverse" section. Figure 15 shows the base of the Cretaceous structure so projected into the "transverse" section. The top section is viewed horizontally. It is foreshortened for comparison with the shorter transverse section, but the trough shape is not shown. The bottom section shows the same structure as a trough because it is projected down the plunge to the position of the "transverse" section. On either side, where the sections intersect, their elevation is the same. Parts of the perimeter section farther away from the "transverse" section are projected downward over a greater distance, and are therefore lower. Figure 16 is a similar illustration for the Porters Creek Clay. Figure 17 shows both perimeter sections projected into the position of the "transverse" section. Figure 18

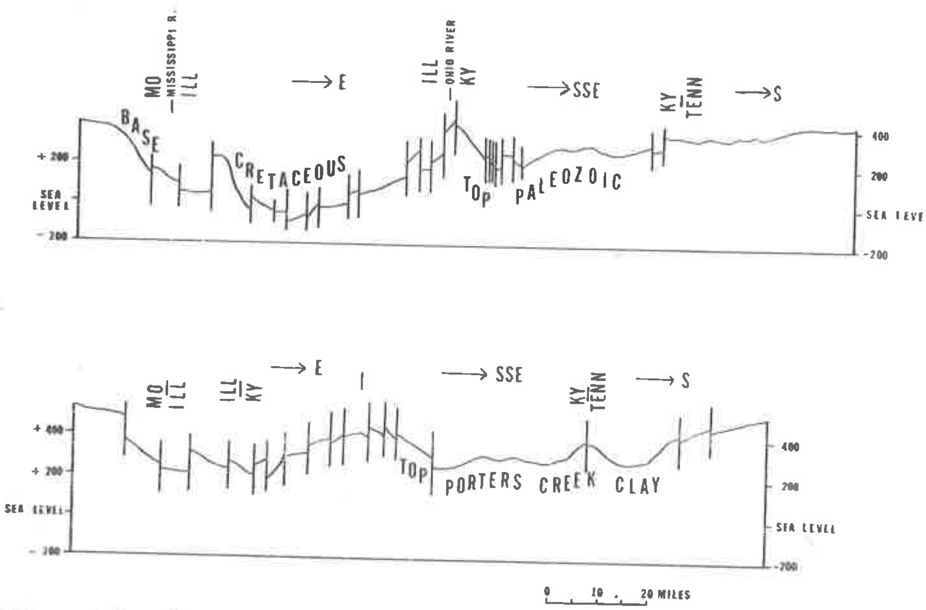

Figure 14. Structure sections of the northeast perimeter of the Mississippi Embayment showing the configuration of the Base Cretaceous-Top Paleozoic surface and the top (Paleocene) Porters Creek Clay at and near its outcrop, from Missouri, through southern Illinois and Kentucky to Tennessee. A (top) is Top Paleozoic; B (below) is Top Porters Creek. These are the datums of the structure contour maps, and are also lines on the cross sections (Figs. 12 and 13). These sections will be projected southward on subsequent illustrations (Figures 15 to 18).

shows the "transverse" section revised to include all the projected faults from the two perimeter sections. It might be considered that this drawing carries the hypothesis of abundant faulting too far; it is, after all, really only a view of faulting at the Embayment head. However, it serves to illustrate the hypothesis that such faulting may be common, perhaps at least as common as the faults on the interpretive maps (Figures 9 and 10).

ALTERNATIVE ABUNDANCE OF FAULTS

There have been three alternative fault abundances illustrated herein. Present field and drill data demand only a few faults (Figures 2, 6, 7) with an average length of about 10 miles or less, and 30 to 40 miles of unfaulted territory between faults. Interpretive contouring

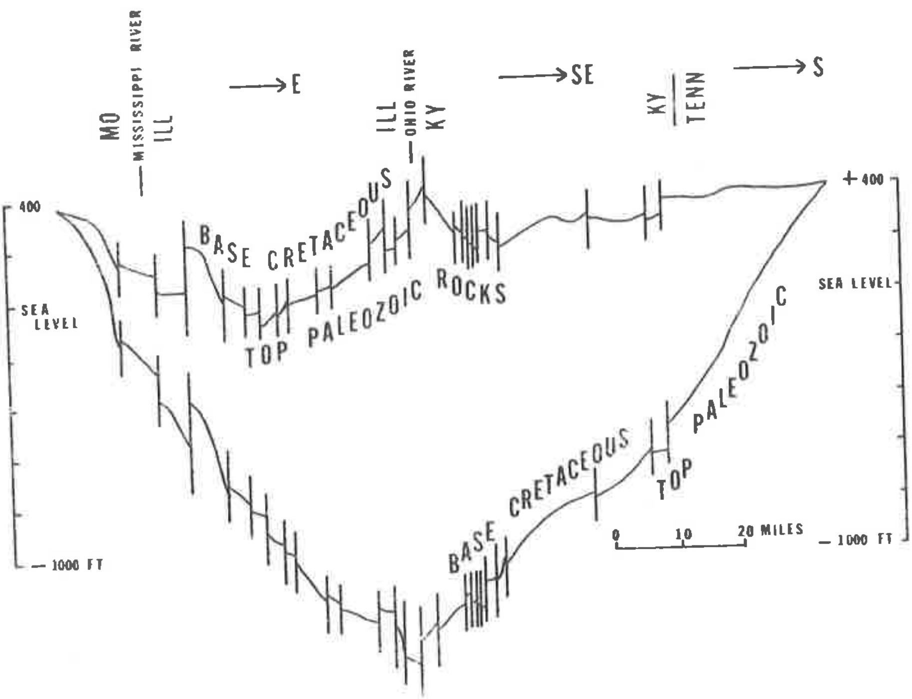

Figure 15. Foreshortened "Perimeter of the Embayment" Base Cretaceous-top Paleozoic rocks section and its projection into the position of the "transverse" section. The top section is as viewed horizontally, and is only foreshortened. The bottom section is projected down the plunge of the embayment axis. The section is drawn this way so it can be compared with the "transverse" section (Figure 13).

suggests many faults 20 miles long or longer with an average distance of about 20 miles between faults. Projection from the faulted embayment head results in an estimate of a fault about every 5 miles. Future drilling and field work in the vicinity of the "transverse" section may determine which of the three is the best approximation.

SEISMIC SOLUTIONS COMPARED WITH INTERPRETIVE FAULTS

The seismic analyses of Street, Herrmann and Nuttli (1974) permit an independent test. It is however necessary to entertain the hypothesis that the faults along which earthquakes occur extend from focal depth to the earth's surface. It is of course logically possible for the inferred near-surface faults of this paper to be correct, but not agree

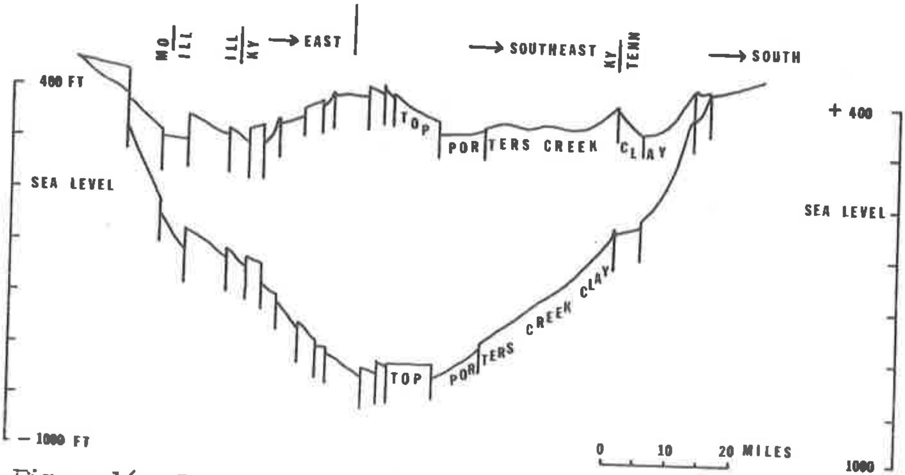

Figure 16. Foreshortened "Perimeter of Embayment" Porters Creek Clay cross section, and its projection into the position of the "transverse" section. The top section is as viewed horizontally, and is only foreshortened. The bottom section is projected down the plunge of the embayment axis. The section is drawn this way so it can be compared with the "transverse" section.

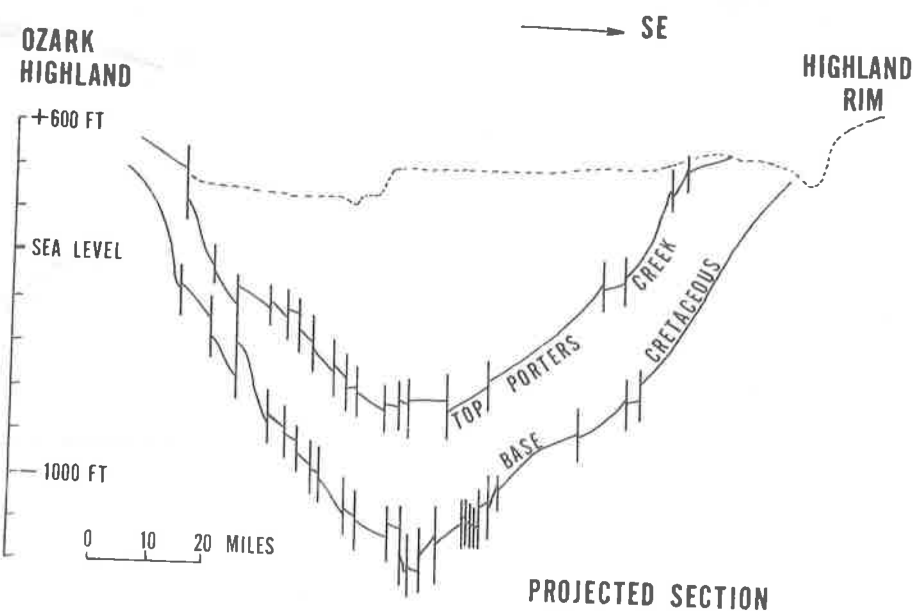

Figure 17. Both perimeter sections projected to position of transverse section.

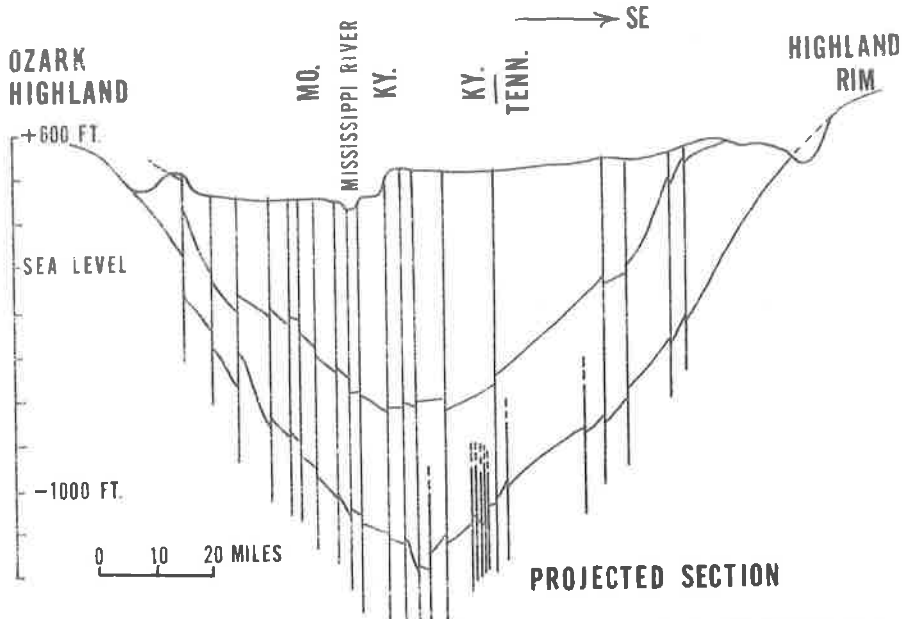

Figure 18. Revised "transverse" section including all faults projected from the two perimeter sections; as though all the faults at the embayment head also occur this far south. This illustrates the hypothesis that even more faults occur in the embayment than can be inferred from sparse drill data.

with seismically active deep faults; however, if they do agree, both the reality and the hypothetical extension of the faults to focal depth should be concluded to be possible.

Seismic data permit estimates of the orientation of fault planes at depth. Twenty-two[1] earthquakes in the area of this report between 1962 and 1974 have been analyzed by Street, Herrmann, and Nuttli (1974). For each earthquake two possible fault planes satisfy the data. If the calculated fault plane having the closest fit to our nearest inferred fault is indeed the correct one, then there is considerable agreement (Figure 19). Table 1 gives the comparative data for the "seismic fault planes" having closest fit to the interpreted faults of this report.

Correlation between earthquakes and faults is a problem because of the uncertainty in locating both. The first factor is an uncertainty of about 5 km in the location of epicenters (Nuttli, personal communication). The second factor is uncertainty in positioning faults

[1]These investigators studied 38 earthquakes, but 16 of these are not in the head of the Embayment area of this report.

Table 1. Relationship Between Seismic [1] and Geological [2] Faults

Event No. (3)	Distance in km between epicenter and nearest geological fault	Angle between strikes of nearest geological & seismic fault (degrees)	Distance in km between epicenter and a nearly parallel geological fault ($\leq$ 15°)	Angle between strikes of seismic fault and nearly parallel geologic fault (degrees)
1	8	15	8	15
3	2	28	8	8
4	3	73	9	10
5	3	7	3	7
6	8	29	10	5
7	8	5	8	5
8	0	0	0	0
9	3	14	3	14
10	9	19	10	9
11	3	3	3	3
12	18	33	22	7
14	2	16	7	5
17	15	3	15	3
26	15	17	26	10
27	5	17	15	0
29	15	11	15	11
30	2	2	2	2
32	6	35	39	12
33	16	4	16	4
35	0	21	13	9
37	15	12	8	3
38	8	3	8	3
No. of Events is 22	Av. Distance is 7 km.	Av. Angle is 17°	Av. Distance is 12 km.	Av. Angle is 6°

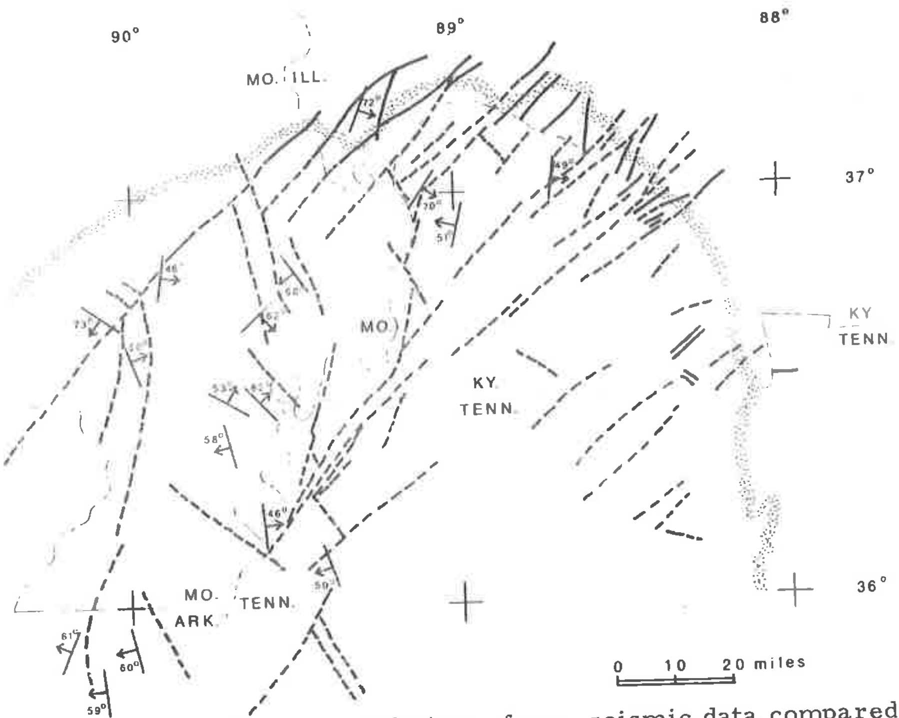

Figure 19. Fault plane solutions from seismic data compared with the inferred fault pattern of this paper. The steeper-dipping fault of the two possible solutions for each earthquake is shown by a dip and strike symbol at the epicenter location. Dashed lines are hypothetical faults. Solid lines are faults mapped on the surface by previous workers.

between wells and in lineation swarms; this is estimated by the writers to be about 3 km. A third factor is the inclination of faults that have one position near the surface and another at focal depth; for example, in a fault with 72° dip and 6 km focal depth, the offset would be about 2 km. In the preceding example, if all of these uncertainties were additive, a 10 km difference would result.

Table 1. Footnotes.

(1) A "seismic" fault is a location and orientation of a possible earthquake producing fault determined by pressure and tension data from seismograph records. Here it is the one of two possible faults that most closely correspond with "geological" fault.

(2) A "geological" fault is one inferred from interpretive contour mapping using lineations and well data, as shown on Figures 9 and 10.

(3) Assigned by Street, Herrmann and Nuttli (1974).

Fourteen of the 22 seismic events agree in location and strike with geologically inferred faults. The other 8 agree in strike (14° difference) with the nearest geologically inferred fault.

Because selection of one of the two alternate fault planes introduces a bias, we will also use the contrary hypothesis that seismic fault plane solutions are random with respect to inferred faults. Then 44 alternate solutions exist to be compared with 22 nearest faults. Now, a random series of the smallest angles between seismically and geologically inferred faults should average 45°. The actual average is 33°. Thus it appears that even unselected seismic calculations relate earthquakes to geologically inferred faults. However, it may be that chance has resulted in a difference averaging 12° less than the expected 45° average.

CONCLUSIONS

A northeast and a lesser northwest series of faults appear to cut Cretaceous and younger strata in the northern Mississippi Embayment. This conclusion is from outcropping faults, manipulation of contour maps based on formational elevations in wells, and on topographic lineations. An agreement in location and strike between these geologically inferred faults and seismically inferred fault planes at depth reinforces the fault pattern and opens up the possibility that many of these faults extend to earthquake focal depth and are still active today.

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PARTICLE SIZE-VELOCITY RELATIONSHIPS FOR SWASH SAND TRANSPORT AT CAPE HATTERAS, NORTH CAROLINA

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ABSTRACT

A fluorescent tracer study was conducted at Cape Hatteras, North Carolina in order to quantify the relationship between particle velocity and grain size in swash sediment movement. Experimental sites were established on either side of the Cape at a 1.5 km distance from the Point. Samples were collected at a 1 minute interval for 30 minutes at a position 20 m downdrift from the point of tracer release. The sampling line consisted of a 10 m transverse section of the beach foreshore, stretching from the uppermost swash penetration to the zone of breaking waves.

Four size classes of 100,000 grains each were dyed different colors and introduced into the mid-swash zone. The rate of transport for each size fraction was determined by plotting particle recovery versus time. Results indicated an inverse size-velocity relationship with larger particle velocities being obtained for the northern portion of the cuspate foreland. Some swash surges were recorded in the particle-recovery distribution as well-marked peaks.

INTRODUCTION

Fluorescent tracer particle studies have proven to be an inexpensive and quick means of investigating sand transport in the nearshore marine circulation zone. The present study utilized tracers in order to quantify swash zone size-velocity relationships on either side of Cape Hatteras, North Carolina (Figure 1). This giant cuspate foreland is located on some 2,500 meters of marine sediments. Its shape has been

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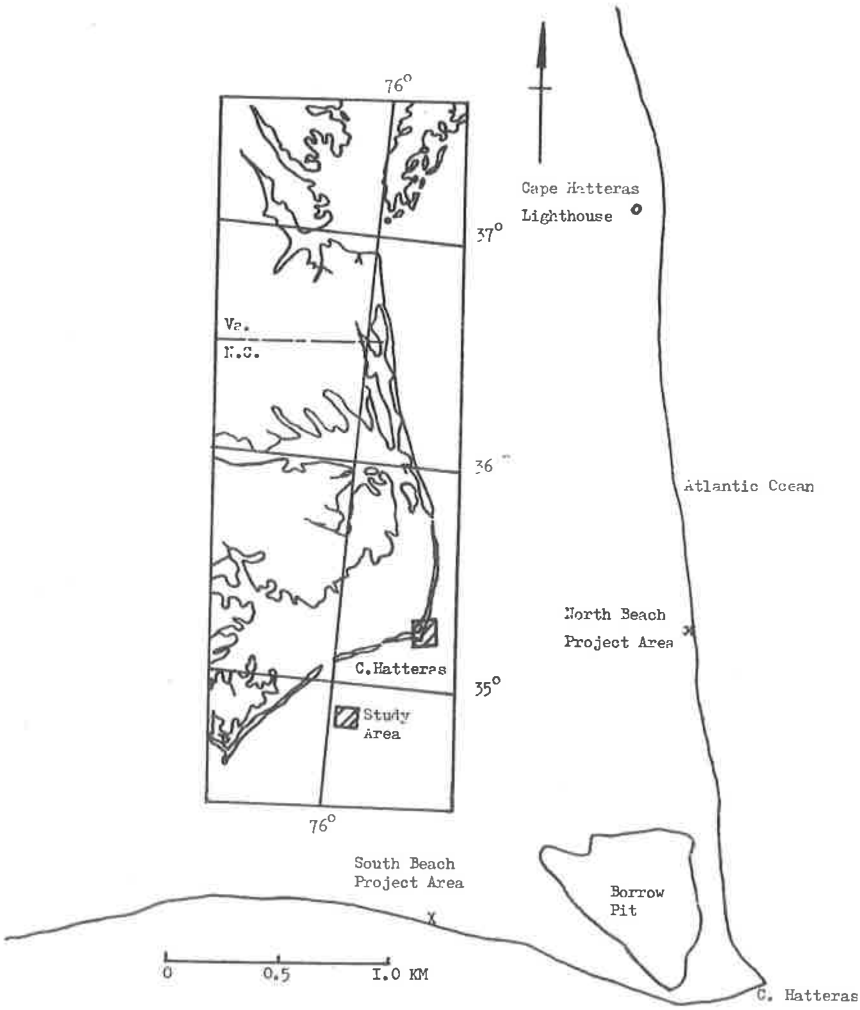

Figure 1. Location map of the investigated area.

attributed to many factors, viz, nodal points of Gulf Stream eddies, wave refraction, an accumulation zone where two curved spits meet, etc. In essence it appears to require for its formation: (a) high rate of sediment input in the littoral drift areas, (b) high energy conditions and (c) stabilizing or shielding points around which sand can accumulate (Swift and all, 1972, p. 525).

Yasso (1965) showed that an inverse size-velocity relationship was normal for sandy beach foreshores. Therefore smallest sized grain particles should arrive before the larger ones. If this is true, there should be a time differential between the arrivals of differently sized sediments.

Two experimental areas were marked off 1.5 km in a northward and southward direction from the "Point" of the Cape (Figure 1). These were designated northern and southern beaches, respectively, and a time-integration sampling program (Crickmore and Lean, 1962) was utilized in order to quantify the relationships between sand size and velocity. In this sampling procedure, samples are taken at a specified time interval along a section transverse to the direction of flow. Quantification of the fluorescent tracers recovered enabled the time distribution of peak particle concentration to be calculated.

If a quantity of marked sand grains is infected into the swash, the grains will be recovered at the rate $dA_o(t)/dt$ proportional to the amount of tracer $A_o(t)$ remaining at the site. If A is in initial amount of tracer and r is the erosion rate, then $A_o(t) = Ae^{-rt}$, indicating an exponential removal of tracer with time. Crickmore and Lean (1962) have indicated that this is not always true because frequently:

1. Tracers are released at a higher rate than predicted during the later stages of recovery.
2. It is difficult to define the extent of the zone of motion.
3. Depositional processes occur as well as erosional.
4. The tracer motion depth is difficult to determine.

The dominant longshore beach drift off the northern beach is southeast, while on the southern beach it is northeast, hence the zone of accumulation. Estimates of the total amount of littoral drift differ, but Langfelder, and all (1968) have postulated southward values of 2.20 million cu yds/yr and northward values of 0.26 million cu yds/yr. In the past two years, sand has been dredged from a borrow pit located in the Cape and pumped some 5 km northwards. This beach nourishment program has resulted in the average mid-tidal sediment size decreasing slightly from 1.15 Φ to 1.25Φ. For this project, mid-tide modal grain size values for the northern and southern experimental areas were found to be 0.68 Φ and 1.05 Φ, respectively.

Two main mechanisms produce littoral drift. The first is by waves striking a coastline at an oblique angle which generates a longshore current in the surf zone. The second is swash drifting via the action of a wedge of water spilled onto a beach by an obliquely breaking wave. This pushes sediment obliquely up a beach face. Gravity moves the water and sediment downslope in a direction normal to the beach face giving the well-known "saw-toothed" pattern of swash sediment movement. However, an oblique backwash can also contribute to some longshore transport. Sediment could be removed from the swash zone and deposited seaward, but it "does not contribute much to the longshore sediment transport" (Komar, 1971).

Acknowledgments

The writers wish to thank Terry McDuffie, Rich Wiener, John Curtis, Jack Cowardin, and Dr. O. A. C. Williams in collecting sediment samples.

FIELD PROCEDURE

Sediment samples obtained from the Hatteras foreshore were laboratory dried and sieved using a Ro-Tap machine. The sieve interval used was 0.5 Φ and sieving carried on for 20 minutes per sample (Griffiths, 1967). Sediment particles in the size range -1.0Φ, 0.5Φ, 1.0Φ, 1.5Φ were then selected for color coding, i.e. particles with a grain size of 1.0 Φ passed through the 0.5 Φ sieve but were trapped on the 1.0Φ sieve. One thousand grains from each size range were counted and weighed so that the weight of 100,000 grains of each size range could be determined. The four grain size distributions obtained were then coated with different colors of fluorescent dye (Table 1).

On October 26, 1974, a pilot survey of tracer recovery was conducted at the southern beach. Fluorescine dye was placed in the water and its direction and time taken to travel 20 m noted (northeast at 0.28 m/sec). At 1130 hours, 30 minutes before low tide, 100,000 yellow and green coated grains were introduced into the swash as a single injection. Movement of these coarse grains was easily seen with the unaided eye. Sampling was staged at 1 minute intervals. The first coarse-grained sediment (yellow) traveled the 20 m distance downdrift to the sampling point after only 2 minutes, indicating strong beach drifting. The sampling time interval for the main high tide experiment was accordingly set at 1 minute intervals for a 30 minute duration.

However, when the main experiment commenced at 1715 hours, the current direction had reversed and now flowed southwest at a velocity of 0.24 m/sec (Table 2). This current reversal was attributed to a 20° change in the overall wave incidence angle. This meant a reversal of injection and sampling points for this particular beach. For the northern area, the current direction was southeast all day. Current reversals south of the Cape are not unusual and are one of the reasons that Langfelder, and all (1968) found much lower values of net littoral drift in this area.

At 1715 hours, 15 minutes before high tide (the time was chosen in order to eliminate tidal influence), 100,000 grains of each of the four size ranges were introduced at mid-swash level prior to the uprush of a breaking wave at both beaches. Sea state conditions during the test were as shown in Table 2. Standard field techniques were used in obtaining these sea state parameters. At a 20 m distance downdrift from the point of tracer release, a sampling line was established by placing three stakes in the foreshore transverse to the beach. This 10 m section extended from the uppermost swash penetration to the zone of breaking waves which roughly coincided with the step. Ingle (1966) has shown that 90 percent of sand drifting occurs within this zone. The step zone is a marked break of slope normally located under the wave crest or just shoreward of it. It is formed by the returning backwash colliding with the base of the incoming breaker. Each of three samplers was then assigned a 3 to 4 m length of section from which to collect

Table 1. Fluorescent Color Coding and Grain Size.

Color Code	ϕ Size	Wt/100,000 grains (gms)
Red	2.0 - 1.5	15.0
Blue	1.5 - 1.0	41.0
Green	1.0 - 0.5	128.0
Yellow	-0.5 - -1.0	616.0

Table 2. Sea State Parameters, October 26, 1974.

Southern Beach	Parameters	Northern Beach
SSW	Wave direction	ENE
9	Period, T (secs)	10
415	Wavelength, (m)	512
11	Beach Slope (°)	11
50	Breaker Height (cm)	75
SW	Longshore Current Direction	SE
.24	Longshore Current Velocity (m/sec)	.38

approximately a 1 kg sample. Large plastic bags weighted down with big stones were dragged along the surface to a depth of around 3 cm. These samples were laid out on the beach and marked by a fourth person who also acted as a time keeper. At the end of the experiment, the three samples taken during each time increment were combined. It would have been better to examine the individually-collected samples from the three stretches of beach face, but logistics necessitated this amalgamation. Therefore, the sampling technique yielded a "channel" or composite sample along the 10 m beach instead of a sample specific to any definable portion of the beach face.

Plugs of dyed sand, extending from the surface down to 30 cm in the mid-tide position, were used to determine the depth of sediment disturbance by waves (Williams, 1971). After a complete tidal cycle, the plug was sectioned so that the amount of truncation could be determined, the depth of disturbance being equal to the amount of undyed sand deposited on top of the sand plug. As a small proportion of grains are lost by burial (Ingle, 1966), and the thickness of the moving sediment layer was unknown, this experiment was undertaken to determine the total amount of sediment disturbance over a tidal cycle at the two locations. This value determined the plastic bags sample depth.

The sample size for the 60 composite samples collected in the

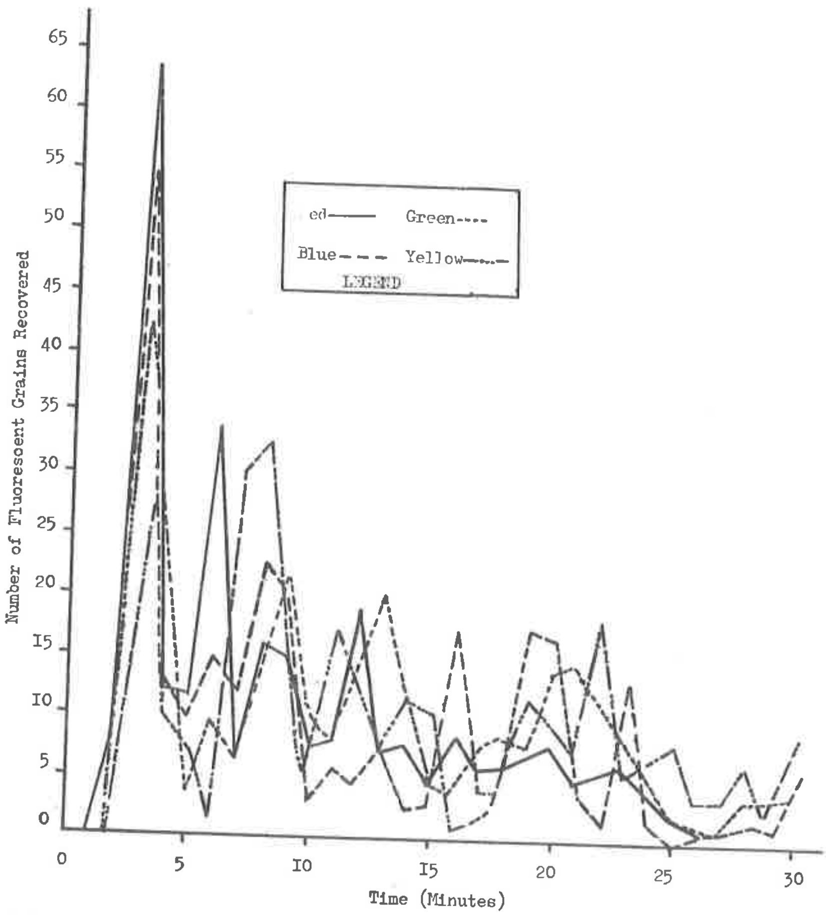

Figure 2. Fluorescent tracer recovery at the northern beach.

field varied considerably. In order to standardize this parameter, larger samples were split down to the smallest obtained sample size by means of a sample splitter. The particles were then spread out on a table to a depth of 2 to 3 millimetres for examination. An ultraviolet light in a darkened room was used to identify and count the individual tracer particles. Figures 2 and 3 show the particle-recovery distribution for the northern and southern sites, respectively. A mechanical analysis was made for all samples from which cumulative frequency curves were drawn. Figure 4 represents the "sweepzone" envelopes for cumulative frequency curves obtained for the north and south beaches. This was carried out in order to determine any significant statistical difference between the sediment populations. In order to give clarity to Figure 4, the curves were not drawn separately.

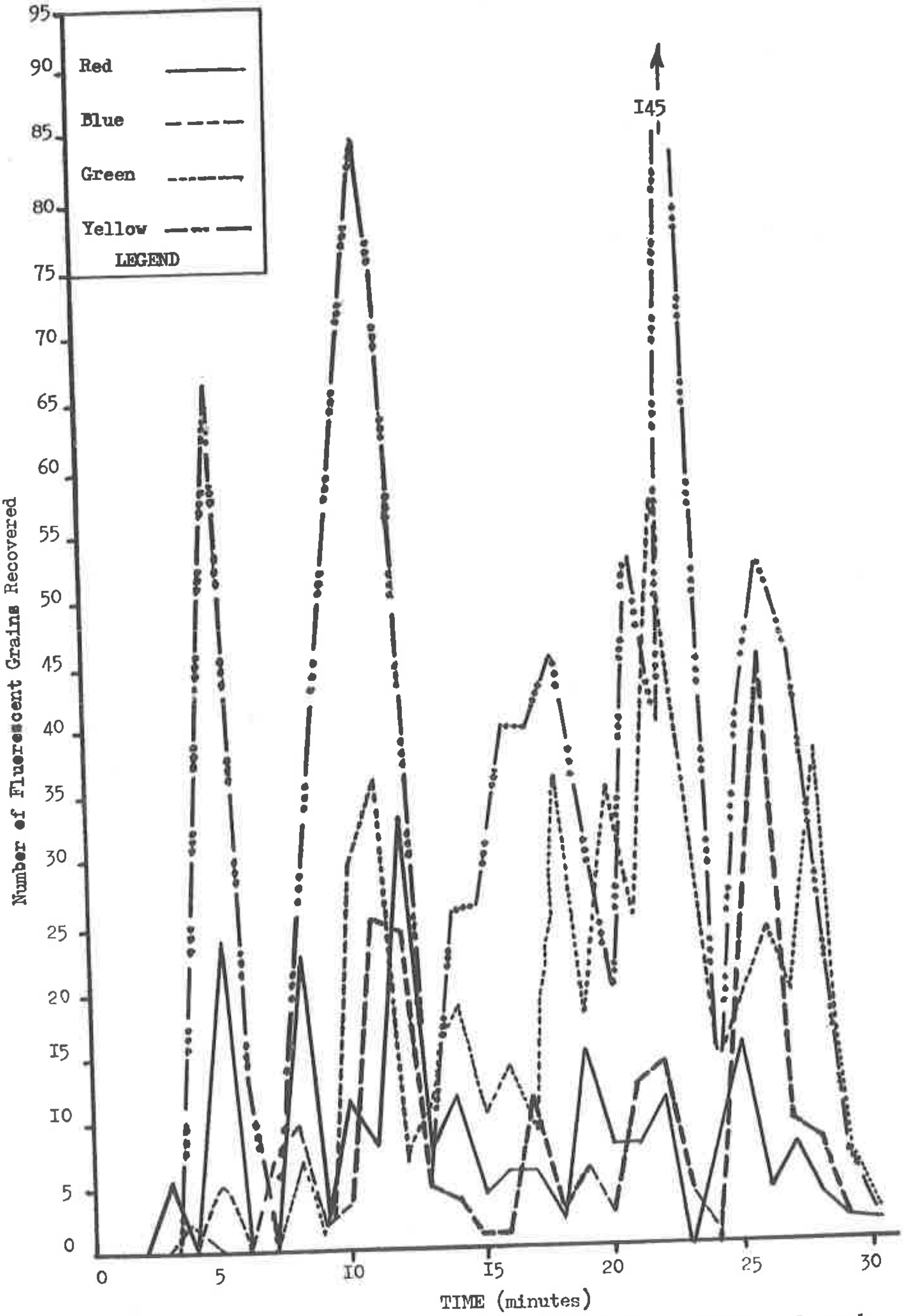

Figure 3. Fluorescent tracer recovery at the southern beach.

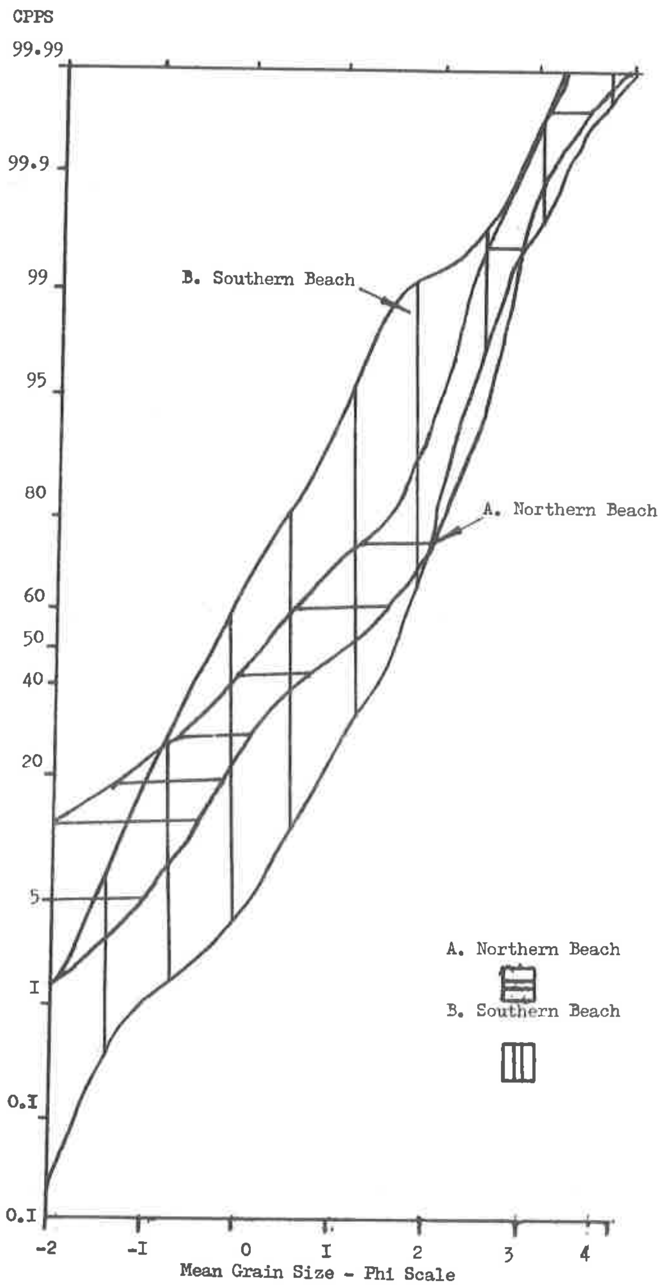

Figure 4. "Sweep zones" of cumulative percent probability scale (CPPS) and mean grain size (Φ scale) of all sediment samples obtained from Northern and Southern beaches (30 frequency curves per graph).

Table 3. Maximum Particle Transport Velocity (cm/sec). (All numbers rounded to nearest whole number).

Southern Beach	Color Code	Northern Beach
11	Red	33
8	Blue	17
7	Green	17
8	Yellow	17

RESULTS

Data from both beaches, based on the 1 minute sampling interval, showed that the smallest sized grains (2.0ϕ- 1.5Φ) did move through the designated distance with the highest velocity, but each succeeding class did not follow in order (Figure 3 and Table 3). The fastest moving grains on both beaches were colored red (2.0ϕ- 1.5Φ), supporting Yasso's (1965) views of the inverse relationship that exists between grain size and velocity (Figures 2 and 3, Table 3). Maximum particle transport velocity is defined by the time of recovery of the first particle for each grain size range.

Peak number of particles obtained have been graphed in Figures 2 and 3, and the peak particle recovery velocities given in Table 4. In each case the proportion of coarser-sized grains to finer-sized grains increased as time elapsed during the experiment. This is in accord with theories first developed for undirectional stream flow, and later applied to beach studies involving grain movement in all directions from the release point (see Ingle, 1966). This suggests an immediate differential movement of grains of varying character. Smaller grains are usually carried in suspension whereas larger grains will be moved along a beach by saltation processes. Grains in suspension have a much greater chance of being moved beyond the breaker zone than particles moving by saltation or traction transport (Grant, 1943).

Data from mechanical analyses showed strong similarities between sediment samples (Figure 4). For the northern beach, median grain size ranged from 0.1 Φ to 1.2 Φ whereas, for the southern beach it was -0.8 Φ to 1.3Φ. Sorting values ranged from 0.21 Φ to 1.32 Φ and 0.34 Φ to 1.21 Φ for the northern and southern beaches, respectively. These values represented the extremes of the population distribution and a much tighter distribution was found for 85 percent of the sediment cumulative curves indicating little difference between samples.

For the northern beach, the largest number of all grain size were recovered during the third minute. In terms of the total number of tracers found, more of the smallest grain size (red) were recovered while the yellows and greens (coarsest material) were retrieved in the

Table 4. Peak Particle Recovery Velocity (cm/sec).

Southern Beach	Color Code	Northern Beach
2.7	Red	11.1
1.3	Blue	11.1
1.5	Green	11.1
1.4	Yellow	11.1

smallest numbers. Velocities obtained were greater than those recorded on the southern beach due to higher energy conditions, and the smallest-sized grains (red) moved twice as fast as the others (Table 3). This finding agrees with Ingle's (1966) work in California. Peak tracer recovery corresponds to an apparent transport velocity of 11 cm/sec for all grain sizes (Table 4). With the high energy conditions encountered at this beach, it was possible that the coarser grain sized materials were moving offshore contrary to the usual sorting concepts characteristic of beach sediments (Grant, 1943). Krumbein (1944) was the first to report selective offshore transport of coarse-grained sediments. However, movement of some grains along on and offshore vectors was probably related to the fact that they were not in equilibrium with the slope and fluid regime at the dump point.

At the southern beach, the coarsest sized material (yellow) was recovered in significantly larger quantities than any other size range. This corroborated the usual beach sediment transport concept that coarse sized grains move on shore, fine sized grains offshore (Grant, 1943, Zenkovitch, 1967). The fastest moving grains (red) had an apparent transport velocity of 2.7 cm/sec (Table 4). A comparison of the graphs for the two beaches indicated an exponential decay rate for the northern beach (Figure 2) while the southern beach exhibited wide variations (Figure 3). The large fluctuations in tracer recovery for the southern beach can be explained by the reasons stated earlier (higher tracer recovery rate during the latter stages, deposition difficulty in determining tracer motion depth and extent of zone of motion). The long-shore current velocity at this beach was 24 cm/sec (Table 2). Rapid increases in the strength of alongshore vectors of grain motion occur only at higher velocities than this (Ingle, 1966). In addition, it was noted that the coincidence of two swash surges resulted in a non-linear interaction where one surge rides over the swash of the preceding surge with an attendant increase in energy conditions. The coarsest material, that is, the yellow colored grains, were transported in greatest quantities during these periods of high influxes of wave energy with minimal transport between times. Orlava (1964) working in a similar wave regime in Europe showed that sediments of grain sizes 2 Φ and -1 Φ moved with velocities of 0.6 and 0.3 respectively of the longshore current's velocity. There is close accord with these figures and the results presented in Table 3 (15 cm/sec and 7 cm/sec respectively for the fine and coarse-grained sediments at the southern beach; 25 cm/sec and 12 cm/sec respectively for the same sediments at the northern beach).

CONCLUSIONS

The time history of a tracer's position as it moves along a beach face can be used to deduce the sand advection rate. The thickness of the moving sediment layer was unknown but over a 30 minute increment during sea level stillstand little material would be buried beyond the 3 cm sampled level. This assumption was verified by the depth of disturbance experiment. The depth of sediment disturbance over the tidal cycles corresponding to the tracer study was only 7 cm and 6 cm net deposition at the northern and southern beaches, respectively. Therefore, scraping the top 3 cm of sediment took in approximately 50 percent of the beach tidal cycles's sediment deposition. Any buried tracer grains would be included within the sample. Average maximum velocities of 33 cm/sec for red grains (2.0 Φ to 1.5 Φ) and 17 cm/sec for others were recorded at the northern beach; 11 cm/sec of red (2.0Φ to 1.5Φ), and 8 cm/sec for blue (1.5Φ to 1.0Φ) and green (1.0Φ to 0.5Φ), and 7 cm/sec for yellow (-0.5Φ to -1.0Φ) at the southern beach (Table 3). Several large swash surges gave well-marked peaks of fluorescent tracer recovery at the southern beach. With respect to transport velocities, much higher values were recorded for the northern beach. Results indicated a consistency in the inverse size-velocity sand grain relationships for medium-coarse sand. However, more results are needed from experiments conducted under a variety of sea state conditions since many complex factors are involved in the transport of sediment along natural beaches.

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SERPENTINIZATION OF THE HOLCOMBE BRANCH DUNITE, NORTH CAROLINA

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ABSTRACT

A geochemical and mineralogical study of the partially serpentinized Holcombe Beach dunite of Madison County, N. C. indicates that the serpentinization process randomly altered the ultramafite. The degree of serpentinization differed randomly throughout the body, indicating that the serpentinization did not progress from exterior to interior. Chemical studies indicate that the MgO/SiO_2, FeO/SiO_2 and $MgO + FeO/SiO_2$ ratios bear no apparent relation to the degree of serpentinization, ruling out a constant volume chemical model for the serpentinization process. Lack of structural deformation in the country rocks likewise argues against a constant composition model. It is concluded that the alteration was a product of several chemical processes acting on the body, yielding constant volume and essentially constant composition serpentinized dunite.

INTRODUCTION

Statement of the Problem

The purpose of this study is to attempt to understand the serpentinization process that has affected the ultramafic rocks of the southern Appalachians. These bodies vary in their degree of serpentinization from less than 10 to 100 percent.

The Holcombe Branch dunite of Madison County, North Carolina (Figure 1), is a partially serpentinized ultramafic mass. Preliminary

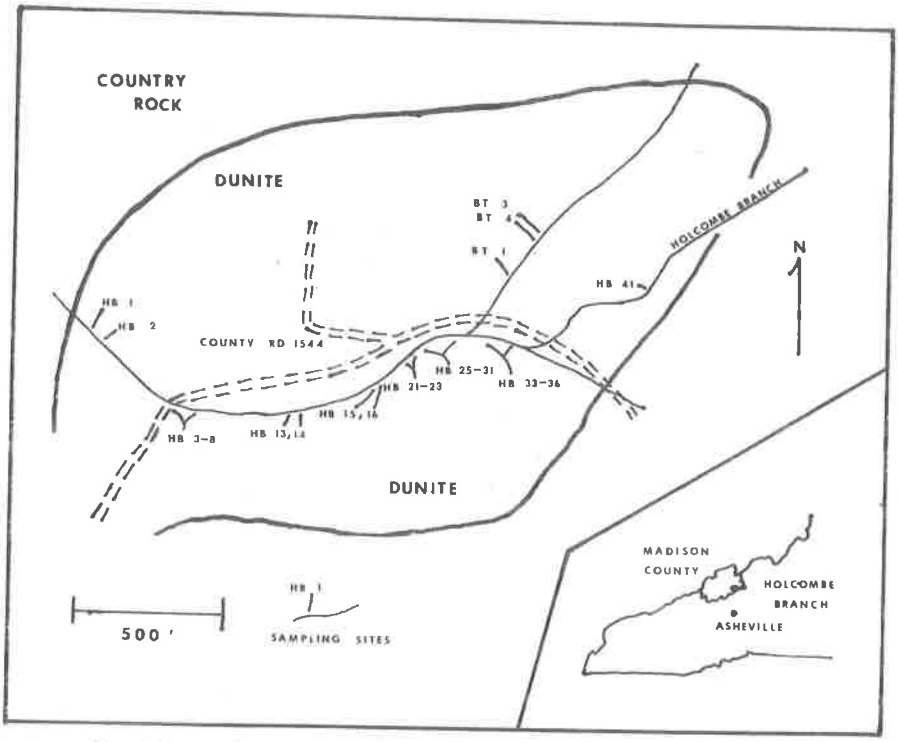

Figure 1. Map of Holcombe Branch dunite showing sample localities and approximate boundary of the dunite body (modified from Hunter, 1941). Inset shows index map of western North Carolina.

studies indicated that the degree of serpentinization of this ultramafic body is variable, which makes it a good example on which to test various hypotheses of serpentinization. Thirty specimens of serpentinized dunite were collected along Holcombe Branch and its largest tributary to detect systematic variations between the degree of serpentinization and both the bulk rock chemistry and proximity to the country rock contact.

Previous Work

The only previous work on the Holcombe Branch dunite was done by Hunter (1941) as part of a study of the economic potentials of olivine deposits in North Carolina and Georgia. He presented a sketch map and a brief petrographic description of the dunite, but did not discuss the serpentinization. Previous studies on the serpentinization process affecting any of the southern Appalachian ultramafic rocks include works by Miller (1953) and Condie and Madison (1969), who studied the

Webster-Addie complex, and Neuhauser and Carpenter (1971), who studied the Bank's Creek serpentinite body of Yancey County, North Carolina.

Acknowledgments

We appreciate the help of Robert G. Coleman, of the U. S. Geological Survey, Menlo Park, who provided the serpentine mineral samples used for X-ray diffraction standards, and a translation (by P. B. Hostetler) of the paper by Shteinberg (1960). The investigation was partially financed through National Science Foundation grant GA-4518 to J. R. Carpenter.

DESCRIPTION OF THE DUNITE

Field Relations

The dunite body is roughly 2500 feet long and up to about 1,000 feet wide. It is elongated in a NE-SW direction, roughly parallel to the local and regional structural trend. Outcrops are generally quite fresh; however, almost everywhere the rock is jointed in three approximately orthogonal directions, and these fractures and the outcrop surfaces are moderately to severely weathered, to a depth of up to a half-inch. The least weathered samples are found in stream exposures, particularly along Holcombe Branch, which transects the length of the dunite mass. Relatively unaltered dunite is yellowish to yellowish-green, granular, and somewhat friable. The more serpentinized dunite is dark to very dark blue-green, quite smooth, and compact. The olivine grain size is small in these rocks, with discrete grains usually not visible to the naked eye.

Chromian spinel occurs throughout the dunite, from disseminated very fine-grained octahedra to massive bands as much as a foot wide in outcrop. Talc is present in several outcrops as a massive alteration product at the edges of the dunite. It also occurs throughout the dunite as veinlet fillings in fractures. Two small abandoned feldspar and corundum prospects are located near the edge of the dunite. The relations of these deposits to the dunite are not known, and were not studied in this investigation.

Petrography

The Holcombe Branch deposit is a partially serpentinized dunite, consisting of the primary minerals olivine, chromian spinel, clinopyroxene, and orthopyroxene. Serpentine, and to a lesser extent, talc and magnesite occur as alteration products. Prior to serpentinization the rock was obviously a dunite which contained about 95 percent olivine, 1-2 percent chromian spinel and minor pyroxene. The degree of alteration of the primary olivine to serpentine varies from 25 to 70 percent throughout the deposit. The distribution of primary minerals and serpentine appears to be random, except for occasional layering of chromian spinel.

Primary Minerals

The observed olivine grain size ranges from less than 0.5 mm to about 1.5 mm, with rare individuals as large as 2.6 mm. Olivine composition, as reported by Carpenter and Phyfer (1975), is $Fo_{92.2}$. Olivine-olivine grain boundaries (disregarding the serpentinization products) are straight, and the grain boundaries generally meet at high angles. This characteristic is especially obvious in the coarse-grained, least altered samples, and is reminiscent of the textures of annealed metals. In the fine-grained samples this texture is less obvious and is commonly obscured by alteration products. The olivine in this rock is generally remarkably free of strain effects. This is consistent with textures observed in other ultramafic rocks in this general area (Astwood, Carpenter, and Sharp, 1972; Carpenter and Phyfer, 1975), and dissimilar to most alpine-type ultramafites from other parts of the world (Raleigh, 1965; Ave'Lallement, 1967, Loney, et al, 1971).

Other primary minerals are present as accessories in varying amount. Chromite, the most common accessory, occurs in all of the samples as euhedra up to about 0.5 mm, usually finely disseminated, but sometimes in "massive" layers up to a few millimeters thick. Orthopyroxene and clinopyroxene are present as rare accessories, generally in the fine-grained specimens. Their grain size is small, around 0.1 mm, except for a few rare orthopyroxenes as large as 1.5 mm. In some places both pyroxenes occur as small grains at the junction of three olivine grain boundaries. In other instances they appear less interstitial and are sometimes as large as the associated olivine grains.

Secondary Minerals

All samples show sea-and-island serpentinization texture in various stages of development. Serpentine minerals have formed at olivine grain boundaries and in fractures in the individual olivine grains. In almost all cases the serpentine is slip-fiber chrysotile or cross-fiber chrysotile; however, in regions where replacement is rather extensive the serpentine does not appear fibrous and may be composed largely of the platy serpentine lizardite. X-ray diffraction analysis of the light mineral fraction ($\rho \approx 2.8$) of ten samples yielded the serpentine mineral assemblage chrysotile + lizardite in every sample. No antigorite was detected.

The amount of serpentine contained in each sample was determined by measuring the bulk density of the rock. Modal analyses were attempted and considered to yield less reliable results, due to the fact that the serpentine was distributed almost entirely along olivine grain boundaries and fractures. Water-loss-upon-heating was also considered to be a less reliable method due to the presence of other hydrous phases in the samples. Cylinders were cut from the samples and their densities calculated from the measured volumes and masses, after the method of Burch (1965) and Page (1966). Since in most of the samples olivine plus the serpentinization products composed at least 98 percent of the sample, the densities of the samples were taken to indicate the relative proportions of forsterite ($\rho \approx 3.3$) and serpentine ($\rho \approx 2.55$). Even a maximum of 2 percent chromian spine ($\rho \approx 4.5$) would not significantly alter the bulk rock density. Therefore the degree of serpentinization can be expressed as the percentage of serpentine in a serpentine + olivine rock required to yield a given density. Porosity of the dunite can be considered negligible (D. T. Secor, personal communication). Values thus obtained indicate that the amount of serpentine in the Holcombe Branch dunite ranges from 23 to 67 percent by weight.

Very fine-grained magnetite is discernible in transmitted and oblique reflected light in the more extensively serpentinized areas. Brucite is present in trace amounts in most samples as flakes up to 0.1 mm long. It was detectable by XRD in some of the light-mineral separates, but technical shortcomings and brucite's characteristically poor scattering of X-rays made non-detection inconclusive.

Talc is present as large flakes and flake aggregates in small veins cutting across all textural features, and is in most cases associated with serpentine and a carbonate. Locally talc appears to replace partially individual olivine grains. The carbonate, probably magnesite, is present in some samples as veinlets along olivine grain boundaries, alone or side-by-side with the serpentine veinlets.

THE SERPENTINIZATION PROCESS

The serpentinization process is essentially the response of high-temperature, generally Mg-rich minerals to a lower temperature regime in the presence of water. The upper stability limits of serpentine resulting from the hydration of olivine and the hydration of the assemblage olivine + talc are known from extensive experimental and thermodynamic work in the system MgO-SiO_2 - H_2O (Bowen and Tuttle, 1949; Johannes, 1968; King et al. 1967; Kitahara and Kennedy, 1967; Kitahara et al, 1968). Relations encountered in the hydration of olivine are summarized in Figure 2. There is still some doubt as to the lower temperature limit at which serpentine can form. Recent work by Wenner and Taylor (1971), based on oxygen isotope fractionation between coexisting serpentine and magnetite, suggests that serpentinization can take place in the range of 85° to 115° C. Barnes et al (1969, 1972) believe that in

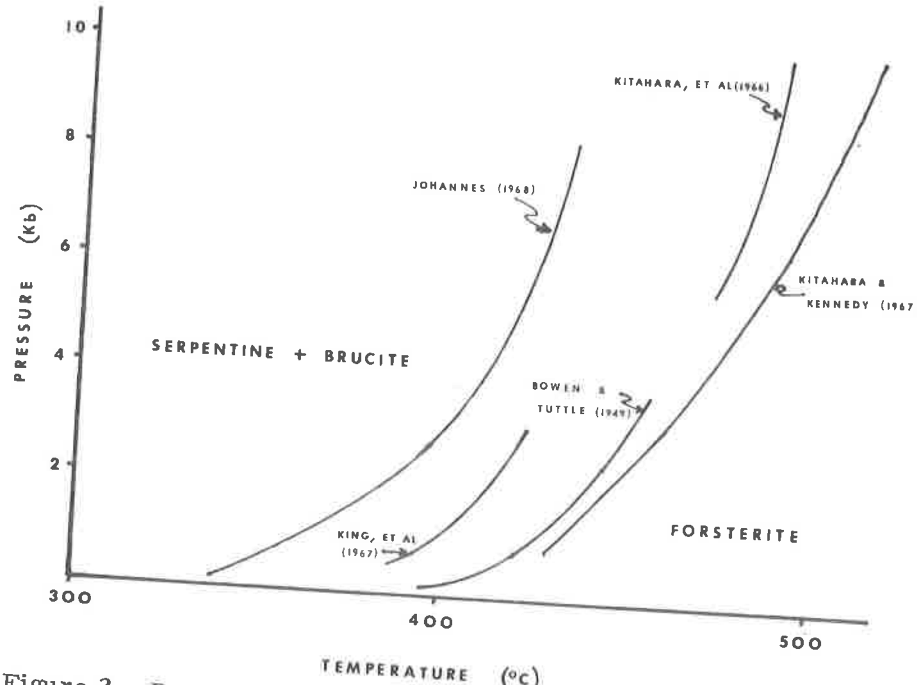

Figure 2. Results of five experimental and thermodynamic investigations of the reaction 2 forsterite + 3 H_2O = 1 serpentine + brucite (after Johannes, 1968).

certain cases serpentine minerals can form under surface temperatures and pressure conditions.

There are three types of simplified reactions by which olivine in ultramafic rocks can be hydrated to form serpentine. One "end-member" family of reactions involves addition of water and removal of cations in aqueous solution from the system without any volume change across the reaction. Another "end-member" family involves addition of water, the removal of aqueous cations, a large positive volume change, and hydration of liberated cations to form another phase besides serpentine. A third "end-member" process requires the addition of silica in aqueous solution, a large positive volume change, but neither removal of cations nor formation of new hydrated cation phases. The case of pure-Mg components is considered first:

(1) Constant volume, cation removal:

$$2Mg_2SiO_4 + 2H_2O \rightarrow Mg_3Si_2O_5(OH)_4 + MgO_{aq}$$

forsterite → serpentine + removed

(2) Constant composition, volume increase:

$$2Mg_2SiO_4 + 3H_2O \rightarrow Mg_3Si_2O_5(OH)_4 + Mg(OH)_2$$

forsterite → serpentine + brucite

(3) Silica addition, volume increase:

$$3Mg_2SiO_4 + 2H_2O + SiO_2 \rightarrow 2Mg_3Si_2O_5(OH)_4$$

If reaction (1) describes the process of serpentinization in a given instance, changes in the whole-rock MgO/SiO_2 ratio are encountered because of the removal of Mg in the aqueous solution. In the case of a pure magnesium olivine the MgO/SiO_2 ratio (weight percent) is 1.34; the completely serpentinized version of that rock, according to reaction (1), would have a MgO/SiO_2 ratio of 1.01.

If the serpentinization process occurs according to reaction (2), no changes in the whole-rock MgO/SiO_2 ratio occur, because MgO is retained and hydrated as brucite. One would expect a constant MgO/SiO_2 ratio (weight percent) of 1.34, independent of the degree of serpentinization.

If reaction (3) describes a given serpentinization process, then changes in the whole-rock MgO/SiO_2 ratio (weight percent) from 1.34 to 1.01 would occur, depending on the degree of serpentinization. In this case, however, the change is due to silica addition rather than MgO removal.

One can probably infer that in nature any combination of the three reactions could occur, and one or more might dominate in a given situation, on any scale. Most workers now agree on this idea.

Addition of iron to the forsterite in reactions (1) through (3) complicates matters somewhat, but more closely approximates the natural situation. Reactions (1) through (3) can again be considered, but now the products are slightly iron-bearing serpentine and brucite, and magnetite. The situation is complicated by the fact that magnetite production requires oxidation of some Fe^{+2} to Fe^{+3}, and therefore O_2 must be considered as an additional reactant. Further, there is evidence that magnetite production increases with increasing f_{O_2} during the later stages of serpentinization (Page, 1966). The general reactions considered here are:

(4) olivine + water + O_2 → serpentine + magnetite + MgO_{aq}

(5) olivine + water + O_2 → serpentine + brucite + magnetite

(6) olivine + water + silica → serpentine + magnetite

These are the iron-bearing equivalents of reactions (1) through (3).

According to Coleman and Keith (1971), a dunite with an olivine composition of Fo_{90} (close to the composition of the Holcombe Branch olivine) reacting with water would yield 83 percent serpentine, 17 percent brucite, and less than 1 percent magnetite. It is assumed in this study that little or no iron was removed from the dunite in aqueous solution, because of the presence of magnetite in all the samples.

The greatest $(MgO + FeO)/SiO_2$ ratio (weight percent, total Fe as FeO) that would be encountered in the "isochemical" serpentinization of Fo_{90}-olivine (reaction 5) is 1.45. The lowest possible $(MgO + FeO)/SiO_2$ ratio would be approximately 1.21, the situation in which Fo_{90}-olivine is changed to serpentine and magnetite, with removal of excess MgO (reaction 4) or addition of SiO_2 (reaction 6). Inasmuch as

Table 1. Comparison of the data used in the least squares regression analyses. Oxide ratios which exceed the theoretical maxima and minima for processes discussed in the text probably reflect the presence of minerals, such as brucite, talc, or carbonate, in larger amounts than were apparent by thin-section examination (see text).

Sample	Measured Density	Calcul'd % Serp'n (see text)	Distance from HB-1	MgO/SiO_2 (wt pct)	FeO/SiO_2 (wt pct)	$\frac{MgO+FeO}{SiO_2}$ (wt pct)
HB-1	2.97	44	0	1.31	0.196	1.506
HB-2	3.03	36	53	1.26	0.198	1.458
HB-3	3.02	37	371	1.19	0.214	1.404
HB-4	2.91	52	379	1.07	0.207	1.277
HB-5	3.06	32	390	1.22	0.206	1.426
HB-6	3.13	23	403	1.26	0.201	1.461
HB-7	3.11	25	422	1.19	0.200	1.390
HB-8	3.12	24	437	1.23	0.203	1.433
HB-13	3.00	40	722	1.16	0.186	1.346
HB-14	3.13	23	764	1.12	0.194	1.314
HB-15	2.96	45	1037	1.07	0.180	1.250
HB-16	2.83	63	1054	1.04	0.179	1.219
HB-21	2.96	45	1182	1.16	0.193	1.353
HB-22	3.00	40	1194	1.31	0.189	1.499
HB-23	3.00	40	1199	0.95	0.190	1.140
HB-25	2.97	44	1226	1.07	0.194	1.264
HB-26	2.86	59	1250	1.04	0.189	1.229
HB-27	2.89	55	1266	0.97	0.185	1.155
HB-28	2.95	47	1284	1.32	0.207	1.527
HB-29	2.95	47	1300	1.30	0.176	1.476
HB-30	2.97	44	1305	1.28	0.192	1.472
HB-31	2.84	61	1315	1.12	0.192	1.312
HB-33	2.97	44	1468	1.28	0.188	1.468
HB-34	2.99	41	1491	1.28	0.187	1.467
HB-35	3.00	40	1501	1.42	0.195	1.615
HB-36	3.09	28	1544	1.35	0.198	1.548
HB-41	3.07	31	2111	1.38	0.197	1.577
BT-1	2.83	63	---*	1.30	0.187	1.487
BT-3	2.80	67	---*	1.24	0.208	1.448
BT-4	2.87	57	---*	1.24	0.200	1.440

*Samples from largest tributary of Holcombe Branch; distance from HB-1 not measured.

fO_2 increases as serpentinization progresses, producing increasingly more magnetite, a direct relationship between the amount of Fe^{+3} and degree of serpentinization would be expected. However, no determinations of Fe^{+3} were made as a part of this study.

RESULTS

To obtain information on chemical variation with degree of serpentinization, emission spectroscopy was used to obtain MgO/SiO_2,

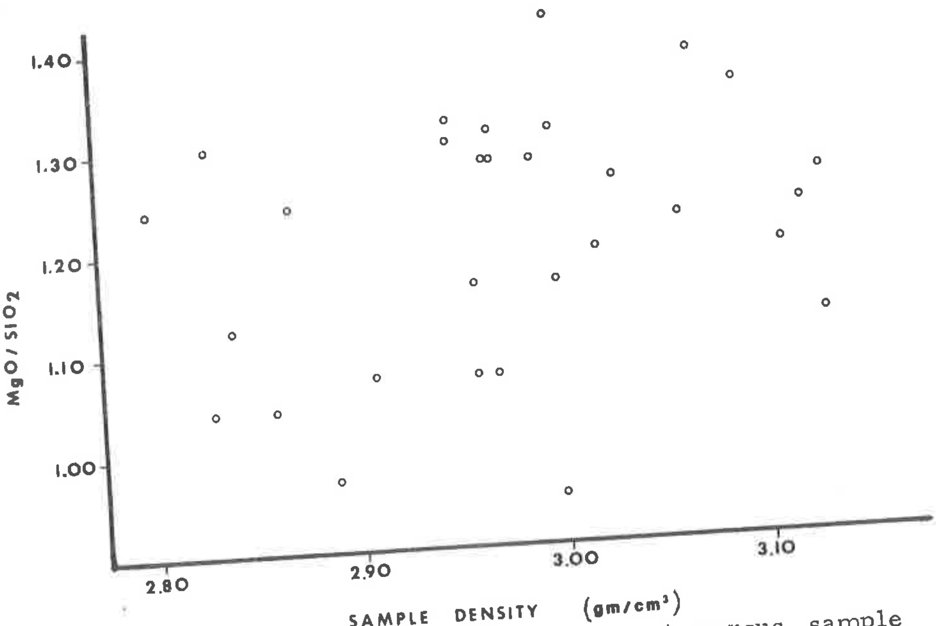

Figure 3. Plot of MgO/SiO_2 in weight percent versus sample density, showing the lack of a recognizable trend between variation of MgO with respect to SiO_2 and extent of serpentinization of the rocks.

FeO/SiO_2, and $(MgO + FeO)/SiO_2$ (weight percent, total Fe as FeO) for each of the thirty samples (Table 1). Each sample was run in triplicate or quadruplicate and the results averaged. These oxide ratios were plotted as a function of density, or degree of serpentinization, in an attempt to deduce which of reactions (1) through (6) predominated during serpentinization of the Holcombe Branch dunite (Figures 3-5). The plots show no obvious trend. Least squares regression analyses of the three ratios as functions of density were run for the five general equations below:

(a) $y = ax + b$

(b) $y = c + d\log x$

(c) $y = ef^x$

(d) $y = gx^h$

(e) $y = i + jx + kx^2$

Correlations were low enough that the relationships between rock chemistry and degree of serpentinization must be considered random (Table 2).

In order to test whether degree of serpentinization is related to the proximity of the dunite-country rock contact, bulk rock densities were plotted against distance from one edge along Holcombe Branch (Figure 6). No systematic relationships between degree of serpentinization and proximity to the edges of the dunite were noted. Regression

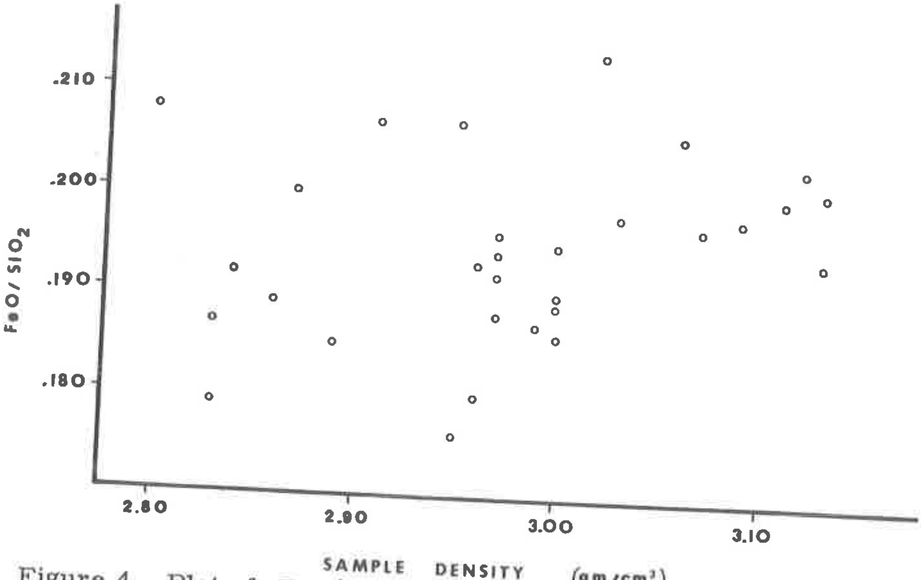

Figure 4. Plot of FeO/SiO_2 in weight percent versus sample density, showing the lack of a recognizable trend between variation of FeO with respect to SiO_2 and extent of serpentinization of the rocks.

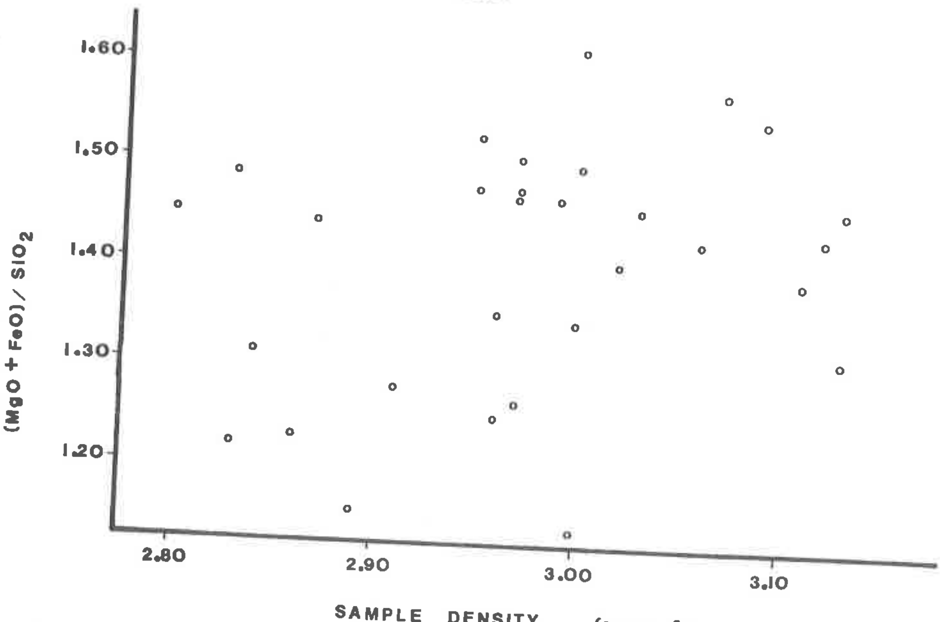

Figure 5. Plot of $(MgO + FeO)/SiO_2$ in weight percent versus sample density, showing the lack of a recognizable trend between variation of the sum of MgO and FeO with respect to SiO_2 and extent of serpentinization of the rocks.

Table 2. Correlation coefficients of data comparisons calculated from least squares regression analyses for five modal equations. Note uniformly low values. Symbols: ρ = density of sample: d = distance along Holcombe Branch from first sample (HB-1); MgO/SiO_2, etc., are ratios of certain oxides in each sample. See text for further explanation.

$y=f(x)$	no. of samples	$y=ax+b$	$y=c+d\log x$	$y=ef^x$	$y=gx^h$	$y=i+jx+kx^2$ *
$\rho=f(d)$	27	.27	.12	.27	.12	----
$MgO/SiO_2=f(\rho)$	30	.28	.28	.28	.28	----
$FeO/SiO_2=f(\rho)$	30	.28	.27	.28	.28	----
$(MgO+FeO)/SiO_2=f(\rho)$	30	.29	.30	.30	.30	----

*Correlation coefficients not calculated for $y=i+jx+kx^2$. However, visual inspection of plotted regression equations suggests poor correlation with data.

analyses were done for the same five general equations referred to above, and all correlations were poor (Table 2). It must be concluded that, based on the methods employed, the pattern of serpentinization in the dunite is random. The presence of several percent of talc and carbonate in a few of the samples moderated the measured density, which yields an artificially high percentage of olivine on calculation, but these few inaccuracies tended to reduce rather than exaggerate the scatter of the data.

CONCLUSIONS

Differences in the oxide ratios reported in Table 1 seem to argue clearly against a simple constant composition model for the serpentinization process in the Holcombe Branch dunite. The lack of correlation between oxide ratios and degree of serpentinization, moreover, tends to negate the simple constant-volume model of reaction (1) and, especially, reaction (4). Three conclusions might be drawn from this:

(1) the data are not valid.

(2) the simple models presented in reactions (1) through (6) are inadequate and must be either discarded or modified.

(3) more than one chemical process has acted on the parent ultramafic mass.

Conclusion (1), we believe, must be rejected. The duplication in the chemical analyses in this study argues that both the accuracy and the precision of the spectroscopy data are better than the scatter reported here.

Conclusion (2) also has to be rejected on the grounds that these models seem to work for other serpentinized ultramafic rocks (see, for example, Coleman and Keith, 1971, and their study of the Burro

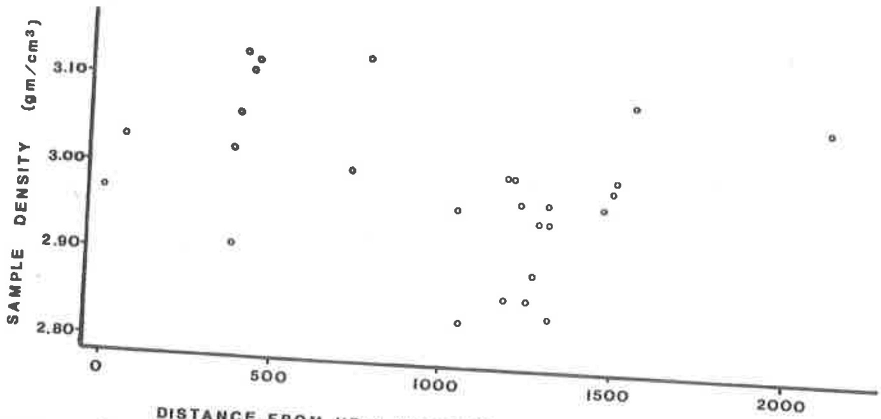

Figure 6. Plot of sample density versus distance along Holcombe Branch showing the lack of a recognizable trend between the degree of serpentinization of the rocks and position within the dunite body.

Mountain peridotite).

Thus we are left with only one possible conclusion, that more than one chemical process has been acting on the Holcombe Branch dunite. We cannot ascertain if the different processes were acting on different parts of the body at the same time, or on the entire body at different times, or on different parts of the body at different times.

Other structural, mineralogical and chemical support the conclusion that multiple processes were acting on the body. A detailed examination of the foliations of the county rock immediately adjacent to the dunite was not possible due to poor exposure quality, however no obvious major distortions in the foliations were observed. If the simple constant composition (equation 2) or silica addition models had predominated during the serpentinization process, one might expect a large volume increase in the serpentinized dunite and a concomitant distortion of the county rocks. For the "pure Mg" system, either of the models mentioned above would have resulted in 50-70 percent volume increase had the serpentinization proceeded to completion. The overall degree of serpentinization of the Holcombe Branch body is about 50 percent therefore one would expect roughly an overall 25-35 percent volume increase. Intuitively it seems that this degree of volume expansion would have resulted in visible or measurable distortions in the country rock. Also, there is no evidence of widespread MgO or SiO_2 metasomatic activity (with the exception of the talc zone at the dunite-siliceous country rock contact), which is implied by the application of the simple constant volume model. Finally, the spotty distribution of brucite among the samples suggests that the simple models requiring either the formation or the non-formation of brucite are inappropriate.

The lack of any correlation between the degree of serpentinization and proximity to the edges of the dunite body argues against a simple model of incomplete hydration working its way in toward the center of an ultramafic body from the edges in contact with the enclosing country rocks. It seems reasonable to suspect that the serpentinization process might have been governed by some pre-serpentine fracture pattern in the dunite, although this possibility can not be demonstrated conclusively. Earlier it was mentioned that on a thin section scale, serpentinization occurred along olivine grain boundaries and along fractures. Perhaps this might serve as evidence to support a "fracture controlled serpentinization" model on an outcrop scale.

The conclusions thus drawn differ from those made concerning the serpentinization of the Burro Mountain peridotite (Coleman and Keith, 1971). We believe, however, that the conclusions presented here are valid and that the differences between the processes that affected Holcombe Branch and Burro Mountain are real. This simply, once again, points up the observation made by many workers that the serpentinization processes are many and complex and differ from body to body, depending on prevailing local conditions.

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U. S. NATIONAL COMMITTEE FOR INQUA ANNOUNCES TRAVEL SUPPORT PROGRAM FOR TENTH INQUA CONGRESS IN ENGLAND

The U.S. National Committee of the Internation Union for Quaternary Research (INQUA) is seeking funding for a travel support program to ensure that the United States will be represented by an adequate number of qualified scientists at the X International Congress of INQUA, to meet in Birmingham, England, August 16-24, 1977. Funds for this purpose, now being solicited from a number of government agencies and private institutions, will be coordinated by the U. S.

Applicants for travel grant support should request application forms from Mr. W. L. Petrie, USNC/INQUA, National Academy of Sciences, 2101 Constitution Avenue, N. W., Washington, D. C. 20418. Four completed application forms, together with four copies of the abstract of the paper submitted to INQUA must be received by the Academy Office no later than December 1, 1976. Grant awards may be made as late as August 1, 1977, depending on funds received. If possible, some advance indication of tentative selections will be communicated by April 1, 1977.

The purpose of the International Union for Quaternary Research (INQUA) is to bring together on a world-wide basis scientists in all disciplines concerned with the history of man's environment, and with the processes by which environment and man's relation to it have evolved. Included among these disciplines are: archaeology, botany, climatology, ecology, geochemistry, geography, geomorphology, geophysics, hydrology, paleontology, limnology, oceanography, palynology, physical anthropology, soil science, tectonophysics and zoology.

The National Academy of Sciences is the adhering body to INQUA on behalf of American scientists. The U.S. National Committee, under the chairmanship of Dr. R. Goldthwait, Ohio State University, plans U. S. participation in INQUA activities. One of the functions of the Committee is to arrange for travel support of U. S. scientists attending the international congresses of INQUA held at four-year intervals. Further information about the X Congress may be obtained by writing to Dr. W. G. Jardine, Secretary-General X INQUA Congress, Department of Geology, University of Glasgow, Glasgow G12 8QQ, United Kingdom.

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