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SOUTHEASTERN GEOLOGY

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THE CONTINENTAL MARGIN SOUTH OF CAPE HATTERAS, NORTH CAROLINA: SHALLOW STRUCTURE*

by

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ABSTRACT

Continuous seismic profiles, drill hole data and dredge hauls show that the geologic history of the continental margin south of Cape Hatteras has been complex, marked by carbonate deposition, erosion by the Gulf Stream, and differential subsidence. The Blake Escarpment east of the Blake Plateau is believed to represent a chain of algal banks that flourished during the Cretaceous. A thick sequence of shallow-water carbonate sediments accumulated on the Blake Plateau behind the algal sandbanks to the east. Soon after the banks died during Late Cretaceous submergence of the area the Gulf Stream extended its course across the Blake Plateau. During the Tertiary, the locus of deposition shifted westward to the area near the present shelf-break, where 600 to 1000 meters of shallow-water sediments were deposited. South of Latitude 32° sediment prograded against the westward margin of the Gulf Stream throughout the Tertiary. Farther north, where the Gulf Stream lies farther offshore, outbuilding extended beyond the eastern margin of the Blake Plateau. Vertical uplifts followed by erosion by the Gulf Stream during the Pleistocene modified this sedimentary framework.

INTRODUCTION

Within the past few years continuous seismic profilers have been used extensively on the continental margin off the east coast of the United States in order to determine its structure. Using such devices Moore and Curray (1963), Uchupi and Emery (1967), and Hoskins (in press) were able to show that most of the continental margin was formed by a combination of upbuilding on the shelf and progradation on the continental slope. Seismic surveys of the Gulf of Maine by Uchupi (1966b) revealed eroded pre-Triassic, Triassic, Cretaceous, and Tertiary rocks covered by a thin veneer of Pleistocene glacial sediments. Profiler recordings from Georges Bank complemented by bottom samples

*Contribution no. 1918 of the Woods Hole Oceanographic Institution.

indicate that the bank has a core of gently seaward dipping Cretaceous and Tertiary strata that are partially truncated by the continental slope south of the bank (Emery and Uchupi, 1965).

Low energy, high resolution seismic profilers have also been used to determine the shallow structure of Cape Cod Bay (Hoskins and Knott, 1961), the southern margin of Georges Bank (Roberson, 1964), the shelf off New York (Ewing, Le Pichon, and Ewing, 1963; Ewing, Luskin, Roberts, and Hirshman, 1960; Knott, Hoskins and Weller, 1963; and Knott and Uchupi, 1966), and the Straits of Florida (Jordan, Malloy, and Kofoed, 1964; and Malloy and Hurley, 1966). Seismic recordings clearly show that the present side slopes of the Straits of Florida are depositional and erosional, not structural in origin (Rona and Clay, 1966; Uchupi, 1966a). Seismic reflection profiles, sediment cores, and seismic refraction data were used by Ewing, Ewing, and Leyden (1966) in a study of the structural relations of the Florida peninsula and the Blake Plateau.

The present study describes the results of a seismic profiler survey of the continental margin between Cape Hatteras, North Carolina, and Cape Kennedy, Florida. Information from these recordings, supplemented by data from bottom samples and from bore holes drilled on the margin, makes possible the reconstruction of the geologic history of this segment of the continental margin.

Acknowledgments

Since 1962 the U. S. Geological Survey in cooperation with the Woods Hole Oceanographic Institution has been investigating the sediments, stratigraphy, and structure of the continental margin off the east coast of the United States. All but two of the recordings described in this report were obtained during this investigation. The writer wishest to thank the officers and crew of the R/V GOSNOLD, and A. R. Tagg, and R. K. Paul for the maintenance of the electronic equipment, and for standing watches during the seismic cruises. For the loan of profiles TW8 and TW9 the writer wishes to thank J. R. Curray of Scripps Inst. Oceanography. Appreciation is also extended to John Schlee and K. O. Emery for critical reading of the manuscript.

TOPOGRAPHIC SETTING

The continental margin between Cape Hatteras and Cape Kennedy consists of the following provinces: (1) continental shelf, (2) Florida-Hatteras Slope, (3) Blake Plateau, (4) continental slope, (5) continental rise, and (6) Blake Basin and Blake Ridge (Figure 1).

The continental shelf has a gradient of less than 0° 10' (3m/km) and ranges in width from 130 km south of Charleston to less than 10 km off southern Florida. Depths along the shelf-break from Cape Kennedy

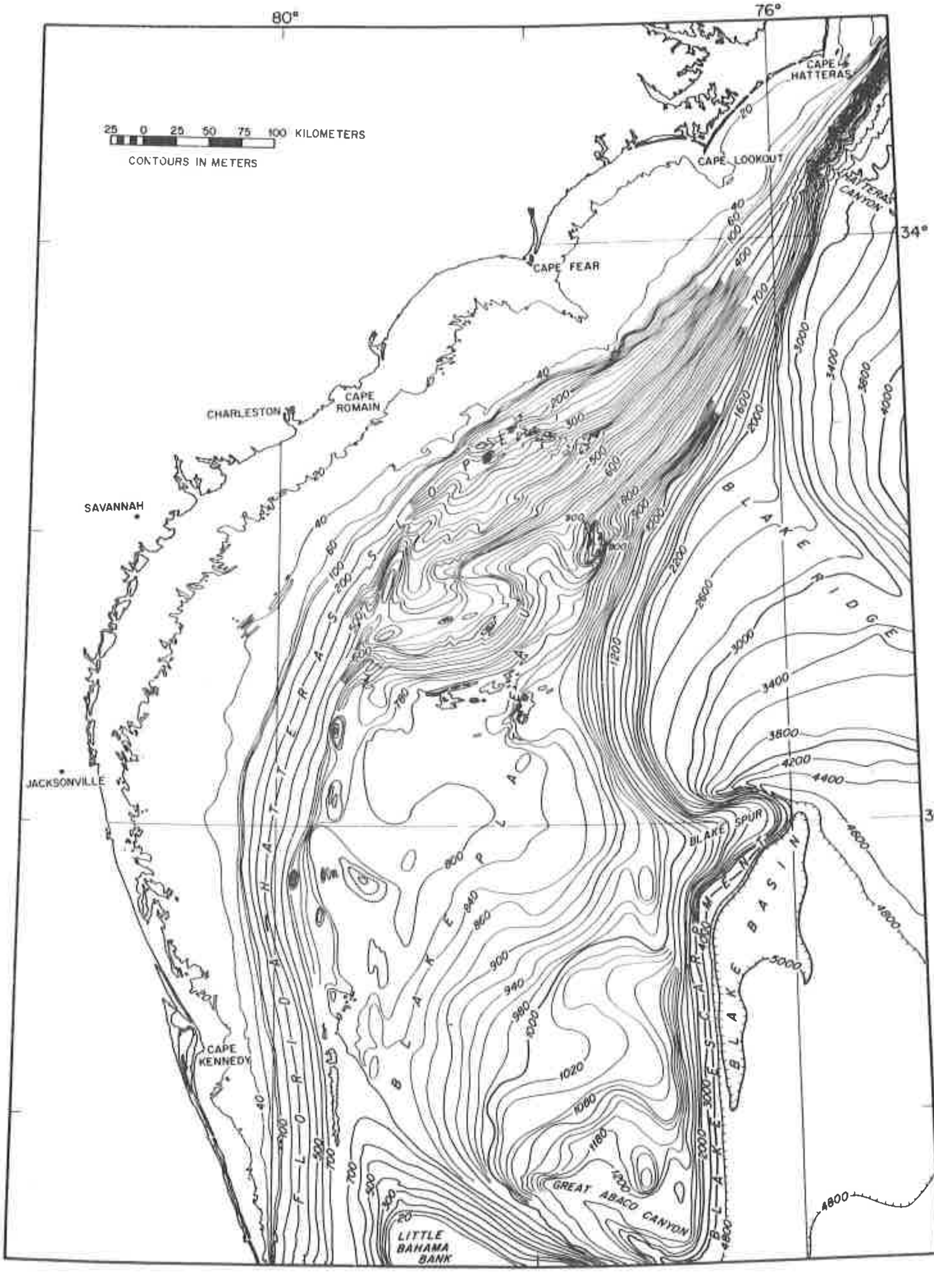

Figure 1. Bathymetry of the continental margin between Cape Hatteras, North Carolina and Cape Kennedy, Florida. Based on sounding from the U. S. Coast and Geodetic Survey hydrographic surveys, and from a chart by Pratt and Heezen (1964, Fig. 1).

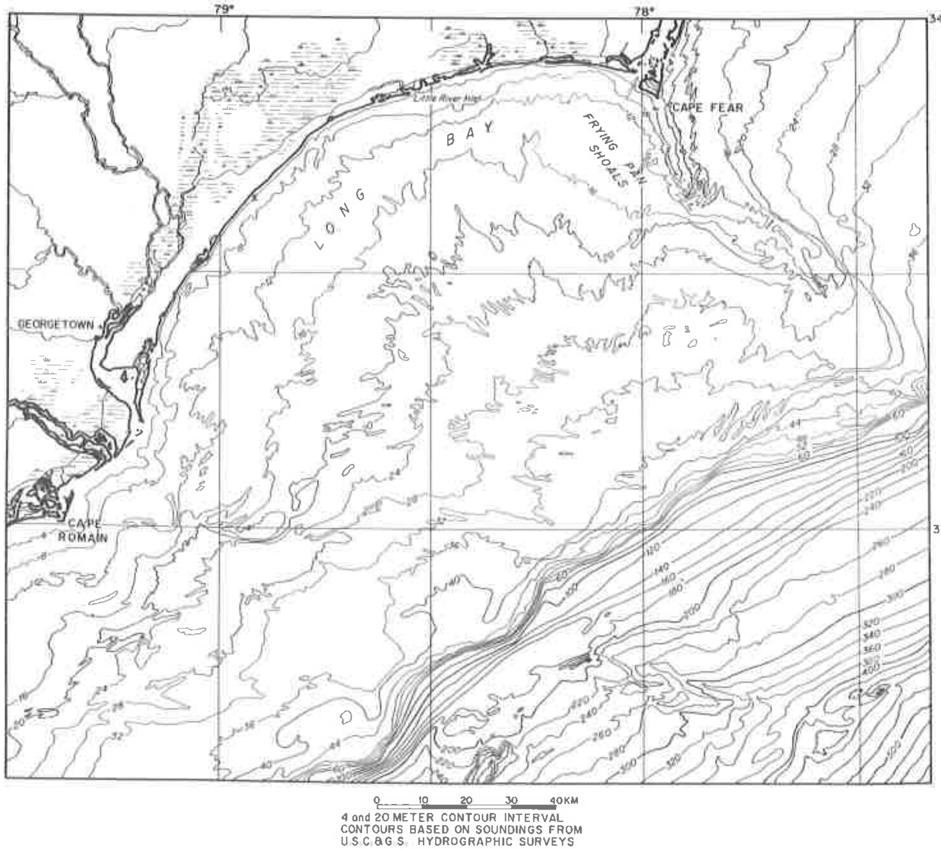

Figure 2. Bathymetry of a segment of the continental margin between Cape Fear and Cape Romain. Based on soundings from the U. S. Coast and Geodetic Survey hydrographic surveys.

to Cape Romain are between 70 and 50 meters, shoaling to 40 meters seaward of Cape Hatteras and 9 meters off southern Florida (Uchupi and Tagg, 1966). Topographically the shelf's surface is quite irregular, consisting of cuspate projections off Capes Kennedy, Fear, Lookout, and Hatteras, sand ridges generally aligned at right angles to the present shoreline (Figure 2), algal ridges and banks, and terrace like features. These topographic irregularities are believed to have formed in response to eustatic lowering and rising of sea level during the late Pleistocene Epoch (Uchupi, in press). Probably the most striking of these relict features is a ridge near the shelf-break that extends discontinuously from Cape Hatteras to southern Florida. The best surveyed segment of this ridge is that between Cape Fear and Cape Lookout. Here the ridge is at a depth of 80 to 100 meters and is located along the western margin of the Gulf Stream (Menzies and others, 1966). The ridge appears to have been formed by the growth of Lithothamnion during a lower stand of sea level 19,000 years ago.

Seaward of the shelf and separating it from the Blake Plateau is the Florida-Hatteras Slope, a relatively smooth slope with a gradient that rarely exceeds 1° (7m/km). Its relief ranges from 700 meters east of Cape Kennedy to less than 10 meters off Cape Lookout. This slope extends only a few kilometers northeast of Cape Lookout before it blends with the Blake Plateau.

The most prominent province of the continental margin south of Cape Hatteras is the Blake Plateau, a 228,000 km^2 platform with an average depth of 850 meters (Pratt and Heezen, 1964). It forms an intermediate surface between the oceanic basin to the east and the Bahama Banks and the continental shelf to the south and west.

North of Latitude 32° the Blake Plateau is relatively smooth, dips seaward with a gradient greater than 0° 30' (>9m/km), and forms a transitional zone between the continental slope to the north and the broad and flat plateau proper to the south. The plateau is widest south of Latitude 30°. This segment of the plateau is relatively smooth, except for a rough zone near the base of the Florida-Hatteras Slope. This irregular zone extends from the Straits of Florida to Cape Romain. Between Latitudes 30° and 32° the jagged topography extends across the plateau. Within the area of rough topography are large, broad and generally flat-bottomed hollows with reliefs in excess of 40 meters. Around the margins of these lows and occasionally within them are innumerable conical hills that add to the relief of the depressions and accentuate the roughness of the topography. Bottom samples and bottom photographs indicate that the conical hills are coral mounds (Stetson, Squires, and Pratt, 1962).

The line of depressions along the western margin of the Blake Plateau coincides with the axis of maximum current of the Gulf Stream. Consequently, Pratt (1966) and Pratt and Heezen (1964) have suggested that hollows are erosional in origin and were formed by the Gulf Stream. Seismic profiles of this survey and those of Ewing, Ewing, and Leyden (1966) clearly show that the hollows are erosional, not structural in origin. If the lows along the base of the Florida-Hatteras Slope delineate the present position of the axis of maximum current of the Gulf Stream, those running east-west along Latitude 31°, and the ones beneath the Florida-Hatteras Slope as revealed by the seismic profiles probably indicate former positions of the stream's axis (Figure 3). As the depressions have reliefs in excess of 40 meters, the Gulf Stream must have occupied each of these areas for a considerable length of time.

The continental slope east of Cape Hatteras has a gradient of about 5° (87m/km), is 2000 meters high, and is cut by innumerable small gullies. The gullies coalesce toward the bottom of the slope to form Hatteras Canyon (Figure 1). Near Latitude 30° the slope swings northeasterly to form a promontory which Uchupi (in press) has named the

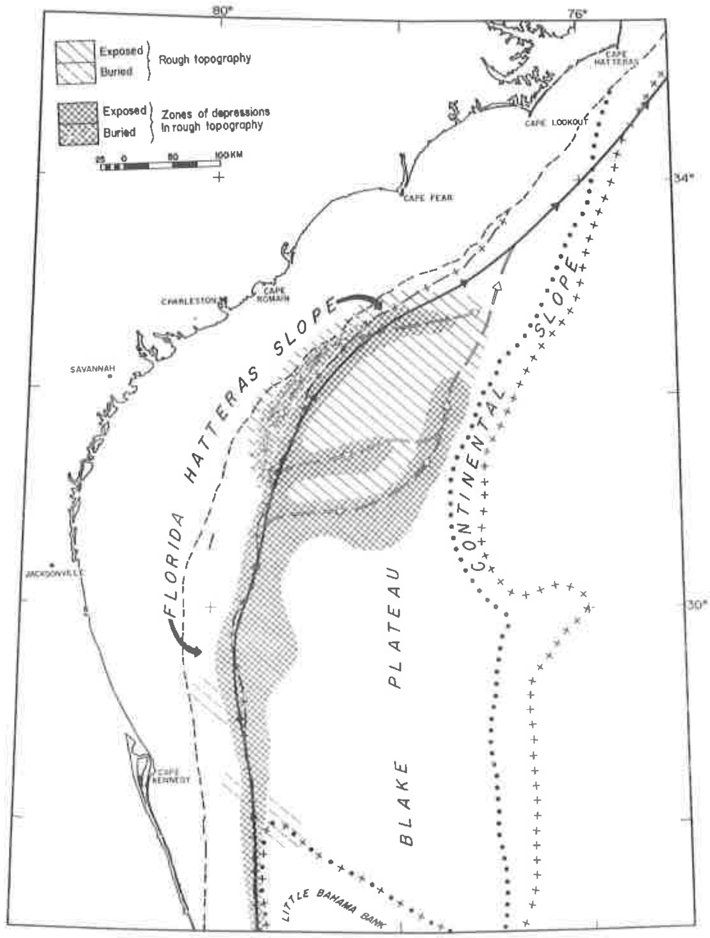

Figure 3. Present and former positions of the Gulf Stream axis of maximum current. Solid arrows indicate its present position; open arrows previous positions.

Blake Spur; this feature has been interpreted by Ewing, Ewing, and Leyden (1966) as a slump block. South of the Blake Spur is the Blake Escarpment with a gradient greater than 10° (>176m/km). The escarpment is the steepest segment of the continental slope off the east coast of the United States.

East of Cape Hatteras the continental slope is flanked on the seaward side by the continental rise, a broad sedimentary apron. This apron pinches out toward the southwest and is replaced by the Blake Ridge—a broad sedimentary ridge, and Blake Basin—a flat-bottomed

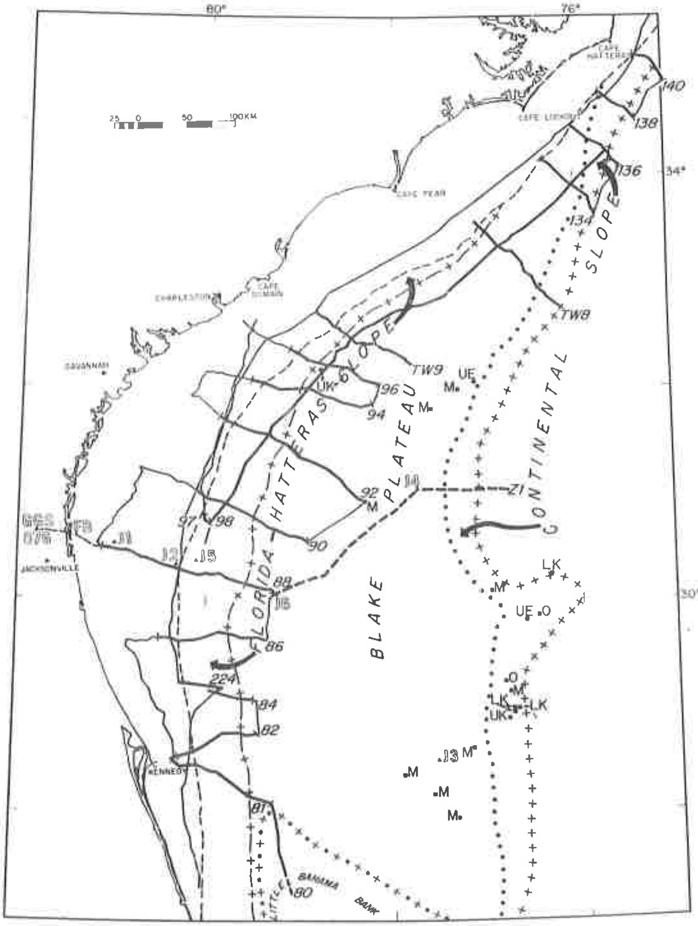

Figure 4. Location of seismic profiles and positions of rock samples and holes drilled by JOIDES (Bunce and others, 1965). Ages of rock samples are as follows: LK-Lower Cretaceous; UK-Upper Cretaceous; UE-upper Eocene; O-Oligocene; M-Miocene. J1-J6 JOIDES drill holes. Ages of most rock samples are from Ericson, Ewing, and Heezen, (1952), Ericson, Ewing, Wollin, and Heezen (1961), and Heezen and Sheridan (1966). The Upper Cretaceous rocks southwest of Cape Fear were dredged by R. M. Pratt.

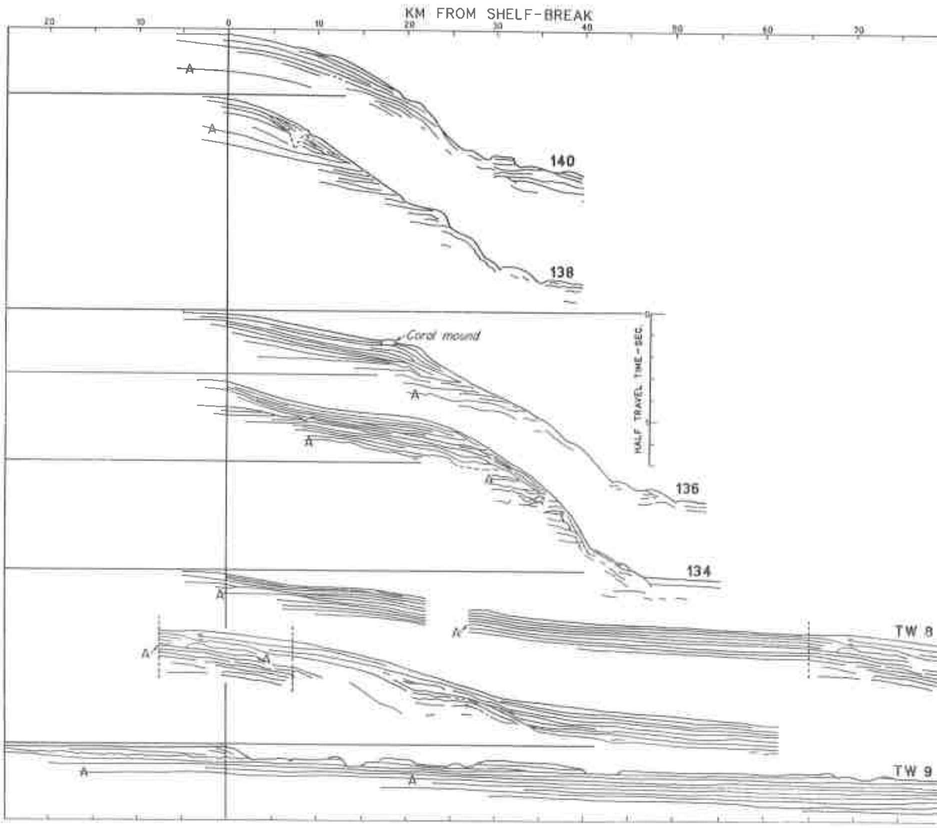

Figure 5. Continuous seismic profiles of the continental margin between Cape Hatteras and Latitude 32°. Modified from Uchupi and Emery (1967). Reflector A marks top of Cretaceous.

enclosed depression with a maximum depth of 5046 meters (Pratt and Heezen, 1964).

STRUCTURE

Methods

The profiles recorded during this study were obtained with a 10,500 joule sparker source programmed to discharge at 5.0 second intervals, or at an interval of 12.5 meters when the ship moves at a speed of 9 km/hour. Echoes from the bottom and sub- bottom were detected by a 5-hydrophone array towed 20 meters behind the ship, and recorded aboard ship on a Precision Graphic Recorder (PGR). Locations of the seismic profiles are shown in Figure 4. The sections shown in Figures 5 and 6 were drawn from the recordings obtained in the field.

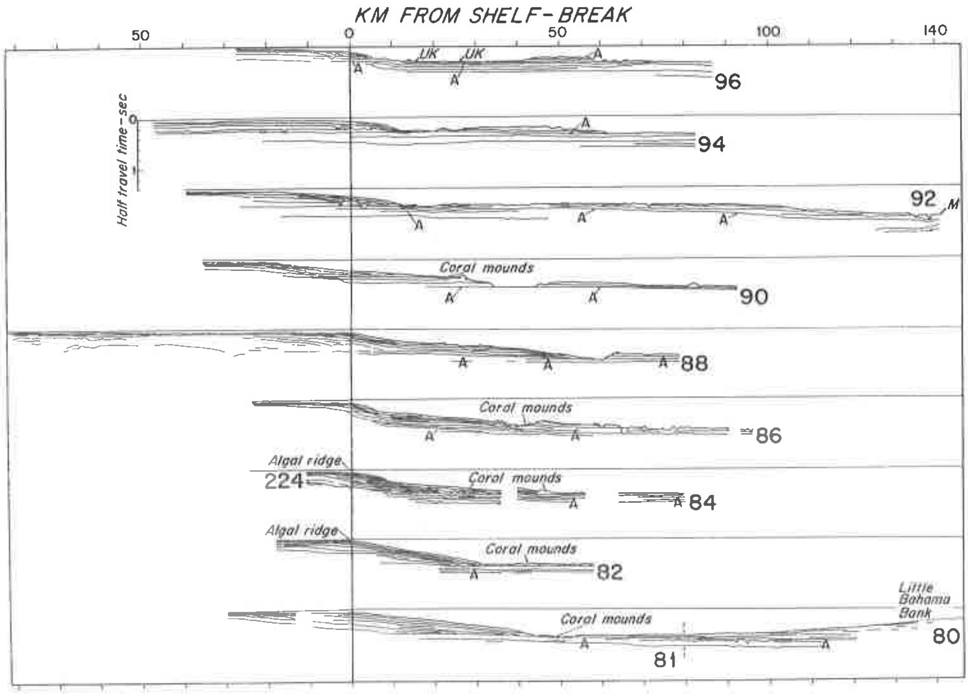

Figure 6. Continuous seismic profiles of the continental margin between Latitude 32° and Cape Kennedy. Modified from Uchupi and Emery (1967). Reflector A marks top of Cretaceous.

Description of Profiles

The seismic profiles in Figures 5, 6, 7, and 8 show in detail the sub-surface morphology of the continental margin south of Cape Hatteras. In profile 140 off Cape Hatteras the reflecting horizons generally parallel the present surface of the shelf, and continental slope, although they have undergone some modification by erosion. Along 138, discontinuities are truncated midway down the slope. A buried canyon with a relief in excess of 400 meters occurs within the shallow reflectors in profile 138.

The remainder of the profiles extend across the shelf onto the Blake Plateau, and four of them cross the plateau to the base of the continental slope. Profiles 136 and 134 are in the transitional zone between the broad plateau to the south and the continental slope to the north. In both of these profiles as in 138 the deeper discontinuities are truncated by the continental slope. The shallower layers have prograded over these reflectors the width of the plateau, a distance ranging from 20 km in profile 136 to more than 100 km in TW8. The hill located at the top of the continental slope in profile 136 (Figure 5) is believed to be a coral mound.

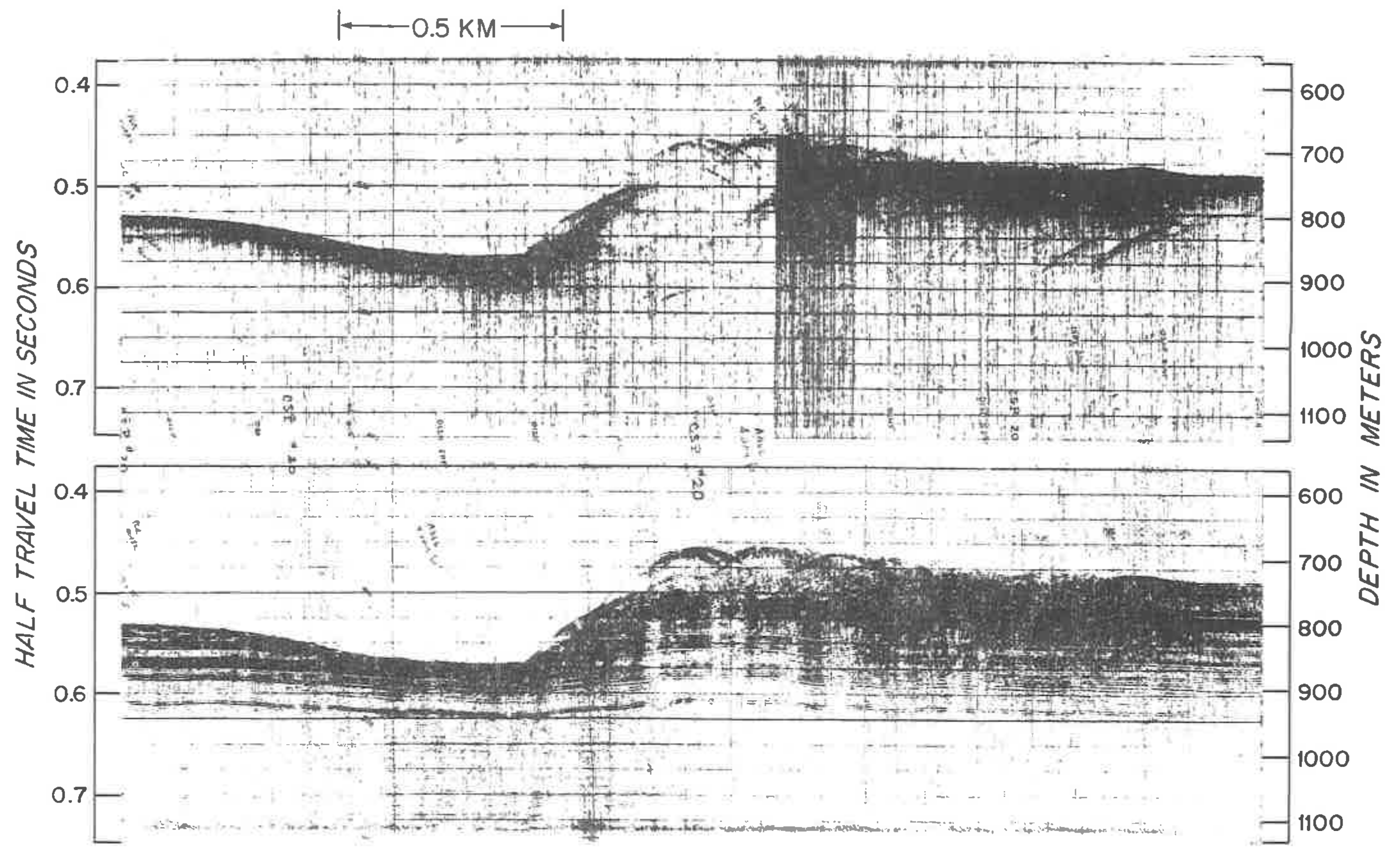

Figure 7. Seismic profile across one of the depressions on the Blake Plateau. Position: Latitude 31°50', Longitude 77°30'. The top panel is recorded at a frequency of 9600-2400 cycles per second. Lower panel at 225-67.5 cycles per second. Photograph courtesy of T. R. Stetson of Woods Hole Oceanographic Institute.

The erosional origin of the rough topography of the Blake Plateau is clearly seen in profile TW9, and is even more evident in Figure 7. The latter profile, made with a 1,000-joule Edgerton Sonar Thumper, shows the structure within the top 100-200 meters of sediment. As indicated by this profile the deepest reflector extends across the depression without any disruption. The irregularities along the periphery of the depression are the coral mounds mentioned previously. These coral mounds also occur in profiles 86, 84, and 224 farther south. Along 224 sediment outbuilding toward the east has buried three of them. In profiles TW9, 96, and 92 the Blake Plateau's rough surface extends beneath the shelf and Florida-Hatteras Slope for a distance of 20 km. In profiles 94 and 92 two deep hollows, similar to these along the western margin of the Blake Plateau, occur beneath the slope. If these depressions were formed beneath the axis of maximum current of the Gulf Stream, they indicate that the axis formerly was west of its present position. After the hollows were cut, the stream appears to have migrated eastward, and progradation of the shelf and slope buried the depressions and irregular surface. In profiles 90, 88, and 86 the Florida-Hatteras Slope has undergone considerable erosion, but seaward progradation after the erosional cycle has been negligible.

In contrast to the profiles farther north, there is little evidence of erosion of the Florida-Hatteras Slope along 82 and 81-80. Only the reflecting horizons on the Blake Plateau reveal any evidence of erosion. Progradation along these profiles has been at least 35 km, and along 81-80, deposition from Little Bahama Bank has been as great as that from the mainland. In profile 80, the reflector above horizon A has undergone considerable erosion. The topography of this reflector is similar to that of the present surface of the Blake Plateau and it too may be a result of erosion by the Gulf Stream. The age of this reflector was previously suggested to mark the top of the Cretaceous strata (Uchupi, 1966a), but it probably delineates the top of the Paleocene. If erosion of this reflector is due to the Gulf Stream, then the stream established its present course and invaded the Blake Plateau during late Paleocene or early Eocene times.

Profile 98 (Figure 8) parallels the Florida-Hatteras Slope, and extends across the rough zone of the Blake Plateau between Latitudes 31° and 32°, and the transitional zone to the continental slope off Cape Lookout. Shallow layers in the rough zone have undergone considerable erosion, so much so that Cretaceous sediments have been exposed. The surface reflectors also appear to be slightly warped with the axis of the structural high located near the 300 km mark. This structural ridge is probably the seaward extension of the Cape Fear Arch.

The profile in the lower panel in Figure 8 is a composit of profiles 88 and Z-1. It extends along the line of JOIDES drill holes; 88 from hole 1 to hole 6, and Z-1 from hole 6 to 4 and from there eastward to the base of the continental slope. In Figure 9 taken from a report by Emery and Zarudzki (in press) the reflecting horizons of the composite

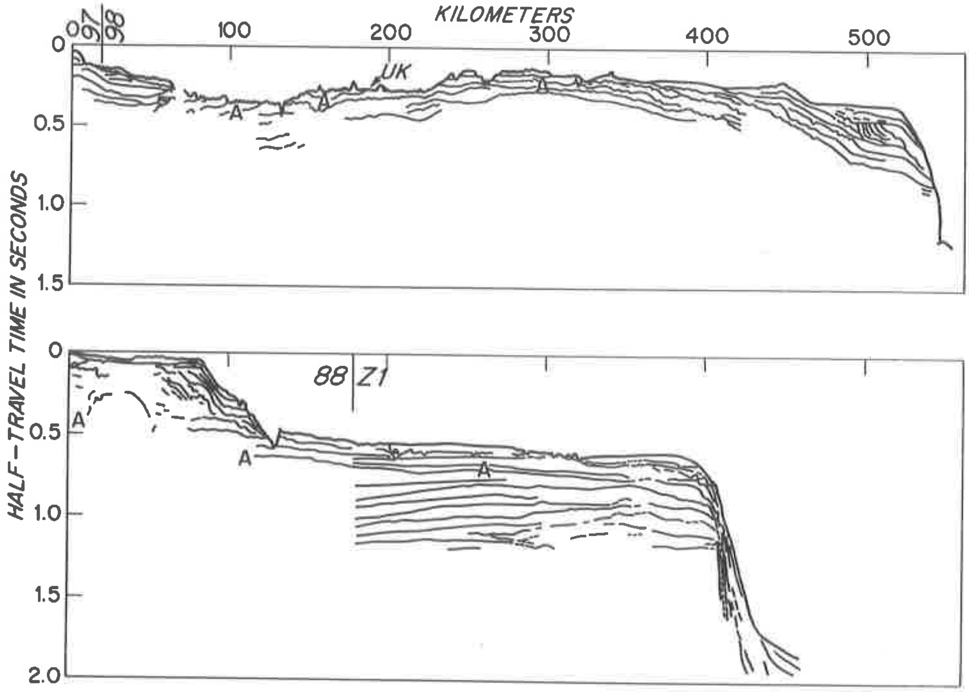

Figure 8. Seismic profiles at the base of the Florida-Hatteras Slope, and along the line of JOIDES holes. The figure in the lower panel is a composite of two figures from a report by Emery and Zarudzki (in press).

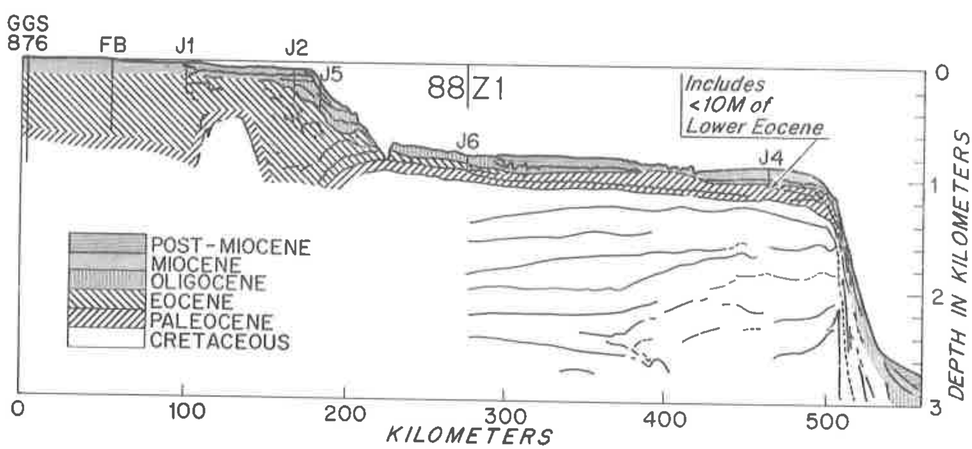

Figure 9. Seismic profile across the shelf, Blake Plateau and continental slope. Profile has been corrected for sound velocity and stratigraphic data from the JOIDES holes have been superimposed on the cross-section. From Emery and Zarudzki (in press).

profile have been corrected for sound velocity, and the stratigraphic units encountered on the JOIDES holes (Bunce and others, 1965) have been superimposed on the profile. As can be seen, some of the reflecting horizons correspond to stratigraphic horizon. Stratigraphic data from the JOIDES drill holes, and the occurrence of Cretaceous fossils southeast of Cape Fear, suggest that the reflector marked A delineates the top of the Cretaceous. This horizon is probably correlative with reflector A from the deep sea (Ewing and others, 1966). On profile 88 the probable extension of reflector A on the shelf appears to be slightly warped. The discontinuities within the Tertiary section generally parallel the present surface of the shelf, but are truncated by the Florida-Hatteras Slope. The Tertiary sediments beneath the shelf also thin very rapidly along the western margin of the Gulf Stream. This decrease in the Tertiary section near the Gulf Stream is further evidence that the stream probably extended its course into its present location sometime during the early Tertiary Period.

The shallow reflectors on the Blake Plateau dip gently seaward and thin in the same direction. They have undergone considerable erosion; these unconformities are in horizons down to reflector A, but are more pronounced at the surface where many narrow depressions have been cut through the Miocene sediments to expose the older deposits. Horizons beneath reflector A dip toward the west, and terminate to the east against a rise near the Blake Spur. This rise south of Blake Spur has been interpreted by Ewing, Ewing, and Leyden (1966) as ancient shallow algal banks. The recovery of oolitic calcarenite and coarse algal limestone of Lower Cretaceous age (Heezen and Sheridan, 1966) from the Blake Escarpment appears to verify this interpretation. Both the uppermost Cretaceous and the lower Tertiary deposits as shown by Figure 9 appear to be slightly warped.

Geologic Map

Information from the sparker profiles described in this report, supplemented by data from seismic refraction profiles (Sheridan, Drake, Nafe, and Hennion, 1966; Hersey, Bunce, Wyrick, and Dietz, 1959; Antoine and Henry, 1965), dredge hauls (Ericson, Ewing, and Heezen, Ericson, Ewing, Wollin, and Heezen, 1961; Heezen and Sheridan, 1966) and bore holes (Bunce and others, 1965) has been used to compile the geologic map shown in Figure 10. This map shows the distribution of pre-Pleistocene strata if the Quarternary sediments were removed.

Shallow-water calcareous sand and algal limestone of Lower Cretaceous age appear to crop out half-way down the Blake Escarpment. The occurrence of these deposits has led Heezen and Sheridan (1966) to postulate that algal banks flourished along the eastern margin of the Blake Plateau during Early Cretaceous time. The seismic profiles described in this report, and those by Ewing, Ewing, and Leyden (1966), and Emery and Zarudski (in press) indicate that the banks terminate slightly north of Blake Spur. Absence of the algal banks farther north

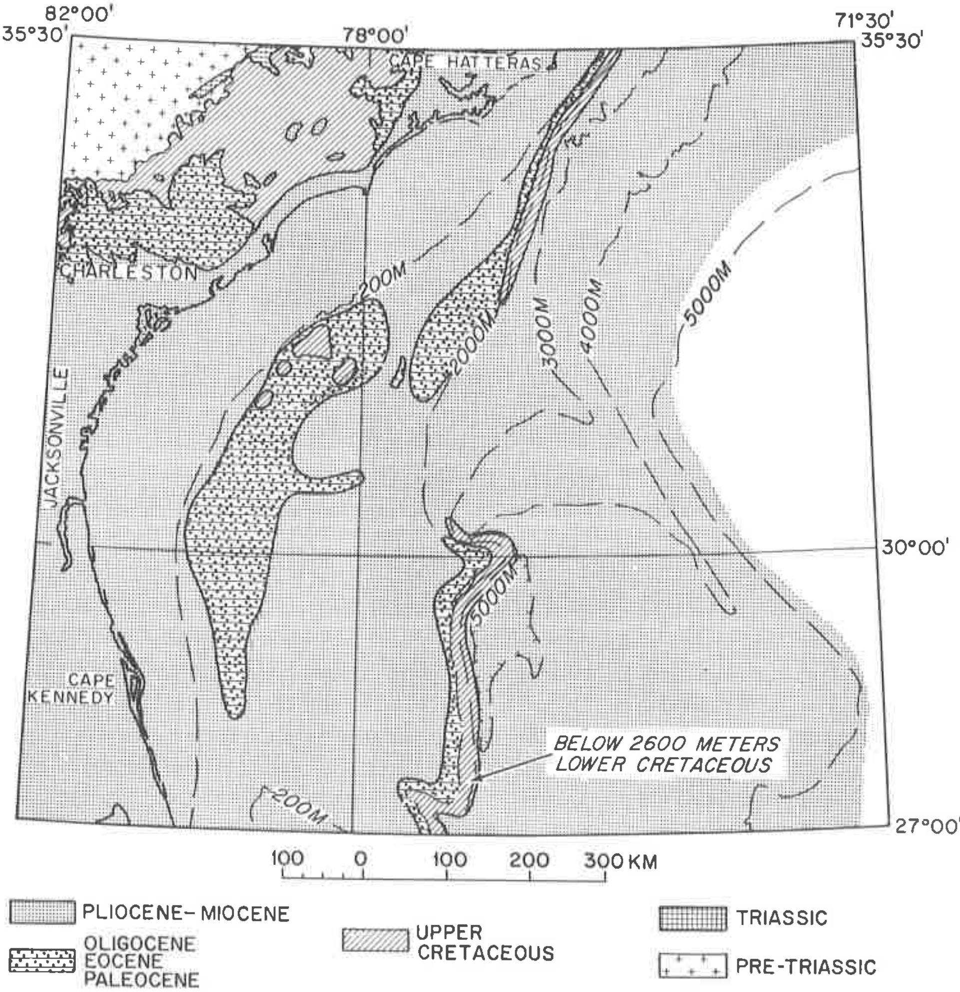

Figure 10. Geologic map of the continental margin south of Cape Hatteras, if the Quaternary sediments were removed.

may be due to their removal by erosion. More probable, however, is the belief that the banks never extended farther north in the past. Some ecologic factor or factors probably prevented the algal banks from flourishing north of the Blake Spur.

Upper Cretaceous sediments occur on the upper part of the Blake Escarpment, on the continental slope northeast of the Blake Spur, and in four isolated patches southwest of Cape Fear. Strata of Paleocene to Oligocene age are present along the eastern and western margins of the Blake Plateau, and in the plateau's rough zone between Latitudes 31° and 32°, and throughout most of the Florida-Hatteras Slope. Miocene-Pliocene sediments occur on the continental shelf and the rest of the Blake Plateau.

Erosion of the Blake Plateau

Erosion by the Gulf Stream has greatly modified the present surface of the Blake Plateau and probably past surfaces. Denudation is most extensive along the western margin of the plateau, and between Latitudes 31° and 32°; so much so that Cretaceous strata have been exposed southwest of Cape Fear. The youngest rocks cut by this erosional surface are Miocene in age, suggesting that the erosion occurred in Miocene or post-Miocene time. Bottom current measurement by Pratt (1963) in the depressions along Latitude 31° and near the base of the Florida-Hatteras Slope yielded speeds of less than 40 cm/sec. Currents of this magnitude are probably competent to prevent filling of the depressions, but not enough to erode them. The low current speeds and the presence of some partially filled hollows indicate that they are not being eroded at present.

Possibly much of the rough topography and enclosed hollows were formed during the Pleistocene. Increase in the erosive capability of the Gulf Stream at that time was probably in response to eustatic lowering of sea level. This drop of sea level may have decreased the average depth of the Blake Plateau by as much as 20 percent. The Pleistocene eustatic lowering of sea level was accompanied by slight increase in the average salinity of the oceans, and a marked change in the latitudinal water temperature gradient. Both of these factors probably resulted in the intensification of the ocean circulation, and may also have helped increase the erosive capability of the Gulf Stream. Possibly erosion by the Gulf Stream in Miocene or post-Miocene time may have been in response to vertical uplift, rather than eustatic lowering of sea level.

The depressions along the base of the Florida-Hatteras Slope, along Latitude 31°, and beneath the slope mark the positions of the Gulf Stream axis during the cycle of erosion. During times of glacially lower sea level, the stream was forced to flow around the seaward extension of the Cape Fear Arch. In interglacial periods, when sea level was higher, the stream impinged on the Florida-Hatteras Slope eroding the slope westward and forming the depressions beneath the slope. As the stream migrated back to the east, progradation by the slope toward the east buried the hollows formed by the stream.

Recent work by Manheim and Richards (1966) on the interstitial waters from the cores obtained in the JOIDES drilling indicate that fresh and brackish waters are present on the shelf sediments. These waters appear to seep out near the base of the Florida-Hatteras Slope. Such underwater springs probably aided in the formation of the depressions along the base of the slope. However, such "spring sapping" was probably not an important factor in the formation of the hollows till the Gulf Stream formed a channel to the top surface of the fresh-brackish water aquifers.

Depth To Top of Cretaceous and Isopach Map of Tertiary Sediments

The horizon marked A in the profiles is believed to be located at or near the Cretaceous-Tertiary boundary. Identification of this reflector is based principally on the Cretaceous fossils recovered southwest of Cape Fear and on the stratigraphic data from JOIDES drilling (Bunce and others, 1965). On the Blake Plateau this reflector is within or at the base of the 1.7 to 1.9 km/sec. layer determined by seismic refraction (Sheridan, Drake, Nafe, and Hennion, 1966; Hersey, Bunce, Wyrick, and Dietz, 1959). On the shelf reflector A appears to correspond to the base of the 3.0-3.7 km/sec. horizon as determined by seismic refraction methods (Sheridan, Drake, Nafe, and Hennion, 1966). Depth to the top of the Cretaceous and thickness of the Tertiary sediments on the Blake Plateau and Florida-Hatteras Slope were computed from the travel time measurements in each seismic profile by assuming velocities of 1500 m/sec. in the water and 1800 m/sec. in the sediments. Contours for most of the shelf are based mainly on seismic refraction data (Sheridan, Drake, Nafe, and Hennion, 1966; Hersey, Bunce, Wyrick, and Dietz, 1959; Antoine and Henry, 1965) and interpolation from well data on land (Maher, 1965). Figure 11 shows the trend and depth of the surface marking the top of the Cretaceous. This surface dips gently to the southeast with its greatest declivities off Cape Hatteras. Continuity of the surface is disrupted by two broad ridges, one southwest of Cape Fear, and the other off Cape Kennedy.

An isopach map of Tertiary sediments is shown in Figure 12. Tertiary sediments appear to be absent from most of the Blake Escarpment, from four small areas on the Blake Plateau, and from the continental slope off Cape Hatteras. Southwest of the Blake Spur Tertiary strata range in thickness from 200 meters along the eastern margin of the Blake Plateau to 1000 meters off Cape Kennedy and in Little Bahama Bank. North of the Blake Spur the Tertiary thins to the northwest to less than 200 meters near the Florida-Hatteras Slope.

The isopach map in this report differs from one recently published by Ewing, Ewing, and Leyden (1966). Their map shows the Tertiary ranging in thickness from 300 meters near the Blake Escarpment to 1200-1600 meters along the western margin of the Blake Plateau. It is suggested that the reflector they believe to be the top of the Cretaceous is too deep. For example: at Latitude 32° 10.6', Longitude 76°46', and Latitude 32°08.8', and Longitude 78°35.6' they report that the Tertiary is thicker than 800 meters; Cretaceous fossils have been recovered in both areas. The preservation, size and number of fossils recovered at both sites preclude the possibility that the fossils were transported to the areas where they were dredged. At JOIDES hole 4 (see Figure 6 for location) they indicate a thickness of 800 meters. This hole penetrated 178.3 meters into the bottom, the lower 89.9 meters of which were in the Paleocene. Based on the Ewing, Ewing, and Leyden (1966) map this

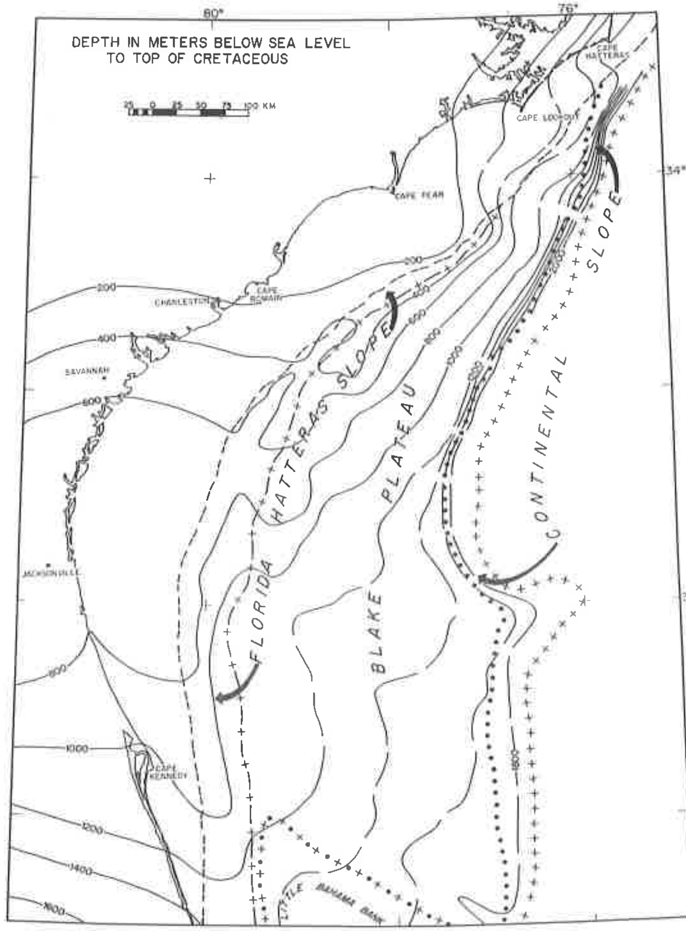

Figure 11. Depth below sea level to top of Cretaceous strata in meters. Based on seismic reflection profiles described in this report. Additional information from reports by Bunce and others (1965), Ewing, Ewing, and Leyden (1966), Hersey, Bunce, Wyrick, and Dietz (1959), Antoine and Harding (1965), Antoine and Henry (1965), and Maher (1965).

would mean that the thickness of the Paleocene is almost 7 times the thickness of the rest of the Tertiary. At JOIDES hole 3 this same map shows the Tertiary to be about 300 meters thick. The drill hole penetrated 178.3 meters below the sea floor and ended in the Lower Eocene. From the Ewing map this would mean that the rest of the Eocene together with the Paleocene is 112.7 meters thick. At JOIDES hole 6 the

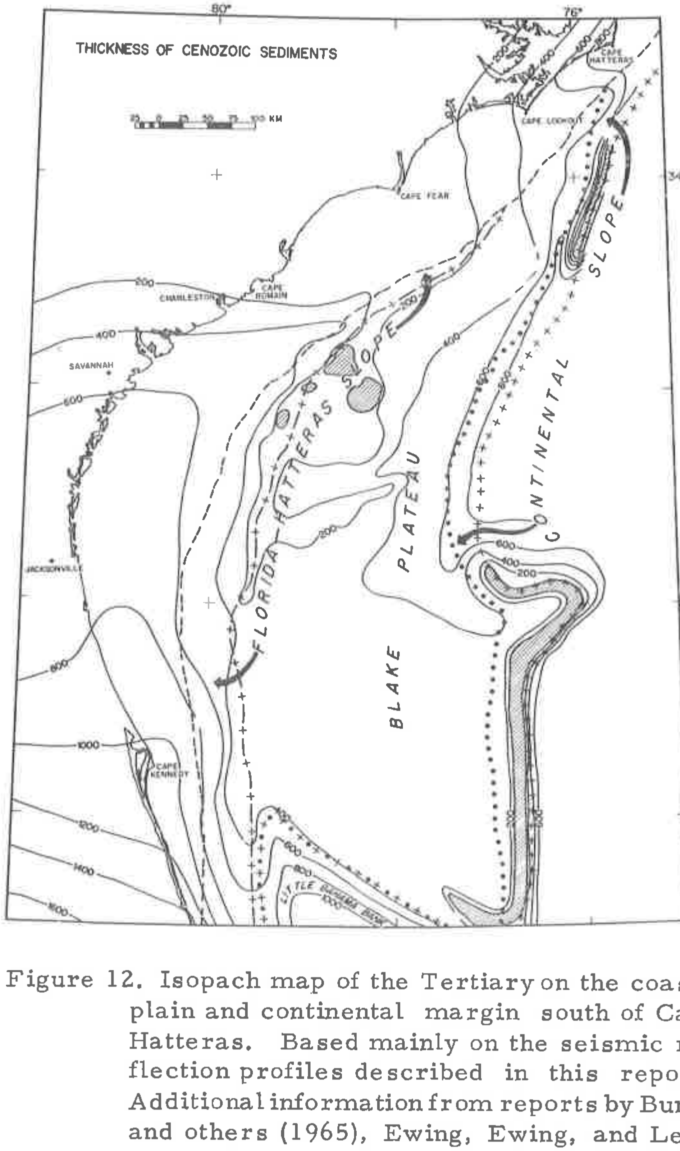

Figure 12. Isopach map of the Tertiary on the coastal plain and continental margin south of Cape Hatteras. Based mainly on the seismic reflection profiles described in this report. Additional information from reports by Bunce and others (1965), Ewing, Ewing, and Leyden (1966), Hersey, Bunce, Wyrick, and Dietz (1959), Antoine and Henry (1965), Antoine and Harding (1965), and Maher (1965). In hatchured areas Tertiary sediments are absent or less than 10 meters thick.

Ewing isopach map indicates a Tertiary section 1100 meters thick. Paleocene sediments were encountered in this hole 116.7 meters below the bottom. Based on their map this would indicate that the Paleocene is 983.3 meters thick, or almost 8 times the thickness of the rest of the Tertiary section on this site. The occurrence of Cretaceous fossils on the Blake Plateau and stratigraphic data from the JOIDES holes amptly demonstrate that the reflector which Ewing, Ewing, and Leyden (1966) believe to be the top of the Cretaceous is too deep. However, if they are correct, then the Paleocene equals or is 8 times thicker than the rest of the Tertiary.

CONCLUSION

The seismic data described in this report when complemented by information from dredge hauls, boreholes, and seismic refraction data makes possible the reconstruction of the sedimentary framework of the continental margin south of Cape Hatteras (Figure 13). Throughout the Cretaceous calcareous banks flourished along the eastern margin of the Blake Plateau. These algal banks extended from Little Bahama Bank to slightly north of the Blake Spur. Shallow water carbonate sediments accumulated during the Lower Cretaceous Period west of the banks.

The organic banks appear to have died during the Cretaceous, as the Tertiary deposits on the Blake Plateau are all deep water calcareous oozes. During the Tertiary the locus of deposition shifted to the west near the present shelf-break. The shift of deposition to the west may have been in response to the Gulf Stream that extended its course to the Blake Plateau at this time. South of Latitude 32° deposition of shallow carbonate Tertiary sediments decreases very abruptly along the western margin of the Gulf Stream. Sediments of equivalent age beyond the Gulf Stream on the Blake Plateau have a total thickness of only about 300 meters or one-third the thickness of the shelf sediments. Apparently the Gulf Stream has prevented transport to the east and accumulation of shallow-water detritus on the Blake Plateau. In contrast, north of Latitude 32° where the Gulf Stream is farther offshore, progradation from the west has extended across the width of the plateau.

The Tertiary deposits on the continental margin south of Cape Hatteras appear to be disrupted by two broad ridges, one southwest of Cape Fear, and another off Cape Kennedy. A flexure also occurs off Georgia and northeast Florida on the shelf (Bunce and others, 1965). Other profiles on the shelf north of Cape Hatteras (not described in this report) reveal similar structures.

During the Pleistocene, a decrease in the average depth of the Blake Plateau by as much as 20 percent, caused the plateau to undergo considerable erosion by the Gulf Stream. The greatest amount of scour occurred at the base of the Florida-Hatteras Slope, and between Latitudes 31° and 32°. Erosion was so great that Cretaceous sediments were exposed and the Florida-Hatteras Slope retreated westward as

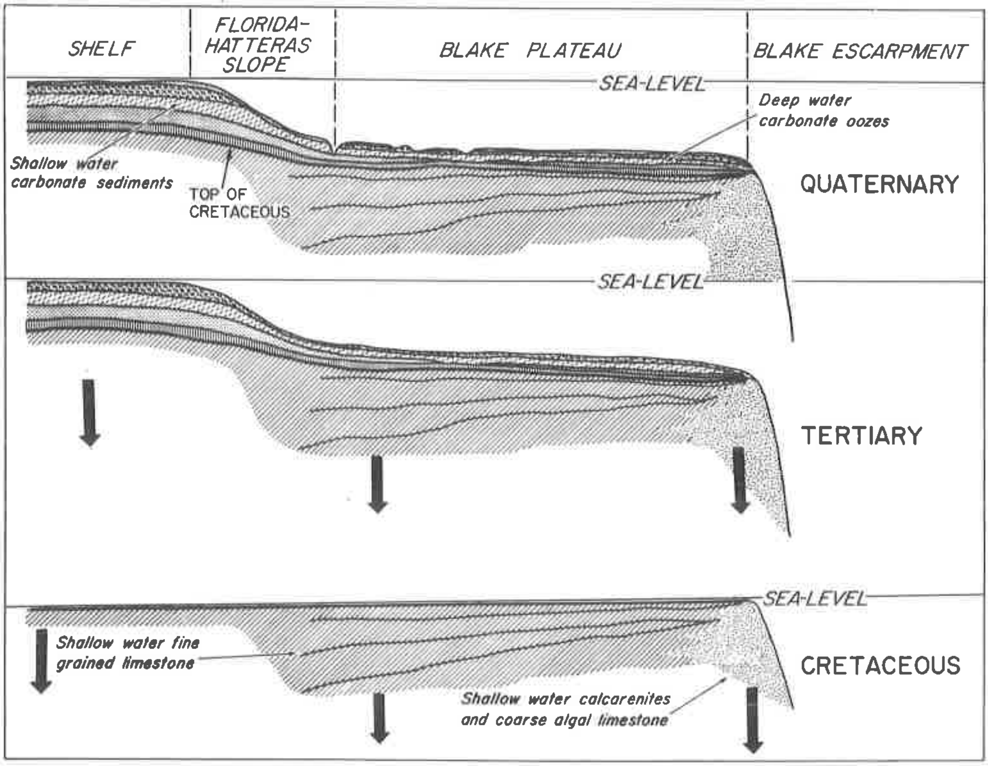

Figure 13. Schematic diagram showing the evolution of the continental margin south of Cape Hatteras. Profiles are not corrected for isostatic adjustment due to sediment loading.

much as 20 km. During the erosion cycle the Gulf Stream axis of maximum current appears to have meandered across a zone over 100 km wide. Present and past positions of the stream's axis are marked by depressions having relief in excess of 40 meters. Sediments removed from the Blake Plateau and Florida-Hatteras Slope were transported to the northeast and deposited on the plateau's smooth segment north of Latitude 32°. As the Gulf Stream shifted to the east, outbuilding of the Florida-Hatteras Slope buried some of the depressions.

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THE PHYSIOGRAPHY OF SEQUATCHIE VALLEY AND ADJACENT PORTIONS OF THE CUMBERLAND PLATEAU, TENNESSEE

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ABSTRACT

Topographic forms in Sequatchie Valley and adjacent portions of the Cumberland Plateau, Tennessee, reflect relative resistances of rock units to erosion. The landscape is considered to be in dynamic equilibrium, and the region to have undergone downwasting continuously since the end of the Paleozoic. Most unconsolidated sediments in Sequatchie Valley are probably of Tertiary and Quaternary age. Three terrace levels are present in Sequatchie Valley. Remnants of a high terrace are preserved on cherty clay residuum forming a ridge above the Knox Group along the axis of Sequatchie Valley. Two lower, younger, terraces are preserved in lateral valleys on either side of the Central Ridge, and the Sequatchie River and several of its larger tributaries locally are intrenched into the Central Ridge. Alluvial fans, which occur at the mouths of large gorges or gulfs in the Cumberland Plateau, contain cobbles and boulders of Pennsylvanian sandstones derived from plateau uplands. The highest alluvium in Sequatchie Valley is Tertiary, or perhaps Early Pleistocene, and the two lower terrace levels are probably Pleistocene in age.

INTRODUCTION

Sequatchie anticline is an isolated Valley and Ridge-type structure situated in gently dipping and only locally deformed rocks of the Cumberland Plateau of Tennessee and Alabama (Figure 1). The structure, which parallels the regional trend of the Appalachians for a distance of more than 200 miles, is reflected along most of its length by Sequatchie Valley (Browns Valley in Alabama), and by mountains at its ends where unbreached Pennsylvanian sandstones arch over the anticlinal crest. The anticlinal structure is superficial and is bottomed by the Sequatchie Valley fault. Footwall rocks extend southeastward beneath the Sequatchie Valley fault along the projection of the regional dip. Sequatchie Valley drainage is tributary to the Tennessee River, which flows through part of the valley, and is intimately related to the development of that river.

> In Tennessee the relief and pronounced linearity of the valley and the consistently rugged valley walls are the most striking topographic features of the structure. Maximum elevations of the eastern valley wall, which range from 2000 feet in the southern portion of Sequatchie Valley to 3000 feet at the head of the valley, are generally 100 to 400 feet greater than those of the western escarpment. The relief from the top of the Cumberland Plateau to the floor of the valley is consistently greater than 1000 feet throughout the length of the Sequatchie Valley in Tennessee (Milici, 1963, p. 815-816).

This report is largely an application of the principles described by Hack (1960, 1966) to many observations made in detailed geologic mapping in the southern Cumberland Plateau of Tennessee. Observations concerning the physiography of the region support the concept of dynamic equilibrium, rather than that of the geographic cycle and peneplain concept (see Hack, 1960, 1966). Hack (1966) has shown that the concept of peneplanation is not required to explain topographic features of the region. In this paper the concept of dynamic equilibrium is further supported by additional examples of the relationship of topography to bedrock and surficial geology in Sequatchie Valley and on adjacent portions of the Cumberland Plateau (Table 1).

Acknowledgments

Richard G. Stearns and C. W. Wilson, Jr., of Vanderbilt University, J. M. Kellberg of the Tennessee Valley Authority, and Stuart W. Maher and Robert J. Floyd of the Tennessee Division of Geology critically reviewed and edited the manuscript. The study of the Sequatchie Valley region was performed under direction of William D. Hardeman, State Geologist.

PHYSIOGRAPHIC DEVELOPMENT OF SEQUATCHIE VALLEY AS AN AREA IN DYNAMIC EQUILIBRIUM

Breaching of Sequatchie Anticline

The development of the Tennessee River and Sequatchie Valley are intimately related; the river enters Sequatchie Valley near Chattanooga, and flows southwestward along the structure in Alabama for about 50 miles before resuming a generally northwestward course. The Tennessee River probably eroded headward across Sequatchie anticline during the Mesozoic and initiated formation of the Sequatchie Valley. Inversion of topography from anticlinal hills and mountains

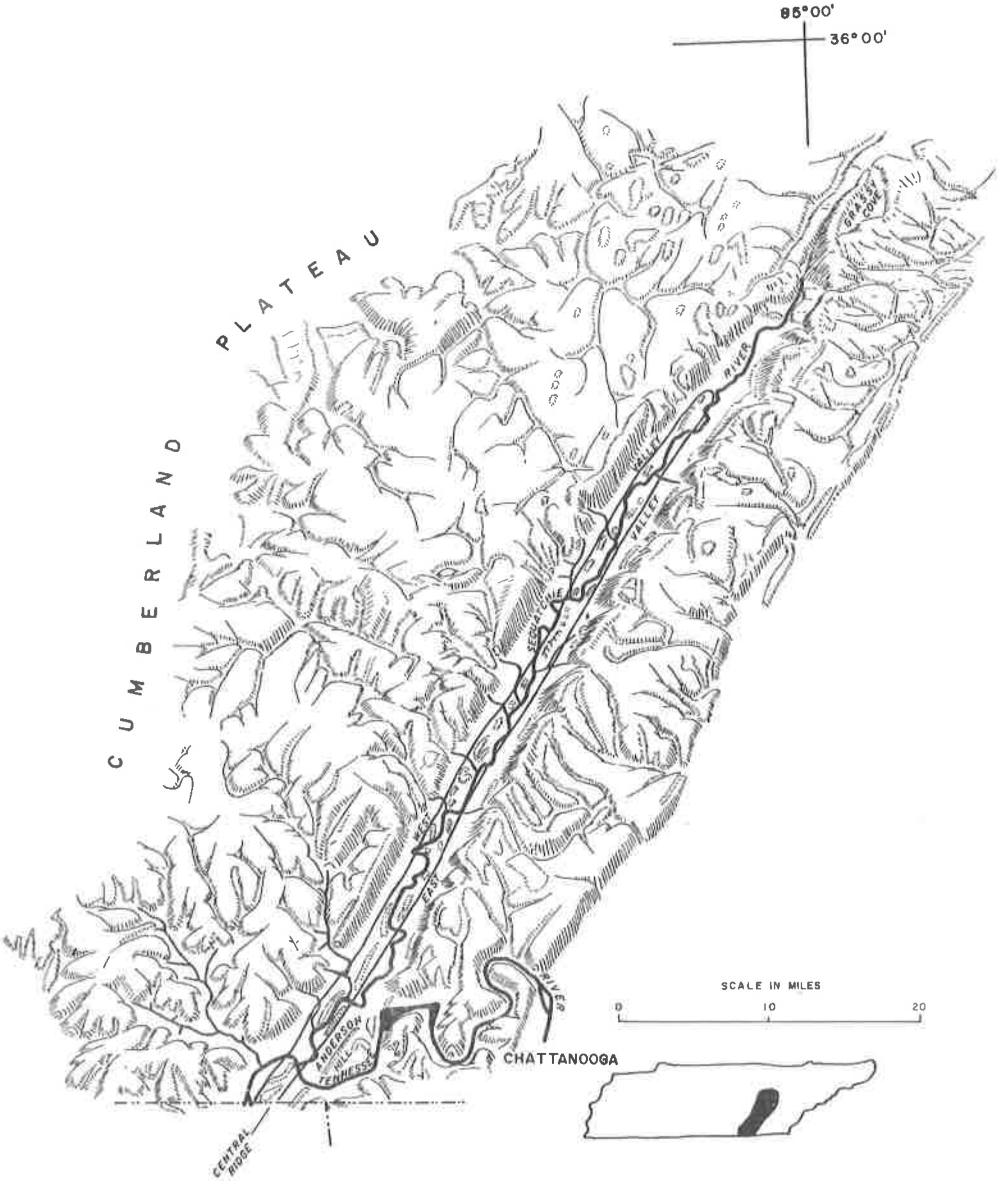

Figure 1. Physiographic diagram of Sequatchie Valley.

A Central Ridge of cherty clay residuum overlies the Knox Group, and separates West and East valleys underlain by Mississippian and Ordovician limestones and shales. Anderson Hill is a hogback of Fort Payne Chert. Crab Orchard Cove is along the crest of the anticline about 5 miles northeast of Grassy Cove.

System	Group	Formation	Lithology	Physiographic Grouping
PENNSYLVANIAN	Crab Orchard Mountains	Rockcastle Conglomerate	Sandstone and conglomeratic sandstone; 200 feet thick.	UNIT 6- Forms upper Cumberland Escarpment and Cumberland Plateau uplands
		Vandever Formation	Shale, siltstone, sandstone, and conglomeratic sandstone; 300 feet thick.	
		Newton Sandstone	Sandstone; 80 feet thick.	
		Whitwell Shale	Shale, siltstone, sandstone; 75 feet thick.	
		Sewanee Conglomerate	Sandstone and conglomeratic sandstone; 100 feet thick.	
	Gizzard	Signal Point Shale	Shale, siltstone and sandstone; 60 feet thick.	
		Warren Point Sandstone	Sandstone and conglomeratic sandstone; 100 feet thick	
		Raccoon Mountain	Shale, siltstone and sandstone; 150 feet thick.	
MISSISSIPPIAN		Pennington Formation	Variegated shale, sandstone and limestone; 300 feet thick.	UNIT 5- Forms relatively gentle slopes about half way up slopes of Cumberland Escarpment
		Bangor Limestone	Limestone; 200 feet thick.	UNIT 4- Forms steep ground at base of Cumberland Escarpment
		Harselle Formation	Calcareous sandstone; 50 feet thick.	
		Monteagle Limestone	Limestone; 250 feet thick.	
		St. Louis Limestone	Limestone; 100 feet thick.	
		Warsaw Limestone	Limestone; 80 feet thick.	
		Fort Payne Formation	Siliceous limestone; 200 feet thick.	UNIT 3- Forms chert hills and hogback in Sequatchie Valley
DEV.-MISS.		Chattanooga Shale	Black shale; 30 feet.	UNIT 2- Forms lowlands of eastern and northern Sequatchie Valley
SILURIAN		Brassfield Formation	Limestone and shale; 100 feet thick.	
ORD.	Richmond	Sequatchie Formation	Limestone and shale; 100 feet thick.	

Table 1. Stratigraphy of the Southern Cumberland Plateau, Tennessee (Stratigraphic terminology from Hardeman, Miller, and Swingle, 1966)

Table 1. -- Continued

System	Group	Formation	Lithology	Physiographic Grouping
ORDOVICIAN	Maysville	Leipers Formation	Shaly limestone; 60 feet thick.	
	Eden	Inman Formation	Limestone and shale; 50 feet thick.	
	Nashville	Catheys Formation	Limestone, 300 feet thick.	
		Bigby-Cannon Limestone	Limestone; 150 feet thick.	
		Hermitage Formation	Shaly limestone; 80 feet thick.	
	Stones River	Carters Limestone	Limestone; 150 feet thick.	
		Ridley Limestone	Limestone and shaly limestone; 275 feet thick.	
		Pierce and Murfreesboro Limestones	Limestone and shaly limestone; 500 feet thick.	
	Knox		Siliceous limestone and dolomite; up to 1,000 feet exposed	UNIT 1 - Forms cherty hills along axis of Sequatchie anticline.

(such as those that persist at both ends of the structure) to the anticlinal valley began when Pennsylvanian formations were breached and the Pennington shales were exposed along the crest of the structure at a level of 4,000 to 5,000 feet above the present valley floor. The valley subsequently developed by headward erosion of drainage tributary to the Tennessee, and by underground solution of structurally elevated limestones where overlying formations were breached.

At the northeast end of Sequatchie Valley (Head of Sequatchie, Vandever quadrangle), the Sequatchie River rises primarily from several springs that have their source in large areas of interior drainage farther to the northeast--Grassy, Swaggerty, and Crab Orchard Coves, and underground solution and erosion will eventually join the coves to Sequatchie Valley (Lane, 1952, 1953; Wilson, 1965, p. 42-43), (Figure 2). The coves, which are large sinkholes or uvalas situated on Sequatchie anticline, are bottomed by horizontal or gently dipping Mississippian carbonates and are rimmed with Pennsylvanian sandstones, conglomerates, siltstones and shales. Mississippian shales form the slopes of the coves. Remnants of an absorbed cove are visible in northern Sequatchie Valley. The breached uvala, termed herein Litton Cove for the small community at its southern boundary, is topographically distinct from Sequatchie Valley to the southwest (Figure 2). Litton Cove is separated from Sequatchie Valley proper by Cedar Ridge, a series of low limestone hills that are normal to the valley axis. Southwest of the boundary hills Sequatchie Valley is extensively alluviated,

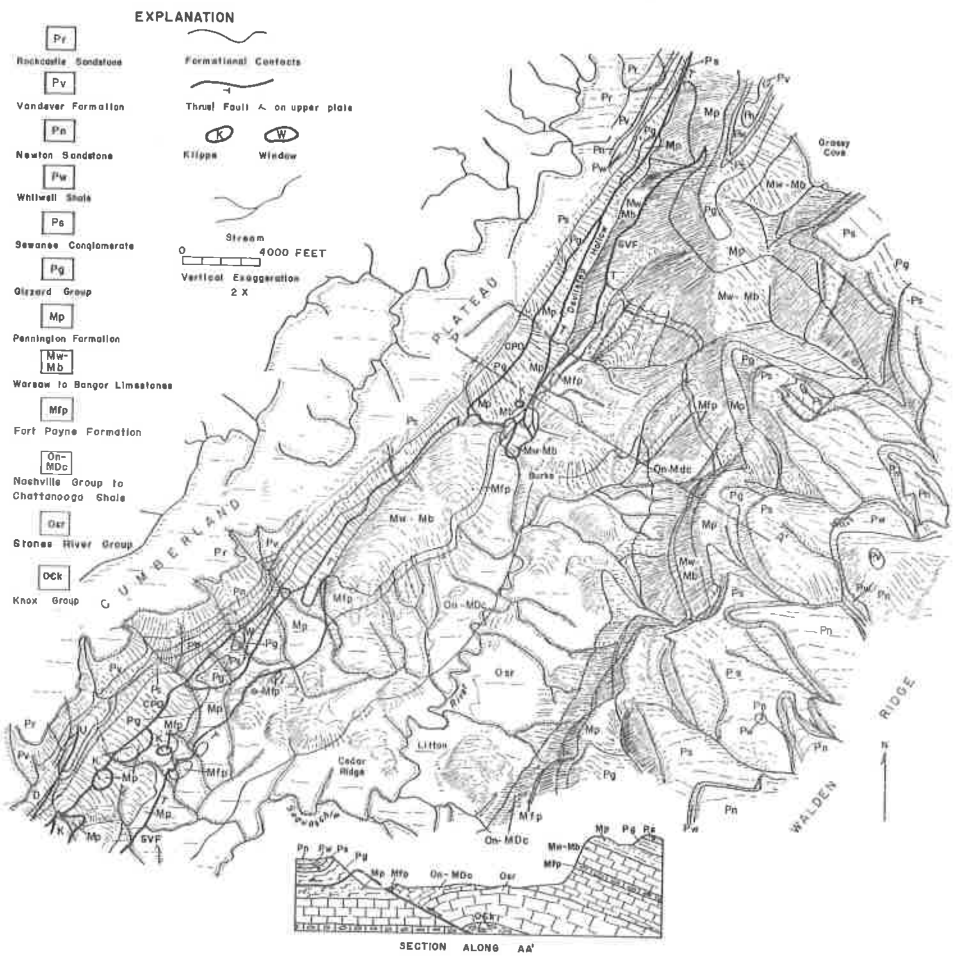

Figure 2. Relationship of physiography to bedrock geology in northern Sequatchie Valley, Tennessee (modified from Swingle, 1964; Milici, 1960, 1965; and Milici and Coker, 1967). SVF is the Sequatchie Valley fault; CPO, is the Cumberland Plateau overthrust.

and consists of open rolling topography. Northeast of the boundary hills, in Litton Cove, the topography is much more rugged and consists largely of limestone hills covered with thin soil.

Relationship of Topography to Bedrock Geology

Chert-armored hills. The Knox Group and Fort Payne Formation weather to cherty soil that forms low but steep hills in the floor of Sequatchie Valley. The writer believes that steepnesses of the hill slopes are sufficient to facilitate removal of chert so that chert hills in an area of topographic equilibrium are lowered at the same rate as adjacent lowlands, and relief remains approximately constant as long as the rocks of contrasting physical characteristics are exposed under the same climatic conditions. Slopes change in areas not in equilibrium so that chert (or other rock) is removed at rates faster or slower than the equilibrium rate until equilibrium slopes and reliefs are attained. Such hills have been described by Hayes (1899, p. 17) and Hack (1960, p. 87-89).

The Knox Group is the oldest rock exposed in the area and forms a ridge of residual cherty clay that extends from north of Pikeville, Tennessee, where the Knox plunges northeastward under Middle Ordovician limestones, southward into Alabama. The ridge, which rises as much as 400 feet above the Sequatchie River, separates Sequatchie Valley longitudinally into two lateral valleys: West Valley, largely underlain by Mississippian limestones and shales (overridden by the Knox and Stones River Groups along the Sequatchie Valley fault), and East Valley, underlain by Middle and Upper Ordovician limestones. Northeast of Knox exposures, Sequatchie Valley is generally underlain by Middle and Upper Ordovician limestones thrust over Mississippian shales, and the threefold subdivision of the Valley disappears because of similar erodibility of rocks exposed (Figures 1 and 2).

The Fort Payne Formation forms prominent hills and ridges in Sequatchie Valley, where it is involved in structures along the toe of the western valley wall, and where it is exposed in gently to moderately dipping beds along the toe of the eastern valley wall. Numerous Fort Payne klippen are preserved between the Cumberland Plateau and Sequatchie Valley fault traces, and generally below the western wall of the valley. Fort Payne is thrust over Pennington shales and Pennsylvanian shale, siltstones, and sandstones along the trace of the Sequatchie Valley fault in northern Sequatchie Valley. Differential etching of resistant cherty carbonates thrust over the more easily eroded shales has formed low hills that stand approximately 200 feet above the surrounding valley floor, i.e., the Fort Payne has acted as a protective cap that inhibited erosion of underlying softer rock. In contrast, klippen of Mississippian limestones and shales thrust over similar lithologies have no topographic expression (Figure 3).

Valley wall profiles. The southeastern and northwestern walls of Sequatchie Valley have different profiles that reflect the difference in the rocks exposed.

Typically, the southeastern valley wall has a concavo-convex profile (Figure 4). Mississippian limestones at the base of the southeastern valley floor generally underlie steep ground from 200 to 300 feet above the valley floor. The slope gradually decreases in the next 400 feet of vertical ascent; this zone begins near the top of the thick Mississippian limestone section and continues through Pennington shales. Near the top of the valley wall, about 300 to 400 feet below the

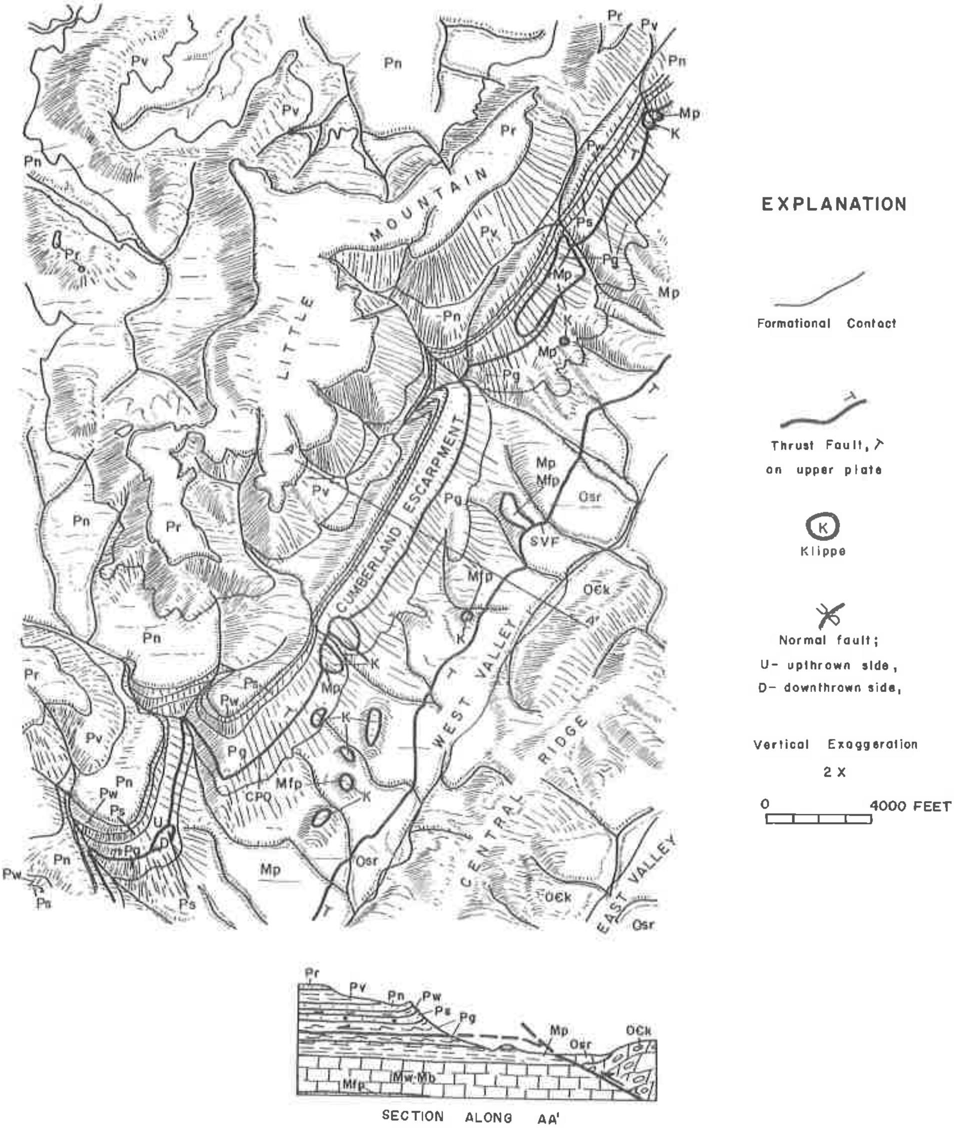

Figure 3. Relationship of physiography to bedrock geology in central Sequatchie Valley, Tennessee (modified from Garman and Milici, 1967).
Pr--Rockcastle Conglomerate; Pv--Vandever Formation; Pn--Newton Sandstone; Pw--Whitwell Shale; Ps--Sewanee Conglomerate; Pg--Gizzard Group; Mp--Pennington Formation; Mw-Mb--Warsaw to Bangor Limestones; Mfp--Fort Payne Formation; Osr--Stones River Group; OЄk--Knox Group. The trace of Sequatchie Valley (SVF) is in floor of the west valley, and the trace of the Cumberland Plateau overthrust (CPO) is along the Cumberland Escarpment.

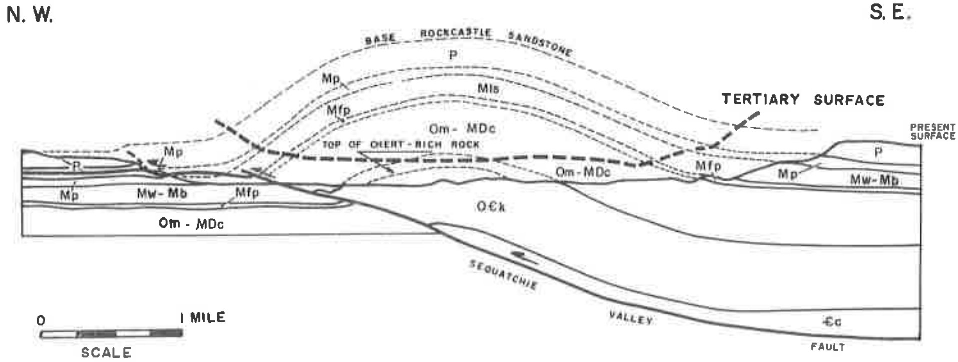

Figure 4. Cross-section of Sequatchie Valley (see Figure 3 for location of section). Modified from Milici, Coker, and Garman (in press).
P--Pennsylvanian; Mp--Pennington Formation; Mw-Mb--Warsaw to Bangor Limestones; Mfp--Fort Payne Formation; Om-MDc--Murfreesboro Limestone to Chattanooga Shale; OЄk--Knox Group; Єc--Conasauga Group.

rim of the Plateau, the slope steepens once again where Pennsylvanian shale, siltstones, and sandstones form the Cumberland Escarpment.

The northwestern wall of Sequatchie Valley is characterized by complicated geologic structures throughout much of its length; however, its generally concave slope reflects more resistant Pennsylvanian rocks above the less resistant Pennington shales at or near the valley floor. In southern Sequatchie Valley, Tennessee, the Mississippian limestone section is exposed at the toe of the northwestern wall and its profile resembles the profile of the southeastern wall.

Surficial Deposits

Introduction. Unconsolidated accumulations in Sequatchie Valley are probably all of Tertiary or Quaternary age. Residual clays are common above carbonates in the floor of Sequatchie Valley, and generally contain chert fragments. Colluvium is abundantly developed on the sides of the valley where fragments of Pennsylvanian sandstones as large as several tens of feet in size have broken from massive rim rock. Pleistocene alluvial fans formed at the mouths of large gulfs or gorges in the Cumberland Plateau and Walden Ridge, and alluvial terraces deposited by Sequatchie River and its tributaries occupy much of Sequatchie Valley.

Terraces. Along its course from Head of Sequatchie to the Tennessee River, the Sequatchie River cuts back and forth across the Central Ridge from West Valley to East Valley. Alluvial terraces of fine-grained sand and silt, interbedded with coarse chert fragments and containing vein quartz pebbles derived from Pennsylvanian conglomerates, are well developed in the particular lateral valley in which the Sequatchie River flows. In places boulders and cobbles derived from Pennsylvanian sandstones are interbedded with fine-grained terrace materials. Alluvial materials deposited by tributary streams also occur in the lateral valley on the other side of the Central Ridge from Sequatchie River but do not form well-defined terraces. Several larger tributaries that extend as far as 15 miles headward into the Cumberland Plateau are incised and flow through the Central Ridge. Smaller Sequatchie River tributaries flow in lateral valleys parallel to the Central Ridge and intersect Sequatchie River below where it crosses the Central Ridge (Figure 1). The smaller tributaries have headwaters in much smaller gulfs into the Plateau and are probably younger. Size of Sequatchie River tributaries decreases generally from the Tennessee River northeastward to the head of the Sequatchie Valley, and is evidence for headward extension of the Valley.

Longitudinal profiles of the Sequatchie Valley floor, the upper limit of alluvial terrace elevations, the upper Central Ridge elevations were prepared by projecting Sequatchie River and hilltop elevations onto a line drawn medially along Sequatchie Valley (Figure 5). Two levels of terrace deposits are above the valley floor and below maximum elevations of the Central Ridge. The lower terrace extends from the Tennessee River northeast to near Dunlap; the upper terrace is well defined in the vicinity of Dunlap and to the northeast. Only in central Sequatchie Valley do the two terrace levels coexist (Henson Gap quadrangle). The terraces record climatic changes and intermittent lowering of the Sequatchie Valley floor. The upper terrace has been eroded completely in southern Sequatchie Valley, Tennessee, and the lower terrace probably was never formed in northern Sequatchie Valley. Patches of alluvium are preserved on crests and upper slopes of the Central Ridge, and cobbles of vein quartz and quartzite as much as 6 inches across are preserved both on top of the Central Ridge and in the lower terrace near the Tennessee River in southern Sequatchie Valley, Tennessee and in northern Alabama (Browns Valley). The cobbles were derived from Precambrian rocks of the Blue Ridge Province.

Alluvial fans. Alluvial fans of sandstone boulders and cobbles derived from Pennsylvanian formations are well developed in Sequatchie Valley at the mouth of large gorges, or gulfs, in the Cumberland Plateau and Walden Ridge. Sandstone blocks fell away from the Cumberland Escarpment, moved downslope as underlying material was sapped, and were deposited in fans by running water. Downslope movement of sandstone blocks and formation of alluvial fans probably are related to frost action and increased runoff during glacial times.

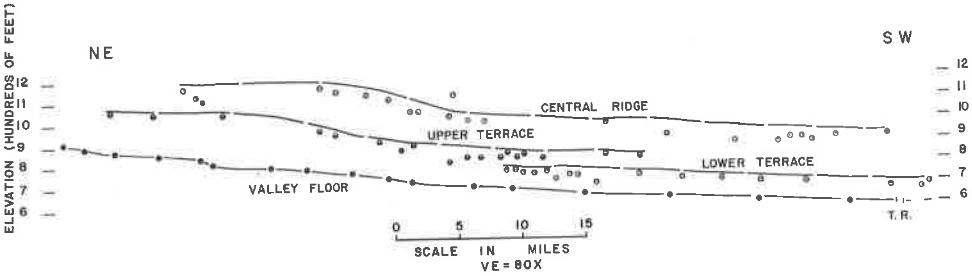

Figure 5. Longitudinal profiles in Sequatchie Valley, Tennessee. Two alluvial terraces are between upper elevations of the Central Ridge and the Valley floor. Remnants of an upper terrace are preserved on top of the Central Ridge in some areas. T. R. is the Tennessee River. Points along the Sequatchie River profile were plotted at 20-foot intervals of elevation. Points along profiles of the Terrace levels and Central Ridge mark elevations of highest hills for each physiographic feature.

RELATIONSHIP OF TOPOGRAPHY TO BEDROCK GEOLOGY IN THE CUMBERLAND PLATEAU

Physiographic features in the southern Cumberland Plateau of Tennessee follow surface geology in minute detail and ". . . features inherited from the past are not in evidence (Hack, 1966, p. C 7)." Pennsylvanian formations of the Gizzard and Crab Orchard Mountains Groups are exposed in the Cumberland Escarpment and Plateau. In general the base of the Cumberland Escarpment is formed by the Gizzard, while the Sewanee Conglomerate forms the rim-rock in Sequatchie Valley and around much of the southern Cumberland Plateau in Tennessee. The Sewanee is generally 100 to 160 feet thick, and forms prominent bluffs at the top of the Escarpment. In places, thinner sandstones of the Gizzard form subsidiary ledges a hundred or so feet below the Sewanee, and in Tennessee near the Georgia and Alabama State lines the Warren Point Sandstone is at the top of the Plateau and forms the upper bluffs. The youngest formation preserved in the area is the Rockcastle Conglomerate, and as much as 100 feet of that formation occurs on tops of hills northwest of Sequatchie Valley (Figure 3). Three structural environments will be used to illustrate the mutual adjustment of shale and sandstone formations to the processes of downwasting acting in the Cumberland Plateau region: (1) the broad Sampson divide, (2) horizontal beds of the Little Mountain area, and (3) moderately to gently dipping beds of the Walden Ridge monocline. Detailed adjustment of Plateau surfaces to rock type is explained by the concept of dynamic equilibrium, and the postulate of a Cretaceous peneplane to explain these features is unnecessary (Hack, 1966, p. C 5).

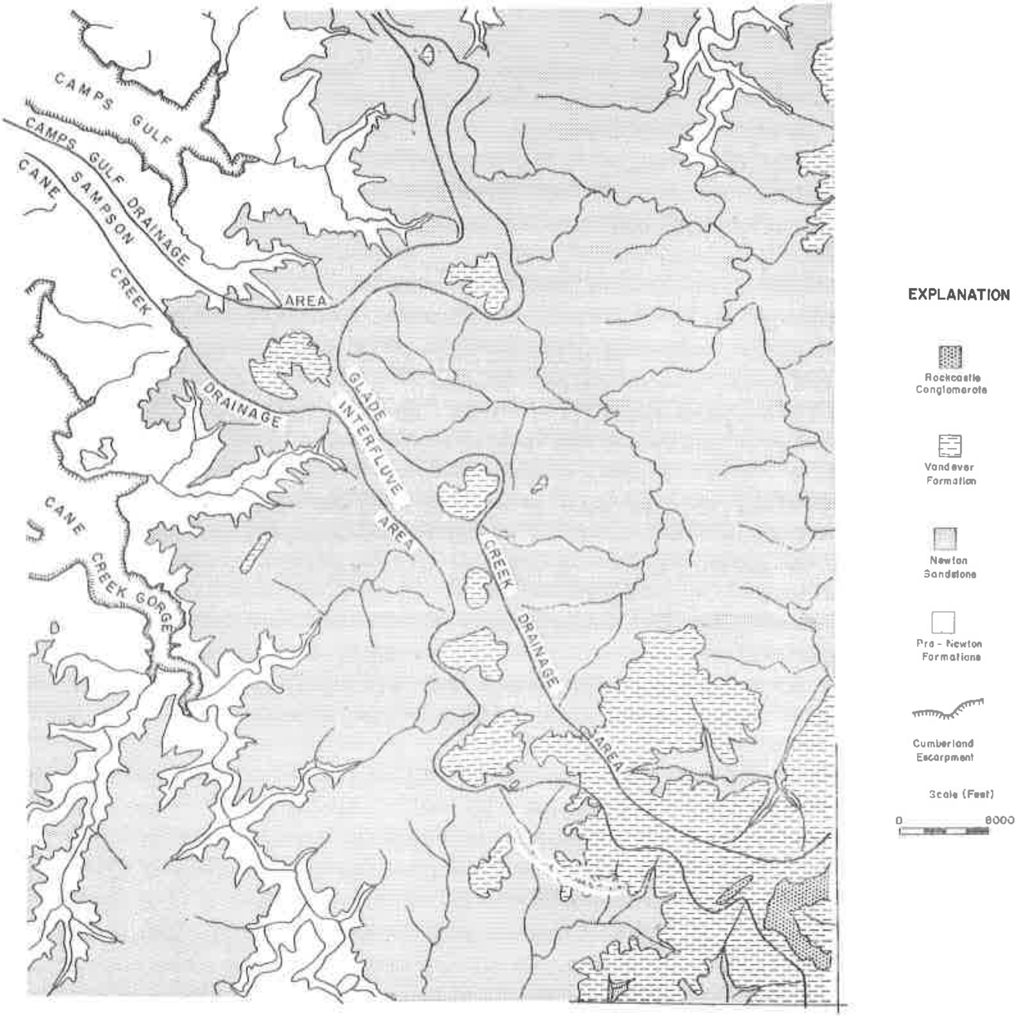

Figure 6. Geologic sketch map of Sampson Divide (modified from Milici, in press).

Much of the upland surface of the Sampson quadrangle, Van Buren and Bledsoe Counties, Tennessee, is underlain by the Newton Sandstone (Milici, in press), (Figure 6). Older formations are exposed in Camps Gulf, Cane Creek Gulf, and along Glade Creek; and the younger Vandever Formation is preserved in the southeastern part of the quadrangle because of the regional dip. Lower beds of the Vandever form hills northwest of the main outcrop that are as much as 100 feet high. The outliers mark the drainage divide between north- and west-flowing tributaries to Glade Creek. In other words, the Newton Sandstone has been stripped, and low hills of Vandever shales and siltstones are remnants on the least eroded portion of the stripped surface.

At Little Mountain, an area of nearly flat-lying beds, uplands are underlain by completely stripped Rockcastle sandstone, slopelands by Vandever shales and siltstones, and bottomlands by Newton sandstones now being stripped of overlying shales. Northwest of Little Mountain the Newton underlies much of the Cumberland Plateau surface, and shales, siltstones, and sandstones of the Vandever Formation occur in isolated knobs on the Newton (Figure 3).

Resistant sandstone surfaces are more markedly stripped on Walden Ridge where dips near Sequatchie anticline are much steeper than the regional dip of the Cumberland Plateau west of the anticline. For example, Pennsylvanian formations on the southeast flank of Sequatchie anticline dip nearly 10° to the southeast, and broad slopes have developed on the Sewanee, Newton and Rockcastle Formations. Northwest-facing reverse slopes of much smaller areal extent have developed in intervening shale formations (Figure 2).

SUMMARY

Sequatchie anticline formed by thrust-faulting near the end of the Paleozoic. During much of the Early Mesozoic the anticline was marked by a prominent ridge of Pennsylvanian clastics, perhaps similar to the present Crab Orchard Mountains at the northern end of the structure. The Tennessee River probably formed as a consequent stream on the Mesozoic coastal plain, eroded headward, and cut through the Sequatchie ridge about where the modern river traverses Walden Ridge.

Sequatchie Valley was initiated when Pennsylvanian strata were breached and the valley was extended by headward erosion of the ancestral Sequatchie River, and by solution of elevated limestones and assimilation of uvalas (large sinkholes) that developed on the crest of the anticline. These processes are continuing today. Remnants of a now-joined uvala are present in northern Sequatchie Valley, and isolated uvalas such as Grassy and Crab Orchard Coves become smaller and smaller to the northeast.

Sometime during the Tertiary(?) Sequatchie River developed a wandering course over Middle Ordovician and Mississippian limestones at an elevation above that of the Central Knox Ridge. All limestones in the valley floor during the Tertiary(?) were about equally resistant to erosion, and all of Sequatchie Valley was topographically similar to the region northeast of the present Knox outcrop. Tributaries to Sequatchie River drained the Cumberland Plateau and Walden Ridge and established courses across the flood plain of Sequatchie River. During the Late Tertiary and Pleistocene Sequatchie River was incised at least 400 feet, and cherty carbonates of the Knox Group were exposed. Some of the winding pattern of the river was preserved and several of the larger pre-existing tributaries to Sequatchie River were able to maintain their courses across the Central Ridge as residual chert accumulated in the middle of the valley. Pauses in downward movement of the Sequatchie River, which may have resulted from Pleistocene climate or sea-level fluctuations, are marked by two levels of alluvial terraces and remnants of a third. Two alluvial terraces in places interfinger with boulders of Pleistocene alluvial fans.

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POTENTIAL USES OF FLOW NET ANALYSIS IN WATERSHED ENGINEERING[1/]

by

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ABSTRACT

Flow nets are graphical representations of flow patterns, and are a means of analyzing and interpreting the water table and piezometric maps. When mathematical interpretations are not practical, flow nets often provide a solution for groundwater flow in watersheds. Saturated flow conditions are constantly changing; hence, the flow nets are constantly changing and necessitate flow computations at a predetermined, groundwater elevation change to compute groundwater flow from, or into, watershed areas.

Factors that need to be considered and analyzed before a flow net is constructed for a watershed are: (1) Surface runoff from other watersheds in the area with similar geologic and groundwater conditions; (2) surface runoff from subwatersheds within the watershed area; (3) geological conditions within the watershed area (structure, stratigraphy, and lithology); (4) groundwater conditions within the area (water table, flow direction and gradient, number of aquifers, aquifer boundaries, and permeability); and (5) variations in groundwater elevation within the watershed area with time.

After these factors are considered and boundary conditions established, flow nets can be drawn for the watershed area. They will permit analysis of groundwater movement from, or into, the area, assuming that reasonably accurate values of transmissibility and aquifer thickness are known.

Flow nets were used to analyze the groundwater phase of the hydrologic cycle for the Meridian Formation, Marshall County, Mississippi. This study points to the need for groundwater information before water balance studies can be performed on a watershed, and briefly outlines the method of this type of analysis.

1/ Contribution from the Southern Branch, Soil and Water Conservation Research Division, Agricultural Research Service, USDA, in cooperation with Georgia Agricultural Experiment Stations, North Carolina Agricultural Experiment Station, Florida Agricultural Experiment Station, Water Resources Center of the Georgia Institute of Technology, Central & Southern Florida Flood Control District, and Middle South Georgia Soil Conservation District.

INTRODUCTION

Flow net analysis, a method usable in the groundwater phase of the hydrologic cycle, permits watershed engineers to quantitatively evaluate the water entering and leaving the drainage basin within each aquifer. This example of flow net analysis is limited to the Meridian aquifer lying above the Wilcox Group. The Wilcox in this area has a low permeability; therefore, water within the Meridian moves laterally rather than into deeper aquifers. Hence, this study is confined to the Meridian aquifer for water balance studies of this area.

Flow nets, which are graphical representations of flow patterns, offer reliable assistance, and often provide a solution when mathematical solutions are not practicable. Graphical analysis of flow patterns was first developed by Forcheimer (1930). A flow net is composed of two families of lines, or curves:

1. Streamlines, or flow lines, indicating the path followed by water as it moves in the direction of decreasing head.
2. Equipotential lines which intersect streamlines at right angles and represent contours of equal pressure head in the aquifer.

Theoretically, flow patterns contain an infinite number of flow lines and equipotential lines. Flow net construction, however, makes use of only a few of the lines. Equipotential lines are selected so that the total drop in head across the system is evenly divided between adjacent pairs of potential lines. Similarly, the flow lines, normal to the equipotential lines, are selected so that the total quantity of flow is divided equally between adjacent pairs of lines.

The hydraulic gradient is determined by the drop in head between two equipotential lines in an aquifer, divided by the distance traversed by water moving from a higher to a lower potential line. This assumes that the water moves in a straight line, which is often not the case under natural geological conditions; some error may, therefore, be introduced into the analysis. In general the water will follow the path of least resistance, or the shortest path between equipotential lines. Thus, it follows that the direction of water movement is everywhere analogous with paths that are normal to the equipotential lines.

A flow net constructed on the above principles is assumed to be rectangular, and the ratio of the mean dimensions of each rectangle is constant. This must be kept in mind at all times during construction of the net. The net becomes a system of squares if the sides of the rectangles are equal. Flow nets, however, involve curved flow paths and the geometric form is somewhat curvilinear, and true squares or rectangles are seldom attained.

Flow net construction of the groundwater pattern is difficult. It requires a thorough knowledge of the geological conditions and experience in construction of nets. Casagrande (1937) listed the points below, which are helpful in flow net sketching:

(1) Study the appearance of well-constructed flow nets and try to duplicate them by independently re-analyzing the problem they represent.

(2) In the first attempts at sketching, use only four or five flow channels.

(3) Observe the appearance of the entire flow net; do not try to adjust details until the entire net is approximately correct.

(4) Frequently parts of a flow net consist of straight and parallel lines, which result in uniformly sized true squares. By starting the sketching in such areas, the solution can be obtained more readily.

(5) In flow systems having symmetry (for example, nets depicting radial flow into a well), only a section of the net need be constructed, as the other part or parts are images of that section.

(6) During the sketching of the net, keep in mind that the size of the square changes gradually; all transitions are smooth, and where the paths are curved, are of elliptical or parabolic shape.

Taylor (1948) suggests a somewhat different technique. He recommends that a trial flow line be drawn and that the entire system be completed as if the trial line were correct. If the completed system is not correct, then the initial line is corrected and the entire system resketched. This method permits a more accurate analysis, in that the whole picture of the flow net is reviewed and analyzed as one picture.

DISCUSSION

Before attempting to construct a flow net, it is necessary to establish the boundary conditions and describe them. For steady flow, with particular boundary conditions, only one flow net exists. However, flow nets in watershed analyses are constantly changing and are a function of time. This is not always the case in drainage studies, drawdown studies, pipe flow or flow-around barrier. Flow net changes, in these cases, are allowed to reach equilibrium, or are drawn for a specific time; and the analysis necessary is completed and not continued over long periods of time as in watershed analysis.

Therefore, we in watershed engineering need to consider saturated flow conditions as constantly changing. Hence, to obtain accurate subsurface flow volumes, computations must be made whenever an appreciable streamline or equipotential line change is noted. These changes can be measured in observation wells, and groundwater maps drawn when a predetermined elevation change is noted. This change is defined as any change affecting the groundwater flow into, or from, the subsurface boundaries of the watershed.

Now let us itemize some of the important factors that must be considered before an analytical flow net can be constructed for a watershed:

1. Streamflow from other watersheds in the region with similar geological, hydrological, and groundwater conditions. This study may indicate basins where lower- or higher-than-average flow exists, as is the case with Pigeon Roost Creek (Sta. 34) (Figure 1). In this example it can be noted that in years of low annual runoff, the total surface runoff from Pigeon Roost Creek does not differ as greatly as in years when higher average flow occurs.

2. Streamflow from subwatersheds within the study area. Flow variation within the watershed usually indicates loss or gain areas (Figure 2). This figure shows runoff variations within the Pigeon Roost Creek watershed. Gaging stations 4 and 12 have lower annual runoff than other gaging stations in the area. Conversely, gaging stations 5, 35, and 32 have higher runoff than other stations in the area. Possible explanation for this flow variation is geological conditions within the area. All stations, except 12, are located on the southern boundaries of the watershed. Flow variations can be partially explained due to presence, or absence, of a relatively impervious boundary, within the Meridian, below the watershed, and subsurface drainage in this area is diverted from the watershed. In other words, the surface and subsurface watershed boundaries do not coincide.

3. Geological conditions within the area--structure, stratigraphy, and mineralogy. These conditions point to possible explanations for flow variation, and where these flow changes can occur. This is exemplified by the geologic map of Pigeon Roost Creek showing a fault bounding the southeast side of the watershed, which serves to explain the lower flow rates from this specific area (Figure 3).

4. Groundwater conditions within the area (depth, flow direction, number of aquifers, slope, aquifer boundaries, recharge, pumpage, permeability, and variations that do not conform to the surface watershed boundaries).

5. Groundwater configuration variations within the area must be ascertained (confined and unconfined), and a flow net constructed when variations of a predetermined magnitude occur.

After these factors are considered and boundary conditions established and described, a flow net can be drawn for the basin (Figure 4) for a specific time. This flow net will permit the analysis of subsurface water movement from, or into, a watershed area--assuming that reasonably accurate values for the permeability and aquifer thickness are known. Hence, water balance studies for watershed areas can be made if, in addition, accurate surface hydrologic information is available.

The following sample calculations are for Pigeon Roost Creek watershed, Marshall County, Mississippi, and illustrate methodology.

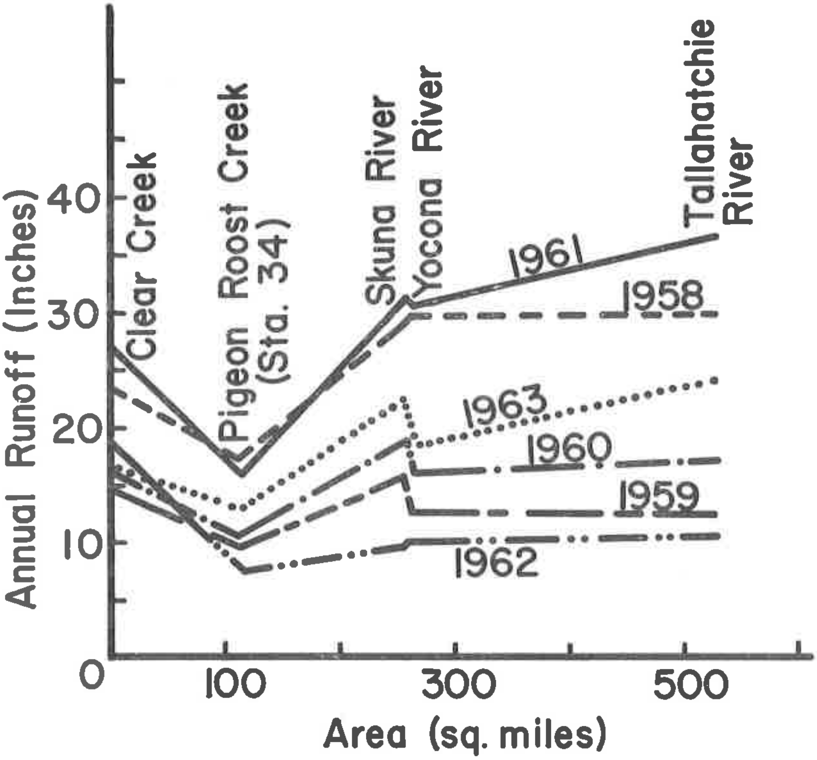

Figure 1. Annual runoff of Pigeon Roost and nearby watersheds.

FLOW EQUATIONS

$dq = K \frac{dh}{ds} dm$ (rectangles) for unit thickness of aquifer (Todd, 1960)

Assume ds = dm, the equation reduces to

- dq = Kdh (squares) for unit thickness of aquifer
- dm = distance between flow lines
- ds = distance between equipotential lines in direction of flow
- K = coefficient of permeability (L/T)
- dh = head loss between flow lines
- dq = constant flow between two adjacent flow lines
- T = 180 ft. thickness of aquifer (Meridan Sand)
- K = 50 ft/day (Meridian sand)

TOTAL FLOW per day through the Meridian formation

$dq = K \frac{dh}{ds} dm\ T$ (rectangle)

$dq = K\ dhT$ (square)

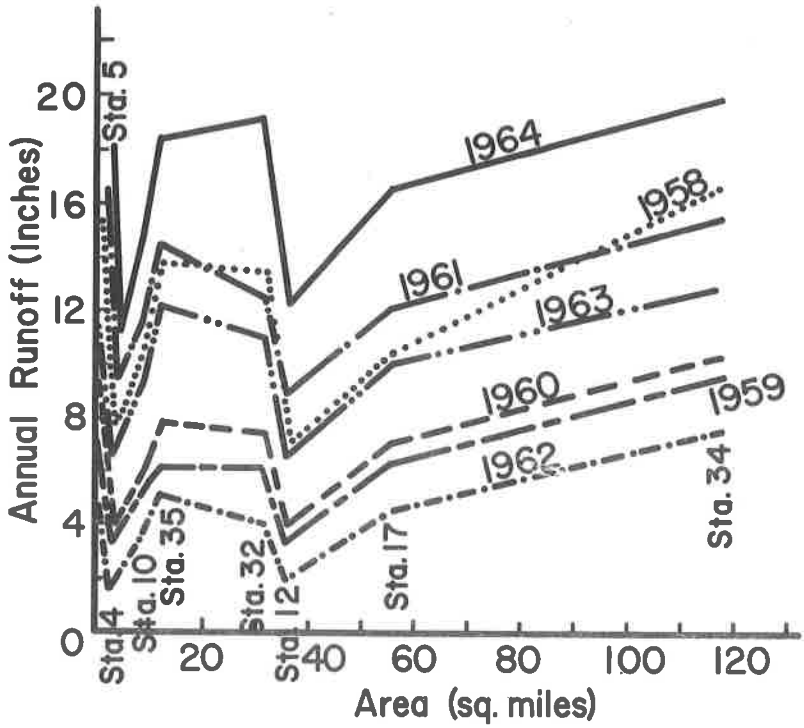

Figure 2. Annual runoff of Pigeon Roost Creek watershed.

Using the equipotential groundwater map, flow lines were drawn for the watershed, culminating in a series of squares or rectangles (flow net) around the watershed boundary. Flow directions were determined for each bounding rectangle to permit analysis of subsurface flow entering or leaving the watershed. To aid in summing the total flow, an identifying number was assigned to each square or rectangle along the boundary. The distance between the equipotential and streamlines were measured (scaled), and the flow computed through each area (Figure 4), using the applicable flow equation. Computations are simplified where the bounding flow net areas are squares.

COMPUTATIONS

Assumptions:

1. Uniform permeability, both horizontally and vertically (homogenous material)
2. Uniform thickness of aquifer (Meridian Formation)

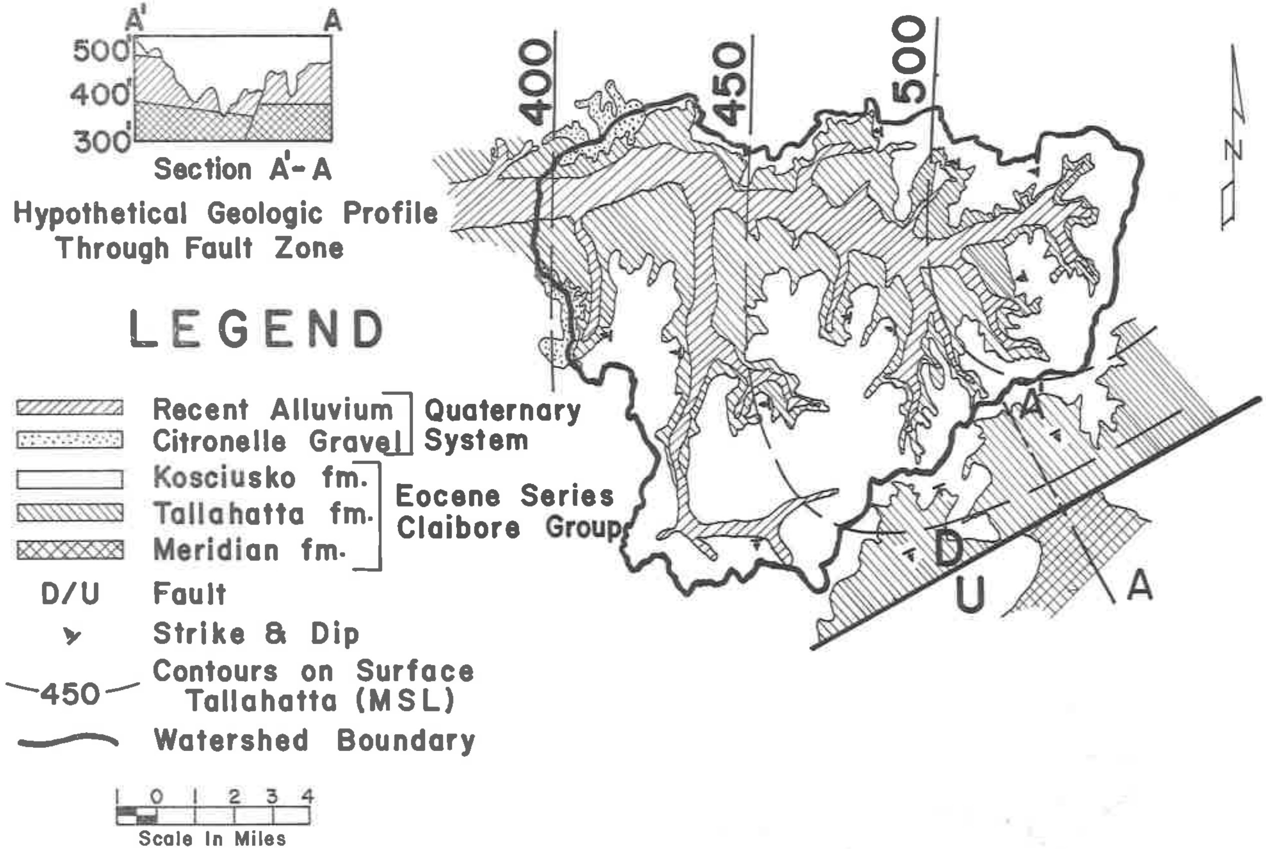

Figure 3. Geologic map, Pigeon Roost Creek watershed.

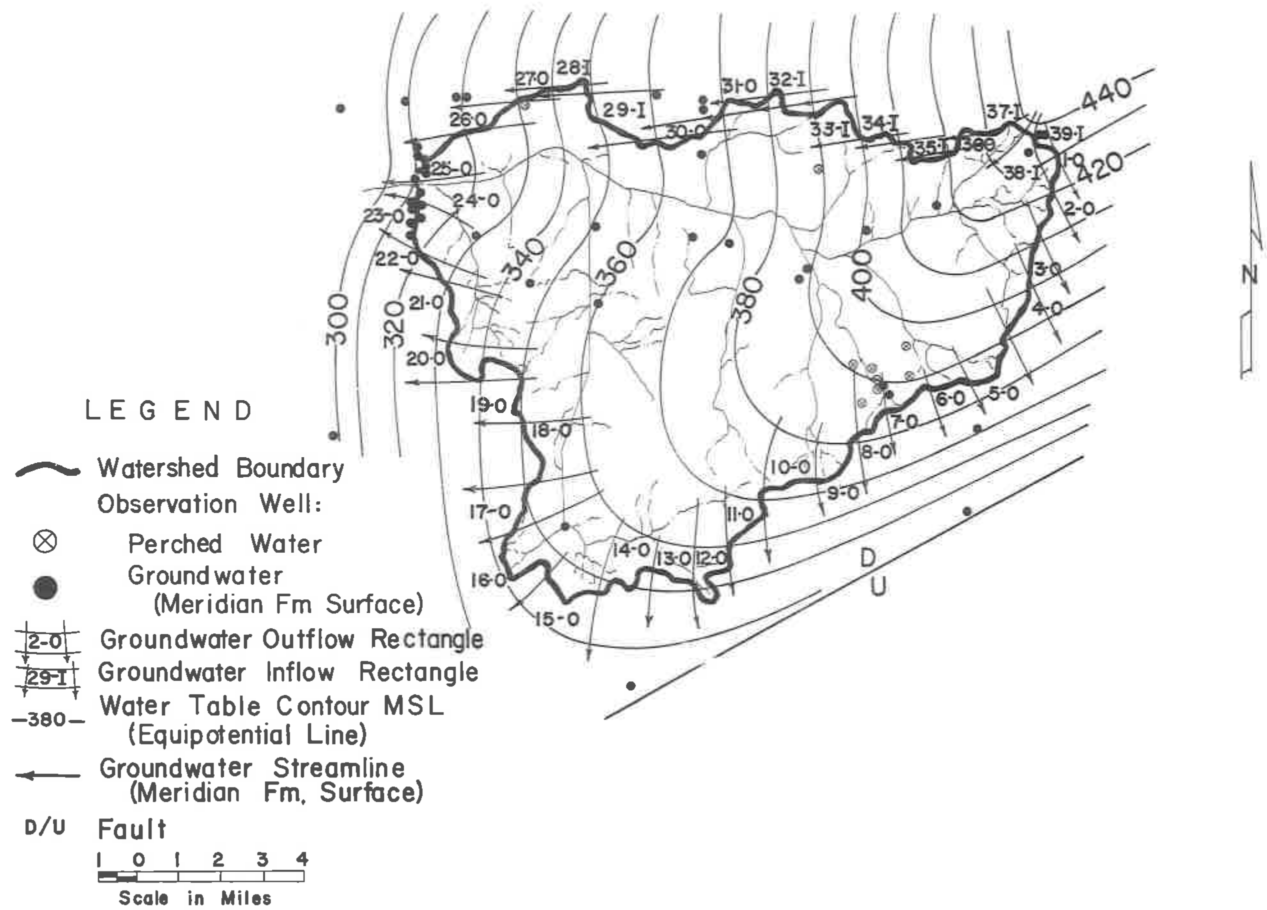

Figure 4. Groundwater flow net, Pigeon Roost Creek watershed.

Outflow from the watershed (rectangles)

Area No. (Figure 4)

1. $dq = 50 (\frac{10}{78}) 45 \times 180 = 51.8 \times 10^3 \text{ ft}^3/\text{day}$

$dq = \frac{\text{ft}}{\text{day}} (\frac{\text{ft}}{\text{ft}} \text{ or unity}) (\text{ft}) (\text{ft}) = \text{ft}^3/\text{day}$

(Square)

9. $dq = 50(10)(180) = 90 \times 10^3 \text{ ft}^3/\text{day}$

$dq = \frac{\text{ft}}{\text{day}} (\text{ft}) (\text{ft}) = \text{ft}^3/\text{day}$

Summation of all groundwater outflow areas from the watershed = 2150×10^3 ft^3/day

$2150 \times 10^3 \text{ ft}^3/\text{day} (x) 365 \text{ days/year} = 784.6 \times 10^6 \text{ ft}^3/\text{year}$

$(\text{WATERSHED AREA}) = 117 \text{ miles}^2 = 3,261 \times 10^6 \text{ ft}^2$

$$\frac{784.6 \times 10^6 \text{ ft}^3/\text{year}}{3,261 \times 10^6 \text{ ft}^2} = 0.2405 \text{ ft/year} = 2.886 \text{ area inches/year}$$

Summation of all groundwater inflow areas to the watershed = $438 \times 10^3 \text{ ft}^3/\text{day}$

$438 \times 10^3 \text{ ft}^3/\text{year} (x) 365 \text{ days/year} = (160 \times 10^6 \text{ ft}^3/\text{year})$

$$\frac{160 \times 10^6 \text{ ft}^3/\text{year}}{3261 \times 10^6 \text{ ft}^2} = 0.049 \text{ ft/year} = 0.588 \text{ area inches/year}$$

(OUTFLOW) —(INFLOW) = WATER LOSS (Pigeon Roost Creek watershed)

2.886 area inches/year minus 0.588 area inches/year = 2.298 area inches/year water loss

These computations assume that the reference flow net represented the mean condition during the year, which may or may not be precisely the case. However, they provide a usable estimated groundwater loss. Hence, the groundwater outflow, minus the inflow, from Pigeon Roost Creek Watershed through the Meridian Formation is 2.298 area inches for the calendar year 1962.

The type of analysis depicted above assumes a homogenous and isotropic medium. For aquifers that are nonhomogenous but have areas of homogenous flow, a single system of squares cannot be used. A net can be constructed in which the length of the sides of the rectangle is proportional to the differences in transmissibility. The flow lines in this case would be refracted, according to the tangent law, when flow from one subarea enters another. An analysis of this type is extremely difficult and requires a measure of the transmissibility in each subarea. Bennett and Meyer (1952) used such a system to determine the quantity of flow from an area using pumpage tests in each subarea. This method provides a more realistic value of the flow from a watershed area than pumping tests alone, which represent only a small area of the aquifer. This type of analysis is expensive and time consuming, but in any case some agreement should be obtained between isolated pump tests and the total flow net analysis.

SUMMARY

Flow net analyses of groundwater systems provide quantitative procedure for analysis and interpretation of groundwater flow in watershed areas. This hydrologic information, in addition to surface hydrology for any watershed area, will permit water balance studies. I believe that flow net analysis of groundwater problems in water balance studies, as applied to watershed engineering, should receive greater attention in the future.

This type of analysis (flow nets) needs to be considered by watershed engineers on a storm, monthly, or whatever basis necessary to determine sufficiently accurate values usable in planning, developing, utilizing, regulating, and controlling pollution of groundwater. Present day computational equipment makes possible flow net analysis with increased frequency and, hence, will provide a more accurate estimate of subsurface flow entering and leaving watershed areas.

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