



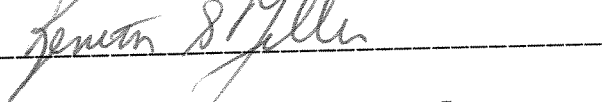
**THE STRUCTURE AND STRATIGRAPHY OF THE  
HOPEWELL FAULT BLOCK, NEWARK BASIN, NEW  
JERSEY AND PENNSYLVANIA**

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## ABSTRACT OF THE THESIS

### The Structure and Stratigraphy of the Hopewell Fault Block, Newark Basin, New Jersey and Pennsylvania

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The Hopewell fault is a predominantly normal fault striking northeast, dipping southeast, having a dip separation of 2 - 3 kilometers, and is located in the Newark rift basin, New Jersey. This study aimed to define more precisely the geometry and relative age of the Hopewell fault by examining the transverse folds in the sedimentary rocks of the hanging wall. A 14km x 5km area was mapped, and all structural data available was recorded, including bedding, joints and small faults. The deposits are Triassic Passaic Formation and, as seen in core and outcrop, consist predominantly of red mudstone and minor sandstone (playa lacustrine deposits) cyclically alternating with purple, gray and black shale (deeper-water lacustrine deposits). The cyclical alternations were produced by fluctuating lake levels driven by climatic changes with Milankovich periodicities. The non-red units were traceable across the study area. Their distribution reveals four synclines separated by three anticlines. These folds plunge approximately 10° northwest. Measured stratigraphic sections of the individual lacustrine cycles were taken at various locations within these folds to determine if there were any systematic variations in thickness relative to position within the fold or any along-strike facies changes within individual non-red units. Analyses of the joint azimuths in the area indicate

two preferred orientations 010° and 040°.

The Hopewell fault strikes both 040° and 085° within the area. The axial orientation of the transverse folds indicates a relationship with those portions of the Hopewell fault striking 045°. This may indicate that the fault was originally segmented and these segments were the result of extension toward 130°. No thickness changes were observed within any of the lacustrine cycles, and facies changes were only observed in two non-red units. This indicates that motion along the Hopewell fault "system" may not have been syndepositional, while motion probably was syndepositional along the original segments. The deposits may have been further deformed by a shift in the local extension direction towards 100° causing the formation of transfer faults striking 085° and linkage of the Hopewell fault "system".

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# Table of Contents

<b>Abstract</b>	ii
<b>Acknowledgments</b>	iv
 <b>I. Introduction</b>	 1
Purpose	1
Structure of the Newark basin	2
Stratigraphy of the Newark Basin	5
The Passaic Formation	10
 <b>II. Methods</b>	 15
 <b>III. The Study Area</b>	 17
Location	17
Description	18
<i>Stratigraphy</i>	18
<i>Facies Changes</i>	21
<i>Structure: Folds</i>	23
<i>Structure: Faults</i>	25
<i>Structure: Intrusives</i>	28
<i>Joints</i>	29
 <b>IV. Discussion</b>	 31
Facies Changes and the Timing of the Folds	31
Development of the Hopewell Fault	36
Timing of the Hopewell Fault	40
Extensional Joints	41
Future Research	43
 <b>V. Conclusions</b>	 43
References	45
Figure Captions	48
Figures	57
 <b>Appendix</b>	 93
Simplified Study Area Maps	96
USGS Topographic Maps (reduced)	101

## **I. Introduction**

### **A. Purpose:**

The study involved three parts. The first part concerned geologic mapping of the purple, gray and black lacustrine marker horizons within the predominantly red Passaic Formation in the hanging wall block of the Hopewell fault. This was done in order to identify and constrain the areal distribution of non-red cyclically deposited lacustrine units as well as define the geometry of transverse folds associated with the Hopewell fault. The field area in question includes the southwesternmost 20 km of the Hopewell fault and that portion of the hanging wall block extending a distance of approximately 5 km from the fault. (Fig. 1)

The second goal of the project was to define better the geometry of the Hopewell fault in this area and to locate any minor intrabasinal faults in the hanging wall that may also be related to its development. By closely examining the Hopewell fault, it was hoped that the origin of the transverse folding could be related to the Hopewell fault itself. The relationship between the fault and the folds could reveal whether the Hopewell fault formed as one single fault, or as several separate fault segments which joined together at a later time through lateral propagation (growth) of fault tips or the formation of transfer faults.

The third and most important goal of the thesis was to determine the extent to which structural factors affected sedimentation in this part of the Newark basin. Specifically, did the growth of the Hopewell fault influence the deposition of the

lacustrine rocks within the Passaic Formation, and if so, to what extent? Previous work had already established that climatic changes driven by Milankovitch periodicities have had a significant impact on paleolake depths and thus the nature of the lacustrine deposits throughout much of the basin (Olsen, 1986). Therefore, it was of interest to examine if and how structural and climatic controls acted in concert to produce the cyclic patterns seen in the Passaic Formation.

A related question concerns whether the Hopewell fault was active during sedimentation. Evidence that one or more of the transverse folds formed syndepositionally would indicate that this was true, provided that the folds are related to movement on the Hopewell fault. Also, by investigating the geometry of the transverse folds, I hope to provide a better view of how these structures form in association with normal faulting.

## **B. Structure of the Newark Basin**

The Newark basin is a large Mesozoic rift structure which formed as the result of continental extension due to the separation of North America and Africa during the Late Triassic and Early to Middle Jurassic (230-175 Ma). It is one of a series of such rift basins which are located along the eastern edge of North America from Nova Scotia to North Carolina (Fig. 2). The basin stretches approximately 200 km from the Hudson River in New York, continues southward through northern and central New Jersey, and then curves westward where it is connected by the narrow

neck to the northern part of the Gettysburg rift basin of south-central Pennsylvania (Fig. 3). The basin is a half graben in which the strata predominantly dip toward the northwest--towards the northwest boundary of the basin (e.g., Schlische, 1992). The northwestern margin consists of a network of predominantly northeast-striking, southeast-dipping normal faults arranged in a right-stepping relay geometry (Ratcliffe and Burton, 1985; Ratcliffe et al., 1986); these faults are referred to as the border fault system (Schlische, 1992). The Ramapo fault in northern New Jersey is one fairly continuous border fault segment that extends from the northeastern edge of the basin approximately 100 km in a southwesterly direction.

Major intrabasinal faults include the Flemington-Furlong, Chalfont, and Hopewell faults (Schlische, 1992) (Fig. 4). The Flemington-Furlong fault extends approximately 45 km in a southwesterly direction from the southwestern end of the Ramapo fault, dips to the southeast and has a dip separation of 4 kilometers (Schlische, 1992). The Hopewell fault also branches from the Ramapo fault, trends south-southwesterly for 20 km, then trends west-northwesterly for 25-30 km where it joins the junction of the Flemington-Furlong and Chalfont faults. The Hopewell fault dips to the southeast and has a dip separation of 2-3 kilometers (Schlische, 1992). Many of these faults exhibit some component of strike slip movement: slickensides found in the fault-zone rocks indicate a component of right-lateral slip (Ratcliffe and Burton, 1988). The large stratigraphic separation on the Flemington-Furlong and Hopewell faults, however, requires a

significant component of dip slip. The geometry of the Hopewell fault and its related structures, as well as the influence of those structures upon sedimentation within the hanging wall of the Hopewell fault block are the focus of this thesis.

Several folds or warps are present in the hanging wall blocks of the border fault system and the Hopewell and Flemington-Furlong intrabasinal faults (Wheeler, 1939; Schlische, 1992), (Fig. 5). Nearly all plunge gently or moderately to the northwest (towards the normal faults), and intersect the faults at high angles. The folds are therefore referred to as "transverse folds." The amplitude of the folding decreases away from the faults, and thus the folds are thought to be related to slip along the faults. In the southwest Newark basin, synclines are generally located near the centers of normal fault segments, while anticlines are located nearer the ends of such segments (Schlische, 1992), (Fig. 6). These folds are thought to be directly related to differential displacement along the normal fault segments, with synclines forming where displacement was greatest (the center of fault segments) and anticlines where displacement was the least (near the ends of fault segments) (Schlische, 1992). The present study seeks to determine the origin of folds in the hanging wall of the Hopewell fault and constrain the time of their formation and by extension constrain the age of formation of the Hopewell fault itself.

Commonly the geometry of bedding in the folds is imitated by igneous intrusions found throughout the Newark basin. This is evident in the areas around the Jacksonwald syncline, the

Sassamannsville syncline, and the folds associated with the Hopewell fault (Fig. 7). These intrusive rocks are comprised mostly of quartz-normative tholeiitic diabase and tend to intrude along planes of weakness, such as bedding planes.

### C. Stratigraphy of the Newark Basin

The Newark basin contains over 7000 m of strata divided into nine formations ranging in age from the middle to late Carnian to Hettangian (Cornet and Olsen, 1985) (Fig. 8). Most of these units show repetition in map view due to the intrabasinal faulting. The strata are predominantly fluvial and lacustrine deposits and the igneous layers are predominantly tholeiitic basalt flows (Olsen, 1980). Below is a brief description of these formations, based mainly on Olsen (1980).

At the base of the section for the Newark basin is the Stockton Formation. It is middle to upper Carnian and about 1800 meters thick in the central Newark basin along the Delaware River. The Stockton Formation generally fines upward on a large scale, but also contains several small-scale fining upward sequences. The lower part of the unit consists of sandstones overlying conglomerates in these smaller sequences; these grade upward into sequences of sandstone fining upward to mudstone near the top of the unit. Much of the sandstone in this unit is massively bedded, with several well-developed cross bed sets and ripple cross-laminations; occasional *Scoyenia* burrows may be found within the mudstone layers (Olsen et al., 1989). The Stockton Formation is interpreted to be a braided to perennial

braided stream deposit with some desiccation features and alluvial fan deposits (Olsen, 1980; Smoot, 1991).

Above the Stockton Formation lies 1100 m of the Lockatong Formation. The upper Carnian Lockatong Formation consists primarily of gray and black, cyclic lacustrine mudstone deposits with significant amounts of carbonate. Near the top of this formation, fine-grained, red, shallow-water lacustrine deposits begin to appear and become more common, although a few can also be observed near the base. The gray and black shales show a cyclicity in the amount of organic matter and the appearance of desiccation structures which indicate regularly fluctuating lake levels (Olsen, 1986). Fossils found within the Lockatong Formation also indicate fluctuating water levels: fossil fish, ostracods and conchostrachans are present in the fine-grained, deep-water deposits and reptile footprints are found in association with the desiccation features in the shallow-water deposits (Olsen et. al., 1989). Similar cyclicity is also evident in the gray and red deposits of the upper Lockatong Formation. Some red units here have dessication cracks like those found in the cycles of the lower Lockatong Formation, but they also contain crumb fabrics which result from repeated wetting and drying (Smoot and Olsen, 1985). The gray rocks also contain crumb fabrics, but they are not easy to see in outcrop.

The Passaic Formation overlies the Lockatong Formation in the central Newark basin and lies directly atop the basement rocks with no intervening Stockton or Lockatong deposits in some areas in the northeastern part of the basin. It consists of

approximately 3300 m of predominantly red mudstone and sandstone, with conglomerate near the basin margins. As in the Lockatong Formation, the Passaic Formation shows a strong cyclicity in its deposits (Fig. 9). There are numerous gray to black mudstones within the Passaic Formation and essentially formed in the same depositional environment as those of the Lockatong Formation. In areas where the Passaic Formation conformably overlies the Lockatong Formation, the boundary between the two is defined as the point where the red units become prevalent versus the gray units (Olsen, 1980). The frequency of non-red units generally decreases upsection (Olsen, 1980). Clusters of gray units have been mapped by McLaughlin (1933, 1943, 1945, 1946, 1948), and a nomenclature has been developed identifying each one by a letter (A, B, C, etc.). Thicker and thus more significant gray units have been named; these include the Warford Brook, Graters, Perkasio, Metlars and Ukrainian members (Olsen, 1980). Thickness changes have been observed within these non-red units, and they generally thicken towards both the center of the basin and the border fault system. A key assumption in this study is that these thickness changes are due to differences in sediment accumulation due to differential displacement along the border faults. If displacement is at a maximum near the center of the border fault, then the basin would have maximum relief and more sediment would accumulate in this particular area.

The Orange Mountain Basalt is the first of the Watchung Mountain basalt flows; it overlies the Passaic Formation and underlies the Feltville Formation in most of the Newark basin. It



consists of at least two distinct tholeiitic basalt flows and interbedded strata and ranges from 100 to 200 meters in thickness. The Orange Mountain Basalt is very similar chemically to the Palisades sill (Olsen, 1980). More recent work has shown that the Palisades sill consists of multiple intrusive units that are geochemically similar to both the Orange Mountain and Preakness Basalts (Husch, 1990). The individual flows of this unit typically have glassy upper and lower contacts, volcanoclastic deposits between them and one large vesicular, columnar cooling unit like those described by Tomeikoff (1940). The volcanoclastic bed which lies between the two flows ranges in thickness from <1 to 4 meters and commonly contains red, gray, or purple siltstone deposits (Olsen, 1980).

The Feltville Formation lies above the Orange Mountain Basalt and below the Preakness Basalt. It consists of approximately 170 meters of red, fine-grained sandstones and gray to white feldspathic sandstones. It also contains one thick limestone unit in its lower half which is black to white and laminated. This limestone is bounded by two gray sandstone units which are 1-7 meters thick. These three beds are laterally continuous throughout the unit. The red units within the Feltville Formation are much like those of the Passaic Formation, and the buff to gray, feldspathic sandstone units highly resemble the deposits of the Stockton Formation (Olsen, 1980).

The Preakness Basalt overlies the Feltville Formation and underlies the Towaco Formation. It is approximately 250 meters thick but can be up to 500 meters thick locally. It is comprised of

at least three significant individual flows, with local concentrations of multiple-flow, vesicular basalt flow sequences and thick beds of angular, vesicular basalt. The lowest of the larger flows is approximately 100 meters thick and contains a complete Tomeikoff sequence with well-developed cooling columns. Above this flow lies a thin red siltstone unit which separates it from the second large basalt flow. In some areas this second flow is thick (up to 250 meters) and is the last flow to appear, while in other areas it is thinner and is overlain by another thin red siltstone unit and then a third basalt flow. The entire unit is typified by a lack of a continuous vesicular portion, a significant thermal metamorphic aureole in the surrounding sediments, and the coarseness of the flows (Olsen, 1980).

The Towaco Formation overlies the Preakness Basalt and consists of approximately 340 meters of red, gray, and black strata deposited in cycles. These cycles are typically symmetrical, with black, calcareous siltstone bounded above and below by fining-upward sequences of gray sandstones. This entire package is bounded above and below by red sandstones and siltstones. The cycles average 35 meters thick and are significantly thicker than any similar cycles found within the Lockatong or Passaic Formations. At the top of the Towaco Formation is a thin (1 meter) volcanoclastic unit containing altered glass, feldspar, and augite in a brown radial natrolite matrix (Olsen, 1980).

The Hook Mountain Basalt is approximately 110 meters thick and is the thinnest of the basalt flows in the Newark basin. It is quite different chemically from the Orange Mountain or

Preakness Basalt, as it is relatively very depleted in  $K_2O$ , Sr, and Rb, and contains a much higher FeO/MgO ratio. It is generally recognized as having two distinct flows: the lower averages 57 meters thick and contains a distinct Tomeikoff sequence with well developed cooling joints whereas the upper flow averages 40 meters thick with no such structures (Olsen, 1980)

. The Boonton Formation is the uppermost unit of the Newark basin section and bears little resemblance to the other units in the basin. It consists of approximately 500 meters of red, brown, gray and black clastic deposits. The lowermost units are comprised of largely dolomitic siltstone. These beds are overlain by alternating thick red sandstones (averaging 35 meters thick) and thinner gray to green siltstones (averaging 2 meters thick). The uppermost units of the Boonton Formation contain the Boonton Fish Bed, a meter of calcareous siltstone, as well as several gray to brown conglomeratic units averaging 0.5 meters thick (Olsen, 1980).

#### **D. The Passaic Formation**

Because the Passaic Formation comprises the vast majority of the deposits within the study area and is the focus of questions posed herein, it warrants a closer examination. As stated above, it has an estimated maximum thickness of 3500 meters and is comprised largely of red mudstone with minor sandstone and conglomerate. It also contains deposits of purple, gray and black shales which represent periods of deeper-water deposition occurring at semi-regular intervals. The Passaic Formation has

been of some interest in that the presence of these cyclic gray units allows the examination of structural controls and climatic effects upon sedimentation within the basin through time.

The Passaic Formation has been continuously cored in five separate drill cores located in Titusville, Piscataway, Somerset, Weston Canal, and Martinsville, New Jersey. (Fig. 10) The upper and lower parts of each core overlap with those of adjacent drill sites, allowing a composite stratigraphic section to be constructed; the Titusville core contains the oldest units, the Martinsville core contains the youngest.

Each core has been examined and described in detail, and a nomenclature was developed, identifying the individual non-red units in the Passaic Formation (Olsen and Kent, 1990; Olsen et al., 1992; Olsen et al., 1993). From the bottom of the Titusville core, where the red units of the Passaic Formation begin to dominate over the gray and black units of the Lockatong Formation, the non-red units, or members, were originally assigned letter names (A, B, C, etc.). Many of these letter names have since been replaced by proper names, usually related to the site of a "type" section of the member. Each "member" contains both the non-red units and the overlying red units. The "A" unit near the base of the Titusville core is also known as the Tumble Falls Member after the location of its type section. Unit "B" is called the Walls Island Member, unit "D" the Warford Brook Member, units "G" and "H" the Grater's Member, and at the top of the Titusville core lies the "N" unit and part of the "O" unit, both of which comprise the

Perkasie Member. The base of the Rutgers drill core section overlaps the Titusville core from units "I" through the Perkasie and continues up through unit "Y" (representing the 18th member of the Passaic Formation). In this section, the "P" unit is named the Neshanic Member, the "V" unit is the Kilmer Member, the "W" unit is the Livingston Member, and the "X" unit is known as the Metlars Member. The Somerset drill core section overlaps the Rutgers core from unit "U" through "Y" and continues up through unit "HH", which is also called the Ukrainian Member. The only other named gray cycle in the Somerset section is unit "GG", which is called the Cedar Grove Member.

The strata of the Passaic Formation are generally classified as shallow-water to moderately deep-water lacustrine deposits with a wide range of lacustrine environments being represented (Olsen, 1986; Olsen et al., 1993). The depth of the lakes at any given time is recognized as the dominant controlling factor of the sediment fabrics and colors of the deposits. The red mudstones which comprise the bulk of the formation are interpreted as shallow-water, playa lake deposits (Smoot and Olsen, 1988). In these units mudcracks are frequently found as well as evaporite mineral deposits and sometimes the fossil footprints of land animals. The red color itself is due to the oxidation of iron to hematite which indicates periods of subaerial exposure. These units reflect times when the lakes became very shallow and/or dried up completely. The darkest gray and black mudstones almost never exhibit any evidence of bioturbation or desiccation, and are typically high in organic content (Olsen et al., 1989). This

indicates that these units were deposited under deep-water conditions. The lakes were deep enough for the water column to become oxygen-depleted near the bottom of the lake, as this condition would prevent the bacterial decay of organic matter being deposited (Olsen, 1990). The purple and light-medium gray units represent times of moderate lake depth in which the water column may have been at least partially stratified and oxygen-poor near the bottom. With increasing water depth and the reduced oxygen levels at the bottom of the lakes, iron begins to become reduced. This also contributes to the gray and black color of the deeper-water deposits.

The various lacustrine environments reflected within the Passaic Formation have been classified according to color, desiccation features, organic content and the presence or absence of laminations or microlaminations. Each unit is assigned a depth rank based on these parameters (Olsen, 1986). A depth rank of one indicates desiccated, playa lake conditions and a depth rank of six indicates very deep, stratified, oxygen depleted water (Fig. 11).

The alternations of red and non-red units within the Passaic Formation indicate that lake depths fluctuated with time in a regular fashion (Olsen 1986; Olsen et al., 1989). The fundamental cyclic unit recognized in Passaic Formation is the Van Houten cycle, which represents one full lake-depth cycle from highstand to lowstand to highstand again. Depth rank curves show that this wet to dry to wet climate trend is exhibited on larger scales. Within succeeding Van Houten cycles, the highstand facies becomes progressively wetter and thicker and then drier and

thinner. This higher-order cycle (compound cycle) consists of five Van Houten cycles. An even larger compound cycle consists of 20 Van Houten cycles. The ratio of 1:5:20 (Van Houten cycle:1st compound cycle:2nd compound cycle) very closely approximates the ratio of the periodicities of the average precession cycle to the first two modes of eccentricity (21,000:109,000:403,000)(Fig. 13).

A more quantitative method for analyzing cyclicity is to conduct Fourier analysis on a depth ranked curve to obtain a power spectrum, which shows the thickness of the most prominent cycles. This process is shown in figure 12 for the Titusville core, located within the study area. To convert these peaks to time, it is assumed that the most prominent high frequency peak corresponds to the precession cycle of 23,000 years. From this value, a sedimentation rate of 0.000253 m/yr is calculated. Applying this sedimentation rate to the other cycles yields periodicities in time of 100,194 years for the 25.31 m-thick-cycle, 240,871 years for the 58.52 m-thick-cycle, and 411,862 years for the 104.04 m-thick-cycle. Thus, the power spectrum contains peaks at 23,000, 100,000 and 411,000 years--corresponding to the Milankovitch cycles of precession and eccentricity. Therefore, it can be concluded that cyclic changes in climate driven by variations in solar insolation had a direct and pronounced effect upon the sediment facies observed within the Passaic Formation.

The Passaic Formation allows detailed examination of the history of the basin's formation. This not only includes clues to climate but also possibly to tectonic activity as well because lake

levels and their deposits may also have been affected by changes in the structure of the basin itself. If faulting around and within the basin occurred during the deposition of the sediments, it should affect the stratigraphy of the synrift units. It has been documented that strata deposited during faulting exhibit thickening both toward the fault itself and towards the center of the fault segment, where maximum displacement occurs (Schlische, 1992), (Fig. 14). In the case of the Passaic Formation, this thickening could be observed in a single Van Houten cycle or more likely a cluster of Van Houten cycles. By measuring the thickness of one cycle at various points from the edge to the center of the basin, one could determine the extent of thickening and thus the extent to which faulting (tectonics) affected the deposition of the Van Houten lacustrine cycles. This phenomenon has already been documented from core data. The lacustrine cycles of the Passaic formation show definite thickening trends toward the axis of the basin itself and in the direction of the border fault system, based on measurements taken directly from the five cored sections (Schlische et al., 1991; Silvestri and Schlische, 1992). Variations in cycle thickness in the study area will be discussed in more detail below.

## **II. Methods**

Mapping of the non-red units within the Passaic Formation was accomplished by walking through every available stream in



the area in order to find outcrop. Roads and railroad tracks were also traveled on foot to supplement the data acquired in streams. At each outcrop, attitudes of bedding, joints, and minor faults (where found) were taken. The occurrence and location of non-red units was recorded on field maps. Stratigraphic sections were measured where outcrop provided good exposure of both upper and lower contacts with the red mudstone. These stratigraphic sections were measured both directly from the outcrop (when possible) and using a modified Jacob's staff (when the outcrop was located on the stream bed). It was possible to follow some of the non-red units across the study area directly from the topographic maps, as they tended to create "strike ridges" in the topography. By comparing the measured sections and bedding attitudes it was possible to identify a single non-red unit at separate outcrops and "connect" or correlate the unit between the two locations and across the map. For example, two stream outcrops of the Perkasio Member in the Rocky Hill quadrangle can be correlated: in both outcrops, bedding strikes easterly, and measured sections of the outcrops show the stratigraphy of the unit to be the same (Fig. 15). Location and identification of the non-red units in the Pennington quadrangle was aided by control from the Titusville drill core. Units found in the core were projected back to the surface and this information was used as a starting point for mapping this area.

The Hopewell fault and other, smaller intrabasinal faults were also mapped in much the same way. The location of such faults was pinpointed by recognizing lithologic changes, fault

gouge deposits, and sigmoidal (deformed) joint sets. In the case of the Hopewell fault, lithologic changes are major, involving a shift from Passaic to Locatong or Stockton Formation deposits, while for minor, intrablock faults, lithologic changes most often involved the presence or absence of a non-red unit. For example, if a non-red unit is found in one stream and strikes east-west, one may expect to find that unit in another nearby stream either to the east or to the west. If there is good exposure in the other stream and yet the non-red unit is not found, then a small-scale fault must lie between the two streams, barring any evidence for folding between the outcrops. Smaller, intrablock faults may also be located by the presence of small-scale drag folds.

By using these techniques in every stream, road and railroad transect, the non-red units of the Passaic Formation were mapped as seen in plates 1-5

### **III. The Study Area**

#### **A. Location**

The area of immediate interest lies in the mid-west part of New Jersey and extends across the Delaware River into eastern Pennsylvania. The Hopewell fault marks the northwestern border of the study area, and mapping undertaken in the hanging wall block extends an average of 6 kilometers from this fault. On USGS 7-1/2' quadrangle maps, the area begins in the Rocky Hill, NJ, quadrangle, extends southwestward through the Pennington, NJ, map and on through the Lambertville, PA, quadrangle. The area

thus encompasses approximately 175 square kilometers. One of the aforementioned drill cores was taken in the area of the township of Titusville, near the edge of Washington's Crossing State Park in the Pennington, NJ quadrangle (Fig. 16).

## **B. Description**

**Stratigraphy--** The areal distributions and bedding orientations of all the non-red units located within the study area can be found by examining the maps included. What follows is a brief description of those specific members that could be identified in outcrops and their locations.

The first units of mention are those which can be projected from the Titusville drill core back to the surface in the study area. The lowest units stratigraphically in this core that was located in the study area are members "G" and "H". In core, these units both contain three gray layers, and this pattern was observed at the southern edge of the study area in Woolsey Creek. Above this is unit "I", which contains many thin purples and grays but is recognized by a single gray unit of moderate thickness. Units "K", "L" and "M" were not located along the projected line from the drill site, but are tentatively placed in streams to the east. The Perkasio member, in core, consists of two thick gray and purple units, and was found in the western tributary of Jacob's Creek in an outcrop consisting entirely of gray siltstone.

To the west in the Lambertville quadrangle, the Perkasio member was identified in one outcrop which consists of a thick layer of gray shale (0.732 meters) above black mudstone (1.22

meters thick) (Fig. 17). The bottom of the non-red unit in this outcrop could not be found. Two units to the south of this may be parts of Member "K", as their position on the map--when corrected for horizontal distance and dip angles--places them approximately 650 meters (+/- 50 meters) below the Perkasio Member stratigraphically. One other non-red unit in Washington's Crossing Park (PA side) could not be positively identified but may be the Metlars Member, based on the measured section of this outcrop. It consists of two individual gray and dark gray shale units separated by a layer of red mudstone.

As previously described, in the Pennington quadrangle good control is provided by the Titusville drill site and the Perkasio, "I", "K", and "L/M" Members can all be identified from the information in the drill core. The Perkasio Member is located parallel to a tributary of Jacob's Creek adjacent to the drill site. It is also located at the mouth of the reservoir in the Baldwin's Wildlife Management area, and is marked by a double gray unit in the bottom of nearby Stony Brook. The Perkasio Member then follows bedding within the Hopewell syncline and can be found again in a northeast-trending tributary of Honeybranch Creek in the Princeton quadrangle.

A measured section also taken from Stony Brook in the Hopewell Valley Country Club--north of the Perkasio member--reveals three separate gray units; the uppermost gray unit also contains a layer of dark gray to black mudstone (Fig. 18). The map position and stratigraphy of this outcrop (in the bottom of the stream) suggest that this is the Metlars Member. Other non-

red units in the Pennington area are difficult to identify, such as the cluster of gray and purple units found in northern Jacob's Creek. Here, the presence of intrabasinal faults has acted to thicken parts of the stratigraphic section, making it difficult to place these units in their proper position within that section.

Finally, in the Rocky Hill quadrangle, the Perkasio Member can be identified both by its measured section and by tracing the unit across the Newark basin from outcrops in the New Brunswick, New Jersey region. In this area, the Perkasio Member contains the thickest single gray unit found anywhere in the study area (approximately 30 meters) and can be easily traced directly across the map (Fig. 19). North of this, a double-gray unit is located and identified as the Neshanic Member. This unit is found everywhere north of the Perkasio member outcrops and everywhere consists of two separate gray shale units. Its map distance to the north of the Perkasio member places it approximately 76.25 meters stratigraphically above the Perkasio member (+/- 1.75 meters). This roughly concurs with core data from the Rutgers drill site.

Farther north, the Metlars Member is again encountered in a large outcrop near the intersection of Aunt Molley Road and Route 518. Here the measured section (from bottom to top) shows a thin (1 meter) gray shale, 10 meters of red mudstone, then a complex gray unit which contains a dark gray mudstone near its base and alternating light and medium gray shales above this (Fig. 20). This second gray unit is 5 meters in total thickness. The Metlars

Member can then be correlated across the Newark basin map from other outcrops near New Brunswick.

In the Rocky Hill quadrangle, there are several smaller, isolated purple and/or gray units which cannot positively be identified. Many of these units disappear over short distances, and most of them are too thin to be identified by comparison with available core data. However, members "I" and "K" might be placed as shown on the map (plate 1) by using their relative location on the map and the dip angles measured at their outcrops to calculate the stratigraphic separation between these units and others like the Perkasio or Metlars Members. Any identification of this sort is tentative as small changes in dip angles translate into large differences in map separation.

**Facies Changes--** For most of the study area, few facies changes were observed along strike within any one non-red lacustrine cycle, with two exceptions. The first example involves a non-red unit which outcrops in the Baldwins Creek system in the Pennington quadrangle (Fig. 21). One outcrop along Burd Road contains only purple chips (float) in a narrow band approximately one meter wide measured parallel to the road surface. These chips are flanked on either side by bright red chips and sparse outcrop of red rocks. In a second location in Baldwins Creek itself, an isolated outcrop of purple siltstone was found in which bedding strikes roughly easterly and dips gently northward. The purple beds project across the map to the west to the location of the purple chips by Burd Rd. This unit appears again along the

railroad tracks north of the town of Pennington itself. The small outcrop here consists of purple siltstone, underlain by a layer of grey shale. Both layers are less than one meter thick.

The three outcrops are all located on the western limb of the Pennington syncline, with the purple and grey outcrop located nearest the hinge region. This might suggest that local lake level was deeper near the hinge region of the Pennington syncline, but these changes may also merely be part of a larger regional trend of deepening.

The second facies change was observed in the Perkasio Member in the Rocky Hill quadrangle. The section shown in Figure 19 is typical of all outcrops of this unit in the Rocky Hill quadrangle. Here, the Perkasio Member consists of approximately 30 meters of medium grey siltstone with little or no color changes from top to bottom. There are some minor changes in grain size. These measured sections of the Perkasio Member are very different from those found in the Rutgers and Titusville drill cores, in which the Perkasio consists of at least four distinct non-red cycles containing purple, grey and black shale units. Each of the cycles in the two cores are separated by red mudstone units. Comparing the outcrops of the Perkasio Member in New Brunswick and Rocky Hill, (Fig. 22) based on these facies changes we can conclude that while lake levels near New Brunswick fluctuated through time and were at some time deeper than they were in Rocky Hill, lake levels near Rocky Hill were consistently of moderate depth during the deposition of the Perkasio Member.

### **Structure--Folds:**

In Pennsylvania, the study area is roughly divided into two very distinct regions by Jerico Mountain. The overall structure can be observed in the topography of this diabase intrusion, which defines a broad syncline. South of the mountain, bedding attitudes are consistent with this type of structure. Strike of bedding changes from approximately  $095\text{-}100^\circ$  in the west, to  $090^\circ$  in the central portion of the Lambertville quadrangle, to  $075\text{-}080^\circ$  in the east. Dips appear to remain relatively constant at around  $17\text{-}22^\circ$  north. In the New Jersey portion of the map, we observe the western half of Baldpate Mountain. North of this, in Moore's Creek, outcrop data suggest the presence of a fold which originates against the Hopewell fault and swings westward towards the Delaware River where it appears to terminate, possibly against Jerico Mountain (Fig. 23). In the Pennington quadrangle, there are at least three intrablock faults and three synclines intertwined to produce the map distribution of non-red cycles here. In the far western portion of the Pennington quadrangle, bedding in both Fiddler's Creek and Washington's Crossing State Park strikes about  $070^\circ$ , although some variations are observed. In particular, bedding attitudes in Fiddler's Creek seem to parallel the trend of the stream itself and define a subdued synform-antiform pair. These folds have axial traces trending towards  $320^\circ$ . The fold axes plunge approximately  $10^\circ$  in this direction. Bedding in the central portion of the quadrangle maintains the generally east-west trend across to Baldwin's Creek and Stony Brook, and dips range between  $15^\circ$  and  $20^\circ$  to the



north. Farther to the east, outcrop availability begins to decline, but there is enough information to suggest the existence of two synclines. The axial trace of the first syncline originates against the Hopewell fault, and runs through the center of Pennington Mountain and passes north of Pennington proper, thus trending approximately  $315^{\circ}$  although the stereonet contour of bedding indicates a trend of  $350^{\circ}$ . This discrepancy is probably due to an insufficient number of bedding measurements. The second syncline is centered around the Honeybranch Creek watershed area. Its axial trace trends approximately  $325^{\circ}$  and the axis plunges  $10^{\circ}$  in this direction. Again it should be mentioned that the lack of outcrop in these areas may be affecting these figures, but the map-scale geometry of bedding appears to concur. Continuing eastward, the lower half of the Rocky Hill quadrangle contains the southern limb of the largest transverse syncline in the Newark basin. This syncline's northern limb is located across the New Jersey/New York border approximately 100 km from this part of the study area. Here, bedding terminates against the Hopewell fault and strikes generally east-west. All bedding in this area dips gently (about  $20^{\circ}$ ) to the north. The units which are found here can be traced northeastward through New Brunswick and to some extent on up into New York State as part of a much larger overall, basin-scale synform. Within this area, however, at least one very minor syncline may be present near the town of Hopewell on the western edge of this map, as bedding at a large outcrop near the intersection of Aunt Molley Road and Route 518 strikes northeast and dips more steeply than at the majority of

outcrops in the area ( $35^\circ$ ) (Fig. 24). This may indicate the presence of a drag folding syncline associated with displacement of the Hopewell fault.

As previously discussed, at least four synclines are located within the study area, and they are separated by three anticlines, which are not as well expressed. The synclines of note are tentatively called, from west to east, Jerico Creek syncline (in the Lambertville quadrangle), the Fiddler's Creek syncline (a subdued structure in the western Pennington quadrangle), the Pennington syncline (central Pennington quadrangle), and the Hopewell syncline (eastern Pennington quadrangle). Bedding measurements from each of these synclines were analyzed to determine the attitude of each individual fold.

Pi-diagrams of the four folds reveal that the Jerico Creek, Fiddler's Creek, and Hopewell synclines all plunge approximately  $10^\circ$  to the northeast (between  $315^\circ$  and  $335^\circ$ ). The Pennington syncline plunges  $10^\circ$  towards the north ( $350^\circ$ ) (Fig. 25).

Each of these folds plunges gently toward the border fault, in keeping with the characteristics of other transverse folds located within the Newark basin. There are differences in the direction of plunge of each fold which may be related to segmentation of the Hopewell fault. This will be discussed later.

### **Structure-Faults:**

The Hopewell fault can be located by using one of two criteria: a change in lithology from predominantly red mudstones

(Passaic Formation) to predominantly grey and dark grey calcareous shales (Lockatong Formation) or a change in bedding attitude from steeply dipping to gently dipping. The latter is a viable method because bedding in rift basins dips more steeply as the normal fault is approached, then quickly becomes shallow again as the fault is crossed. This is indeed the case here: bedding increases in dip to about  $30-35^{\circ}$  adjacent to the fault and then changes abruptly back to  $15-20^{\circ}$  with increasing distance from the fault. An example of an area where the latter method was used is the Pidcock Creek system in the Lambertville quadrangle. Here, available outcrop is scarce. However, three sites helped to locate the position of the Hopewell fault (Fig. 26). Near the mouth of the stream system, at the northwestern base of Bowman's Hill, bedding strikes  $018^{\circ}$  and dips  $42^{\circ}$  west. Such steep dip indicates that the fault must be not far to the north of this outcrop. Approximately 0.5 km farther upstream bedding dips very shallowly northward (about  $10^{\circ}$ ). This would indicate that the fault lies nearby to the south. The Hopewell fault is therefore placed between these two outcrops. A third outcrop, located about 0.75 km south of Buckmanville along Rt. 232, contained many large pieces of float which clearly consisted of fault gouge. The float boulders contained large amounts of recrystallized calcite and possibly gypsum intermixed with red siltstone and mudstone. This fault gouge was very limited in extent along the roadside and therefore should represent the location of the Hopewell fault itself.

The overall map trace of the Hopewell fault begins (in the study area) in the center of the western edge of the Lambertville quadrangle. Here it trends approximately  $080^{\circ}$  until it passes to the north of the western "limb" of Jerico Mountain. It then turns northward at about  $025^{\circ}$  and continues in this direction for approximately one kilometer before again turning eastward. It then trends  $085^{\circ}$ - $090^{\circ}$  on into the Pennington quadrangle, passing just north of Baldpate Mountain.

In the Pennington quadrangle, the Hopewell fault parallels the Hunterdon/Mercer county line until it approaches Pennington Mountain. It then takes a small jog to the north around the northern "finger" of Pennington Mountain and begins to run northeastward. It crosses over into the Hopewell quadrangle near Stony Brook Road and at this point trends  $060^{\circ}$ . The fault continues along this line through the Hopewell quadrangle and into the Rocky Hill quadrangle. It then swings more northerly and finally strikes approximately  $025^{\circ}$ . In the Rocky Hill quadrangle, the Hopewell fault parallels the eastern edge of Pheasant Hill.

There is also evidence for several smaller intrabasinal faults throughout the study area. In Moore's Creek, for example, there are somewhat extensive outcrops in which bedding that dips southwest, the opposite of what is expected. This may indicate the presence of intrabasinal splay faults off the Hopewell fault in this vicinity. There is also evidence for minor faulting in Fiddler's Creek near Route 29. Here, several outcrops in the stream bed exhibit sigmoidal joints running in bands which trend  $195^{\circ}$  and are approximately 0.3 meters in thickness. Similar observations

were made near the mouth of the streams in Washington's Crossing State Park (on the New Jersey side of the Delaware River). These faults may then splay off the Hopewell fault and strike approximately due south, cutting through Baldpate Mountain and running through both Fiddler's Creek and the western edge of Washington's Crossing Park.

Moving slightly to the east, in the northern portion of the Jacob's Creek system we observe significant differences in bedding attitudes between two separate tributaries. Bedding strikes  $035-045^{\circ}$  in the western tributary, and  $065-070^{\circ}$  in the eastern tributary (Fig. 27). Evidence for the existence of a fault in the eastern tributary includes slickensides and highly fractured rocks containing dense, closely spaced joints. These three factors would seem to indicate that drag folding has occurred along this fault splay.

### **Structure-Intrusives:**

In the northern part of the Pennington quadrangle, there are two diabase intrusions which are shaped like three-leaf clovers: Pennington Mountain in the central portion of the map and Baldpate Mountain in the west. The latter extends westward into the Lambertville quadrangle as far as the Delaware River. This clover shape appears to be due to the influence of the surrounding structures. The presence of minor, intrablock faults may explain this. If the intrusives are treated as phacoliths like those found within the Jacksonwald syncline, deformation by movement along the intrablock faults could produce the

morphology observed. The significance of this possibility will be discussed below.

The diabase intrusion known as Jerico Mountain in the Lambertville quadrangle, however, was emplaced near the hinge of a transverse syncline and therefore conforms to the shape of this fold. It can therefore be concluded that these magma bodies intruded through weaknesses between bedding planes and their morphology was therefore dictated by the overall preexisting structure of bedding, not folding during emplacement.

**Joints-** Over 1600 joint measurements were taken over the course of the mapping; most readings consist only of a strike direction, as much of the available outcrop is under water in stream beds. For the purposes of this study, which aims to analyze opening directions and local tectonics, these are sufficient.

The study area was divided into five regions, and joint data from each region were plotted on rose diagrams. The first region contains that part of the study area found only in the Rocky Hill and Hopewell quadrangles. The rose diagram of joint azimuths reveals that the majority of the joints strike between  $010^{\circ}$  and  $020^{\circ}$  or north-northeast (Fig. 28). This direction parallels the strike of the segment of the Hopewell fault which runs through this region and indicates that the extension direction was locally perpendicular to the fault.

The second area covers the Honeybranch Creek system, the town of Pennington itself, and the study area included in the Princeton quadrangle. The rose diagram for this area shows that

the joint directions are more varied, but the majority of joints trend between  $010^{\circ}$  and  $060^{\circ}$  (Fig. 29). However, there is no significant "spike" within any one particular  $10^{\circ}$  bracket.

The third area runs adjacent to the Hopewell fault from the town of Hopewell to the western edge of the study area and also includes the areas around Pennington, Baldpate and Jerico Mountains. The rose diagram shows no distinct trend of joint azimuths. Measurements are dispersed from  $270^{\circ}$  through  $090^{\circ}$  with only minor concentrations around  $020^{\circ}$ - $030^{\circ}$  and  $060^{\circ}$ - $080^{\circ}$  (Fig. 30).

The fourth area lies in the southwestern part of the Pennington quadrangle covered in the study. Here, the rose diagram shows the most concentrated set of joint data of all the regions. The vast majority of joint azimuths here trend between  $010^{\circ}$  and  $020^{\circ}$  (Fig. 31)

. The fifth area contains the very southwestern portion of the study area, specifically the Jerico Creek system. The joints in this area again trend largely to the northeast and cluster around the  $040^{\circ}$ - $050^{\circ}$  bracket with a bit of variation (Fig. 32).

#### IV. Discussion-

##### Facies changes and the timing of the folds-

I have reported that in two distinct cases, facies changes were observed in individual non-red lacustrine cycles along their map traces. These facies changes may reveal information about

the timing of local tectonic activity: namely, the formation of the transverse folds observed in the Passaic Formation of the Hopewell fault block.

Schlische (1993) described the formation and growth of a normal fault and its associated rift basin. He stated that a "growth" fault will increase the length of its map trace as it increases its displacement. This is based on analyses of normal faults comparing the trace length of the fault to its maximum displacement. Plots of length vs. displacement reveal that smaller faults have smaller maximum displacements, and larger faults have larger maximum displacements, suggesting faults grow both in length and in displacement as faulting proceeds. This causes the rift basin to grow both wider and deeper through time as the normal (border) fault continues to grow. The resulting geometry is that of a broad syncline which plunges gently towards the border fault.

The formation of such a basin would act to divert rivers into the basin itself, and it will begin to fill with water and sediment. If sedimentation occurs as movement along the fault continues, the basin deposits will develop certain geometries unique to this situation: the deposits will increase in thickness as one moves towards the border fault, younger deposits will onlap older deposits near the edge of the basin, and the deposits will increase in dip towards the border fault (Fig. 33). Also, if the fault is growing while sedimentation is taking place, then the basin and thus the syncline will grow during sedimentation, becoming wider and deeper through time. Given constant rate of faulting and



sedimentation, this will begin to have an affect on the sediment facies once the basin is deep enough to support a lacustrine environment: a single lacustrine unit's lithofacies (in this case primarily color) will be different depending upon location in the basin. Deeper-water facies will be present near the center of the fault segment, and shallow-water facies near the basin's edges.

Initially, however, a rift basin may not be able to support a lacustrine system. When the rift first develops, the basin is both shallow and small. It quickly fills up with sediment and what it cannot accomodate gets passed out of the basin. Later, as the basin continues to deepen and widen, the sediment supply will no longer be sufficient to keep the basin filled and water (a lake) will begin to fill in the remaining space. It is at this stage that the deposits are the most "sensitive" to tectonically induced changes in basin depth and different facies are created within a single unit (Schlische and Olsen, 1990; Schlische, 1991, 1993) (Fig. 34).

The primary control upon deposition we are concerned with when discussing the Passaic Formation is that of water depth. As discussed in a previous section, water depth can dictate the color, organic content, mineralization, and other characteristics of the sediments being deposited such as bioturbation and desiccation structures. For this discussion, I will focus mainly upon the color of the rocks.

If we consider a lacustrine environment in a rift basin having the geometry of a gently plunging syncline, it is easy to see that such a lake should have its greatest depths where the topographic relief is also greatest. In this case, that area includes

both the hinge of the syncline and the point of maximum displacement along the fault. Movement in any direction outward from this point and from the fault will be towards shallower water--towards "drier" depositional environments. Any single unit deposited under these conditions is likely to have the deeper-water deposits described by Olsen (1990) near the center of the basin, the moderate-depth deposits ringing them, and the shallower-water deposits outlining the edges of the basin.

If movement on the fault ceases at this point, lacustrine deposition would continue until the basin is filled, when fluvial deposition would take over (Schlische and Olsen, 1990; Schlische, 1991). Between these two points in time, however, water depth would still have the same controls over the deposits, and any change in facies would be predominantly attributable to changes in lake depth. It has been demonstrated that paleolake depths were indeed affected by climatic changes driven by Milankovitch orbital periodicities and that water depth changed through time in a cyclic fashion. Olsen (1986) and Olsen et al. (1989) showed that, in fact, many of the Milankovitch periodicities were directly represented in the stratigraphy of the Passaic Formation. Thus, the non-red lacustrine cycles found in the Passaic formation were at least partially brought about through climatic controls.

This situation will be significantly altered if faulting does not cease at or near the onset of sedimentation. Synsedimentary faulting means synsedimentary formation of the transverse synclines. This process would cause water levels to remain deeper for longer periods of time near the hinge of the syncline

than near the limbs. In this case, a constant water budget (constant climate) and constant sediment supply would create a single unit having deeper-water deposits in the hinge region and shallower-water deposits in the limbs. Also, the overall thickness of any one sediment package (in this case the lacustrine cycle between two subsequent non-red deposits) would increase as measured sections were taken from the limbs inward to the hinge of the syncline. Here, the affect of normal faulting on rift basin deposits is not only to cause thickening towards the fault, but also towards the axis of the transverse fold. It also creates facies changes in the deposits from limb to hinge to reflect the transition from a shallow to a deep-water environment. While this has been demonstrated on a basin-wide scale, a question arises as to whether the same affects can be observed on a smaller scale in association with smaller normal fault systems or intrabasinal faults such as the Hopewell fault. If both stratigraphic cycle thickness and lateral facies changes occur on a basin scale (100s of kilometers) (Schlische, 1991), do they also occur on a fault block scale (10-20 kilometers or less)?

In fact, two such facies changes are observed in the Hopewell fault block. The first location where facies changes are observed is in the Pennington syncline. As described earlier, the outcrops of a single non-red unit exhibit increasingly deeper-water lithofacies as one moves eastward, towards the hinge of the Pennington syncline. This indicates that at the time this unit was being deposited, water levels were deeper in the hinge area of this fold than at the western limb, which would seem to concur

with the synsedimentary folding hypothesis. However, lack of outcrop prevented me from determining whether the same trend occurred for the eastern limb of this fold. Therefore, there are two possible conclusions: 1) the Pennington syncline was active at some time during lacustrine deposition, causing facies changes in this unit, or 2) the "deepening" observed is part of a larger-scale, regional trend. Not much can be said conclusively for the other transverse folds in this area, as available outcrop was insufficient to either positively trace one single unit across any one fold or closely examine any facies changes therein.

The second location involves the Perkasio Member of the Passaic Formation in the southern part of the Rocky Hill quadrangle. Here, measured sections of the Perkasio reveal that it consists of approximately 30 meters of gray mudstone and siltstone. This is very unlike the sections measured in any of the drill cores mentioned and unlike most typical outcrop patterns of this member observed elsewhere in the Newark basin (Schlische, 1992). One such typical section can be seen in figure 22 taken from outcrop near Route 18 in New Brunswick, NJ. Here, the Perkasio Member has five distinct non-red units in section. These deposits represent five separate lacustrine cycles and five distinct episodes of deepening and shoaling. The outcrops measured near Route 518 in the Rocky Hill area contain only one lithology and represent a constant moderately deep-water environment (Fig. 19). This does not seem to support the argument above for synsedimentary folding, as it clearly shows a trend towards deeper-water conditions from hinge (in New Brunswick) to the

limb (near Hopewell). However, a larger view of this locality shows that these outcrops lie relatively close to two major diabase intrusions which create a "triple-point" with the Hopewell fault. The presence of these intrusive bodies may indicate that this area was one of structural weakness/ increased fault activity. If this is so, it is possible that motion along the Hopewell fault was at a maximum here, near the limb of the transverse syncline rather than at the center, or hinge. Consequently, the area would have been at a topographic lowpoint relative to surrounding areas and would have maintained a deeper-water environment during the time the Perkasio Member was being deposited.

Also, the deposits of the Perkasio Member in the Rocky Hill quadrangle consist predominantly of fine grained sandstone and siltstone, whereas the New Brunswick outcrop of the Perkasio Member consists of predominantly mudstone and shale. This shows a trend of coarsening towards the south, towards the southern limb of a larger, basin-scale synform and towards the Hopewell fault. This may indicate that coarser sediments were shed from the uplifted footwall block of the syndepositionally active Hopewell fault. Additional coarse material may have been funnelled into the basin at relay ramps between overlapping fault segments.

**Development of the Hopewell Fault-** In the discussion above involving the development of a single rift basin, only one fault was responsible for the creation of the rift. In this example, the imaginary fault was considered to have essentially constant

strike, and a smooth displacement gradient from a maximum at the center of the fault trace to a minimum at the ends. The growth of this fault produced a basin geometry with a single, gently plunging syncline. However, the Hopewell fault has neither constant strike nor a single, gently plunging syncline associated with it. Traveling from one end of the Hopewell fault to the other requires many direction changes, and there are at least four small synclines found plunging gently toward it, separated by what appear at the map scale to be subdued anticlines.

To produce a similar structural configuration, Anders and Schlische (1993) considered a rift system consisting of separate faults in proximity to one another. Each fault initially develops its own basin with the plunging syncline geometry, but changes begin to occur as the ends of the fault segments grow toward each other. The fault segments will eventually overlap, causing interference in the area between them. This displacement interference can be accommodated by either a relay ramp or a transfer fault linking the two segments (Fig. 35). In either case, both the areas of maximum displacement akin to both fault segments and the deepest parts of both basins become separated by what is known as an "intrabasinal high" (Anders and Schlische, 1994). It is these intrabasinal highs which resemble anticlines in map view. Although Anders and Schlische (1994) are describing border fault systems, these observations are applicable to intrabasinal faults as well, as these smaller faults often interact similarly.

If the transverse folds associated with the Hopewell fault formed in this way, then the fault probably formed as four

separate fault segments. Examining the fault closely, it consists of three distinct segments which are oriented roughly  $040^{\circ}$  to  $045^{\circ}$ . These parts of the fault are connected by two segments oriented roughly  $085^{\circ}$  to  $090^{\circ}$ . This bimodal trend suggests that either two separate extension directions acted to form the fault or the fault originally consisted of one set of segments (either at  $045^{\circ}$  or  $090^{\circ}$ ) which grew and became connected by the other. In this case, the latter set of fault segments would be considered the transfer faults.

Support for the originally segmented fault hypothesis lies in the attitude of the axial traces and axes of the transverse folds. The Jerico, Fiddler's Creek, and Hopewell synclines all have axial traces lying within  $310^{\circ}$  and  $325^{\circ}$ , roughly perpendicular to those sections of the Hopewell fault which trend  $040^{\circ}$  to  $045^{\circ}$  (Fig. 36). However, the Fiddler's Creek syncline plunges towards a segment of the Hopewell fault which strikes east-west. The Pennington syncline has an axial trace with a more northerly trend (approximately  $350^{\circ}$ ), but may in fact trend towards the northeast. The lack of outcrop in the area of the eastern limb of the Pennington syncline prevented a more accurate documentation of bedding attitudes. Therefore, the pi-diagram for the Pennington syncline may not be correct.

The pi-diagrams suggest that the Jerico and Hopewell synclines developed in association with the northeast-striking segments of the fault. The map suggests this is also true of the Pennington syncline. The Fiddler's Creek syncline, however, trends northwest but terminates against an east-striking fault

segment. This could be due to a change towards a more easterly extension direction causing this fold to form via a different folding mechanism--transtension. In other words, two separate extension events in two different directions, possibly at two different times, acted to form the folding pattern observed. The Fiddler's Creek syncline is also not as well-developed as the other three, and is located in an area where drag folding and a large diabase intrusions are found. Therefore, bedding data collected from the area of this syncline may not be reliable enough to make any conclusions as to its origin. Furthermore, joint data recorded from the entire study area shows that although there were likely two extension directions ( $100^{\circ}$  and  $135^{\circ}$ ), both of them are oriented northwest-southeast, not north-south. Also, north of the study area the Hopewell fault has a general trend of north-south for most of its length.

Given these observations, the most likely scenario begins with northwest-southeast extension. Rifting creates fault segments striking approximately  $040^{\circ}$  to  $045^{\circ}$  (Fig. 37a). These segments grow through time and create basins which begin to fill with lacustrine sediments. Continued movement along the fault planes and deposition of sediments work together to create the Jerico, Pennington and Hopewell synclines (Fig. 37b). Eventually, these fault segments and their associated basins/synclines grow close together, overlap, and interfere. (Fig. 37c) Further displacement is then taken up along east-striking transfer fault segments. Rifting continues, and is now both perpendicular to the original fault segments but oblique to the transfer segments. This



allows displacement to occur along the transfer segments at a smaller scale, and this displacement acts to create the Fiddler's Creek syncline (Fig. 37d).

**Timing of the Hopewell fault-** If indeed the facies changes discussed above are a direct result of syndepositional folding, then would indicate that at least some portion of the Hopewell fault was active during sedimentation. These active segments would likely be those oriented northeast-southwest, as bedding data taken from the folds has shown the fold axes are perpendicular to these segments and are therefore related to them. The main problem with this argument is that there was insufficient data (outcrop) available in the study area to investigate these facies changes more vigorously. Whether the east-west striking segments of the Hopewell fault were active syndepositionally is uncertain.

Another factor supporting syndepositional fault activity is the morphologies of the three prominent diabase intrusives in the Pennington and Lambertville quadrangles. The form of Jerico Mountain in Pennsylvania is very similar to phaccoliths located in the Jacksonwald syncline. Both Baldpate and Pennington Mountains were probably emplaced with a similar morphology, but were later deformed by intrablock faults which now cut across them. This subsequent faulting could produce the landforms observed today. If these three igneous bodies are indeed phaccoliths, then they were emplaced syndepositionally. The fact that they were later deformed by faulting indicates that

tectonic activity was taking place after these bodies were emplaced and cooled, or during the time that the units stratigraphically above them were being deposited. In other words, at least some faulting was syndepositional.

**Extensional Joints-** Strike measurements of extensional joints in the study area have been observed to have a bimodal distribution in rose diagram plots. The vast majority of these joints strike northeast-southwest, and fall into two groups at  $010^{\circ}$  and  $045^{\circ}$  for the area as a whole. Different regions of the area, however, display peculiarities with respect to the attitudes of the extensional joints.

In an area which runs parallel to the Hopewell fault and extends approximately one kilometer to the south of it, the joint distribution is much more evenly distributed across a much wider range of strike directions. This is likely due to the area's proximity to the fault, as motion along the fault can act to rotate previously formed joints away from their initial attitudes. Therefore, it can be said that at least one generation of extensional joints formed before motion along the Hopewell fault ceased.

In the Rocky Hill quadrangle, joint strikes are highly concentrated around  $010^{\circ}$ , with very few around  $045^{\circ}$ . In this area, the Hopewell fault itself is oriented more closely to this attitude and the joints probably formed coincidentally with the fault during a single period of extension perpendicular to  $010^{\circ}$ , or they were never affected by the fault at all.

South of the Rocky Hill diabase intrusion, in the Pennington quadrangle, joint measurements cluster around  $045^{\circ}$  in those areas greater than one kilometer from the Hopewell fault. These measurements show a slightly higher degree of variability than those in the Rocky Hill area probably due to the presence of intrabasinal faults. Since these faults have already been observed to have caused drag folding of the Passaic Formation deposits, it is safe to assume that the joints formed in these deposits have also been rotated to some degree in the areas immediately adjacent to these faults. It may be important to note that in the Pennington quadrangle, more of the Hopewell fault trends east-west than northeast-southwest. The shift in joint attitudes may be reflecting this. If so, then it may be that two separate episodes of rifting took place in different directions at different times. Because the  $010^{\circ}$  trending set of joints most closely parallels an actual segment of the fault, I would conclude that this reflects the initial extension direction (towards  $100^{\circ}$ ). A change in this direction from  $100^{\circ}$  to  $130^{\circ}$  occurred later, and helped to form the transfer faults mentioned above. Continued rifting in this direction would also have allowed the formation of the Fiddler's Creek syncline, as the basin would have continued to open and grow through time.

### **Future Research--**

Future studies should focus on testing the model of fault growth and fault linkage presented in this thesis. One way this could be done is by a close examination of the extensional joints in the area

of the Hopewell fault, the Jacksonwald syncline, or other transverse fold in order to document the existence of offset along cross-cutting joint sets. This would help to determine the nature and relative timing of the change in extension direction believed to have occurred in the Hopewell fault block, as well as document similar changes in other portions of the basin.

Also, an examination of a series of shallow drill hole cores placed strategically near the ends of fault segments, where the Hopewell fault changes orientation, could provide clues as to the development of transverse folds through time. If growth faulting is responsible for the formation of the folds, one would expect the folds to steepen with depth. Such a series of drill cores may also reveal linkage of faults at depth in other areas of the Newark basin where said faults are not linked at the surface.

## **V. Conclusions--**

- 1) The Hopewell fault strikes in two distinct directions in the study area:  $040^{\circ}$  and  $085^{\circ}$ . The orientation alternates between these two directions and implies an originally segmented fault system.
- 2) There are at least three minor intrabasinal faults within the study area, two of which are likely splays off the Hopewell fault.
- 3) There are three well-developed and one poorly-developed synclines in the hanging wall of the Hopewell fault which are

separated by three poorly developed anticlines. Each of the synclines plunges gently (approximately  $10^{\circ}$ ) towards the northwest (around  $325^{\circ}$ ), which is nearly perpendicular to those sections of the Hopewell fault that strike  $040^{\circ}$ . Therefore, the synclines are directly related to these segments of the Hopewell fault.

4) Several non-red lacustrine units of the Passaic Formation are traceable across large portions of the study area. These units include the Perkasio Member. Along-strike facies changes were observed in two non-red lacustrine units, but in neither case could this change be definitely related to local tectonics or differential displacement along the Hopewell fault.

5) Changes in stratigraphic thickness could not be measured within any single Van Houten cycle, mostly due to the lack of sufficient outcrop for such measurements.

6) Joint trends in the area indicate that the dominant extension direction was toward  $130^{\circ}$ , and a subordinate direction existed toward  $100^{\circ}$ .

7) The Hopewell fault is a linked fault system in which the transverse folds developed in association with syndepositional growth of fault segments that ultimately linked together.

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## Figure Captions

Figure 1--Geologic map of the study area and its location within the Newark basin. Also shown are the locations of the Titusville, Somerset, Martinsville and Rutgers drill sites. Black lines represent selected members of the Passaic Formation. Base map modified from Schlische (1992).

Figure 2--Map of eastern North America showing the locations of Mesozoic rift basins along the eastern seaboard. Modified from Schlische (1992).

Figure 3--Geologic map of the Newark basin showing lithology and structures. Study area is indicated by the small box. Modified from Schlische (1992).

Figure 4--Schematic map showing the relative positions of the Flemington-Furlong, Chalfont and Hopewell faults, the study area, and the Titusville, Somerset, Martinsville, and Rutgers drill sites. Base map modified from Schlische (1992).

Figure 5--Geologic map of the study area highlighting the axes of transverse synclines associated with the Hopewell and Flemington-Furlong faults. Base map modified from Schlische (1992).

Figure 6--Geologic map of the Jacksonwald syncline, located in eastern Pennsylvania. Note the intrabasinal "highs" as indicated by the two boxed areas. These intrabasinal "highs" represent areas of minimum displacement along the border fault system, caused by the overlap of two normal fault segments. Modified from Schlische (1992).

Figure 7--Schematic map of the Hopewell fault and the study area showing the locations and names of the diabase intrusives located within the study area.

Figure 8--Composite stratigraphic section of the Newark basin obtained by continuous core. Based on Olsen et al. (in prep.).

Figure 9--Stratigraphic section of the Newark basin with accompanying depth ranks for the lacustrine strata of the Lockatong and Passaic Formations. Note the cyclic nature of the depth rank "curve" for the Passaic Formation in particular. Based on Olsen et al. (in prep.).

Figure 10--Geologic map of the central Newark basin showing the locations of the seven drill sites with the accompanying cross section. The cross section is drawn to show the overlapping coverage between each of the sites.

Figure 11--Stratigraphic section of a single Van Houten cycle with accompanying depth ranks. One Van Houten cycle represents the transgression, highstand, and regression of the lake. Modified from Olsen (1986).

Figure 12--Stratigraphic log, depth rank and power spectrum for the Passaic Formation within the Titusville drill core. Power spectrum of depth rank curve shows most prominent cycles in thickness. To convert these to time, assume that the most prominent peak (0.052 cycles/ft or 19.08 ft/cycle or 5.81 m/cycle) corresponds to the 23,000-yr-long precession cycle. This gives a sedimentation rate of 0.000253 m/yr. Applying this sedimentation rate to the other prominent cycles yields the periodicities in time indicated in the parentheses. Note especially the presence of the ~100-kyr and ~400-kyr eccentricity cycles.

Figure 13--Stratigraphic sections showing how Van Houten cycles are grouped into clusters of compound cycles; ~5 Van Houten cycles comprise the first compound cycle, and ~20 Van Houten cycles comprise the second compound cycle. The ratio, 1:5:20, is nearly identical to the ratio of the period of the precession cycle (21,000, the average of 19,000 and 23,000 years) to the periods of the two eccentricity cycles:  
 $21,000:109,000:413,000 = 1:5.2:19.7$ . This suggests that Van Houten cycles are precession cycles, and that compound cycles are eccentricity cycles. From Olsen et al. (1989).

Figure 14--Geologic map of the Jacksonwald syncline in eastern Pennsylvania. Numbers in bold italics are stratigraphic thicknesses of single lacustrine cycles within the Passaic Formation in meters. Note the cycle thickness increases both towards the border fault system and towards the hinge of the Jacksonwald syncline itself. From Schlische (1992).

Figure 15--Closeup map taken from the Rocky Hill USGS 7' quadrangle showing the locations of three separate outcrops of the Perkasio Member found in the Beden's Brook system. The figure demonstrates how two or more outcrops of similar lithologies may be traced across (interpolated) on the map.

Figure 16--Closeup map of the Pennington quadrangle showing the location of the Titusville drill site and some of the non-red units to the south. These non-red units are placed on the map by projection from the drill site based on core data.

Figure 17--Stratigraphic section of one non-red unit within the Perkasio Member taken from a tributary of Jerico Creek in the Lambertville quadrangle. Exact location of this outcrop is shown on the map itself (plate 1).

Figure 18--Stratigraphic section of the Metlars Member taken from Stony Brook near the Hopewell Valley country club in the

Pennington quadrangle. Exact location of this outcrop is shown on the map itself (plate 2).

Figure 19--Stratigraphic section of the Perkasio Member taken from the Beden's Brook in the Rocky Hill quadrangle. Note the singular lithology present. Exact location of this outcrop is shown on the map itself (plate 5).

Figure 20--Stratigraphic section of the Metlars Member taken from the Rocky Hill quadrangle near the intersection of Aunt Molley Road and Route 518. Exact location of this outcrop is shown on the map itself (plate 5).

Figure 21--Closeup map of the western limb of the Pennington syncline in the Pennington quadrangle. The map shows the locations of outcrops which demonstrate a facies change from shallow to deep water as one moves from west to east.

Figure 22--Comparison of stratigraphic sections of the Perkasio Member in the Rocky Hill quadrangle and New Brunswick. Note the significant changes in lithology.

Figure 23--Closeup map of the Moore's Creek system showing folding patterns. Note the high dip angles as one moves closer to the Hopewell fault.

Figure 24--Closeup map of the location of the Metlars Member outcrop shown in figure 20.

Figure 25--Pi-diagrams of bedding for each of the four synclines located within the study area. Pi-diagram of the Pennington syncline may be inaccurate due to the lack of readings. This was caused by a lack of outcrop in the field.

Figure 26--Closeup map of the Lambertville quadrangle showing how the location of the Hopewell fault was pinpointed in two locations and interpolated between them. In the eastern part of the figure, the fault is situated between two outcrops of Passaic Formation, one which dips steeply and one which dips shallowly. In the west, the fault is located by the presence of large amounts of fault gouge along the side of Route 232.

Figure 27--Closeup map of the Pennington quadrangle showing drag folding associated with smaller intrabasinal faults between Pennington and Baldpate Mountains. Good outcrop in both of the Jacob's Creek tributaries provided ample evidence of both the faults and the folding.

Figure 28--Rose diagram of joint azimuths taken from that part of the study area in the Rocky Hill quadrangle.

Figure 29--Rose diagram of joint azimuths taken from that part of the study area in the Princeton and Hopewell quadrangles as well as the area between the Honeybranch Creek system and Pennington proper in the Pennington quadrangle.

Figure 30--Rose diagram of joint azimuths taken from that part of the study area adjacent to the Hopewell fault from the town of Hopewell to the western edge of the study area.

Figure 31--Rose diagram of joint azimuths taken from that part of the study area in the Pennington quadrangle from the town of Pennington to the western edge of the Pennington quadrangle.

Figure 32--Rose diagram of joint azimuths taken from that part of the study area in the Lambertville quadrangle.

Figure 33--Diagram showing the features associated with synsedimentary faulting. These include an increase in dip angle towards the fault, a thickening of the sediments toward the fault, and the onlapping of younger sediments onto older sediments away from the fault. From Schlische (1993).

Figure 34--Diagrams showing the progression from fluvial to lacustrine environment/sedimentation as an extensional basin grows in size. With a constant sediment influx and water

budget, a small basin quickly fills with sediment and the excess is carried in the runoff (river flow). As the basin lengthens and widens through time, its volume increases rapidly, and the sediment supply cannot keep up. The basin then begins to fill with water (lake) and the deposits within the basin begin to reflect the increase in water depth. From Schlische (1993).

Figure 35--Block diagram showing how a segmented normal fault system acts to create transverse folds. Differential displacement along individual fault segments acts to create synclines with axes trending towards the center of the segments. Where the fault segments overlap, relay ramps develop to accommodate the displacement variations. In these areas, displacement is at a minimum and intrabasinal “highs” or anticlines develop. Modified from Schlische (1993).

Figure 36--Schematic map of the Hopewell fault system showing the trends of the four syncline axes and their relation to segments of the Hopewell fault. Note that three of them are perpendicular to those segments of the Hopewell fault which strike northeast.

Figure 37--Four schematic diagrams showing a hypothetical progression to explain the development of the Hopewell fault and the associated transverse folds. (a) Shows the



initiation of extension towards the southeast and northwest and the formation of four individual normal fault segments. (b) Shows the lateral growth of the four fault segments and the initiation of transfer faulting. This diagram also shows the initial development of three small sub-basins as they begin to fill with sediment. (c) Shows the continued growth of the normal fault segments, their linkage by transfer faults, the merging of the three small subbasins, and small drag folds associated with the interference of two of them. (d) Shows a shift towards more easterly extension, which causes the formation of the Fiddler's Creek syncline.

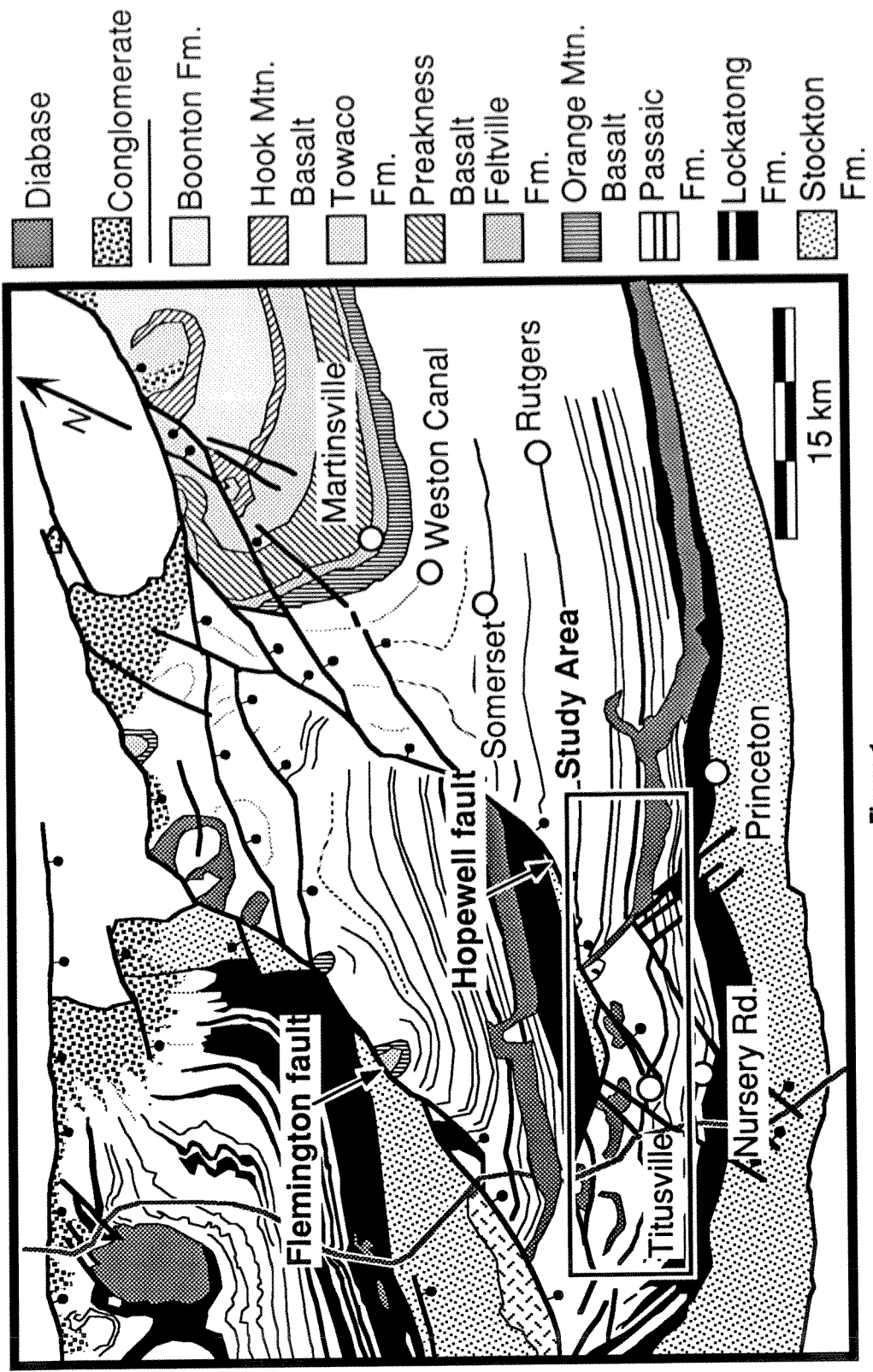


Figure 1

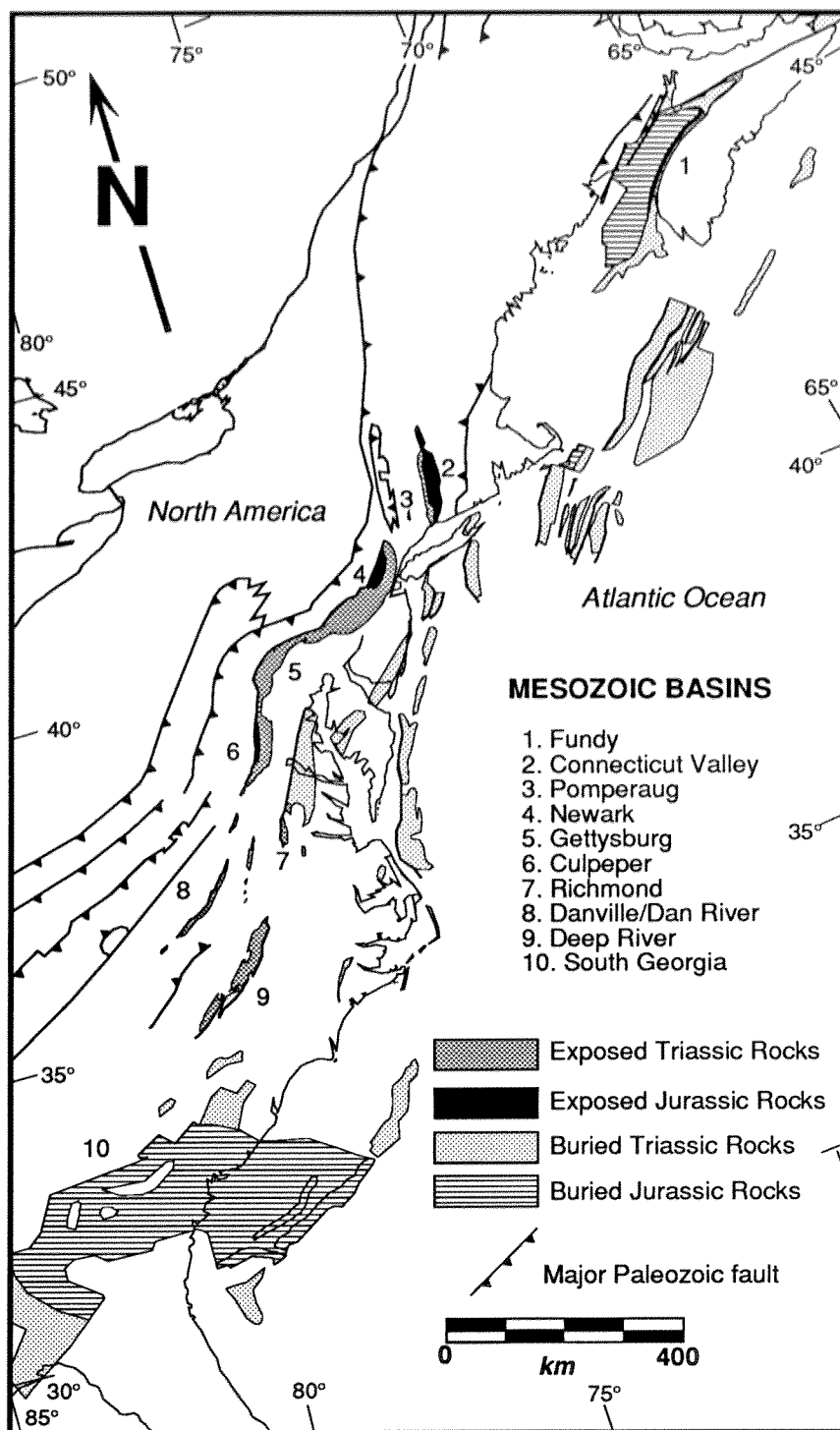
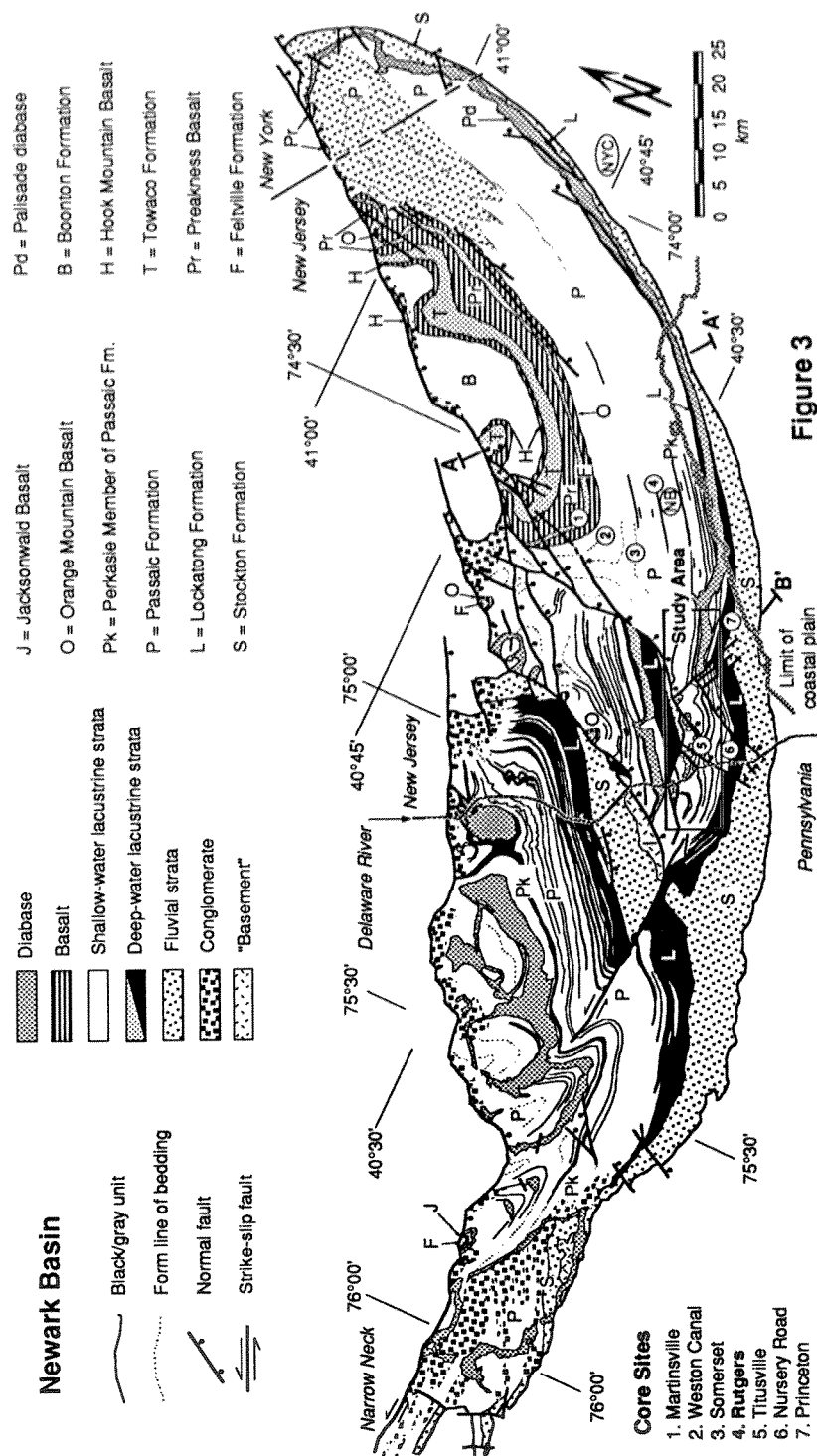


Figure 2



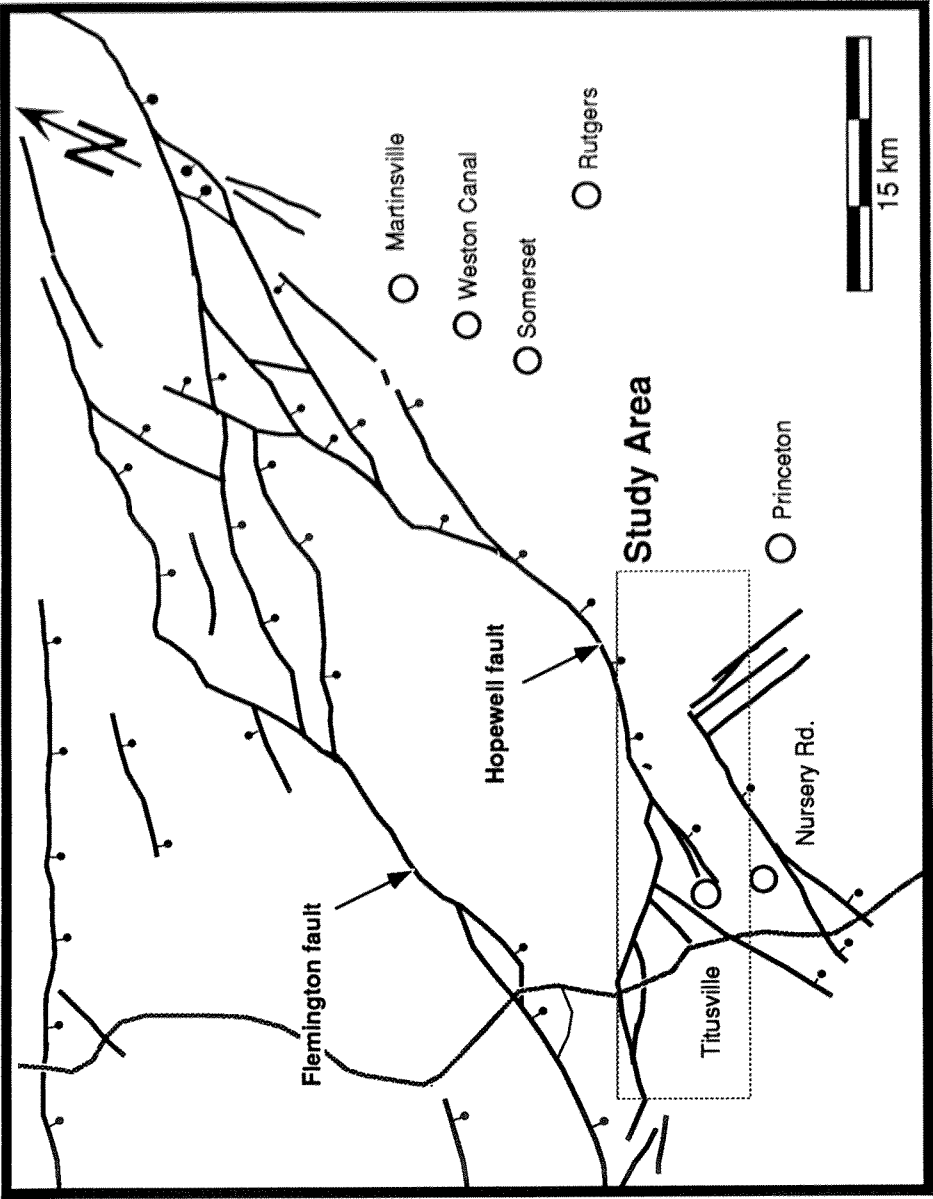


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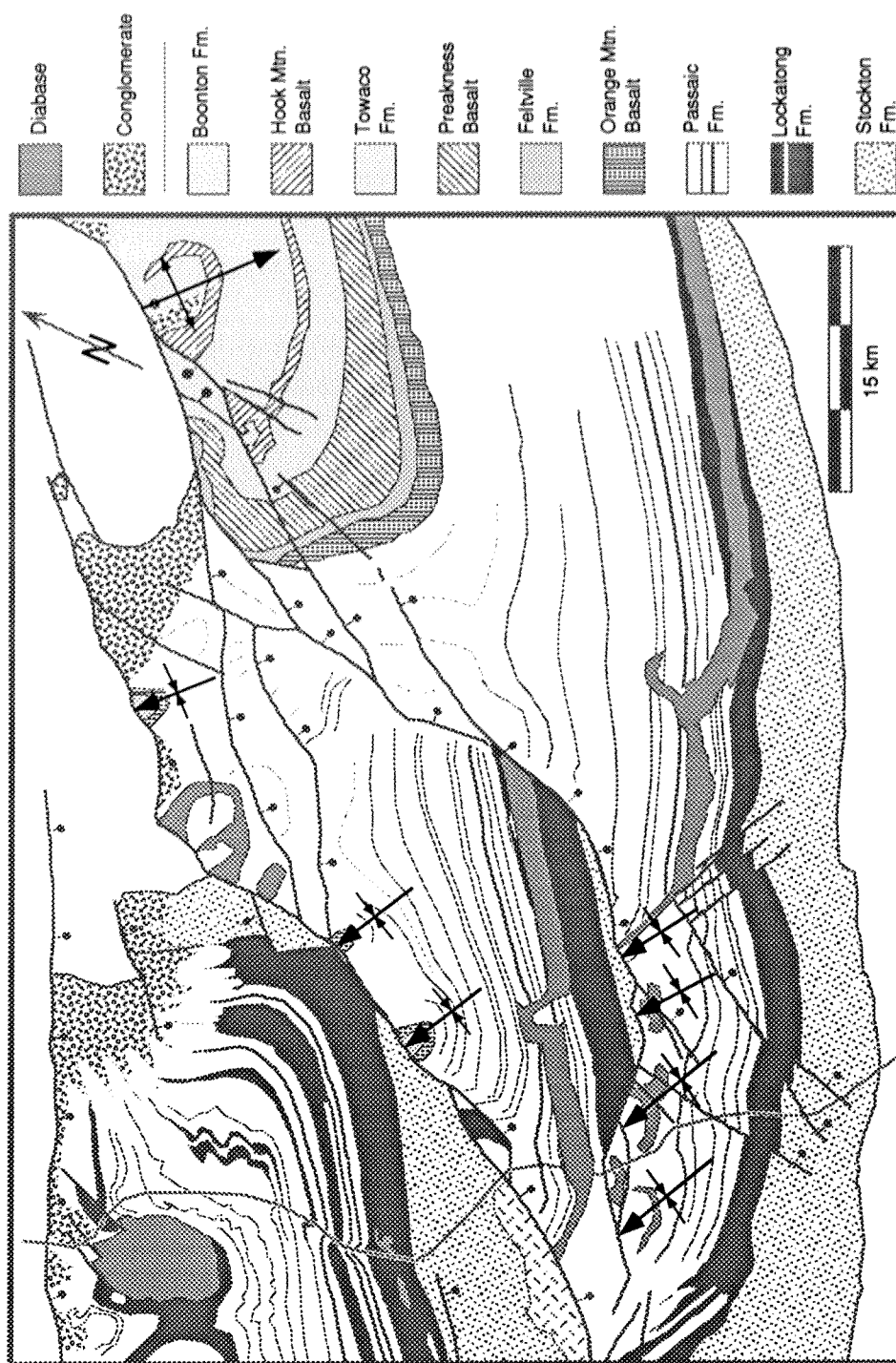


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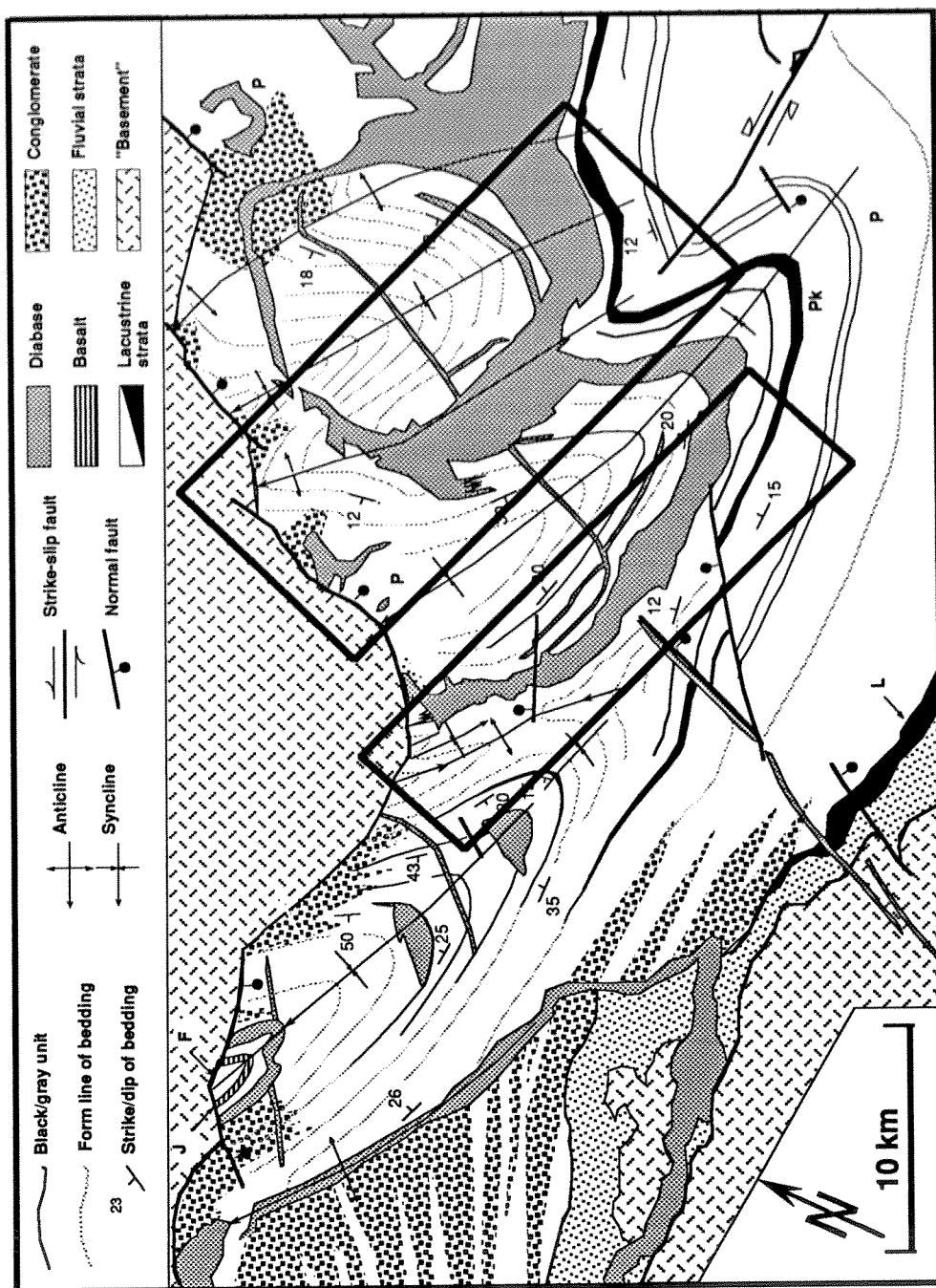


Figure 6

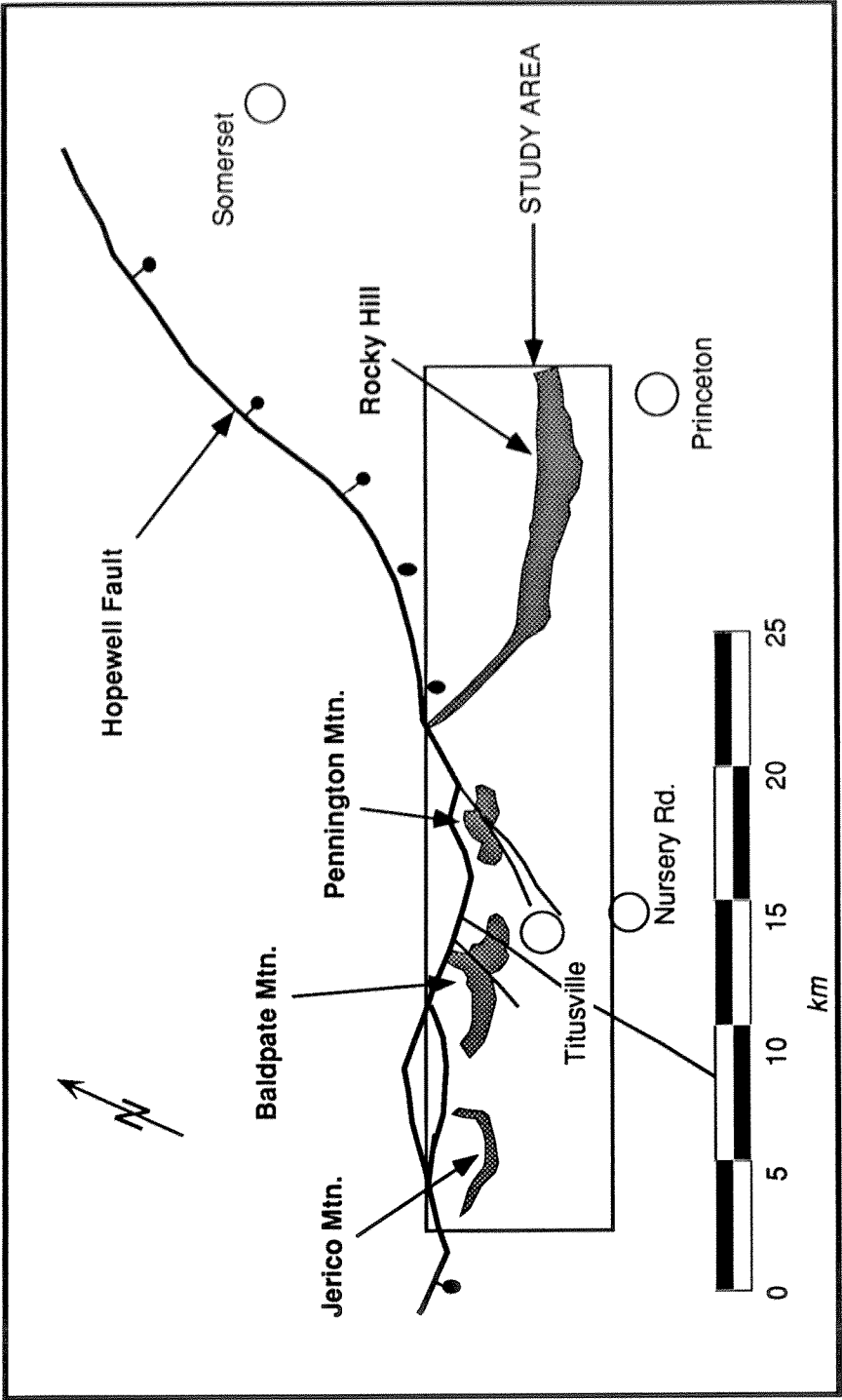
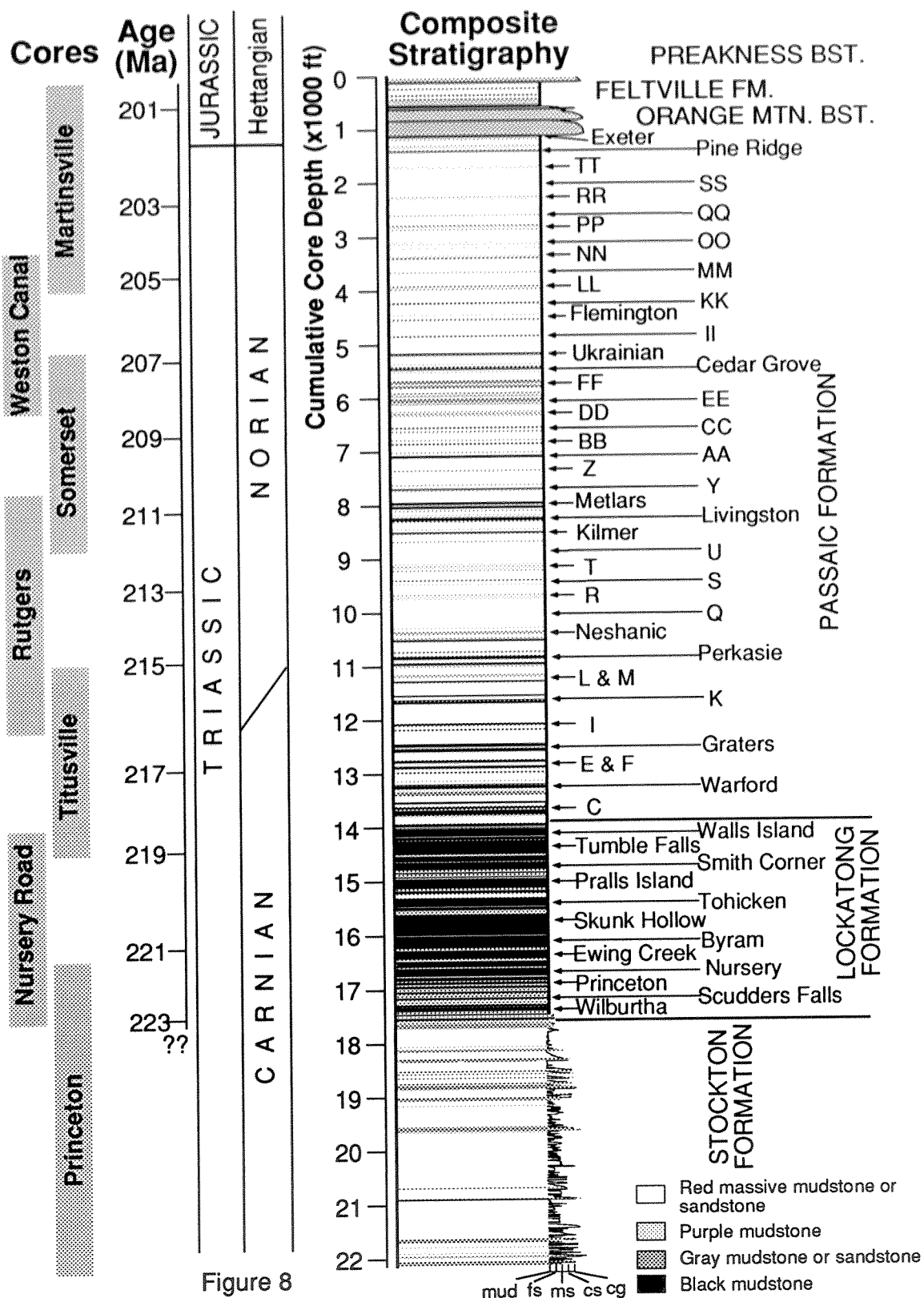


Figure 7





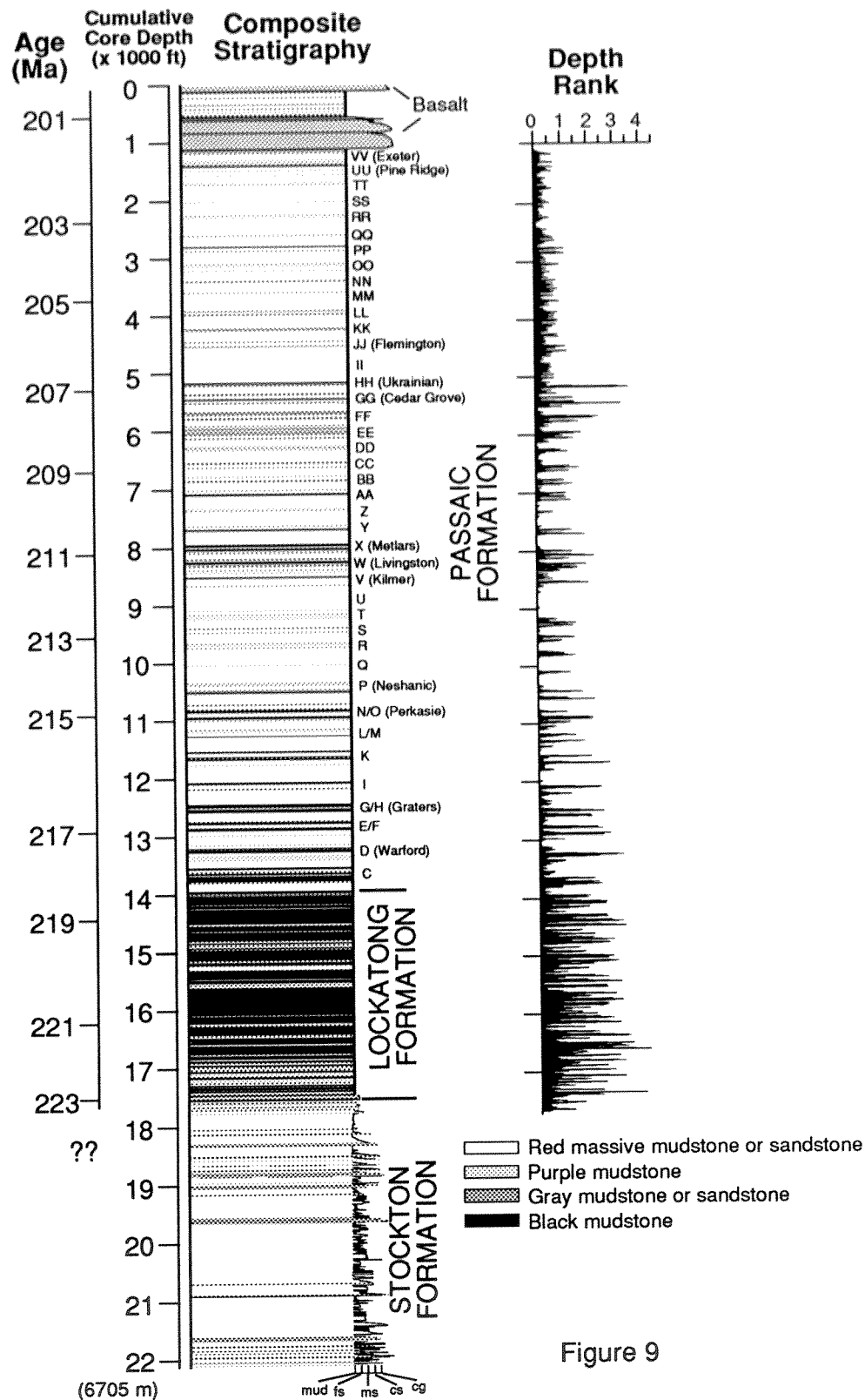


Figure 9

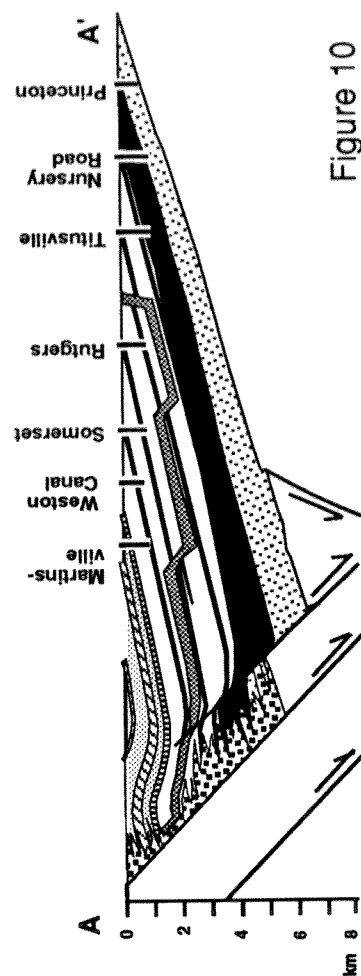
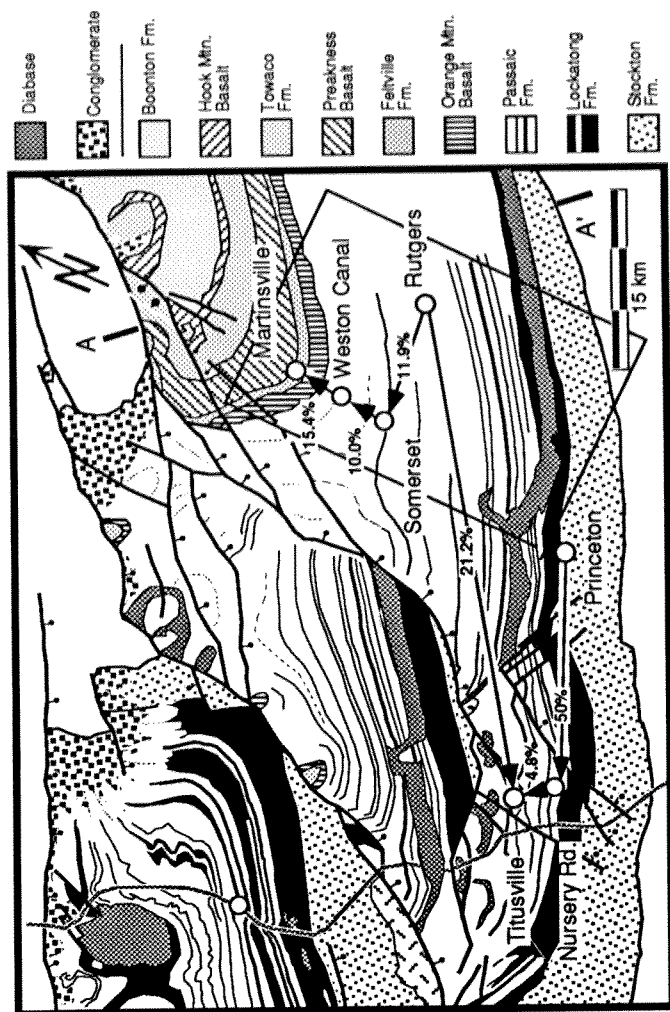
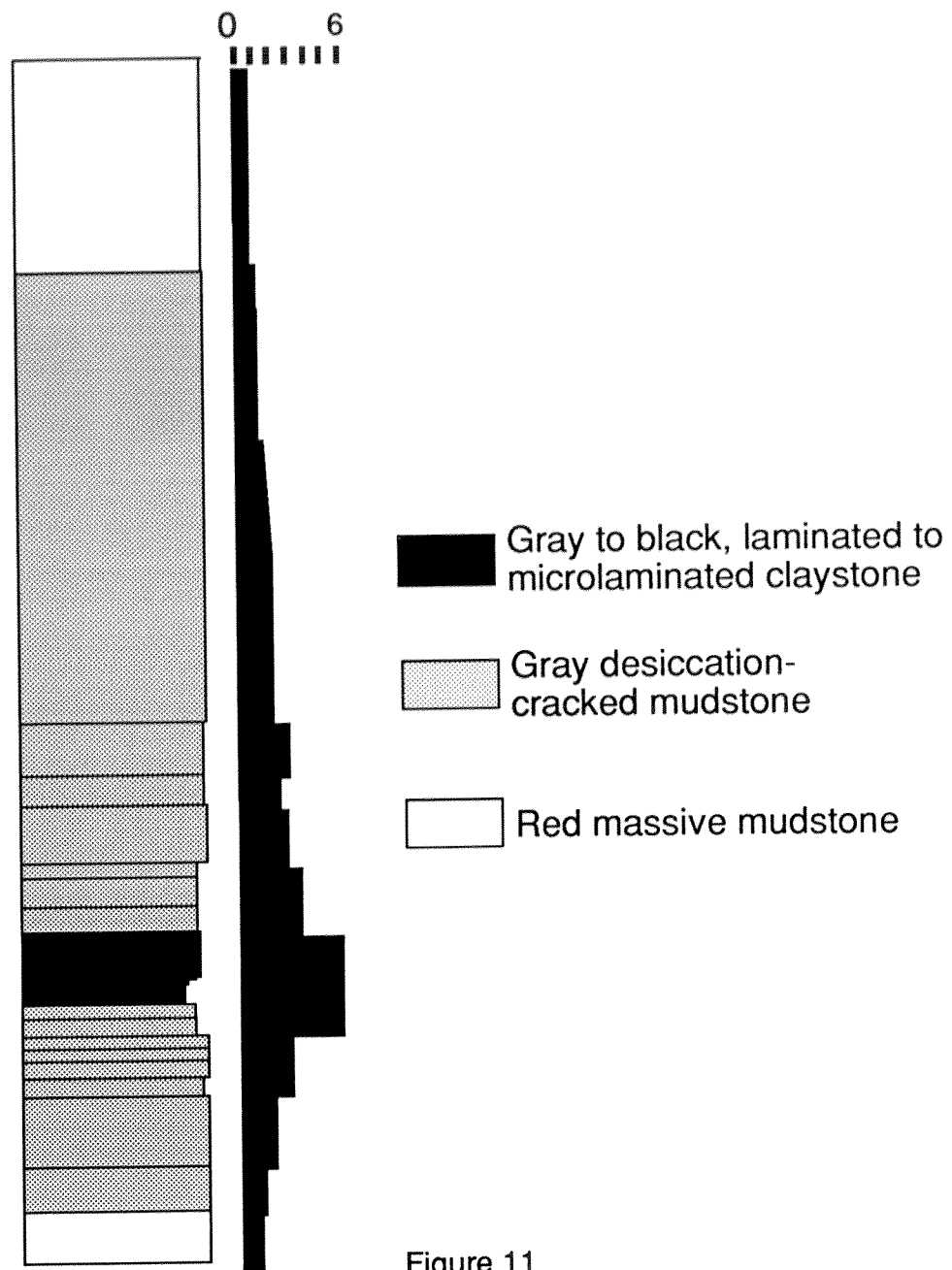


Figure 10



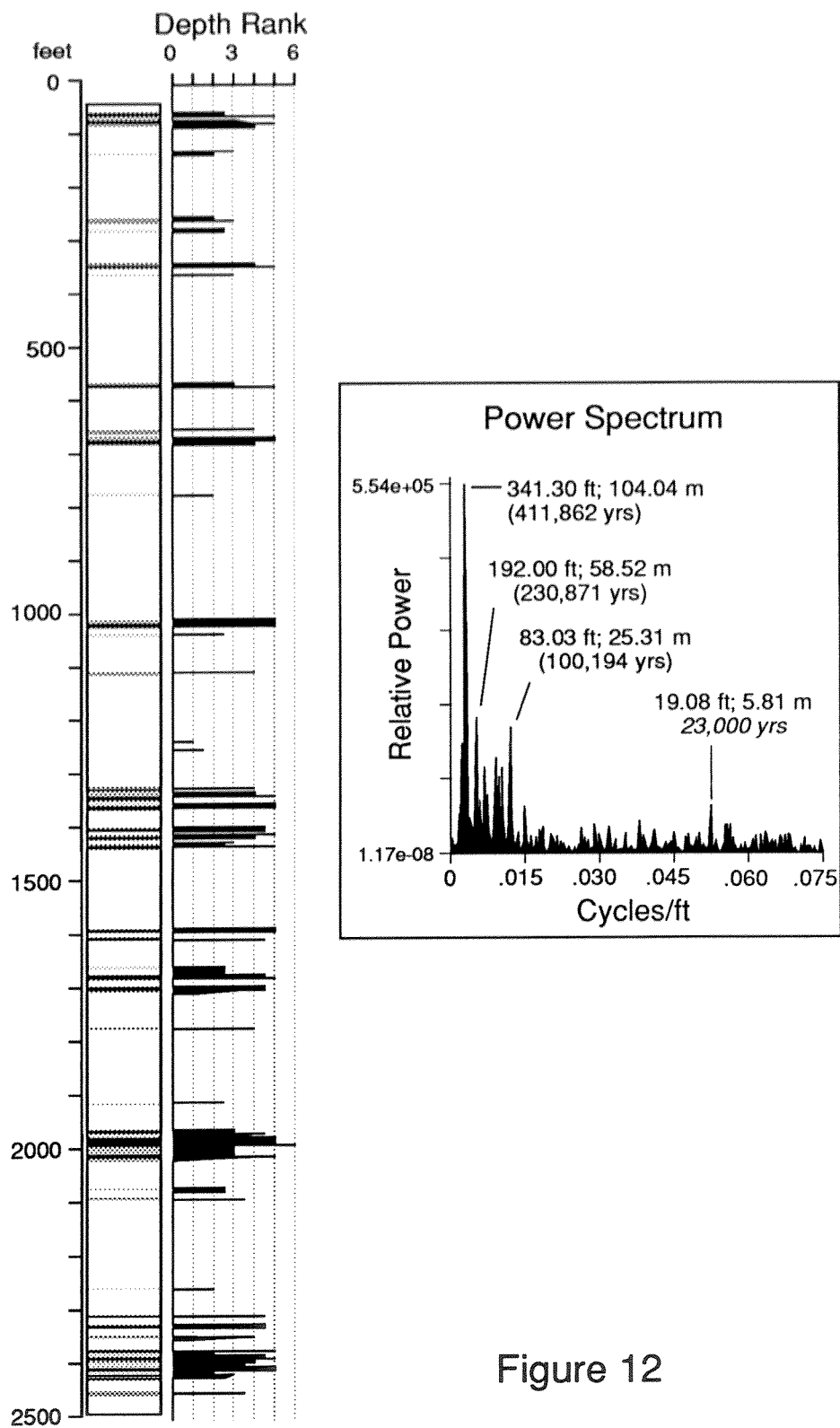


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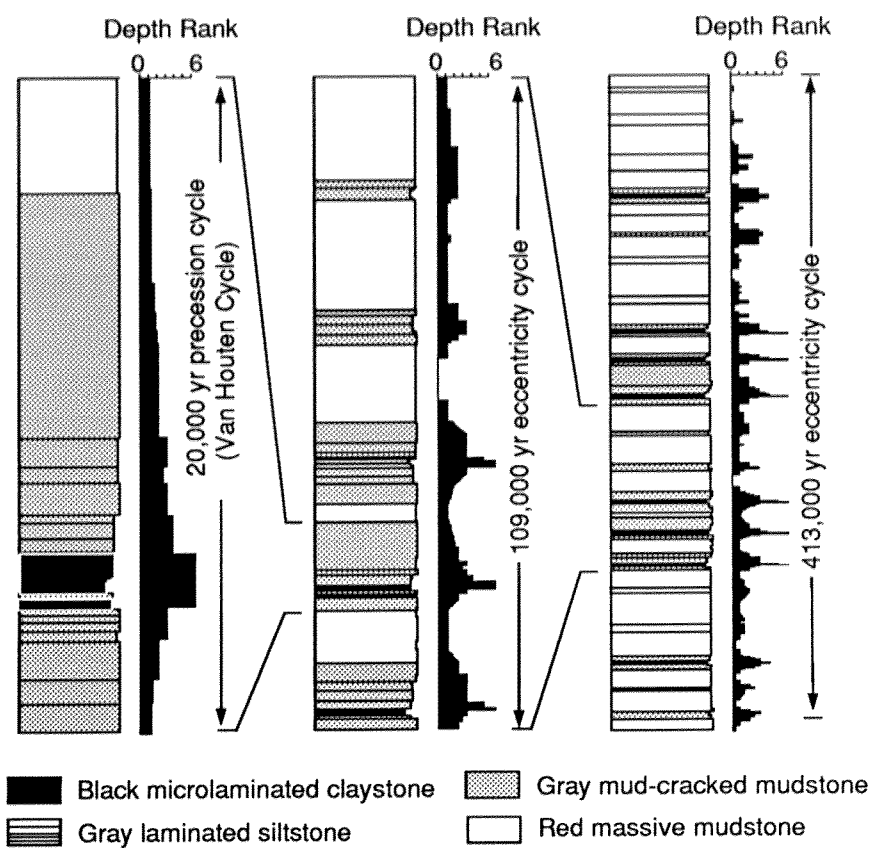


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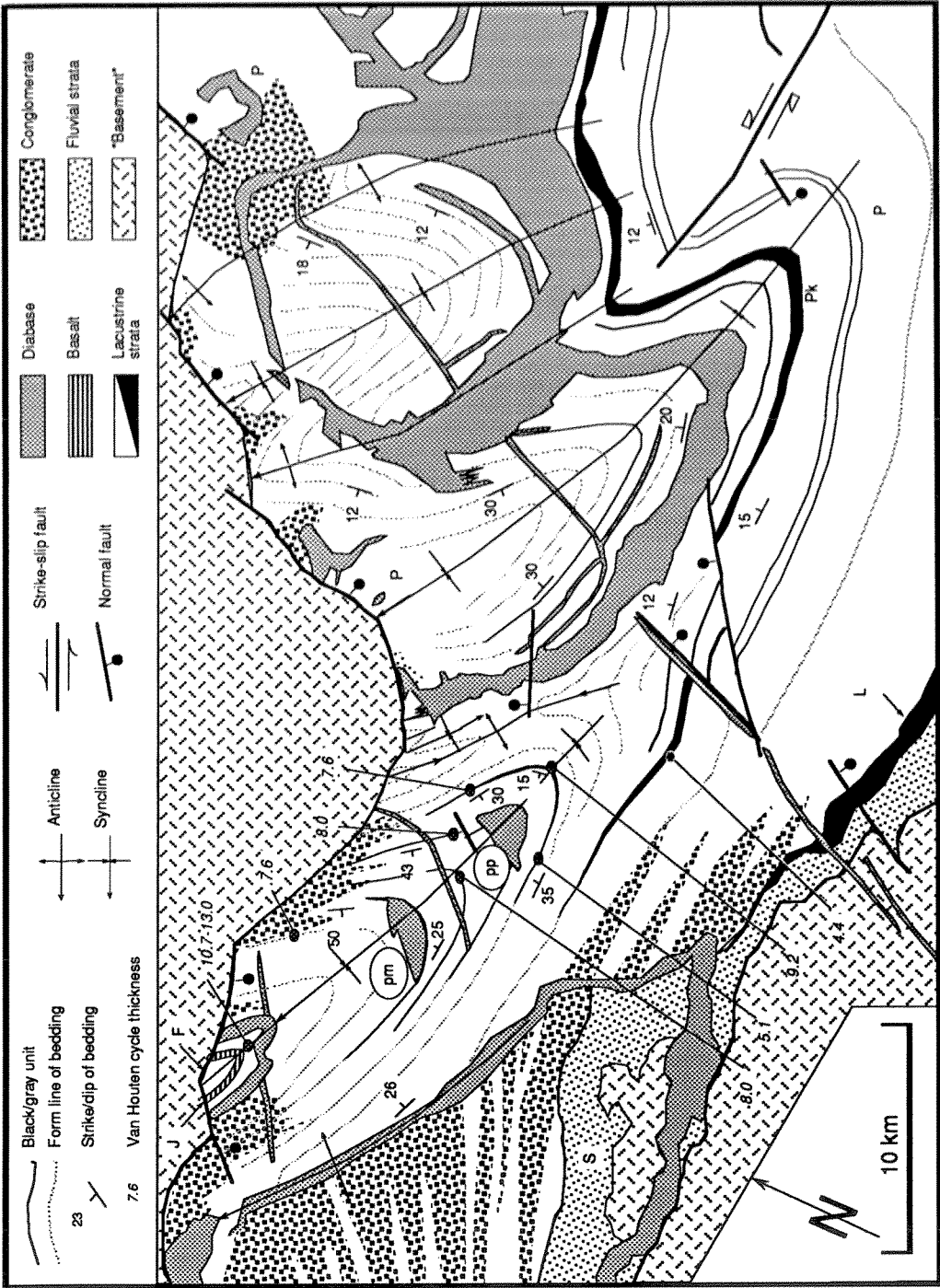


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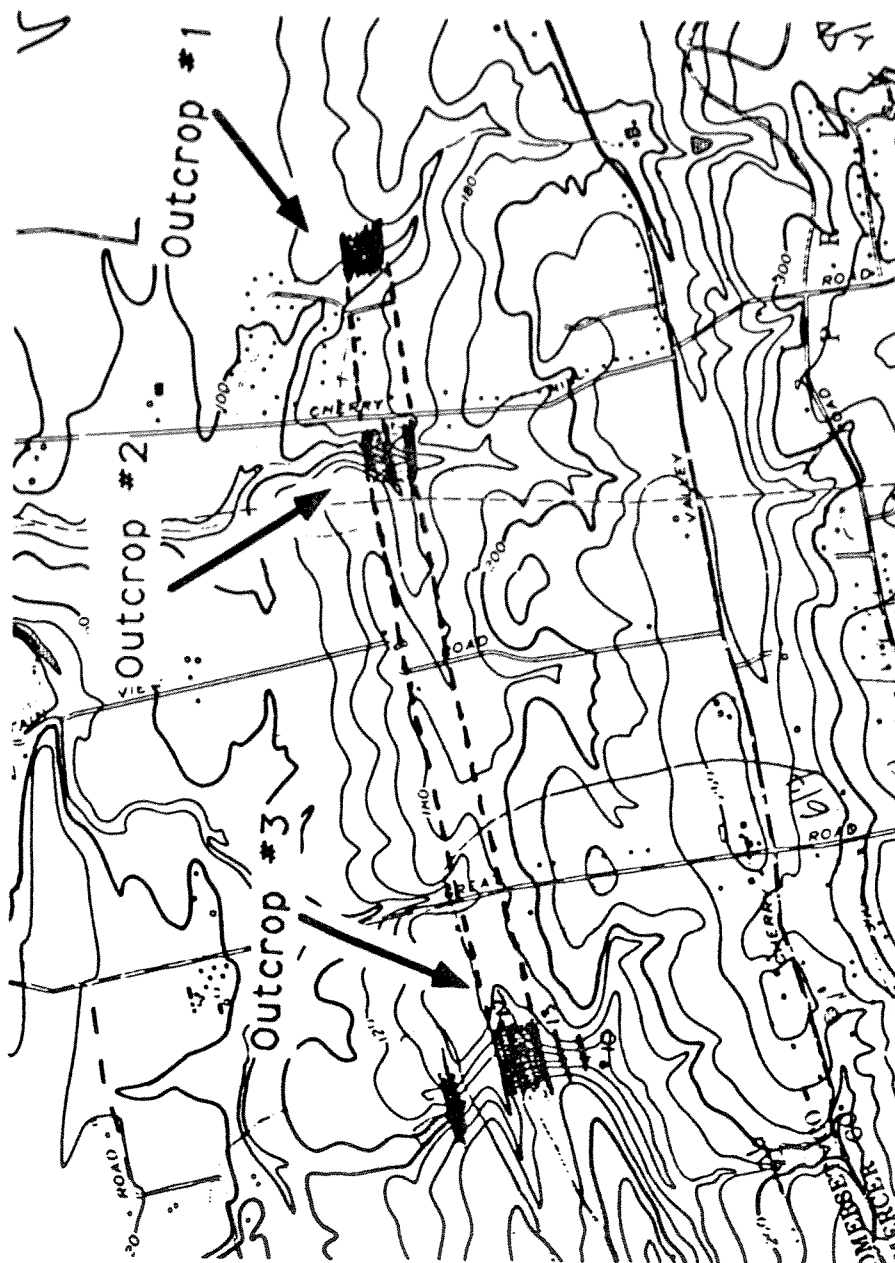


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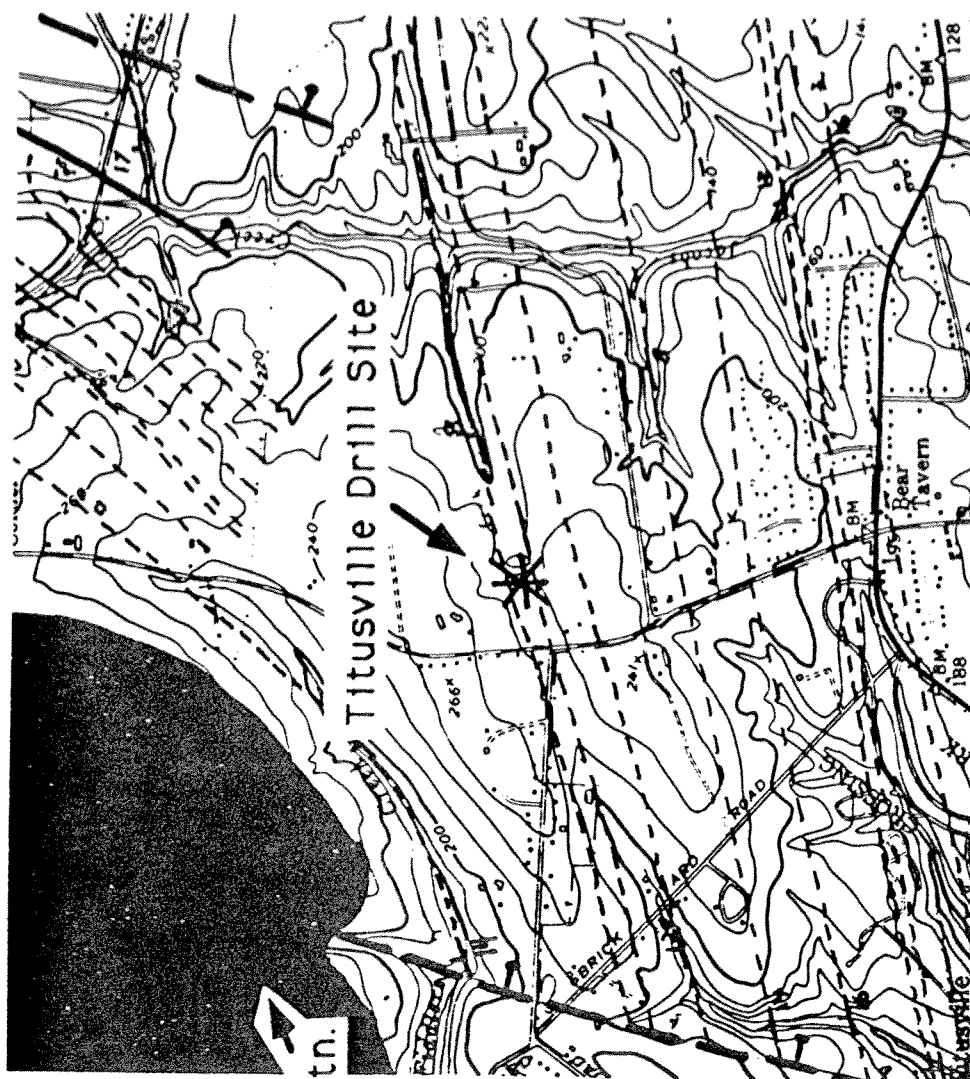


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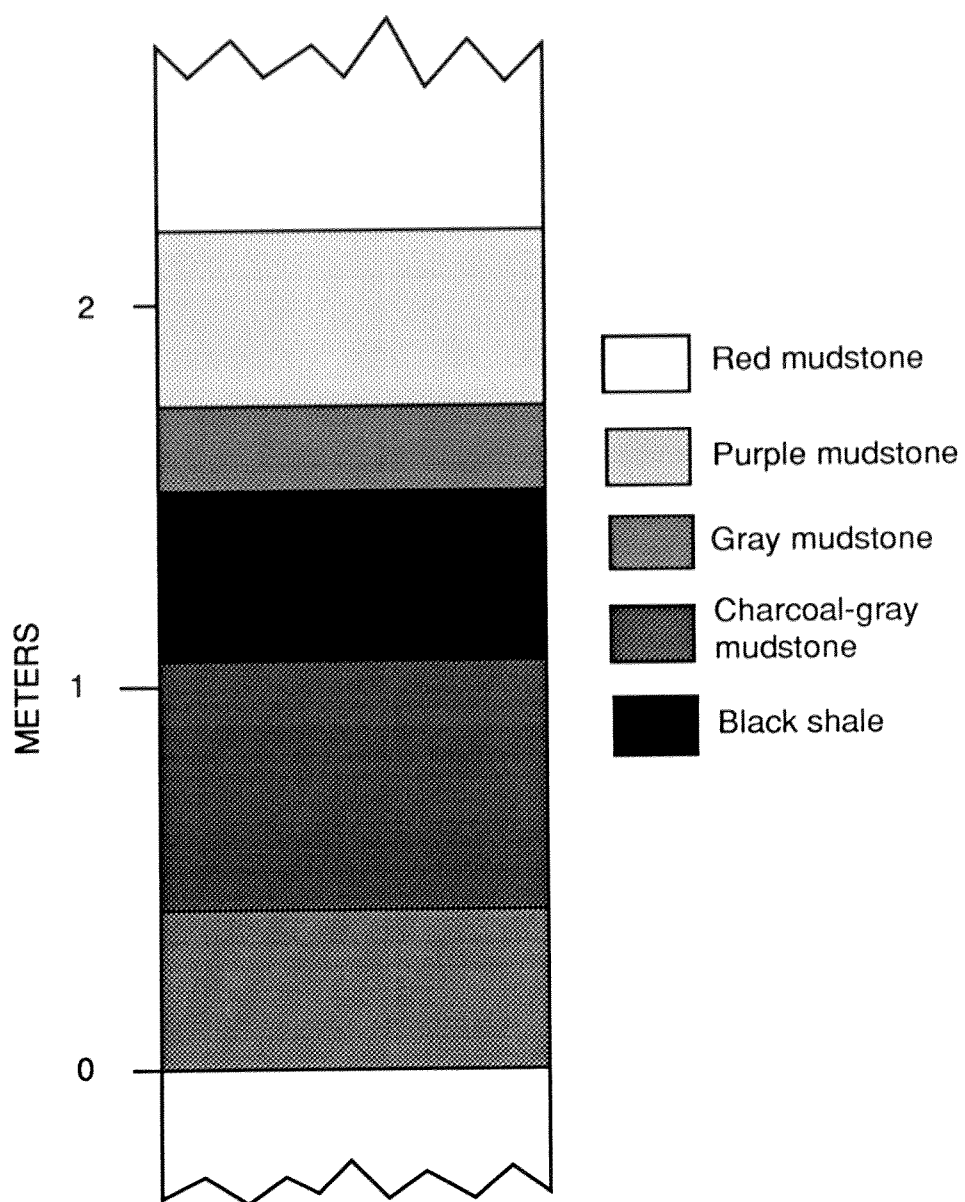


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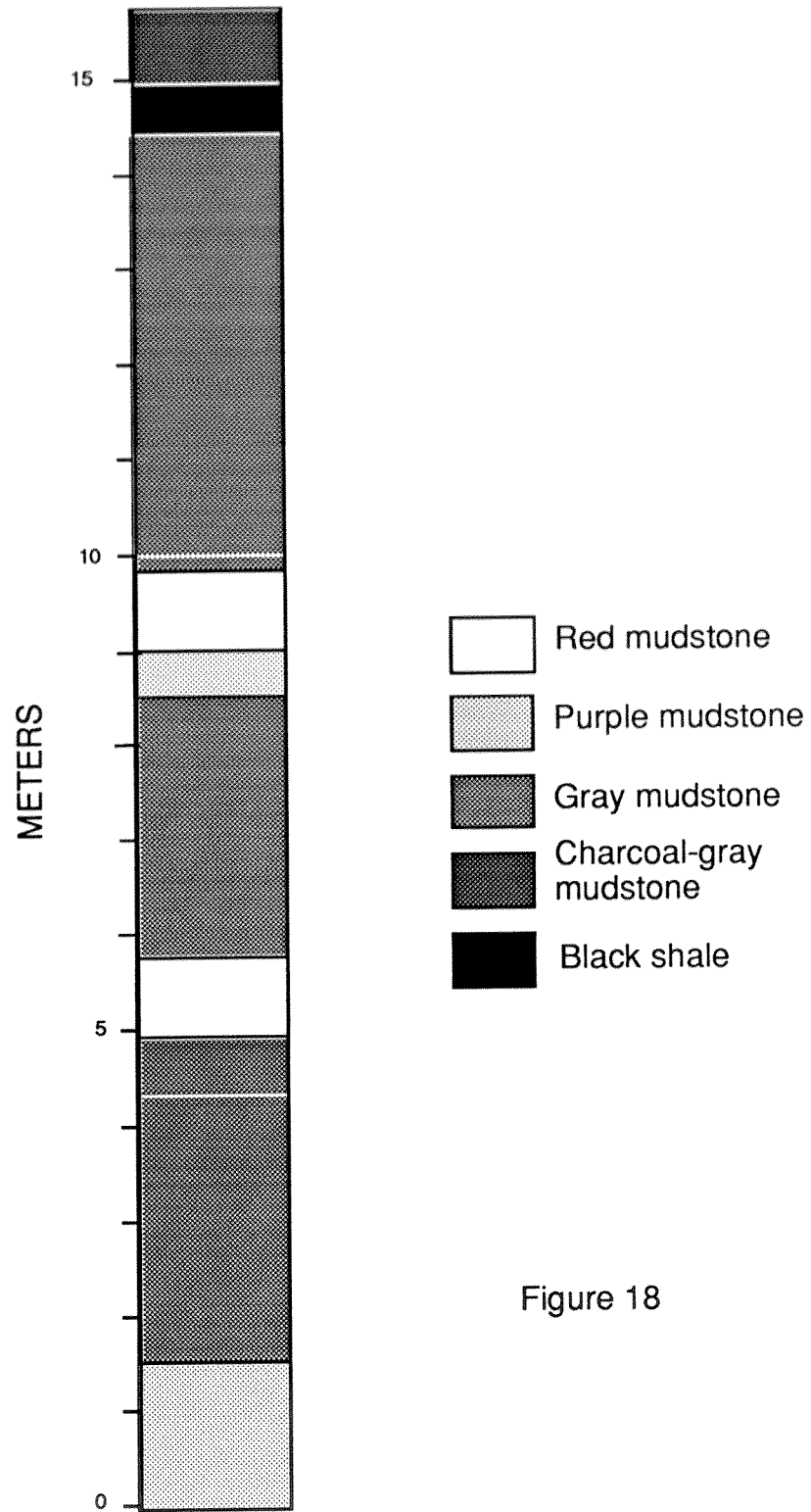


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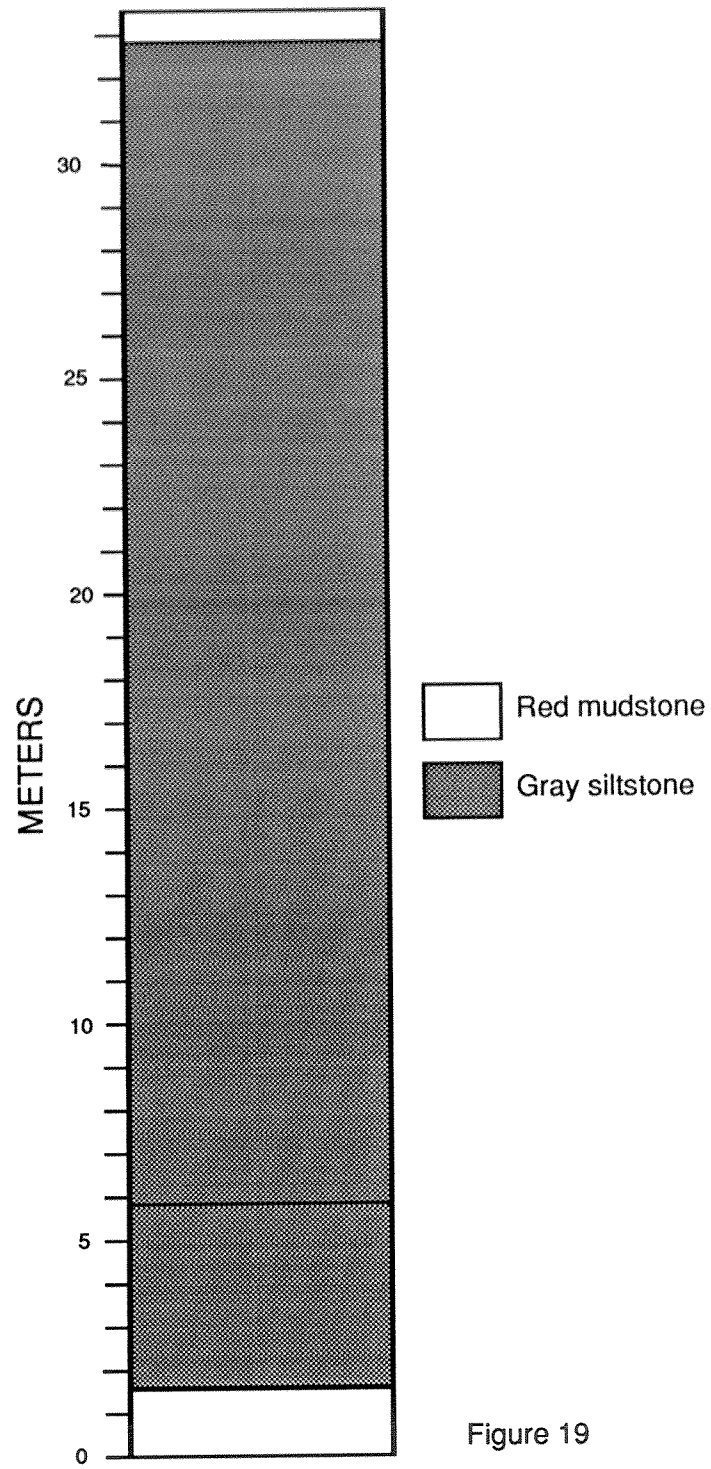


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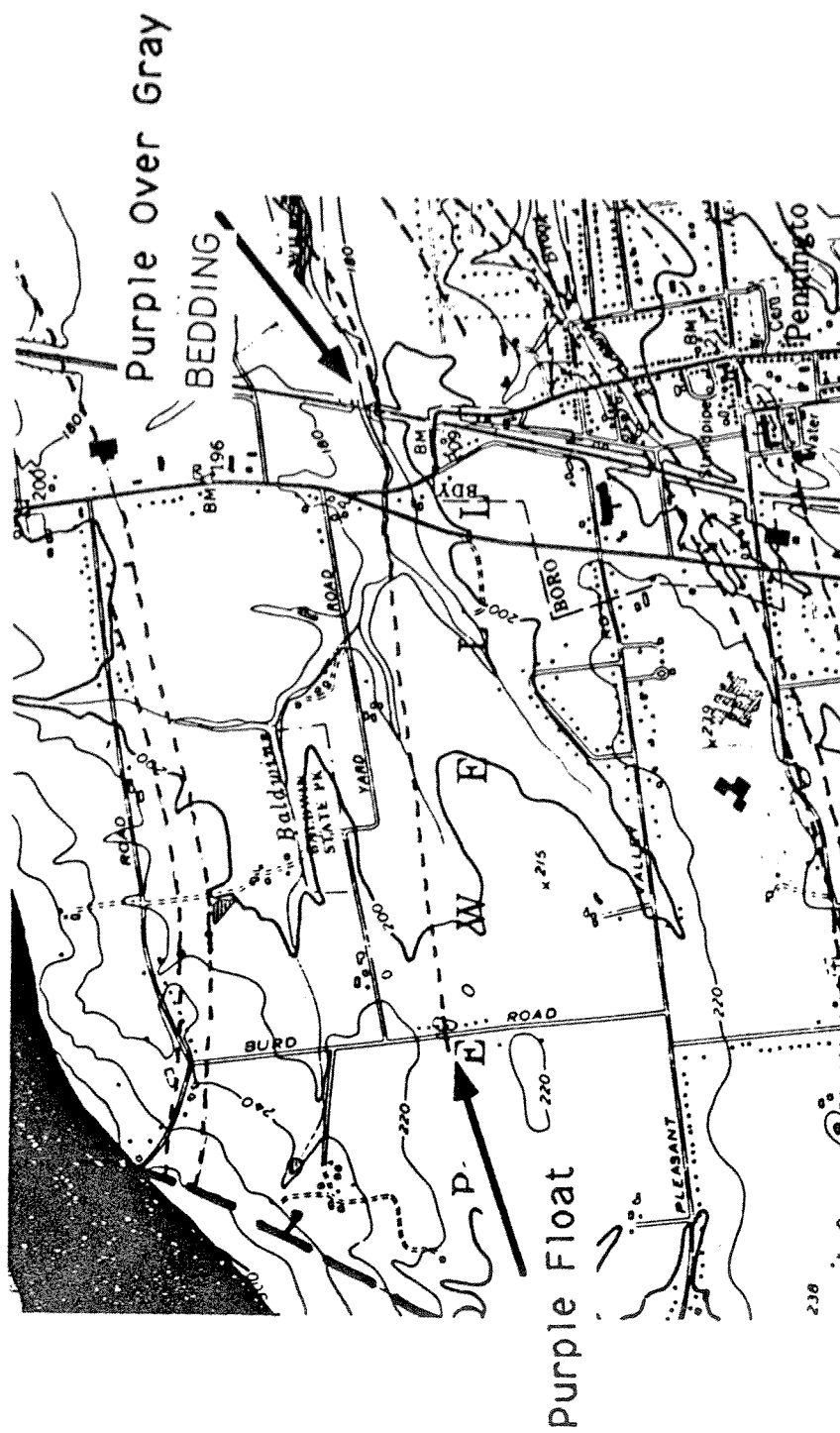
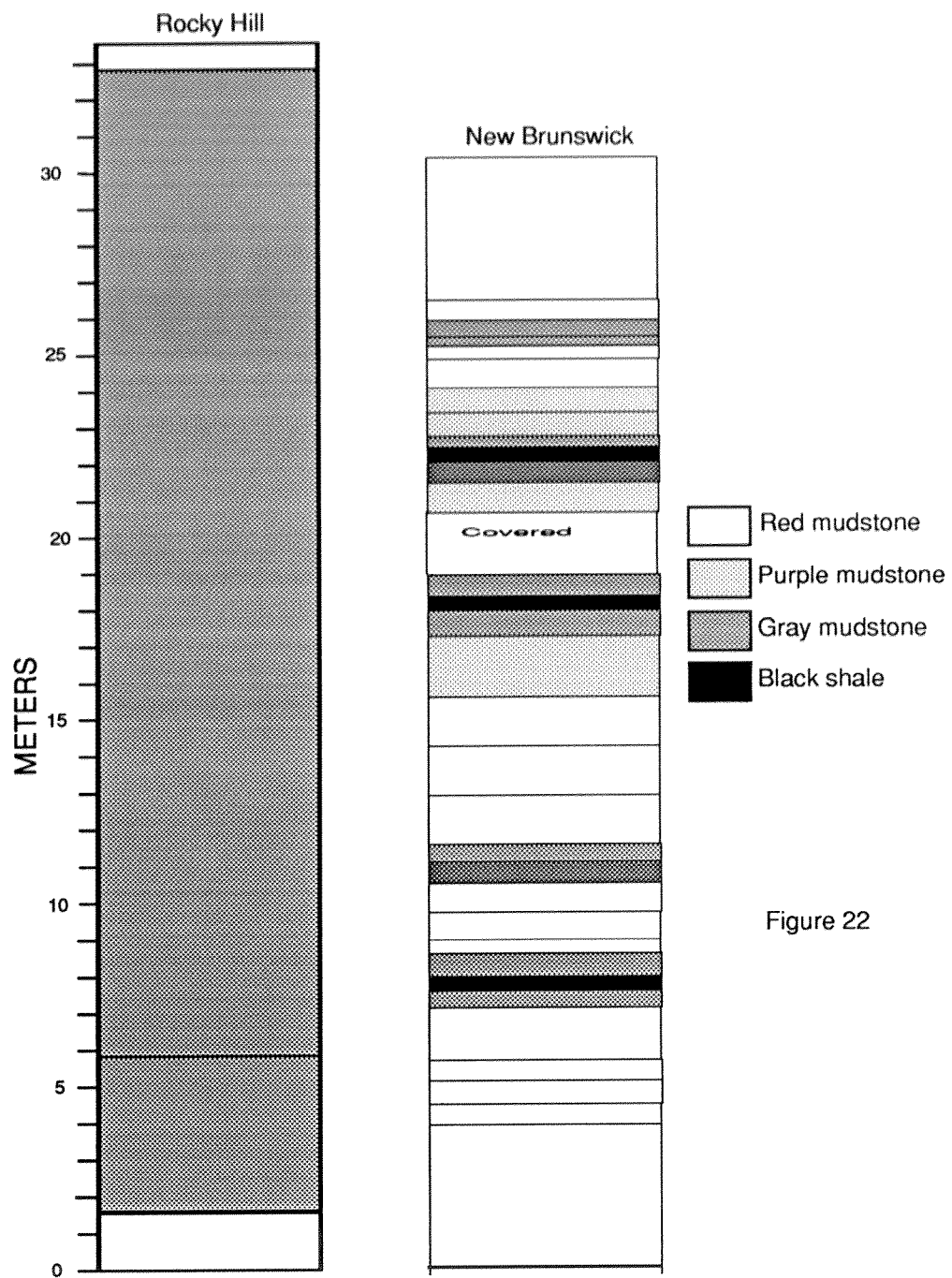


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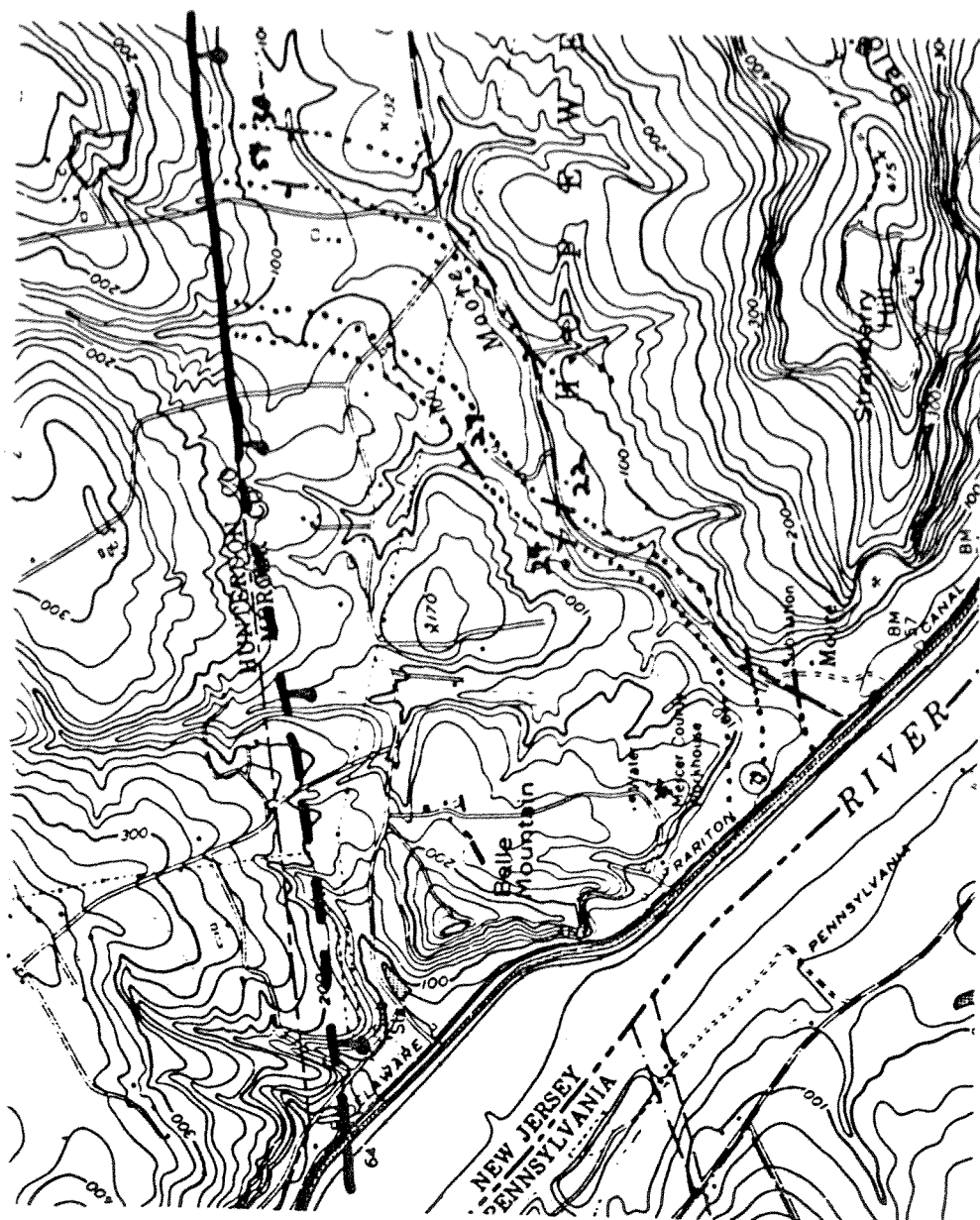


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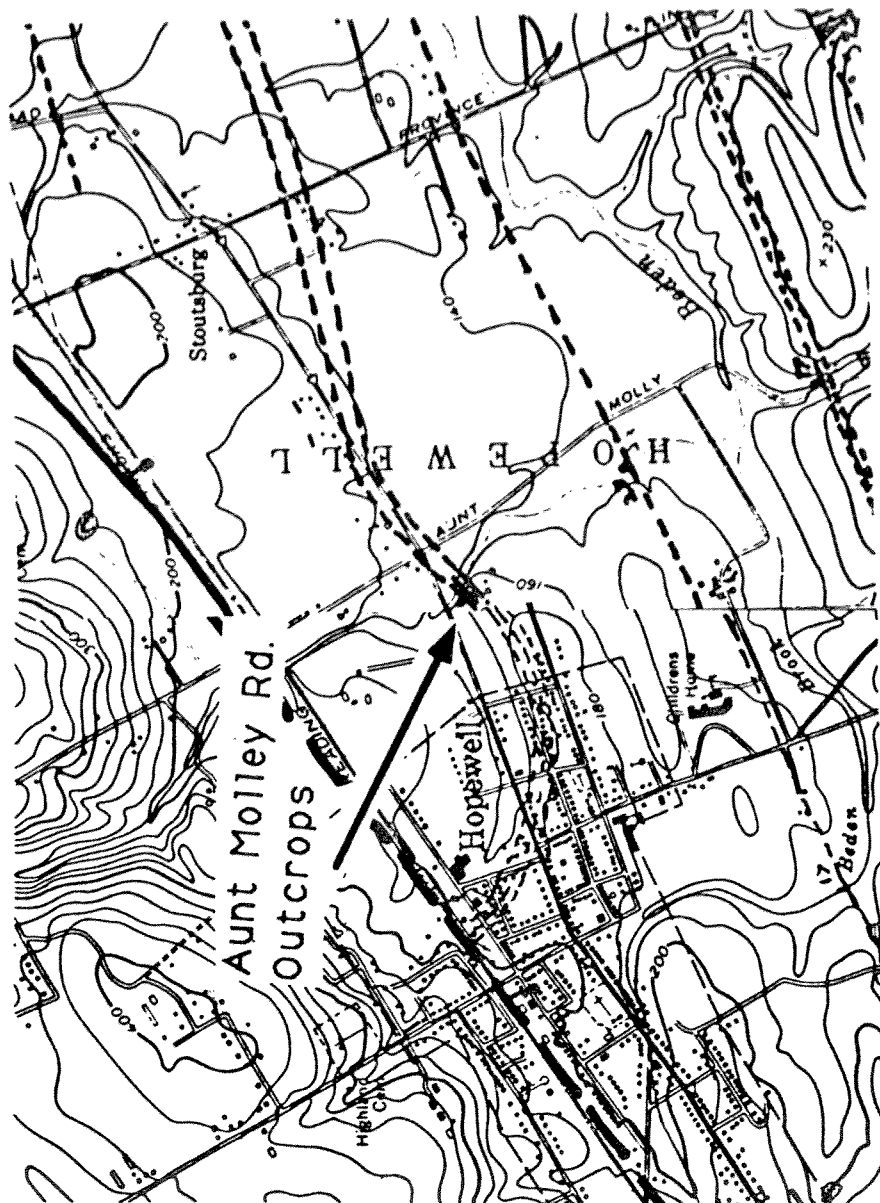


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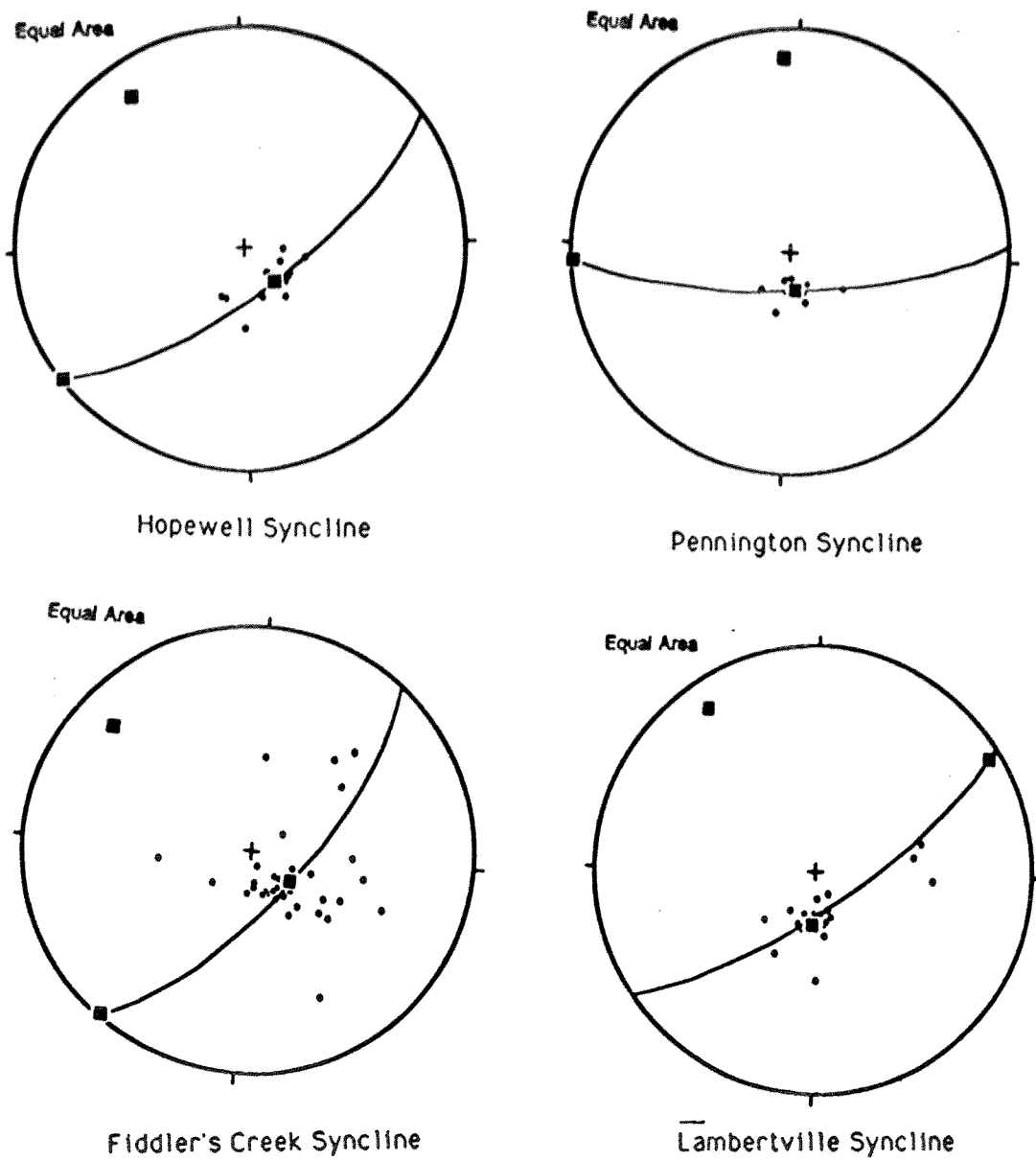


Figure 25

Steeply Dipping Beds

Shallowly Dipping Beds

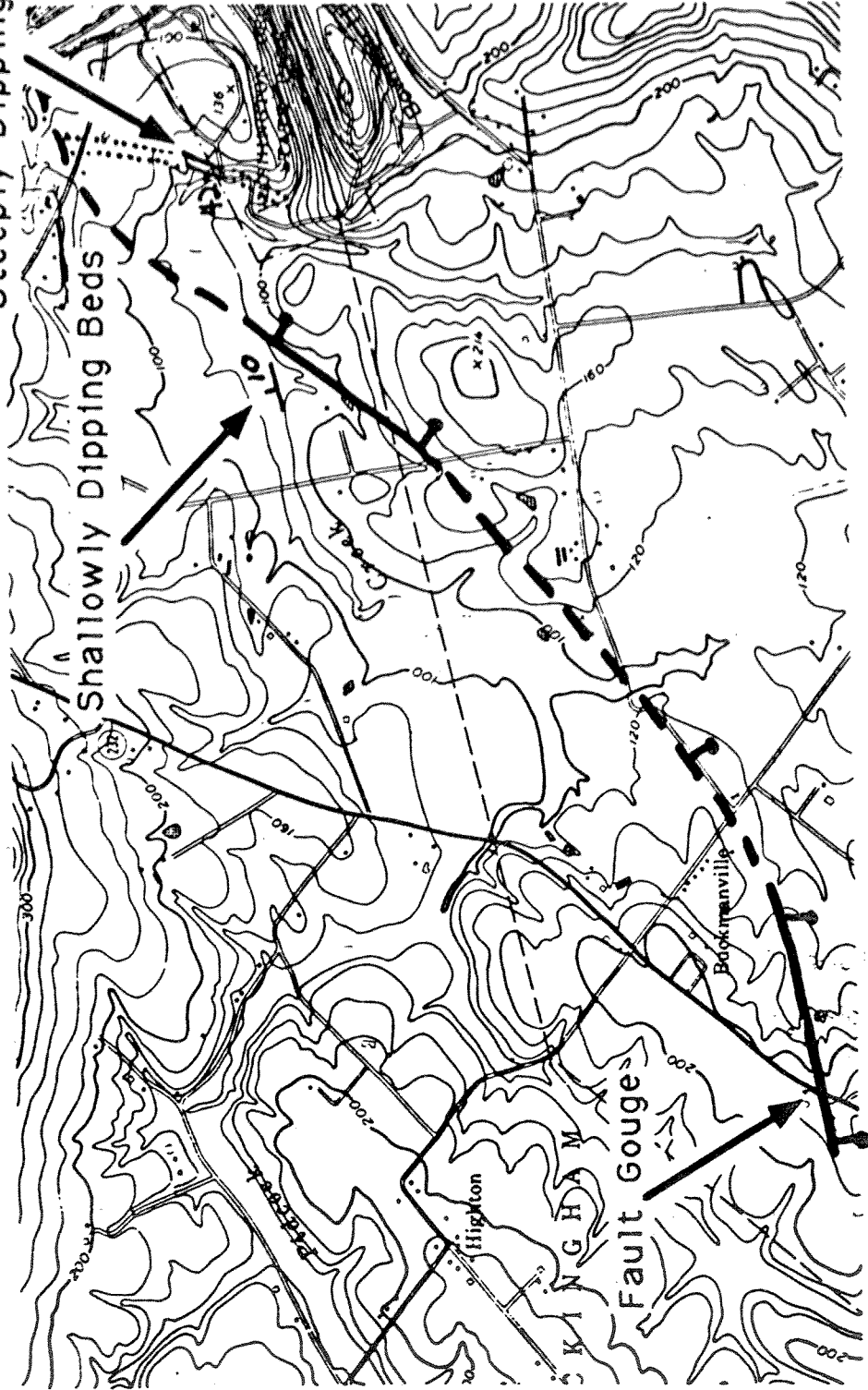


Figure 26

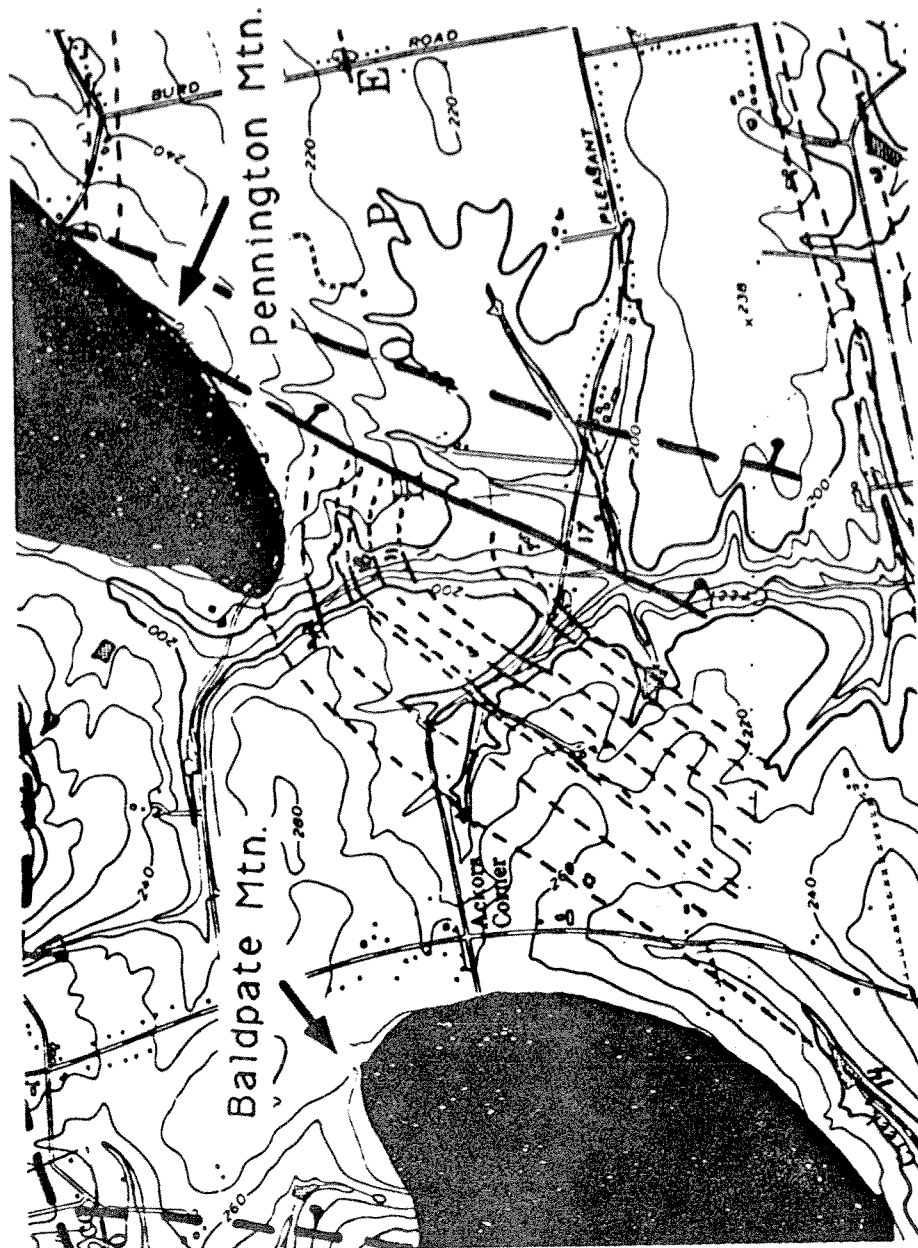


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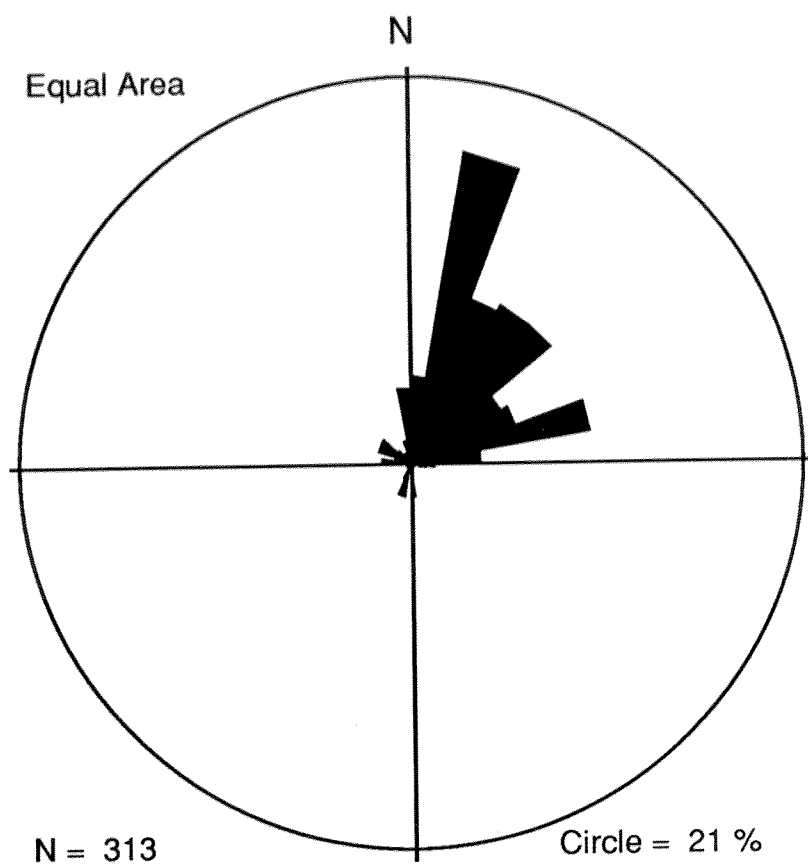


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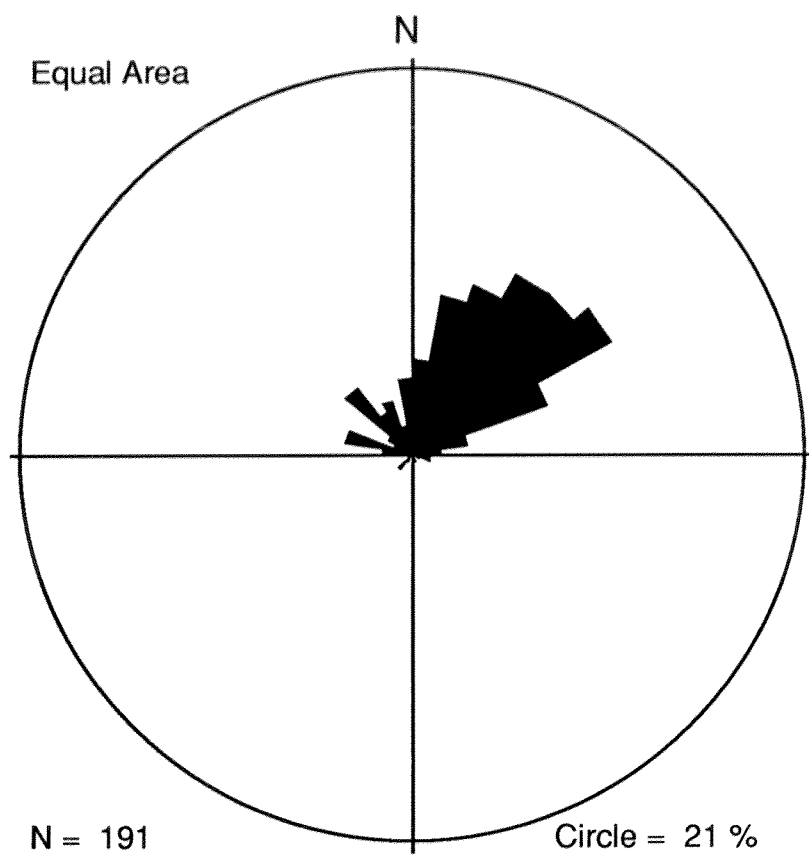


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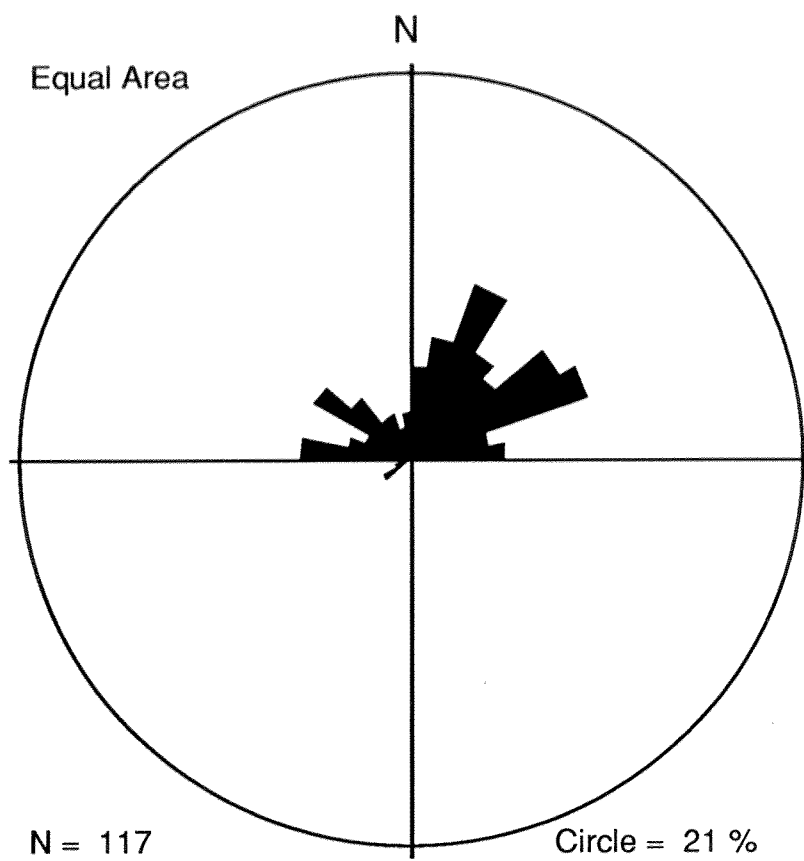


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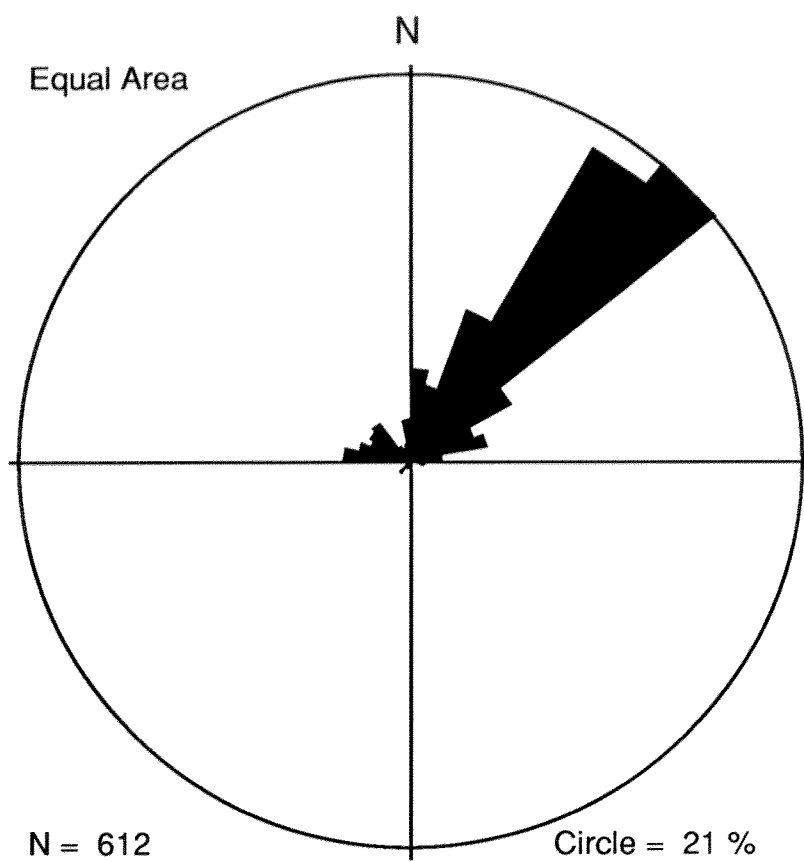


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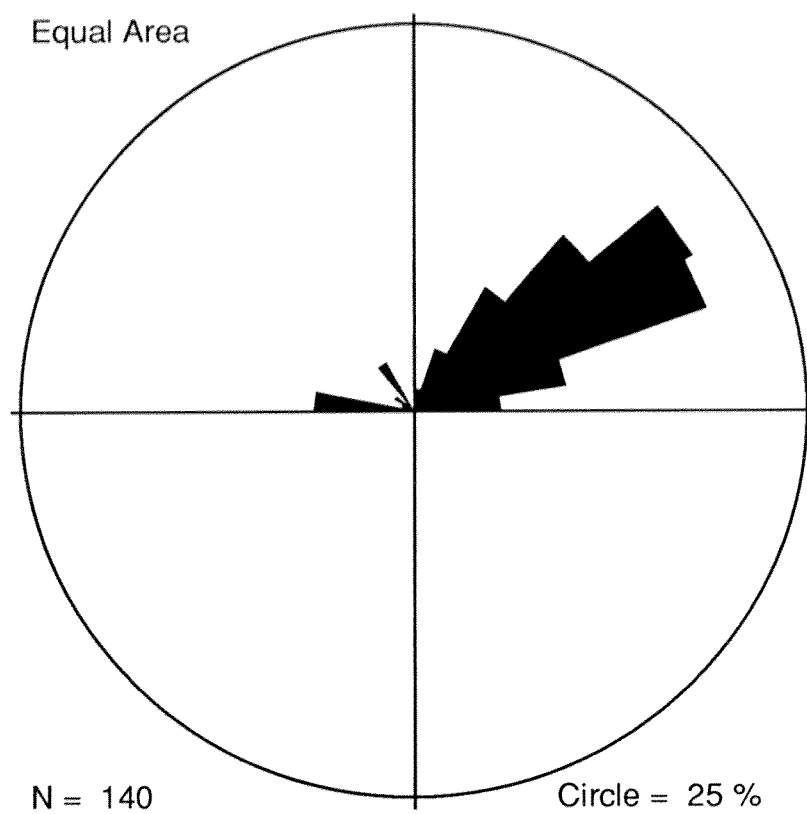


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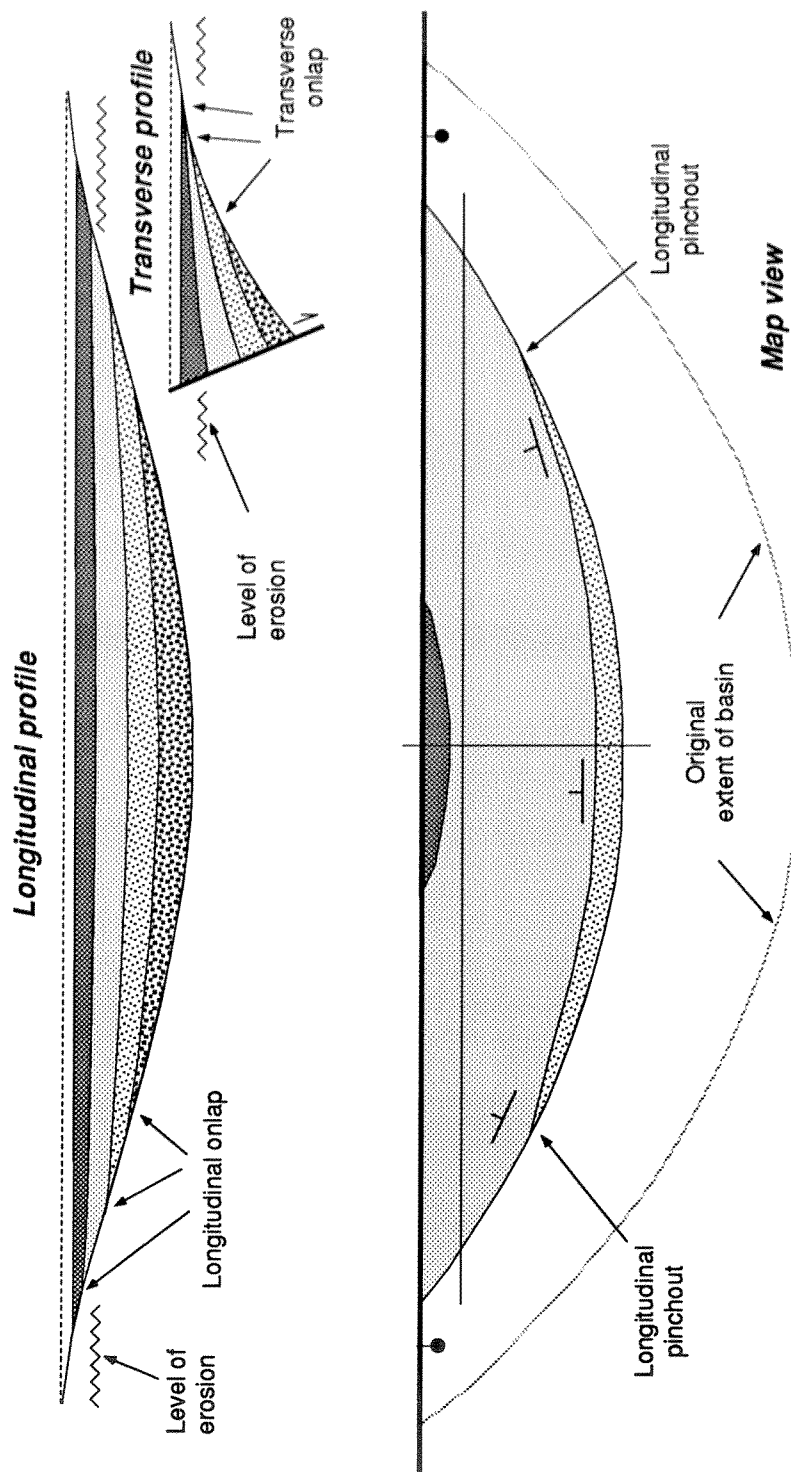


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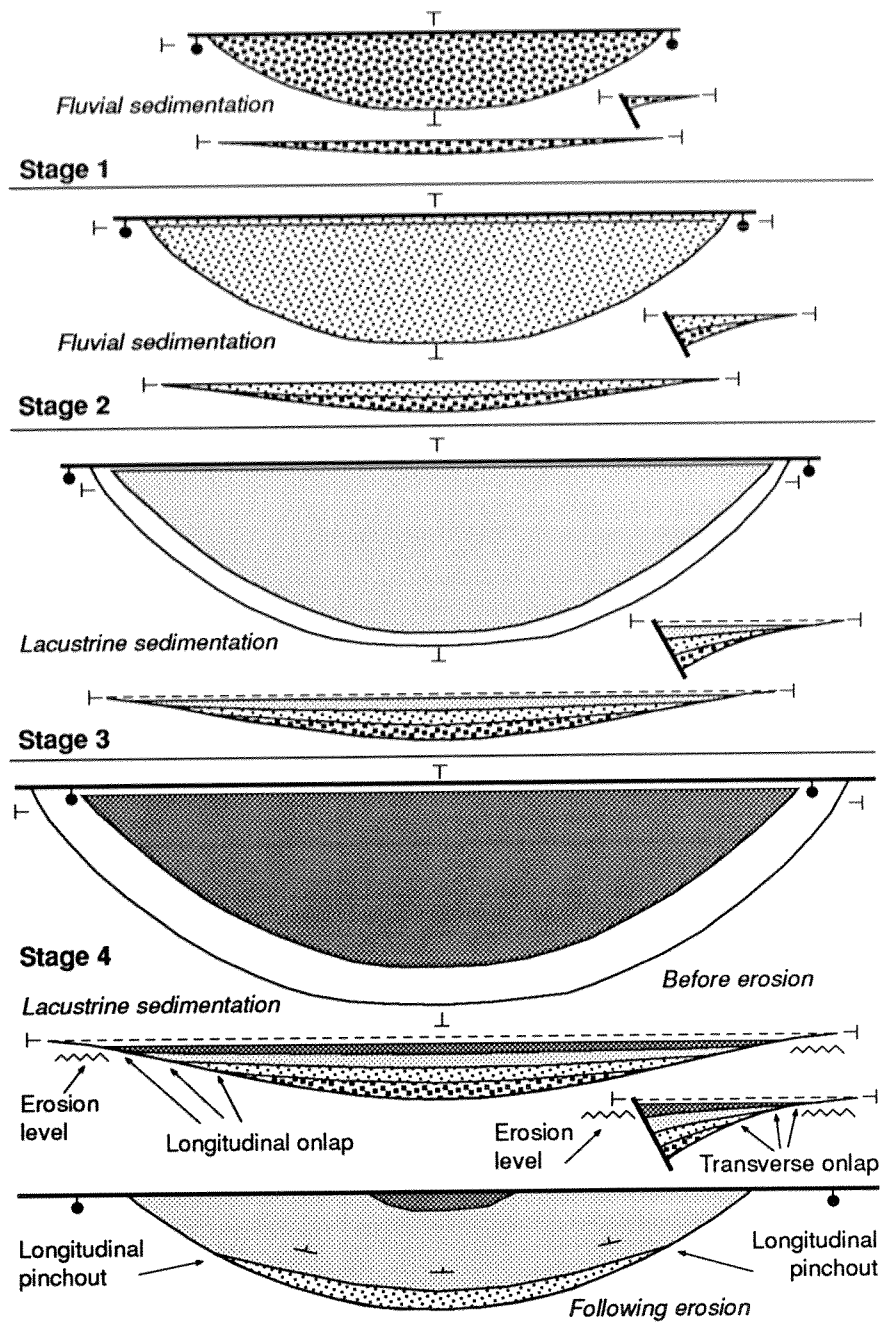


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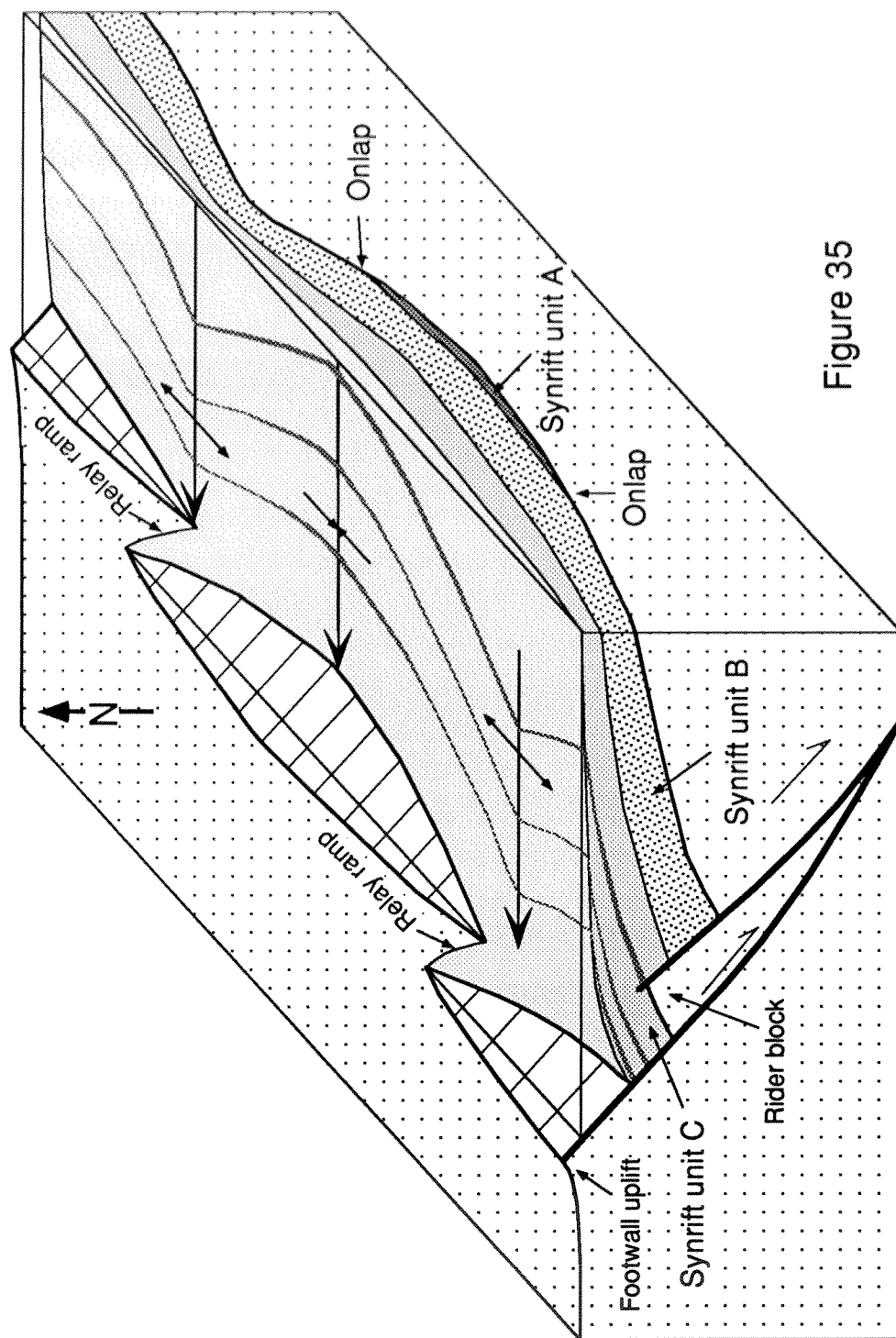


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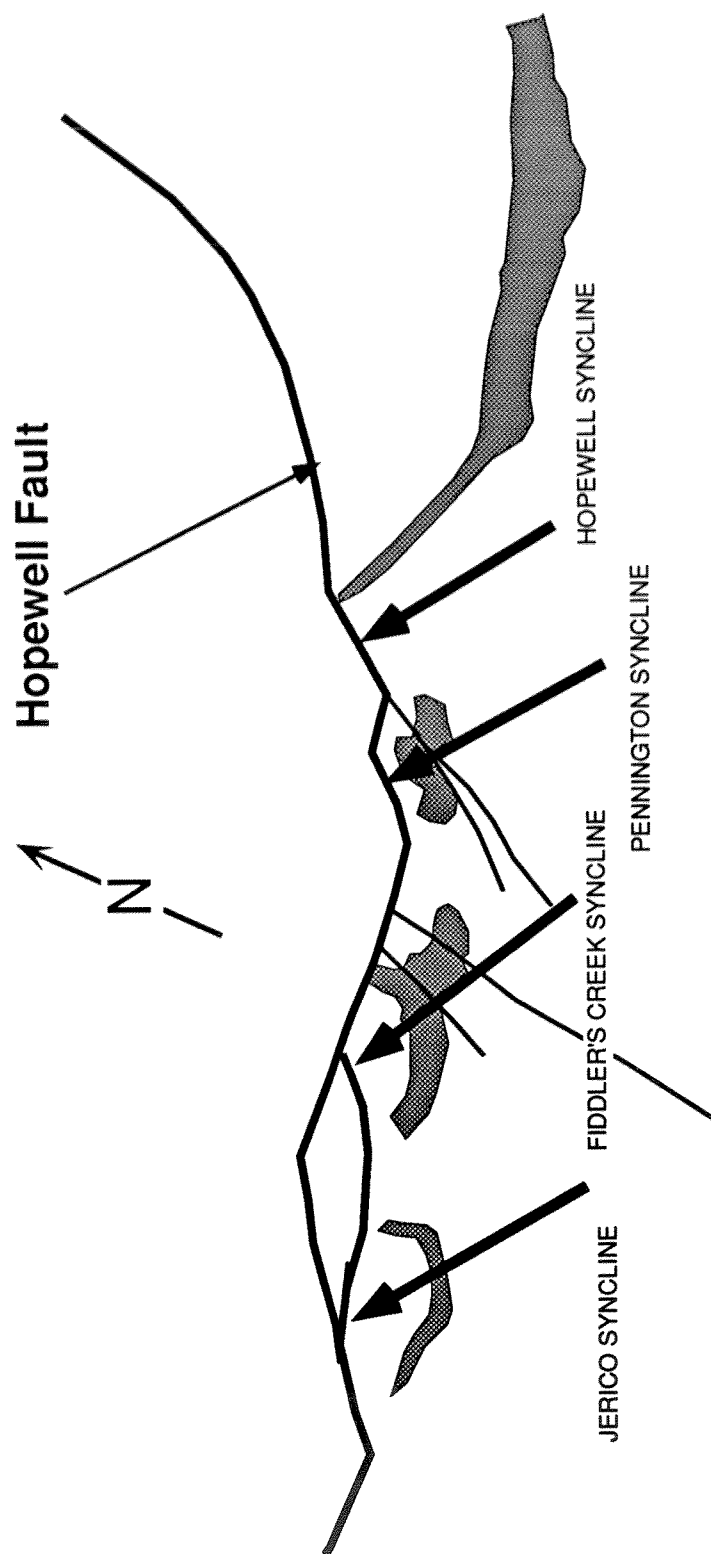


Figure 36

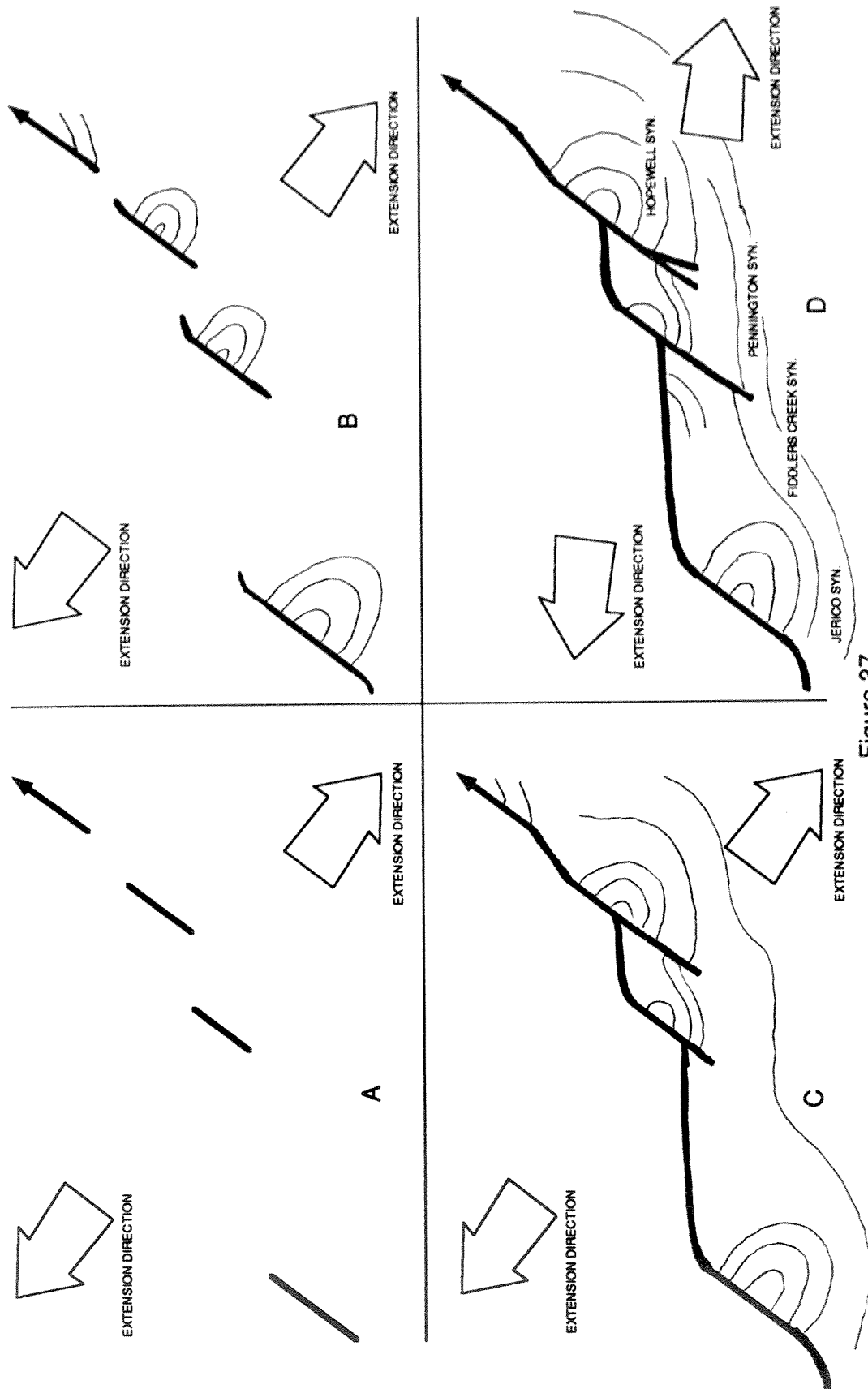


Figure 37

## APPENDIX

This appendix contains the geologic maps of the Lambertville, Pennington, Princeton, Hopewell, and Rocky Hill quadrangles. Figures A1-A5 are simplified geological maps of the quadrangles that have been reproduced to fit on a single page. In addition to the mapping by the author, these summary maps also contain geologic map information culled from other sources. Foldouts of the topographic maps themselves (Figures A6-A10) can be found in the plastic sleeves at the end of the appendix. These maps only contain information obtained by the author.

Figure A1. Geologic map of the Lambertville topographic quadrangle, Pennsylvania and New Jersey. The central portion of the map was mapped by the author. Additional sources of information include: Willard et al. (1959) and Ratcliffe and Burton (1988). Dashed lines indicate proposed correlation with Pennington quadrangle geology.

Figure A2. Geologic map of the Pennington topographic quadrangle, New Jersey. The northern half of the map was mapped by the author. Additional sources of information include mapping by Hugh Houghton, Marjorie Levy, Paul Olsen, Ron Parker, and Roy Schlische as well as Lyttle and Epstein (1987).

Figure A3. Geologic map of the Princeton topographic quadrangle, New Jersey. The northwestern corner of the map was mapped by

the author. Additional sources of information include mapping by Hugh Houghton, Marjorie Levy, Paul Olsen, and Roy Schlische as well as Lytle and Epstein (1987).

Figure A4. Geologic map of the Hopewell topographic quadrangle, New Jersey. The southeastern corner of the map was mapped by the author. Additional sources of information include mapping by Bruce Cornet, Paul Olsen, Roy Schlische, and ShayMaria Silvestri as well as Lytle and Epstein (1987).

Figure A5. Geologic map of the Rocky Hill topographic quadrangle, New Jersey. The southern half of the map was mapped by the author. Additional mapping based on Lytle and Epstein (1987).

Figure A6. Geologic map of the Lambertville topographic quadrangle, Pennsylvania and New Jersey.

Figure A7. Geologic map of the Pennington topographic quadrangle, New Jersey.

Figure A8. Geologic map of the Princeton topographic quadrangle, New Jersey.

Figure A9. Geologic map of the Hopewell topographic quadrangle, New Jersey.



Figure A10. Geologic map of the Rocky Hill topographic quadrangle,  
New Jersey.

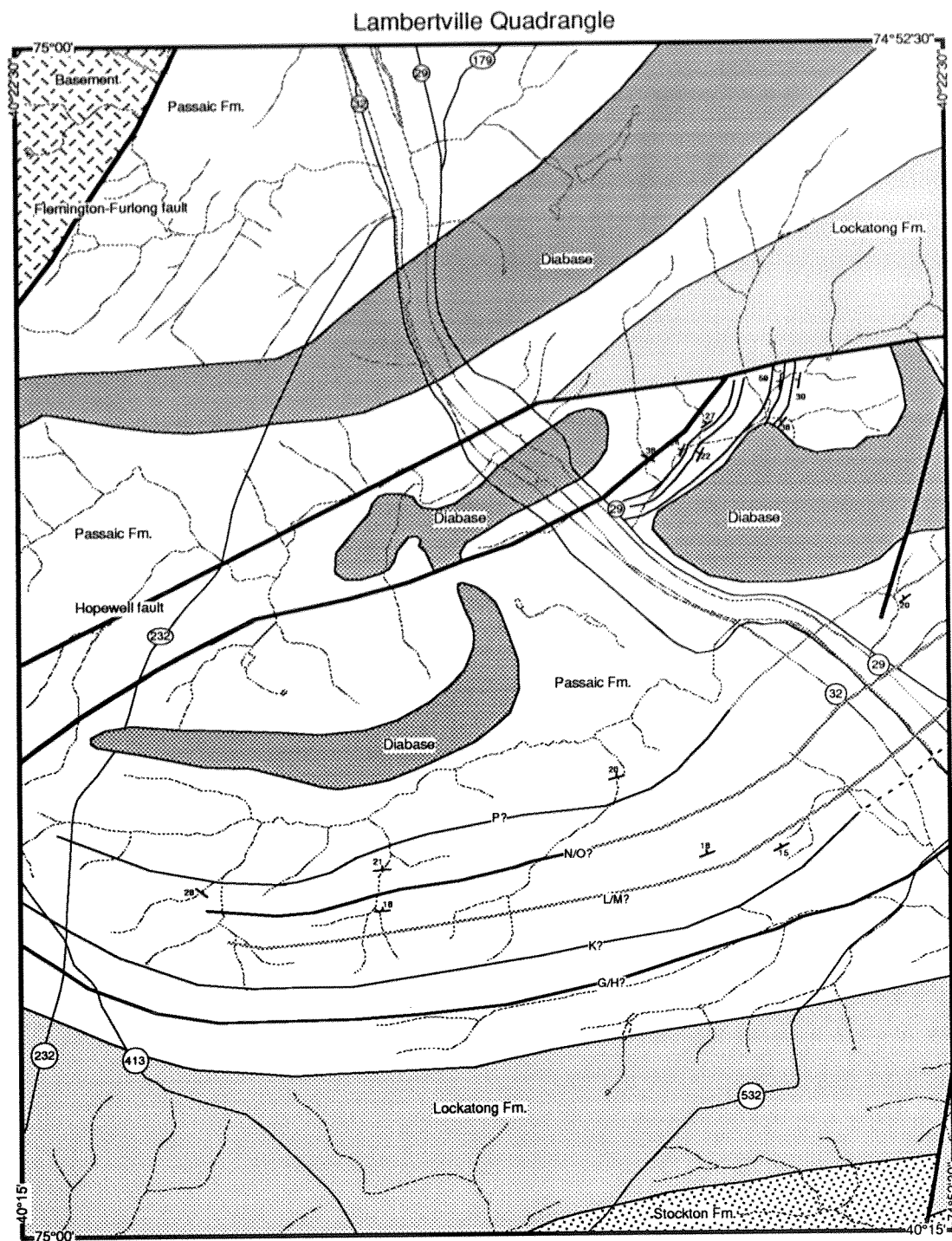
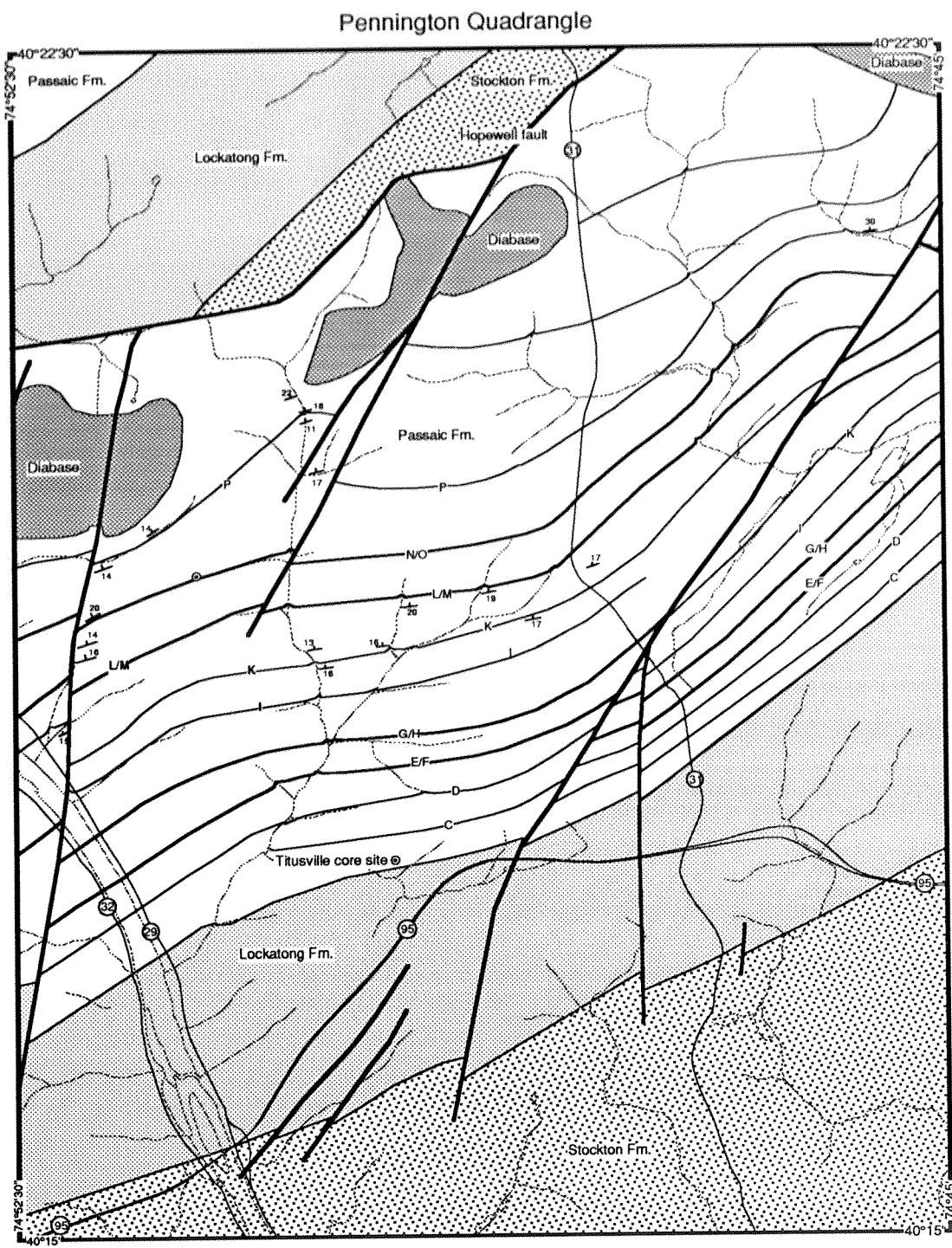
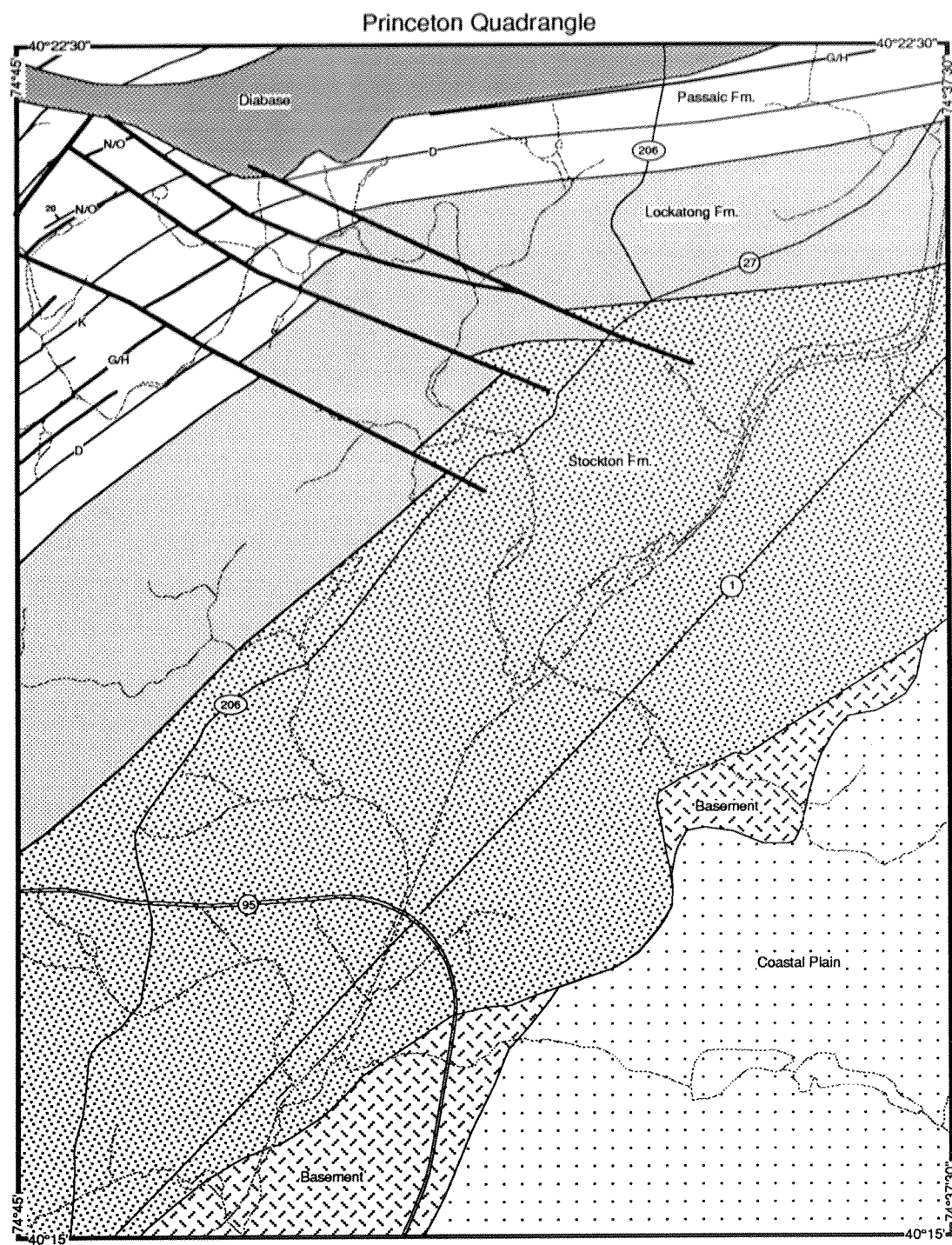


Figure A1





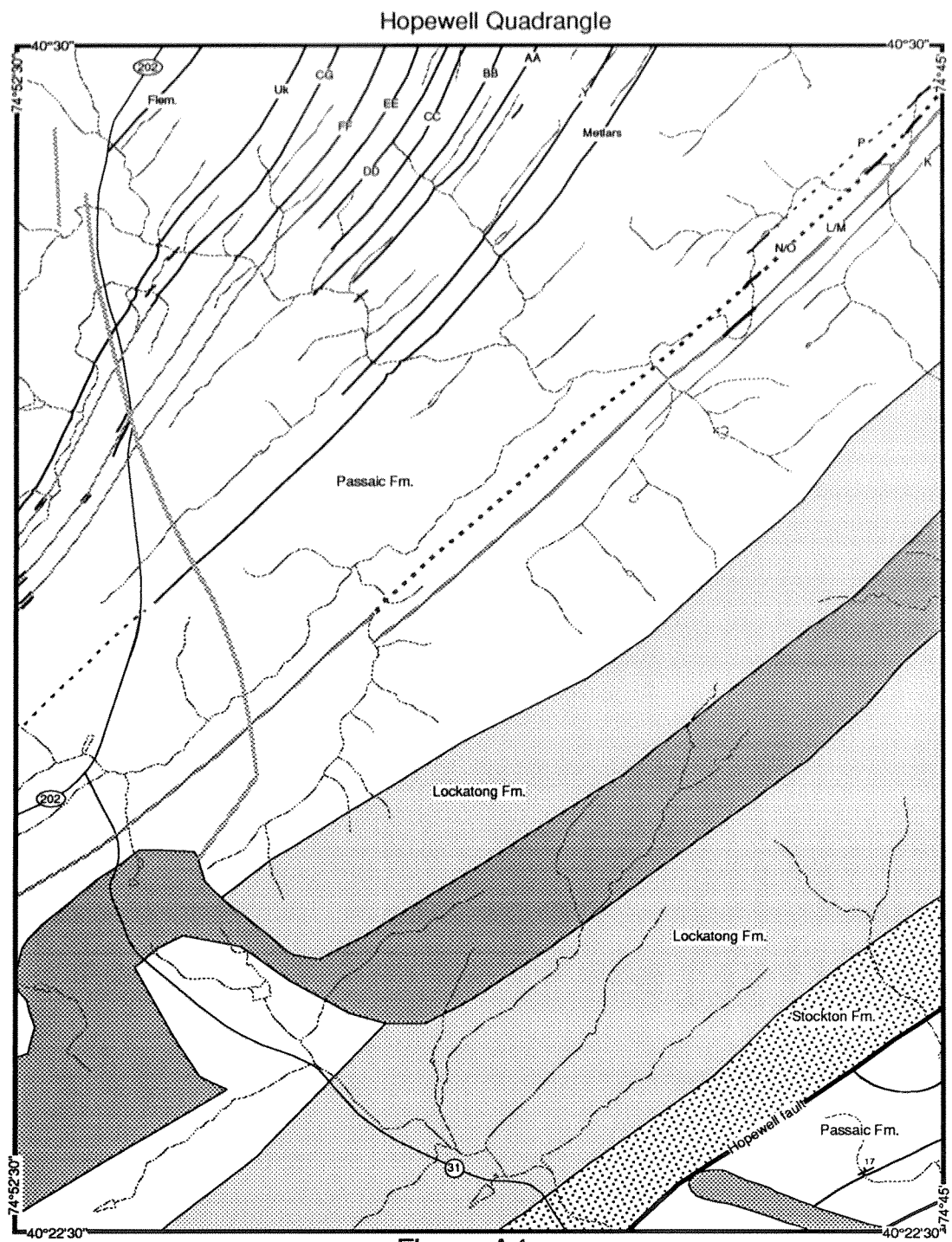


Figure A4

