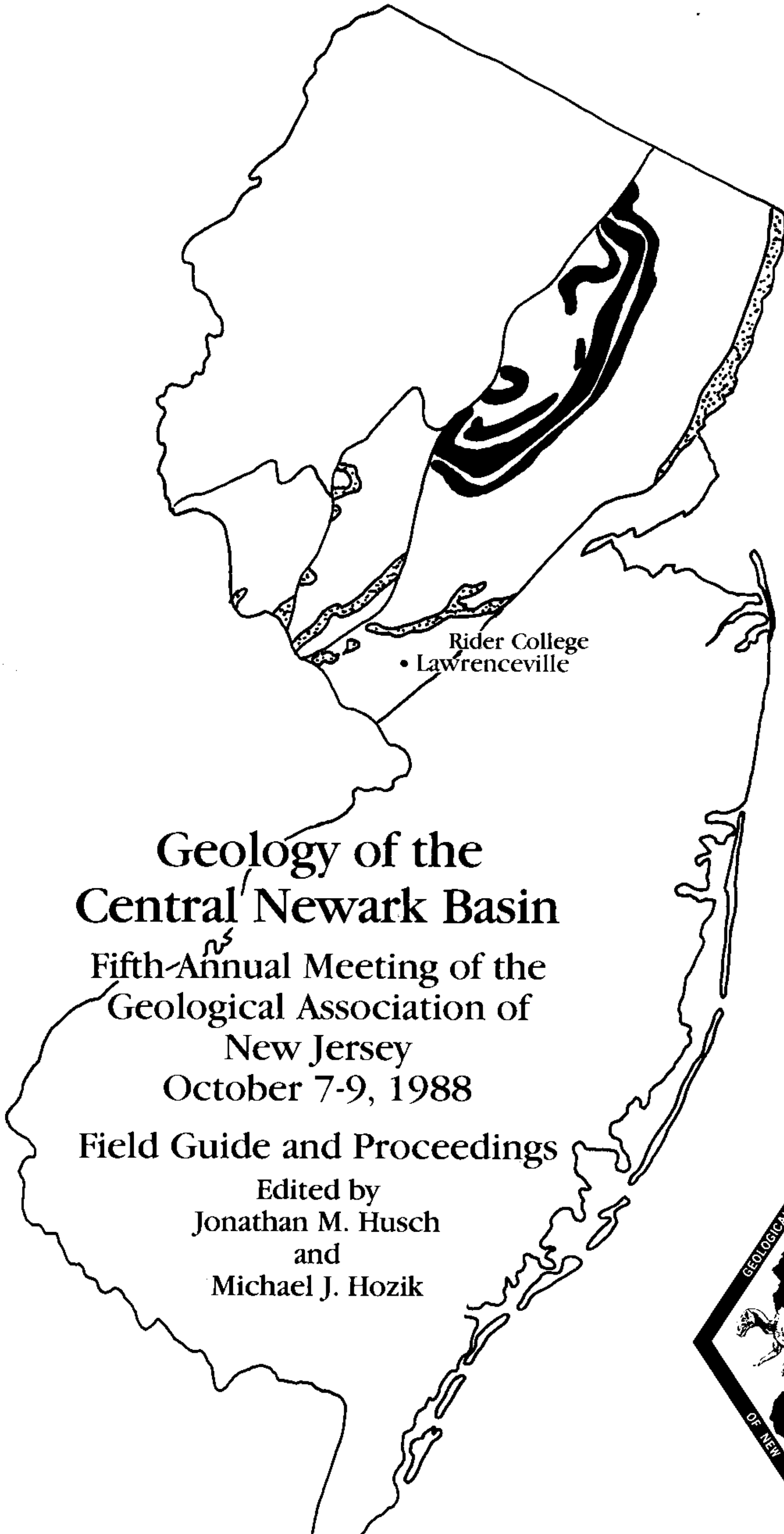

Geology of the Central Newark Basin

Fifth Annual Meeting of the Geological Association of New Jersey
October 7-9, 1988

Field Guide and Proceedings

Edited by
Jonathan M. Husch
and
Michael J. Hozik

GEOLOGY OF THE CENTRAL NEWARK BASIN

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THE GEOLOGY OF THE CENTRAL NEWARK BASIN

TABLE OF CONTENTS

FIELD GUIDES

PREFACE

Organizing the fifth annual meeting of the Geological Association of New Jersey has been a exciting experience overall, punctuated by intervals of immense frustration (panic) when it seemed like none of the pieces would fall into place. It has been most satisfying to watch the association grow and continue as a stable, professional organization and to see our field conference become a tradition. The high quality of papers submitted for this volume demonstrates the respect the Geological Association of New Jersey has acquired over the last several years. We are proud to be associated with this effort.

This volume is a summary of recent thinking about the geology of the Newark Basin, focusing on its central region. In a sense, Franklyn Van Houten laid the groundwork for this conference more than twenty years ago with his classic descriptions of the chemical and detrital cycles in Newark Basin sediments. It is on this foundation that many of the contributors to this volume have built their own work. More recently, Paul Olsen expanded our understanding of the basin-wide significance of these cycles and has made notable progress in deciphering their temporal implications. In 1973, Rodger Faill suggested, contrary to the then-current paradigm, that much of the faulting in the Newark Basin occurred late in its history. This view seems much less heretical today. Given this combined expertise, we are honored to have the opportunity to accompany Van, Paul, and Rodger into the field.

Many contributors have added greatly to the scientific and educational value of this volume. Roy Schlische, in a paper co-authored with Paul Olsen, presents a view of basin tectonics which is sometimes divergent from the ideas detailed by Faill. Regardless of who is closer to the truth, their articles serve to define some of the relevant structural problems which need to be addressed. John Puffer's discussion of the geochemistry of the Watchung basalts in particular, and Mesozoic igneous activity in eastern North America in general, sheds new light on an important and much-debated petrologic topic. Warren Cummings' suggestion of an alternative process for the secondary mineralization of the Watchung basalts makes it difficult not to question older ideas. Mike Fedosh and Joe Smoot's description of the cored stratigraphic section adds information to our understanding of the basin's stratigraphy. This is particularly useful given the frequent lack of extensive outcrop. Karl Muessig and Hugh Houghton's summary of the geologic controls on domestic radon levels emphasizes our fundamental dependence on the geological environment. Finally, Bill Gallagher and Reg Shagham's Earth Science

Teachers' workshops provide useful material for all of us who work with students in the classroom. We are pleased to be able to add the results our own work to this already impressive collection.

None of this would have been possible without the cooperation of a vast number of individuals. Franklyn Van Houten, Paul Olsen, Roy Schlische, Geraldine Hutner, Betty Falkenstein, Mary Kildea, and Patty Coleman-Weeks deserve special mention for their efforts way above the call of duty. We thank all of the contributors for their cooperation, dedication, and geologic insight. We hope you find the results of all our efforts both educational and enjoyable.

Jonathan M. Husch
Co-editor

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LATE TRIASSIC-EARLY JURASSIC DEPOSITS,

NEWARK BASIN, NEW JERSEY

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General Framework

In late Paleozoic time the cratonic blocks of Laurussia and Western Gondwana converged, producing the Alleghenian-Mauritanide-Variscan orogenic uplands between them. During initial breakup of the supercontinent in late Triassic time rifting along the flanks of these orogenes formed tracts of extensional basins (Fig. 1). At least 13 major and 7 minor early Mesozoic rift basins developed in a 2000-km belt that follows the fabric of the Appalachian Piedmont in eastern United States and adjacent maritime Canada.

Throughout the region these basins were filled with reddish-brown fluvial, deltaic, mudflat, and lacustrine facies of a piedmont-valley flat suite with minor tholeiitic extrusive and intrusive rocks. Coeval rift basins filled with rather similar deposits are present beneath the Atlantic Coastal Plain as well as on the Continental Shelf. Essentially correlative redbeds and igneous rocks are found in northwestern Africa and southwestern Europe and on their continental shelves (Manspeizer, 1988). In contrast to the basins and deposits in eastern North America those now on the eastern side of the Atlantic Ocean are somewhat more varied in style and content, and they include evaporites that accumulated in a westward transgression of the Tethys Sea across northern Africa. In fact, this transgression apparently reached the Atlantic margin of North America in Late Triassic time, before opening of the central Atlantic basin had begun.

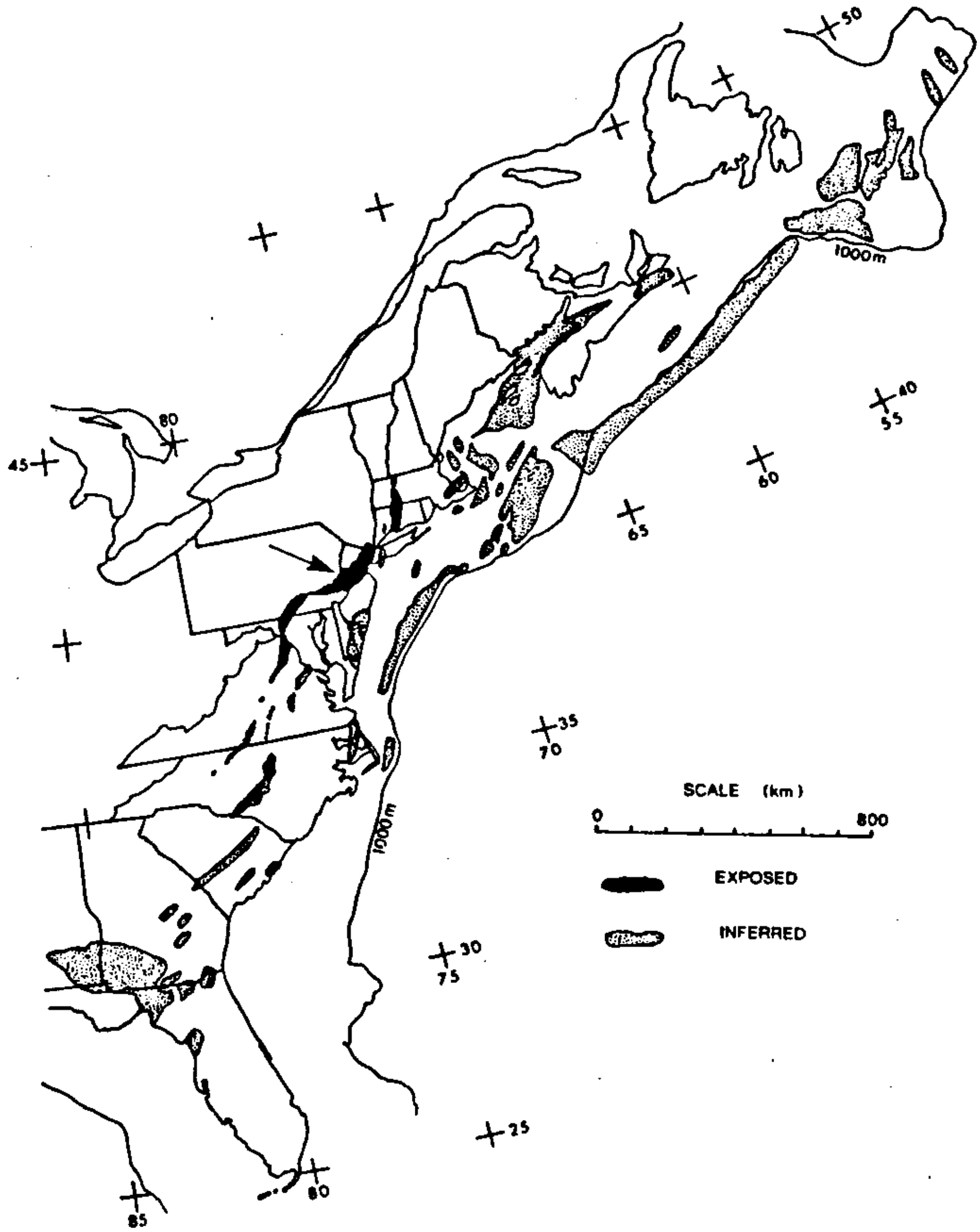

Fig.1. Distribution of Newark Supergroup in eastern North America, after Olsen, 1980. Newark Basin indicated by arrow.

Syn-rift basin filling continued for at least 30 my during Late Triassic and Early Jurassic time (Fig. 2). The associated igneous activity in many of the basins occurred some 20 my after the earliest rifting. Then, at the beginning of the subsequent drift stage, the basins were broken by major faults, intruded by dikes, and extensively eroded before the onset of sea-floor spreading initiated the opening of the Atlantic Basin.

In Late Triassic and Early Jurassic time, eastern North America and adjacent Africa lay near the paleo-equator, drifting northward about 10-15 degrees during this interval (Manspeizer, 1982, p. 910-911). A warm savanna climate prevailed throughout much of the region. In Late Triassic time conditions were moist enough to support coal swamps. The climate was drier in the north, however, and by Early Jurassic time arid conditions in maritime Canada induced accumulation of minor evaporites and eolian deposits.

Newark Basin

Sedimentary Sequence

A deeply eroded remnant of the Late Triassic-Early Jurassic (Carnian to Sinemurian) non-marine Newark Supergroup is preserved in the southwest-trending Newark Basin which extends 225 km from Rockland County, New York, to northeastern Lancaster County, Pennsylvania. These deposits comprise the thickest composite Newark sequence (10,000 m) in eastern North America. As in the other basins, it lies with marked unconformity on early Paleozoic and

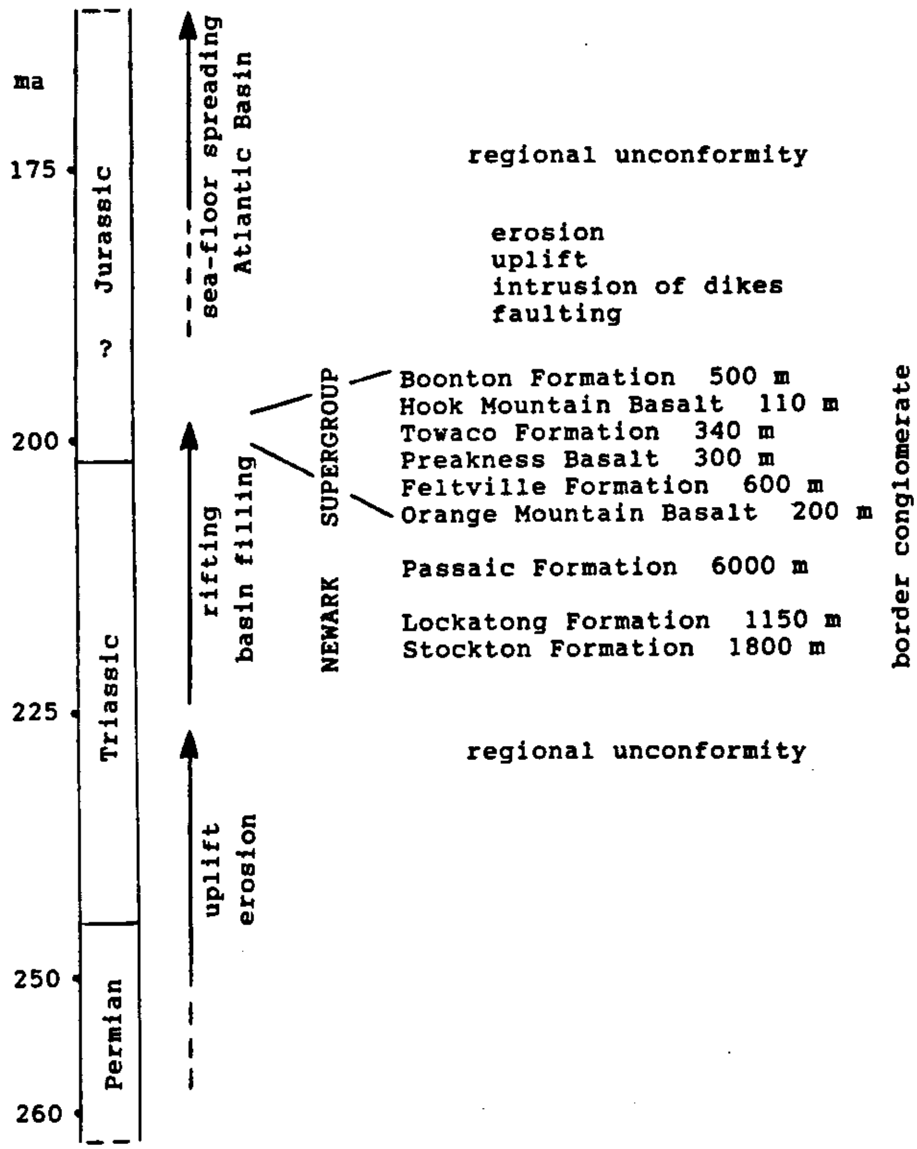

Fig.2. Formations of Newark Supergroup in Newark Basin, with maximum thicknesses. Arrows indicate duration of regional tectonic stages. Question mark locates alternative estimated age of the Triassic-Jurassic boundary.

Precambrian rocks of the Blue Ridge and Piedmont provinces. It is faulted against them along the northwestern margin of the basin and overlaps on them along the southeastern margin. The formations of the Supergroup (Fig. 2,3) are best developed in the central part of the basin. Near the faulted northwestern border, as well as at the northeastern and southwestern ends of the basin, the distinctive formations interfinger with border conglomerates.

Source Areas

The Newark basin fill was derived largely from eastern and southeastern uplands not far southeast of the present outcrop margin. Abundant sodic feldspar in the Stockton and Passaic formations, and the interfingering of these deposits with the soda-rich Lockatong Formation, attest to the derivation of most of these sediments from crystalline rocks exposed east and southeast of the basin. This conclusion is supported by rather scanty paleocurrent data derived mostly from the Stockton Formation (Klein, 1963, p. 805).

Fossil Content

Fossils recovered from deposits in the Newark Basin include leaves from the lower part of the Supergroup that suggest a warm savanna climate. In addition, there are spores, pollen, reptiles (mostly footprints), and fish which help to support local and regional correlation and restrict an episode of rifting and basin filling to Late Triassic and Early Jurassic time. Other fossils indicative of local facies are arthropods, clams, root traces, burrows, coprolites, and fecal pellets (Olsen, 1980, p. 19).

Igneous Rocks and Hornfels

Diabase in the lower part of the Newark Supergroup crops out in the thick Palisades and Lambertville sills, as well as in a few smaller intrusions in central and northeastern

New Jersey and adjacent New York. In eastern Pennsylvania, discordant, dish-shaped diabase sheets are common. Three multiple lava flows in the upper part of the Supergroup are preserved principally in the Watchung Mountains in the northeastern third of the basin. These intrusive and extrusive rocks are olivine-poor quartz tholeiite characteristic of rift valleys developed on continental crust.

Newark sedimentary rocks were metamorphosed locally by both sills and lava flows. The effect of the Watchung lava on the associated reddish-brown mudstone amounted to little more than alteration of its color to greenish gray several centimeters from the contact. In contrast, the effect of the intrusions was significant. The fine-grained metamorphic rocks (hornfels) record the role of temperature, related to distance from an intrusion, and of composition, controlled by the sedimentary rock type involved (Van Houten, 1971). Among the Newark formations the Lockatong was the most susceptible to thermal metamorphism. It has been extensively reconstituted to varieties of very fine-grained biotite-Na-feldspar hornfels marked by the absence of quartz.

Conditions of Deposition

After deep regional erosion of the Alleghenian uplands during a 20-30 my Permian and early Triassic hiatus (Fig. 2) local debris flows, sheet floods, and steep-gradient braided streams built a belt of alluvial fans several kilometers wide along the faulted northwestern border of the Newark Basin. During lulls in aggradation caliche paleosols developed on the fans. At the same time streams from the upland source area to the southeast began to spread Stockton sand across the basin. During this episode the principal drainage presumably flowed along the length of the basin. Widespread accumulation of sand ended with central ponding of the longitudinal drainage, forming a long Lockatong lake with narrow marginal mudflats. Upper Stockton sand in the northwestern interior of the basin built deltas along the southern shore of the lake (Turner-Peterson and Smoot,1985), while its narrow northern shoreline facies interfingered with marginal alluvial fans.

Once established, sedimentation in the Lockatong lake varied within rather restricted limits for several million years. In this quiet setting variations in climate exerted a major control on the lacustrine deposits, producing small-scale "chemical" and "detrital" cycles (see Olsen, this volume).

With the waning of the large Lockatong lake, fluvial and floodplain sediments accumulating along its margins extended across the basin. In much of the interior, broad mudflats with wandering water courses and weak external drainage prevailed for many millions of years. Toward the end of the rifting stage (Fig. 2) continued accumulation of fluvial, floodplain, and recurring lacustrine deposits was interrupted three times by extrusion of extensive basaltic lava flows. During this later filling of the basin the northwestern uplands probably supplied an increased amount of sediment.

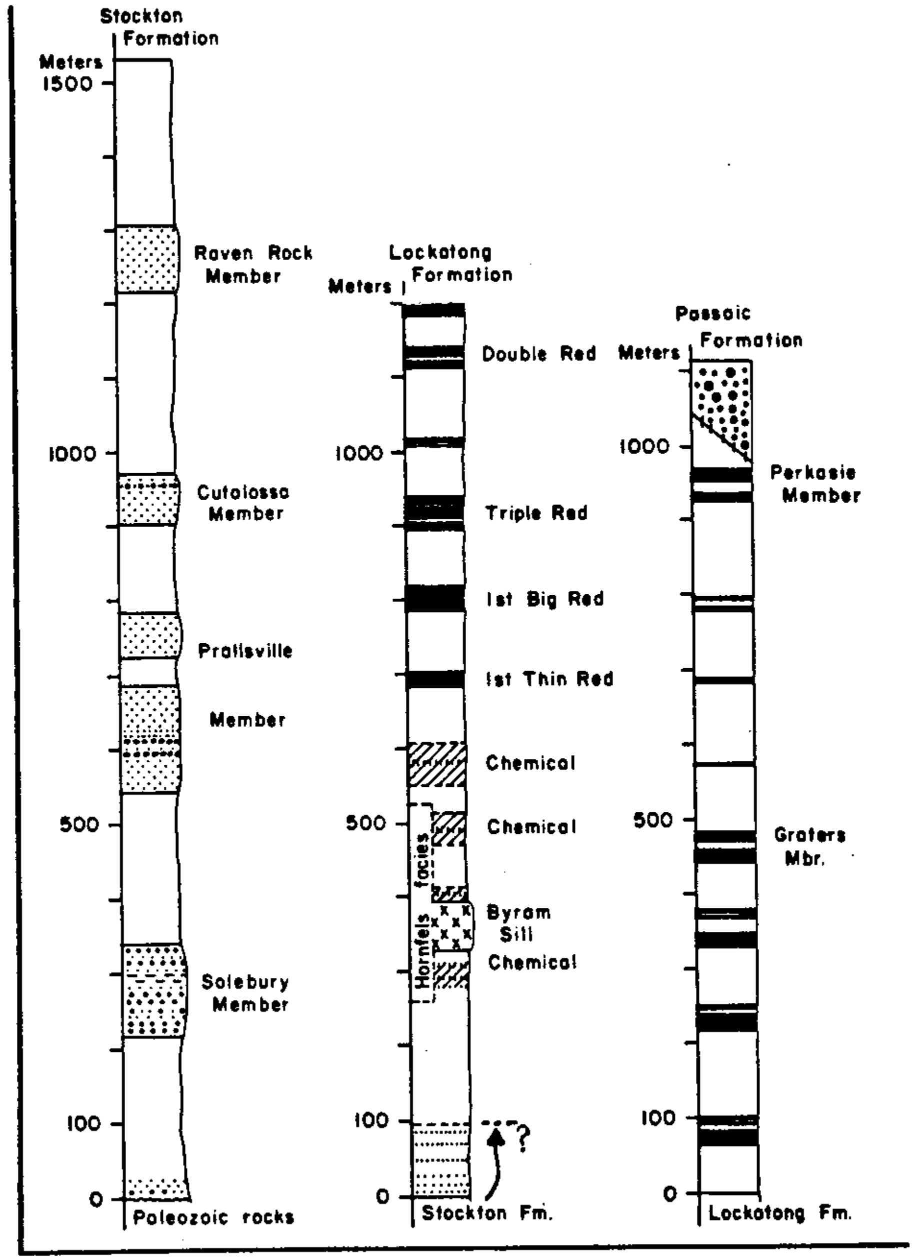

Fig.3. Stratigraphic section of Newark Supergroup in Delaware valley. Black units in Lockatong Formation are grayish-red to reddish-brown analcime-rich chemical cycles; intervals of gray chemical cycles are also indicated. Black units in Passaic Formation are dark gray detrital deposits. Question mark and arrow locate alternative base of the Lockatong Formation.

Delaware Valley Section, West-Central New Jersey

The Late Triassic part of the Newark Supergroup exposed along the Delaware valley includes the Stockton, Lockatong, and Passaic formations and their associated border conglomerates, as well as a diabase sill intruded into the lower Lockatong strata (Fig.3). These formations are repeated in three large northwest-dipping (10-20 degrees) blocks separated by normal faults with displacement of as much as 5000 m (18,000 ft). Description of the sedimentary rocks given here applies mostly to those in the northern fault block, extending from Stockton in the south to Pebble Bluffs northwest of Milford. More detailed information about these formations is in Van Houten (1969) and Olsen (1980).

In the following brief review, and during the field trip, the focus will be on selected aspects of the sedimentation: 1) the southeastern upland source of most of the basin fill; 2) the relatively intense weathering of the source areas required to produce a nearly continuous supply of clay; 3) the evidence of aridity within the basin; 4) the long interval of stability in the rift basin documented in the Lockatong Formation; and 5) the several scales of cyclic sedimentation recorded in the basin fill.

Stockton Formation

The Stockton Formation consists primarily of yellowish-gray to pale reddish-brown fairly well sorted arkose and subordinate conglomerate and reddish-brown siltstone and mudstone distributed in rather distinct units (Fig. 3). These are generally as much as 15-20 m thick. The formation is finer-grained in the upper part, and it grades upward, through tough reddish-brown micaceous siltstone, into the Lockatong Formation.

The conglomeratic beds in the lower part of the formation contain moderately round clasts averaging about 2-3 cm in diameter. These are mostly quartz, with minor amounts of feldspar and metamorphic rocks, and locally with grayish-purple intraformational mudstone clasts.

Bedding in the medium- to fine-grained arkose commonly is outlined by films of reddish-brown clay or rarely by grains of specular hematite (after magnetite). Most of the arkose has a texture of interlocking grains produced by pressure solution from burial below at least 3000 m of overburden. In outcrop much of the sandstone is speckled with intergranular brown patches of limonite after iron-rich carbonate. Albite-oligoclase and subordinate K-feldspar make up 15-40 percent of the grains. Chert, metamorphic fragments, muscovite, biotite, and chlorite are minor consituents.

Lockatong Formation

The Lockatong Formation is composed largely of tough non-fissile dark gray mudstone, with subordinate units of grayish-red to reddish-brown mudstone in the upper part. The deposits in this distinctive formation are arranged in short "detrital" and "chemical" cycles averaging several meters in thickness. Detrital cycles are more common in the lower siltier part of the Lockatong Formation and in long gray sequences; unique chemical cycles are best developed in the upper part and in the thinner grayish-red sequences (see Olsen, this volume; Van Houten, 1987).

Short detrital cycles, 4-6 m thick, consist of a lower black shale succeeded by platy dark gray carbonate-rich mud-stone followed by tough, massive gray calcareous mudstone in the upper part. The massive unit has a very small-scale contorted and disrupted fabric produced by crumpled shrinkage cracks and burrow casts. Some thicker detrital cycles contain a lens of thin-bedded siltstone and fine-grained feldspathic sandstone. On average, these deposits have abundant Na-feldspar, illite and muscovite, some K-feldspar, chlorite and calcite, and a little quartz.

In short chemical cycles, 2-4 m thick, the lower beds are alternating black platy dolomitic mudstone and marlstone disrupted by shrinkage cracks. In the middle, more massive part, layers of tan-weathering dolomitic marlstone are extensively disrupted by shrinkage cracking. The upper part of a chemical cycle is tough gray analcime- and dolomite-rich mudstone brecciated on a microscopic scale. Some thinner cycles are grayish red to reddish brown. In these, thin dark red layers are disrupted by shinkage cracks and broken into intraformational breccia with patches of analcime and dolomite in the cracks. Thicker massive beds are speckled with tiny pseudomorphs of dolomite and analcime after gypsum or glauberite. Mudstone in the upper part of chemical cycles contains as much as 30-40 percent analcime, together with albite, dolomite, and calcite, and illite and minor chlorite.

The Lockatong Formation grades upward into the Passaic Formation through a succession of rather regularly alternating dark gray and reddish-brown units (Fig. 3) that recur at about 100 m (300-325 ft) intervals, and in which the reddish-brown ones are progressively thicker.

Passaic Formation

The Passaic Formation consists of reddish-brown mudstone and siltstone, with subordinate claystone and fine-grained feldspathic sandstone. This central basin facies has no well-developed coarse-grained channel fill like that present near the margins of the basin. Commonly, two kinds of mudstone occur in alternating sequences. One is crumbly bright reddish-brown homogeneous claystone with rare thin layers of siltstone. The other is tougher bioturbated mudstone locally scoured by broad channels filled with bedded fine-grained sandstone and mudstone. Many Passaic strata are marked with small and large tracks and trails. Rarely, molds of glauberite now filled with calcite are abundant in thin lenses.

Widespread units of dark gray mudstone and marlstone recur in the formation at 125-135 m (400-450 ft) intervals, matching those in the upper part of the Lockatong Formation.

Eight such gray units have been identified in the Passaic Formation (Fig. 3). The common feldspathic mudstone and micaceous siltstone contain abundant illite and minor chlorite, quartz (50-75 percent in siltstone, 10-30 percent in mudstone), and less than 15 percent feldspar (Na-type normally predominates). Lithic fragments are rare except near the northwestern margin where they are common. Hematite pigment coats grains and stains the clay fraction. Secondary crystalline hematite (after magnetite) is the opaque mineral grain.

Border Conglomerate

Along the northwestern border of the basin coarse fanglomerate projects southward, interfingering with the upper Passaic Formation. The clast-supported conglomerate is arranged in lenticular, crudely fining-upward units (Fig. 4) with caliche carbonate concentrated in the upper sandy to muddy parts. Clast size increases upward in the section as well as toward the source north of the border fault. Most of the clasts are Paleozoic quartzite. The fewer, smaller ones are dolomite. The dark reddish-brown finer-grained fraction is lithic-rich poorly sorted sandstone and sandy mudstone.

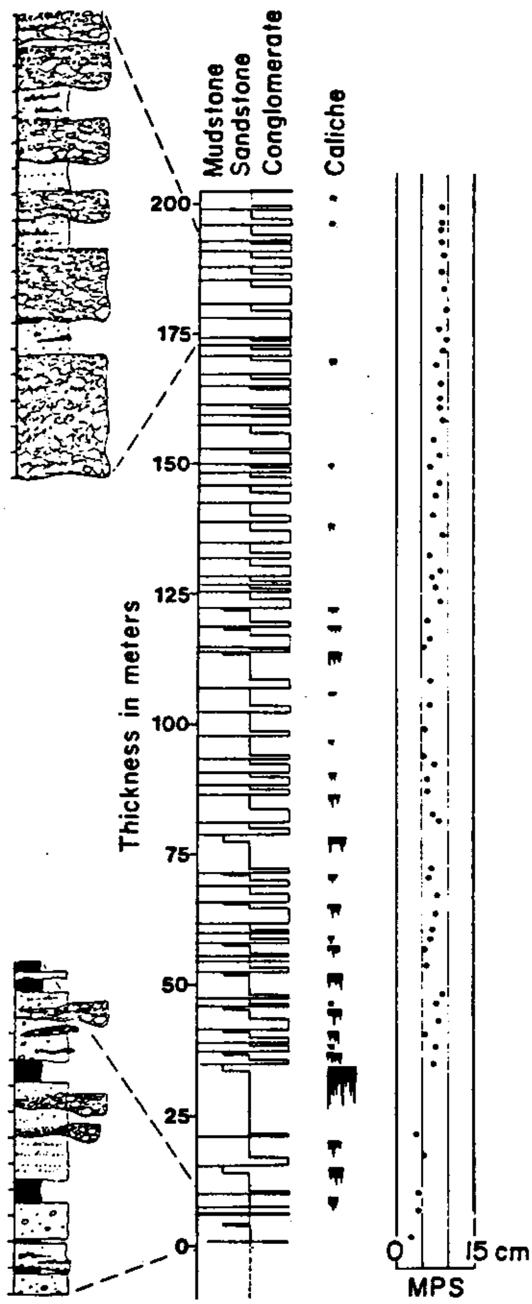

Fig.4. Border conglomerate in the Passaic Formation northwest of Milford, NJ. Columns show patterns of repeated conglomerates, distribution and degree of development of caliche paleosols, and mean maximum particle size (MPS) in beds of conglomerate. Typical sequences in the lower and upper parts of the section are sketched along the left side. From Arguden and Rodolfo, 1986.

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MESOZOIC TECTONICS OF THE NEWARK BASIN, AS VIEWED FROM THE DELAWARE RIVER

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Introduction

The origin and tectonic development of the Early Mesozoic basins in eastern North America have intrigued geologists for more than a century. The isolation of these deposits into numerous separate basins, the continental nature of the sediments, and the monoclinal dip of bedding throughout each basin distinguish these rocks from all the other Appalachian terranes. The basins now are recognized as being extensional features that developed in conjunction with the opening of the Atlantic Ocean, but the nature and sequence of events are yet to be thoroughly elucidated. Additionally, examination of outcrops and considerable subsurface work, such as drilling and seismic surveys, will be necessary to complete our understanding of the tectonic evolution of the Newark and other Early Mesozoic basins.

Historical Development of Tectonic Ideas

Rogers (1858) argued that the monoclinal aspect of the basin bedding reflected foreset deposition on the face of an outward building delta, but this view was not accepted widely. By the 1880s, it was realized that some of the well-known `trap' sheets actually were basaltic lava flows; also, numerous faults had been discovered and traced, both within the basins and at their margins. Davis (1886, 1898) used this new information to construct a two-phase history for the basins: an initial downwarping and sediment filling of the basin; and a subsequent faulting and tilting which brought the basins to their present structural configuration.

But others thought differently. The presence of coarse-grained fanglomerates (suggesting nearby highlands) along faulted down-dip margins convinced Russell (1892) that the two features were related, that the fanglomerates accumulated at the foot of fault scarps. From this, it is but a short step to propose that faulting at the margin occurred throughout basin development, with the down-faulted basin floor providing a continually deepening sink for sediment shed from the adjacent, rising highlands. Barrell (1915) illustrated this with his introduction of the half-graben concept in his summary of the geology of central Connecticut; Kay (1948) formalized this concept as a taphrogeosyncline containing 5 or more km of sediment.

Syndepositional margin faulting, the coupling of a sinking basin and an adjacent rising highland separated by an active fault, has dominated tectonic thinking of the Mesozoic basins ever since. Variations on this theme have appeared in several guises: Bain (1932) suggested a steep reverse fault for the down-dip margin; Russell (1892), Barrell (1915), and Sanders (1963), noting the apparent mirror symmetry between the Newark and Hartford basins, argued they were connected originally in a single "broad terrane" graben; Bain (1957), Manspeizer (1980), and Swanson (1986) proposed strike-slip faults for all of the down-dip margins. But regardless of the sense of movement on the faults, they all use faulting contemporaneous with sediment deposition as the central tenet.

The Listric Fault Model

During the past two decades, geophysical techniques have been brought to bear on the nature of the basin floor and the northwest margin. A gravity profile across the Gettysburg Basin indicated that the deepest part was not near the northwest margin as required in half-graben models, but under the center (Shaub, 1975). The absence of a steep gravity gradient across the northwest margin and the Chalfont and Furlong faults suggested to both Dunleavey (1975) and Sumner (1977) that the northwest margin consists of a series of step-faults. The parallelism of gravity contours with the basin margins and the location of residual gravity lows along the axial portion of the basins suggests the basin profile is v-shaped along its length, with the deepest part in the center (Daniels, 1985).

Deep drilling for gas near the northwest margin has provided some information on the minimum depths of the basin. The well in Bucks County reached a total depth of 3,200 m without crossing into pre-Mesozoic rocks. This depth is the greatest penetration of the Mesozoic rocks in either the Newark or Gettysburg Basins. This depth requires the northwest basin floor to dip at least 22 degrees to the southeast. A subsequent hole in Montgomery County to the west reached 2,046 m depth without leaving Mesozoic rocks; here the basin floor dips a minimum of 17 degrees.

Other recent work in the vicinity of the Delaware River has shed some additional illumination. The conventional half-graben model assumes that the faulted margin is a steep normal fault.

However, Ratcliffe and Burton (1985a) demonstrated with two drill holes and a seismic profile that the northwest margin of the Newark basin, just west of the Delaware River, is a shallow normal fault dipping only 27 degrees to the southeast. This fault, and a companion fault a short distance below it, was a Paleozoic thrust (of the Musconetcong meganappe system) that was reactivated during the Early Mesozoic as an extensional listric fault with movement directed to the southeast, perpendicular to the northwest margin and the basin trend (Ratcliffe and others, 1986).

Listric faults are shallow-dipping normal faults--ideally, they are relatively steep near the surface, and gradually decrease in dip with depth (Fig. 1A). The main difference is in the sense of movement. The thrust fault results in crustal shortening; movement on a listric fault (where the hanging wall moves down) produces crustal extension. In a sense, one can view listric faults as the extensional counterpart of thrust faults.

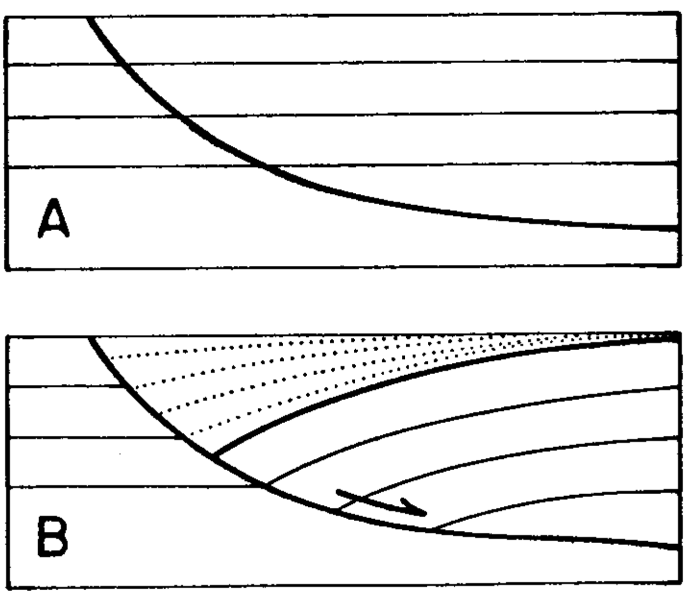

Figure 1. Ideal listric fault. A. Configuration of fault at initiation. Steeper near the surface, the attitude gradually decreases in dip with depth, approaching asymptotically the horizontal. Light parallel lines are reference lines. B. During the displacement (extension), the rocks in the trailing 'toe' rotate because of the curvature and decreasing dip of the listric fault. Dotted lines are sediment layers filling the resulting basin.

One consequence of extensional movement on listric faults is the rotation of the rocks in the 'toe', near the steeper part of the fault (Fig. 1B). If the Mesozoic basins do indeed rest on large listric faults, this would provide a mechanism for imparting the monoclinal dip to the entire basin.

Indeed, other seismic evidence supports this model: the shallow dip of the northwest margin apparently continues southeastward beneath the basin, with the consequence that the deepest part is under its center (Fig. 2). This profile apparently applies to other Mesozoic basins as well (e.g., Hutchinson and others, 1986). On the other hand, some of the correlative basins in northwest Africa apparently are different (Manspeizer, 1980). Bedding in the Newark Basin dips to the northwest for the most part, subparallel to the southeastern basin floor. The simplest tectonic scenario, then, is that the hanging wall ('Buckingham

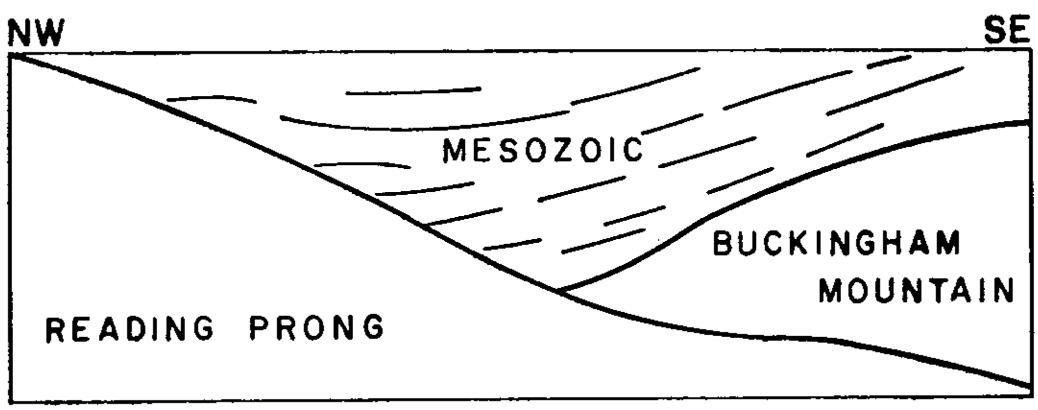

Figure 2. Generalized sketch from a seismic line across the northern half of the Newark Basin in the vicinity of the Delaware River. The heavier lines represent what is believed to be the pre-Mesozoic floor and the fault on which the Buckingham Mountain 'toe' moved. Light lines represent probable bedding.

Mountain' block of Fig. 2) slid southeastward during basin filling utilizing the earlier thrust system as an extensional listric fault (Ratcliffe and Burton, 1985b). The resulting 'hole' at the surface then accumulated sediment from the highlands to the southeast as well as the northwest.

The simplicity of this model is beguiling, because it provides for two important aspects: It creates and continually deepens a basin for accumulating sediment; and it results in a monoclinal dip to the bedding through rotation on the listric fault system. It also gives a mechanism for producing the numerous fanglomerate lobes along the northwest margin, particularly those with large angular carbonate clasts that require a nearby source and a scarp. Syndepositional faulting is still the cornerstone, as it has been for decades. But now the fault at the margin is a listric fault.

Some Caveats with Syndepositional Faulting

Syndepositional faulting has consequences which actually contradict or are not supported by field evidence. The critical question is not the presence or absence of a fault at the northwest margin of the basin; rather, it is the time of faulting. Were the margin faults active throughout basin filling, or did the faulting occur only in the Early Jurassic, at the terminal stage of basin development? Several aspects of the Newark Basin bear on this question.

Monoclinal Dip

Consider first the monoclinal dip of bedding. With extensional movement on the listric fault, the rocks in the `toe' of the hanging wall are rotated because of the curvature of the listric fault (Fig. 1B). If the fault were planar (no decrease of dip with depth), no rotation would occur. It follows that any sediments deposited on top of the toe also become progressively rotated with increasing displacement on the listric fault. And the dip of the younger beds, the ones nearer the fault margin, decreases (as it seems to do in Fig. 2). This pattern also appears to some degree along the Delaware River, where bedding dips are 15 to 20 degrees near the southeast margin, and on the order of 10 degrees near the northwest margin (although locally the dip is higher).

Comparison of Figure 1B with Figure 2 illustrates the appropriateness of the listric fault model. However, some differences exist. In Figure 2, the listric fault is nearly planar (except for a broad sag under the northwest edge of the Buckingham Mountain `toe'). This sketch is based on only one seismic profile and the sag may be just a local anomaly, but the 27 degree dip near the surface does give one pause. This rather low dip and the relative planarity severely reduce the potential for rotating the Mesozoic beds. Another difference between model and `reality' is in the stratigraphic thicknesses--the model calls for appreciable thinning of units away from the fault, but the rocks along the Delaware River show no indication of such a trend to the southeast.

A progressive lessening in bedding dip from the southeast toward the northwest fault margin would support syndepositional faulting. Whereas this may hold to some extent along the Delaware River (but intrabasinal faults complicate the issue), this relation does not exist in other parts of the basin (Faill, 1973). For example, in the 'narrow neck' (which separates the Newark from the Gettysburg Basin) west of Reading, Pennsylvania, bedding dips 40 degrees to the north across the entire basin. Clearly, the strata here must have been tilted as a single package after deposition of all the beds. That is, the movement on the listric fault did not occur incrementally during sedimentation; rather, it occurred only after basin filling, in the Early Jurassic. But then, if the listric fault movement were that late, how did the basin form initially and grow?

The Fanglomerates at the Basin Margins

A second aspect of the Newark Basin (as well as other basins) plays a role in ascertaining the tectonic history. The numerous fanglomerates at the basin margins contain rounded pebbles and cobbles of quartzite and other rock fragments, and/or more angular fragments of limestone and dolomite, in a muddy to silty matrix. These poorly sorted mudflows along the northwest margin often have been cited as evidence of syndepositional faulting. Although active faults can produce such deposits, such fanglomerates are not, in and of themselves, conclusive evidence of active faulting.

The presence of fanglomerates along the unfaulted southeast margin demonstrates this point. A particularly convincing example was well documented west of the Susquehanna River, on the southeast margin of the Gettysburg Basin (Cloos and Pettijohn, 1973). As part of an exploration program for the nearby Thomasville Stone and Lime Company, a corehole was placed just inside the southeast margin of the basin as a test of the carbonates that were thought to closely underlie a thin (30 to 50 m) sequence of basal Triassic beds. To the surprise of everyone, the 'thin' Triassic sequence turned out to be 240 m thick, and included carbonate fanglomerate lithology in the lowest 90 m. Other nearby drilling indicated that a steep declivity must be within 100 m of the core hole. While it is tempting to postulate a fault origin for both the steep slope and the fanglomerate, in fact there is no fault. The core hole intercepted the Salterella zone in the underlying Cambrian rocks at the depth expected by projection from known occurrences. Incidentally, the authors suggested the possibility of a growth fault origin, even though they admitted they could find no evidence of a fault of any kind.

With faulting out of the question, an erosional origin for the fanglomerate becomes more plausible. In the semi-arid climate extant at the time of basin formation, carbonates weather mechanically rather than chemically by dissolution. Therefore, the Paleozoic limestones and dolomites near the basin edge would have developed ridges and cuestas similar to those in the Rocky Mountain foothills of the western United States. The mechanical weathering produced angular fragments which were eroded and incorporated in the numerous mudflows and debris flows during flash floods created by occasional storms. These local deposits overlie and were later covered by the more typical floodplain and mudflat deposits, resulting in the interfingering of the two lithologies at the basin margins that we see today.

The Timing of the Faulting

A third aspect of the basin also puts constraints on the tectonic history. Movement on all the major intrabasinal faults (with one possible exception) postdates the deposition of all Triassic sediments, and possibly some of the Jurassic ones as well. Early Mesozoic faults probably exist in the pre-Mesozoic terranes northwest and southeast of the Newark Basin, but they are understandably difficult to identify. A few faults do transect a basin margin--those that offset the northwest margin are clearly younger than the youngest preserved Mesozoic beds.

The Furlong, Flemington, and Hopewell faults offset Early Jurassic diabase plutons and most of the Passaic Formation. Movement on the Chalfont fault also was late, because it was integrated with the folding along the northwest margin west of the Delaware River. Stratigraphic offset of the Stockton Formation near the east end of the Chalfont fault is about 1 km, but the offset of the Lockatong and overlying Passaic Formations diminishes westwardly (Fig. 3A). Concomitantly, the amplitude of an enclosing anticline increases. In other words, the Chalfont fault and the associated anticline comprise nothing more than a faulted fold. One can see this relation by looking down the

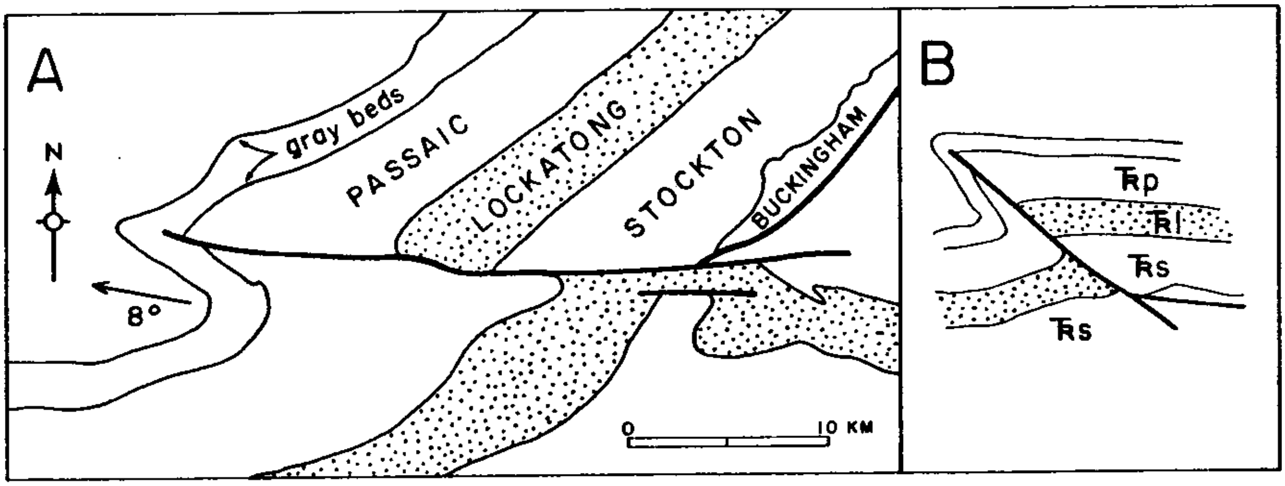

Figure 3. The Chalfont fault. A. Map of the fault and surrounding rock units. The Furlong fault abuts the east-west Chalfont fault, and forms the southeast boundary of the Buckingham window of pre-Mesozoic rocks. B. A down-plunge view of the fold that envelops the Chalfont fault, illustrating the fault-fold nature of the structure. Trp - Passaic Formation; Trl - Lockatong Formation; Trs - Stockton Formation.

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plunge (8 degrees to the west-northwest) of the anticline (as illustrated in Fig. 3B). Displacement on the splay fault (Chalfont) in the lower part converts to folding in the upper part of the structure. This is a compressional (subparallel to bedding), not an extensional structure, which indicates that the maximum principal stress was directed in a subhorizontal north-northeast direction. It follows, then, that the Chalfont probably is a steeply north-northeast dipping thrust fault. Whether this fault and associated fold developed as part of the listric fault movements or as an independent deformation is not known. If forced to choose, I would say it was an independent event resulting from a regional stress reorientation, with the extensional stress, σ_3, redirected from east-southeastward and subhorizontal, to the vertical direction.

The time of fault movement on the intrabasinal faults was Early Jurassic or later, because, in the Jacksonwald syncline west of the Chalfont fold-fault structure, Early Jurassic sediments lying above the basalt flows are folded. Presumably, all the other folds along this margin from Reading to east of the Delaware River were folded at the same time by the same stress system because all the folds are geometrically similar (very open folds), plunge in the same direction (gently to the northwest), and share common limbs with each other.

Looking at syndepositional faulting from another point of view, if the Newark Basin were formed and deepened by continual extension on a listric fault, one might expect that, given probable irregularities on the listric fault surface, evidence of differential vertical movements would appear at the surface. These effects could be in the form of antithetic or synthetic growth faults, or as domes or sags. The faults need not break through the depositional surface, but whatever their form, the relief they produced would seriously affect the sedimentary patterns within the basin, particularly in the mudflat and playa portions which are very sensitive to elevation changes. Even if an event were of short duration (say, 2 or 3 m.y.), its influence would show in the distribution of facies. But the great lateral persistence of the gray shale units, the interfingering of lithologies, and the gradual changes in facies all suggest that intrabasinal movements did not disturb the relative placidity of the depositional surface.

All the faults, then, mapped ones and those seen in outcrop, appear to be post-depositional. No changes in lithology, facies, or bed thickness is associated with any of them, with one possible exception. This fault is at the west end of the Newark Basin, along the east limb of the Morgantown pluton, south of Reading. West of this limb, the conglomeratic sandstone sequences of the Hammer Creek Formation are common and thick; in contrast, east of the limb (but only in the northern part of the basin) they are much fewer and smaller. This apparent large change across such a short distance led MacLachlan (1983) to propose that the east limb of the pluton intruded a syndepositional fault. Apparently, recurrent down-on-the-west movement was sufficient to maintain a topographic barrier that prevented the coarser fraction of the Hammer Creek sediments from spreading to the east into the Newark Basin. In that the conglomerates ae not interrupted in the southern half of the basin indicates the faulting did not begin until late in the basin history--the more extensive conglomerates interfinger with the uppermost gray shales of the Passaic Formation (the ones just below the Quakertown pluton).

The Igneous Rocks

The igneous rocks of the Newark Basin pose less restriction on Mesozoic tectonism than the three other aspects discussed above, but they are a factor. These rocks are of tholeiitic basaltic composition, and occur as dikes, flows, or sheet-like plutons. Geochemically, three groups can be distinguished (Smith and others, 1975): the Quarryville is the oldest type and occurs only as dikes; the York Haven type occurs in all three forms; the Rossville type is the youngest and is present only in dikes and sheets.

The dikes of all three types have intruded the Mesozoic rocks and the underlying pre-Mesozoic basement. For the most part, the sheet-like plutons are roughly saucer-shaped; they usually parallel bedding on their southeast side but, where close to the northwest margin, their northwest portions dip to the southeast, cutting across bedding. The extrusive flows, of course, are confined to the basin, and appear only in the youngest beds. The plutons are similarly confined to the basin, except at two localities where pre-Mesozoic rocks have been rafted above the diabase (at Cornwall, on the north edge of the York Haven pluton, and at the southeastern corner of the Morgantown pluton).

Within Pennsylvania, the York Haven type comprises approximately 80 percent volumetrically; the Quarryville is only 5 percent. Most of the igneous bodies in the New Jersey portion of the Newark Basin are of the York Haven type, including the Palisades sill and the First Watchung flow. The upper two Watchung flows (Puffer and Lechler, 1980) and some of the plutons in western New Jersey (Husch and others, 1984) appear to be separate differentiates of the same York Haven parent magma that were emplaced shortly thereafter.

Compositional variability among closely related magmas may be caused by coeval tectonism. The greater variability of the

Newark Basin tholeiites relative to those west of Reading suggests a 'tectonic tranquility' in the Gettysburg Basin and narrow neck during the York Haven event, and a more disturbed history along the Delaware River and eastward into New Jersey. Temporary ponding of the rising magma in the crust because of increased tectonism and/or stresses would provide more time for additional country rock assimilation (R.C. Smith, pers. commun.).

The presence of the palynological base of the Jurassic System closely below the York Haven flow in the Jacksonwald syncline (Cornet and Traverse, 1975) indicates that the York Haven bodies were emplaced at the time of the Triassic-Jurassic boundary, 185 to 194 ma (Seidemann and others, 1984). None of the magma intruded any of the faults (with the possible exception at the east limb of the Morgantown pluton), not even the faults at the northwest margin. Either the faults were too tightly closed by compressive stresses to be intruded, or else there were no faults to intrude. Many of the plutons, as well as the Watchung and Jacksonwald flows, have been cut by faults, indicating that the faulting postdated all the igneous activity.

The general northeastward trend of the vertical (mostly York Haven) dikes throughout the western Newark Basin indicates that the extension stress, σ_3, was directed subhorizontally to the southeast at the beginning of the Jurassic period, consistent with extension on the proposed listric faults.

Summary of Listric Fault Model, and Caveats

Drilling and seismic data indicate that the northwest margin, towards which the Newark Basin beds dip, itself dips to the southeast at a fairly low angle, and is a reactivated Paleozoic thrust fault. During the Early Mesozoic, southeastward movement of the hanging wall along this fault not only produced a basin within which sediment could accumulate, but also imparted a pervasive northwestward dip to all the beds. Some possible problems with this model are:

1). Monoclinal dip. If the listric fault were active throughout basin filling, then the bedding dip should decrease toward the fault.

2). Fanglomerates. They are not necessarily indicators of syndepositional faulting.

3). Time of faulting. Virtually all the intrabasinal faults are postdepositional.

4). Igneous rocks. Most of these were emplaced at the very beginning of the Jurassic, and they intruded virtually none of the faults.

Tectonic Summary

The relative simplicity of the Mesozoic basins, and the absence of intense folding and regional metamorphism led most workers to relegate this tectonism to a minor event. Until the late 1960s, it was called merely `the Palisades disturbance'; it was even referred to as `the dying gasp of the Appalachian Revolution'. But it now is clear that it was nothing of the sort. The evolution of the plate tectonics paradigm has completely recast our view of these basins. Rather than a final tectonic punctuation, Early Mesozoic tectonism now is recognized as the beginning of a major crustal event, the opening of the Atlantic Ocean, an ongoing process that continues to this day.

Following the Late Paleozoic crustal collisions that assembled Pangea (expressed as the Alleghanian orogeny in the eastern United States), crustal extension along the orogen commenced in the Late Triassic. This extension probably began as a ductile stretching and manifested itself at the surface as a series of isolated depressions along the orogen. The depressions developed across a wide swath that included the Appalachian Piedmont, the (present day) Atlantic continental shelf, and the then adjacent northwestern Africa (Manspeizer and others, 1978; Manspeizer, 1981). Erosion of the surrounding higher lands led to accumulation of sediment in these basins, as reflected by the relatively close correspondence of the sediment mineralogy with the adjacent terranes (Klein, 1962; Glaeser, 1966; Abdel-Monem and Kulp, 1968). Considering the overlap relations around so much of the basins, the basins probably were growing in size with time, and it is possible that nearby basins may have become connected late in the basin filling (e.g., Barrell, 1915). However, this is quite distinct from the broad-terrane hypothesis, in which the present basins are erosional remnants of a single, large graben.

The crustal extension continued throughout the Late Triassic and into the Early Jurassic, at which point considerable volumes of tholeiitic magma rose and were emplaced as largely concordant plutons, lava flows, and dikes (Smith and others, 1975; Philpotts and Reichenbach, 1985). These igneous intrusions and extrusions constituted a major tectonic event towards the end of basin filling. The greater homogeneity of the igneous rocks in the Gettysburg Basin, and the younger age of the sediments in the Newark Basin suggest that post-Early Jurassic erosion may have been more extensive in the Gettysburg Basin, providing us with a deeper view of the Mesozoic basins there than in the Newark Basin (R.C. Smith, personal communication).

How much additional sediment accumulated after the Third Watchung flow can only be surmised, but at some time following the deposition of the youngest preserved beds of the Boonton Formation, deformation of the basin began, producing folds along the northwest margin and the large and small intrabasinal faults. Faulting along the northwest margin also occurred at this time. As the crustal extension continued, its behavior became more brittle. Finally, the crust parted near the midline of the extension zone, far to the east of the Newark Basin, and the Atlantic Ocean was born.

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STRUCTURAL EVOLUTION OF THE NEWARK BASIN

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INTRODUCTION

The Newark basin of New York, New Jersey, and Pennsylvania is an eroded half-graben bounded on its northwestern and northern margins by the SE- to S-dipping border fault system. Synrift strata within the Newark basin generally dip toward the border fault, although they are warped into gently plunging folds in the hanging walls adjacent to the border fault and the two major intrabasinal faults, the Flemington and Hopewell faults (Figure 1). Based on outcrop studies and propreitary seismic reflection profiles of the Newark basin, corroborated by published profiles of coeval synrift basins on the continental shelf (Hutchinson *et al.*, 1986), younger strata generally dip at a shallower angle than older strata, although again complicated by the effects of folding adjacent to the border fault. These observations indicate that sedimentation and hanging wall rotation as a result of slip on the border fault system occurred simultaneously.

The border fault system strikes ENE in the northern Newark basin, NE in the central and southwestern portions of the basin, and ESE in the area west of Boyertown, Pennsylvania, and into the narrow neck between the Newark and Gettysburg basins. The faults appear to progressively step back to the northwest, going from northeast to southwest, such that the border fault system has a relay geometry (Figure 1). The dip of the border fault system decreases from 60^{o} SE in Suffern, New York, to 30^{o} SE and less in Pennsylvania (Ratcliffe and Burton, 1985). However, the border fault again appears to steepen markedly in the area west of Boyertown. The variations in the attitude of the Mesozoic border faults closely mimic that of the Paleozoic thrust faults formed during the Taconian, Acadian, and Alleghenian orogenies. In fact, all along the border fault system, Mesozoic brittle structures, including fault breccia and gouge, overprint but parallel the phyllonitic and mylonitic rocks of the Paleozoic faults. Hence, the border faults of the Newark basin, most of which were active during sedimentation (Arguden and Rodolfo, 1986), represent reactivated Paleozoic structures (Ratcliffe, 1980; Ratcliffe *et al.*, 1986).

According to the Ratcliffe and Burton (1985) model, the direction of fault slip depends on the orientation of (reactivated) faults with respect to the early Mesozoic extension direction. Faults oriented normal to the extension direction should experience pure dip-slip. Faults whose strike is oriented clockwise from the extension direction should experience a component of left-lateral slip, whereas those oriented counter-clockwise from the extension direction should experience a component of right-lateral slip.

We estimate the extension direction to be ESE, normal to the average strike of Early Jurassic diabase dikes (see Figure 1B), which typically form perpendicular to the σ_3 direction. This extension direction produced largely dip-slip on the majority of the border faults and the two intrabasinal faults, all of which strike NE, and left-lateral strike-slip on the E-W trending border fault west of Boyertown. A large component of strike-slip is supported by the observed steepening of the dip of the fault west of Boyertown and by studies of the border fault and related structures in the Jacksonwald syncline (Lucas *et al.*, 1988).

GEOMETRY AND ORIGIN OF FOLDING

In the hanging walls of the border fault system and the Flemington and Hopewell faults, folds whose axes are normal to the associated faults are a common and obvious feature of the Newark basin. Most of the axes are oriented within 15° of perpendicular to the faults (Figure 1B). It appears unlikely that these folds formed as a

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*Figure 1: **(A)** Geologic map of the Newark basin. Regular stipple represents diabase intrusions, irregular stipple represents lava flows. Dotted lines are form lines of bedding, and thin black lines are gray and black units in the Passaic Formation. Abbreviations are: **S**, Stockton Fm.; **L**, Lockatong Fm; **P**, Passaic Fm., **O**, Orange Mountain Basalt; **F**, Feltville Fm.; **Pr**, Preakness Basalt; **T**, Towaco Fm., **H**, Hook Mountain Basalt; **B**, Boonton Fm.; **J**, Jacksonwald Basalt; and **Pd**, Palisades diabase. **(B)** Structural map of the Newark basin and surrounding area, illustrating the close correspondence in attitude between the border faults and Paleozoic thrust faults. Thin double lines represent dikes. Abbreviations are: **r**, Ramapo fault; **h**, Hopewell fault; **f**, Flemington fault; **c**, Chalfont fault; **z**, zone of intense normal faulting (shown schematically); **cl**, Cameron's line; **w**, Watchung syncline; **j**, Jacksonwald syncline; **bd**, Birdsboro dike; **da**, anomalous NW-striking dike; and **db**, dike apparently offset by Chalfont fault. Paleozoic structures after Lyttle and Epstein (1987) and Ratcliffe (1980).*


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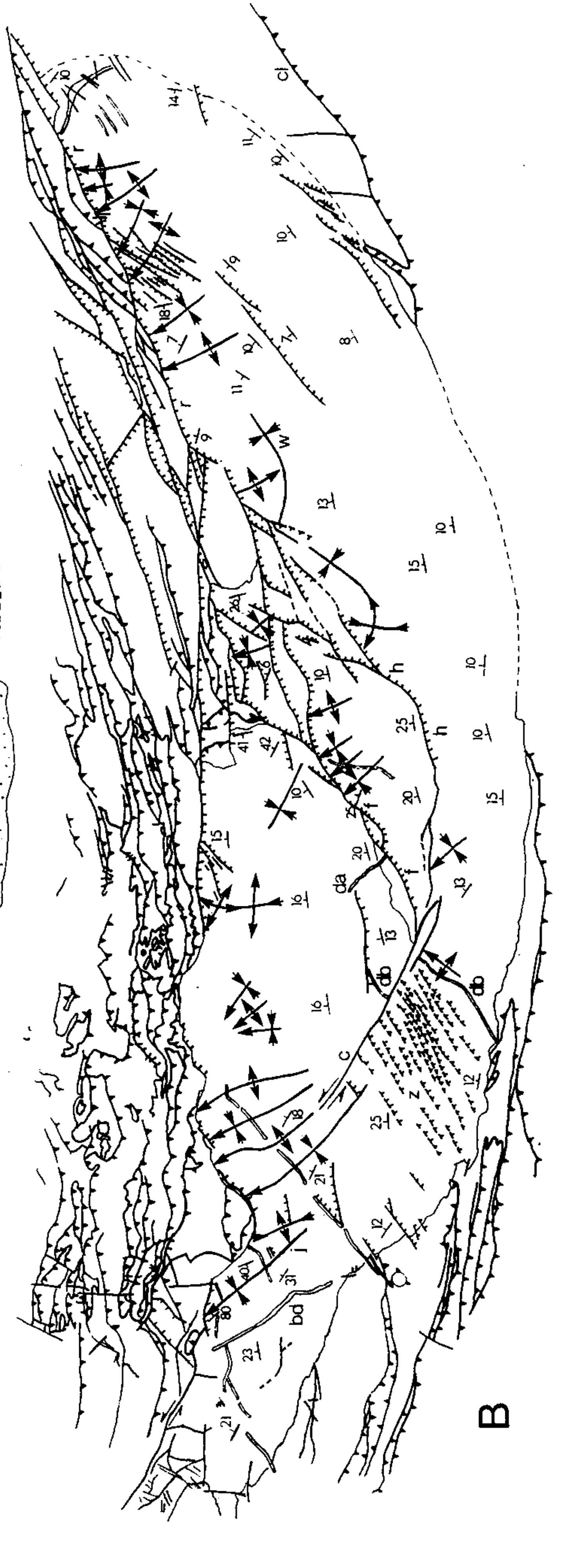
B

result of strike-slip along these faults for the following reasons: (a) folds formed by strike-slip have their axes oriented at 45^{o} or less with the fault (Christie-Blick and Biddle, 1985); (b) the Newark basin folds are not *en echelon;* as are folds formed by strike-slip; and (c) the Newark basin faults experienced predominantly dip-slip. The only exception to this appears to be the Jacksonwald syncline, the axis of which trends at a much lower angle to the border fault, and probably was influenced by a strike-slip faulting, consistent with the attitude of the border fault.

In addition to their transverse nature, the folds die out away from their associated faults in the hanging wall, readily observable in the three lava flows of the northern Newark basin (Figure 1A). The associated faults themselves are not folded, and the folds are not found in the footwall. It therefore appears likely that these folds are intimately associated with the faults and faulting responsible for basin subsidence (Schlische and Olsen, 1987). In the following section, we document the evidence that these folds were growing during basin subsidence and sedimentation.

Our arguments hinge on the contemporaneity of the igneous rocks within the Newark basin. Existing radiometric dates for diabase intrusions continually point to an age of 201 Ma (Sutter, 1988). Dates on the extrusive rocks show a great deal of scatter, but also cluster around 201 Ma (Olsen *et al.*, 1987). Physical relationships suggest that many of the plutons have fed the extrusives: the Palisades diabase has been shown to have fed the Ladentown flows in New York (Ratcliffe, 1988). Furthermore, the pattern and hierarchy of Milankovitch-period lacustrine cycles in the sediments between the lava flows constrain the total duration of the extrusive igneous activity to less than 600,000 years (Olsen and Fedosh, 1988). Hence, it appears likely that all of the igneous rocks in the Newark basin date to 201 Ma, and for the purposes of this discussion, are considered coeval.

In the Sassamansville area of Pennsylvania, diabase plutons are concordant and sill-like in the synclines but discordant in the anticlines, as revealed by distinctive gray to black lacustrine cycles which strike directly into the contact of the diabase (Figure 2). If folding had completely postdated intrusion, it seems likely that the concordance and discordance of the diabase would have been random with respect to the folds. Therefore, some folding probably had occurred prior to or during intrusion of the diabase.

The Late Triassic/Early Jurassic-aged Passaic Formation, a 201 Ma diabase sill, the approximately 201 Ma Jacksonwald Basalt flow, and the Early Jurassic Feltville Formation are folded by the Jacksonwald syncline (Figure 2). Two other diabase bodies are present in the syncline but are restricted to the fold axis. Their geometry suggests that they are phacoliths, having been intruded as accommodation structures in the space created by the buckling of strata. Again, these diabase bodies were intruded during or after the folding of the enclosing Passaic Formation. While the phacoliths were intruded, the coeval Jacksonwald Basalt was extruded onto a nearly flat surface, as there is no evidence of ponding in the synclinal hinge. Folding continued after extrusion, because the Jacksonwald basalt and the overlying Feltville Formation are folded. Paleomagnetic work suggests that much of the folding of the 201 Ma sill occurred after intrusion (Stuck *et al.*, 1988).

Ratcliffe (1980) suggests that the Ladentown lava flows may have been ponded in a synclinal trough developed along the Ramapo fault, but the other lava flows of the northern Newark basin show no such evidence; hence, the majority of the folding occurred after extrusion. The folds developed along the Flemington fault developed late in the history of the basin (see discussion below).

Stratigraphic evidence also indicates that sedimentation and folding within the Passaic Formation were coeval. In Douglasville, Pennsylvania, large bedding-plane outcrops appear to be warped. Certain deep-water lacustrine units appear to have been **only** deposited within the **troughs** of this warped surface. If the warping is tectonic, then it may be evidence of folding during sedimentation. Further work will concentrate on the thickness variation of fixed period Milankovitch lacustrine cycles across the folds. If the folds were forming during sedimentation, we

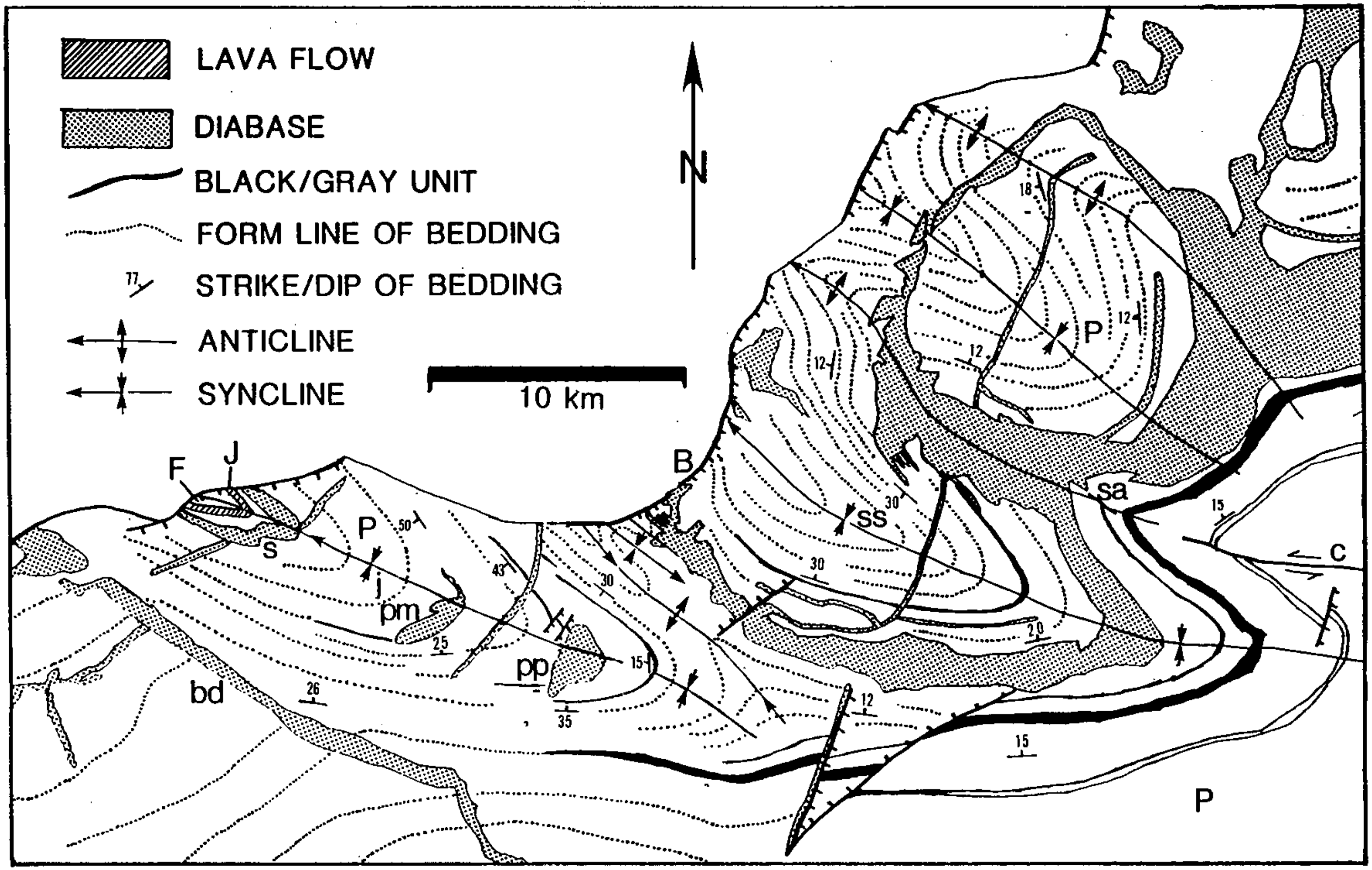

Figure 2: Geologic map of the southwestern corner of the Newark basin in Pennsylvania. Abbreviations are : ***B****, Boyertown, Pa.;* ***P****, Passaic Formation;* ***J****, Jacksonwald Basalt;* ***F****, Feltville Formation;* ***bd****, Birdsboro dike;* ***j****, Jacksonwald syncline;* ***s****, Jacksonwald sill;* ***pm****, Monocacy Station phacolith;* ***pp****, Pottstown phacolith;* ***ss****, Sassamansville syncline;* ***sa****, Sassamansville anticline; and* ***c****, Chalfont fault. Some data from Longwill and Wood (1965, Plate I).*

would expect a a greater cycle thickness in the hinges of the synclines and a lesser thickness on the hinges of the anticlines. Unfortunately, folding itself may induce structural thickening of beds in the hinge and thinning on the limbs (Ramsay and Huber, 1987). We hope to get around this problem by examining previously bedding-plane-perpendicular primary structures for evidence of systematic deformation during the buckling process.

Further evidence of the timing of the folding comes from minor structures from the Jacksonwald syncline. Mudcracks and reptile footprints within thinly bedded mudstones have been stretched parallel to the axis of the Jacksonwald syncline, or shortened perpendicular to the axis, or both. A slight crenulation is present, possibly indicative of microfolding. However, there is no indication of cleavage, nor does the rock break along any preferred orientation. Although thin-section analysis had not been completed at the time of this writing, we suspect that most of the detrital grains will show little, if any, evidence of penetrative deformation and, therefore, tentatively ascribe the observed strain to deformation in incompletely lithified or partially dewatered sediments, again suggesting folding during or immediately after sedimentation. An axial planar cleavage is locally found in the mudstones of the Jacksonwald syncline. This pressure solution cleavage does not fan about the fold axis, indicating that it formed late in the history of the folding (Lucas *et al.*, 1988).

The relationships between compressional and extensional structures is crucial for any kinematic interpretation. The fold axes generally are parallel with the early Mesozoic extension direction and, therefore, are perpendicular to the majority of the NE-striking Early Jurassic-aged diabase dikes. A regionally persistent set of joints parallel these dikes and presumably formed normal to the regional extension direction (Figure 3). Although ubiquitously present in a traverse across the basin, this NE-striking joint set was pervasively developed in hornfels surrounding the diabase intrusions, especially the phacoliths of the Jacksonwald syncline. We attribute the formation of these joints to hydrofracting associated with elevated fluid pressures at the time of intrusion. The density of the jointing diminishes

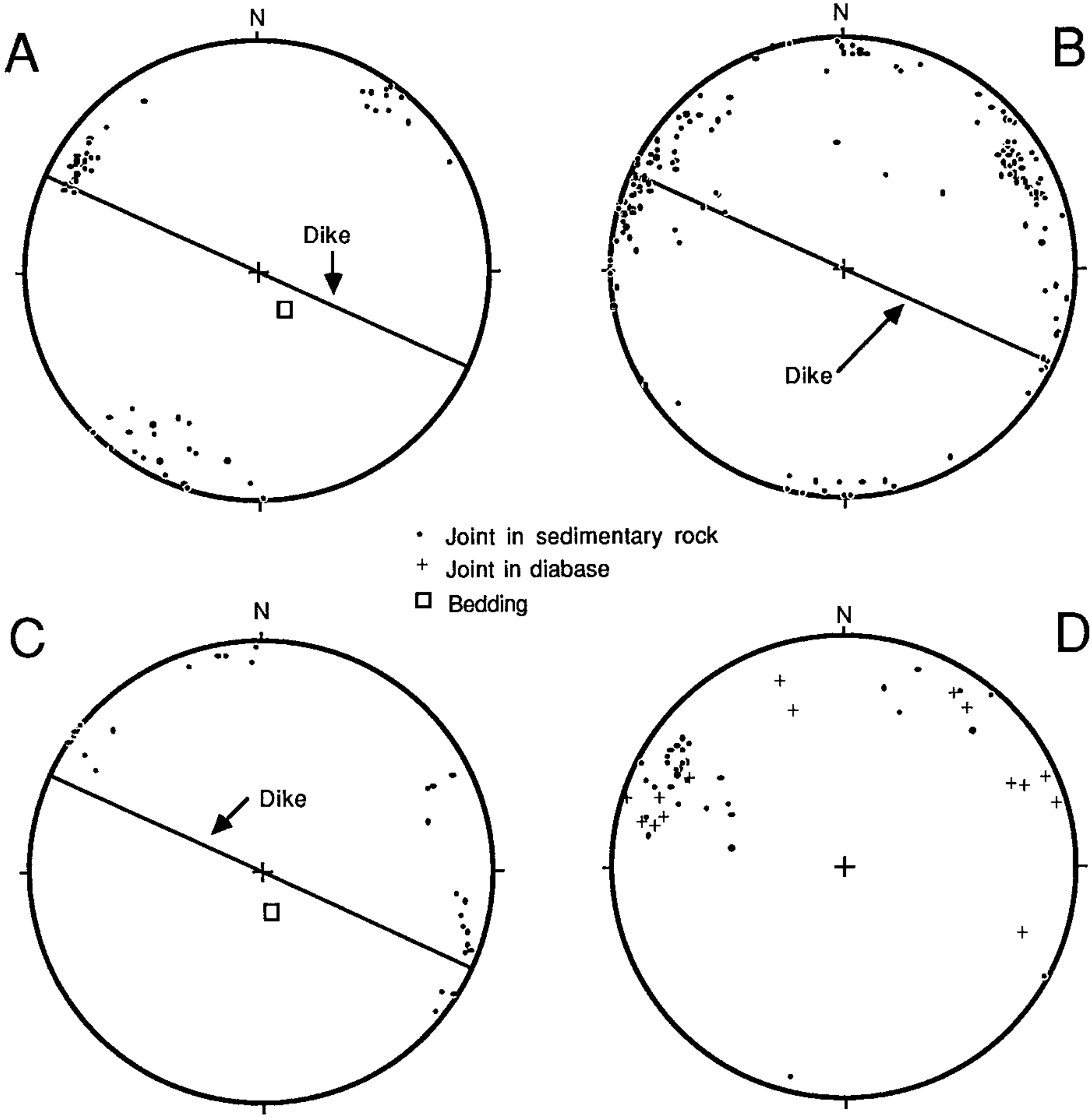

Figure 3: Equal area stereographic projections of poles to joints and bedding from (A) abandoned quarry in Stockton Formation, Rte. 29, near Stockton, N.J.; (B) Lockatong Formation along Rte. 29 near Byram,N.J.; (C) Lockatong Formation exposed in creek NE of Rte. 29 near Tumble Falls, N.J.; (D) contact between Passaic Formation and diabase in Pottstown Traprock Quarry, Jacksonwald syncline, Pottstown,Pa. Note that the NW-striking set of joints, which is subparallel to the anomalous dike, is best developed in A and B structurally below the inferred neutral surface and appears to die out upsection (see C). A majority of the joints in the Jacksonwald syncline strike NE, perpendicular to the regional extension direction.

markedly within the diabase itself, suggesting that the joints could not form in the still molten inner core of diabase. Since there is no fault between the diabase and hornfels, strain compatibility requires that the jointing occurred during intrusion, both as a consequence of regional extension and folding-induced hinge-parallel extension. The regional NE-striking joint set exclusive of those in the hornfels may have formed at the same time as those in the hornfels or any time thereafter as the result of the brittle release of the accumulated strains which resulted from regional extension.

At either end of the basin, the fold axes **parallel** the Birdsboro dike (**bd** in Figures 1B and 2), which separates the Newark basin from the narrow neck in Pennsylvania, and the dike-like extension of the Palisades intrusion in New York. The inferred directions of maximum shortening and maximum extension were parallel immediately adjacent to one another at the same time, an apparent contradiction.

A northwest-striking dike (**da** in Figure 1B) is located approximately 30 km southeast of the border fault in the fault block bounded by the border and Flemington faults. In this region, no folds are present, having disappeared about 10 to 15 km from the border fault. A set of NW-striking joints also is well developed in this area, in addition to the NE-striking set (Figure 3). Hence, heading southeast from the border fault, we pass from a region of fault-parallel shortening with fault-normal extension into a region of fault-parallel extension with fault-normal extension.

In order to explain the origin of the folds and the other coeval structures, we invoke a model inspired by fault-displacement geometries (Shelton, 1984; Barnett *et al.*, 1987). The net slip on a single fault has been shown to be maximized at its center and to die out in all directions. Because the Newark basin is widest and deepest at its center and dies out toward either end, we have applied this displacement geometry to the system of border faults. Neglecting the effects of the later intrabasinal faulting, such a displacement field results in a basin which can best be described as a giant synform plunging toward the border fault system (see Figure 4) . According to the folding model of

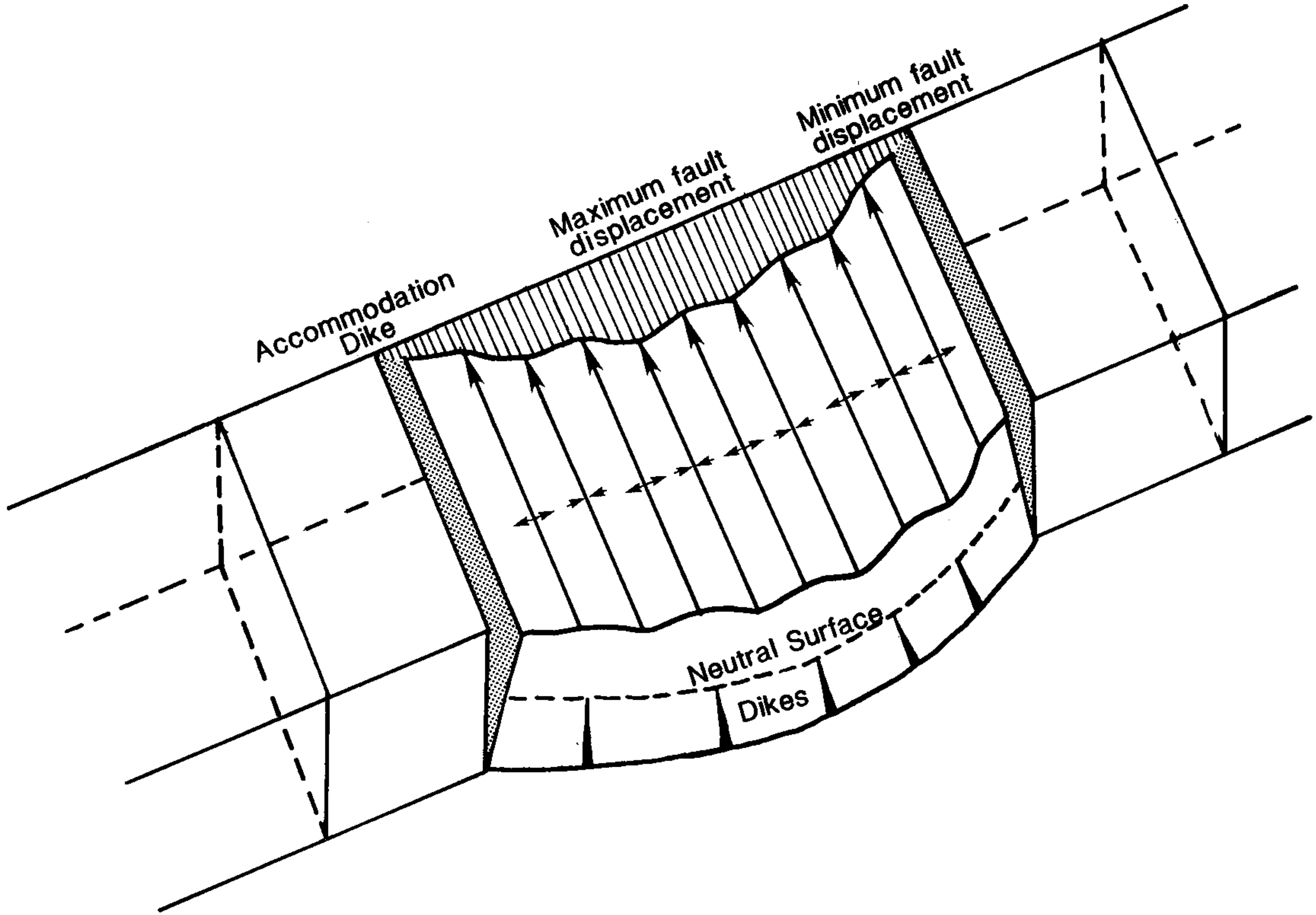

Figure 4: Highly simplified model for generating the compressional and extensional structures of the Newark basin, based on a variable displacement field for the border fault, which produces a synform-shaped basin plunging toward the border fault. In this down-plunge view, the basin fill has been omitted for clarity of presentation. The upper surface of the synform has been shortened and the lower surface extended. Accommodation structures are required on either end of the basin.

tangential longitudinal strain (Ramsay and Huber, 1987), the upper concave surface of the model's plunging synform may experience fault-parallel shortening, producing fault-perpendicular folds, whereas the lower convex surface may experience fault-parallel extension. A neutral surface of no finite strain separates these two regions. Because these structures plunge toward the border fault and the present-day erosional surface is approximately horizontal, both the local folds and extensional structures can be observed in a basin-normal transect. The folds appear to die out away from their associated faults (a) as the neutral surface is approached and (b) because subsidence and therefore shortening also die out away from the fault. The NW-striking dike and the associated joints are easily explained as structures which formed under localized extensional conditions structurally below the neutral surface. This mechanism of fold formation also introduces large gaps at either end of the basin to accommodate the subsidence and shortening within the basin. When filled with igneous material, these structures become the Birdsboro dike and the northern extension of the Palisades sill.

For this mechanism to work, it requires a degree of coupling between layers during the overall synformal downwarping of the basin. If all of the layers were allowed to undergo pure flexural-slip folding with its attendant bedding-plane slip, then the upper surface of the basin's synform never would have experienced shortening. A detachment horizon at some depth also is required. In addition, one end of the hanging wall block containing the basin needs to have been free to move to allow for the shortening of the upper surface. If both ends were pinned, then both the upper and lower surfaces of the synform would have experienced extension.

Although the evidence for some of the requirements of the model are lacking, the model does kinematically explain the origin of all compressional and extensional structures, something which previous models have failed to do. Models calling for post-rift shortening (Sanders, 1963; Swanson, 1982) or syn-rift strike-slip (Manspeizer, 1980; Burton and Ratcliffe, 1985) can be ruled out because the folds were forming during basin subsidence along faults which experienced predominantly dip-slip. Wheeler (1939) proposed that the folds formed as the hanging wall slid down a corrugated fault surface, but the geometry of the faults has not borne this out (*e.g.*, the nearly straight Ramapo fault).

INTRABASINAL FAULTING

On the basis of the style and density of intrabasinal faulting and the nature of the preserved sedimentary record, the Newark basin can be divided into three subbasins, representative cross-sections of which are shown in Figure 5. The New Jersey subbasin comprises the northeastern portion of the basin, includes the Watchung syncline, and contains the thickest preserved accumulation of Jurassic sedimentary rocks. The Delaware River subbasin consists of that portion of the basin north of the Chalfont fault and includes the Flemington and Hopewell fault blocks. The Pennsylvania subbasin forms the remainder of the Newark basin, south of the Chalfont fault and east of the narrow neck. In both the Delaware River and Pennsylvania subbasins, Jurassic sedimentary rocks are only preserved in the structural cores of synclines.

In the New Jersey subbasin, extension was taken up almost exclusively on the moderately to steeply dipping border fault system, allowing for the greatest subsidence of all three subbasins, which resulted in the thickest accumulation and eventual preservation of synrift strata. In the Delaware River subbasin, extension was taken up partly on the shallow-dipping border fault system and partly on the Flemington and Hopewell faults. In part because of the shallow dip of the border fault and in part because of the distributed extension, the Delaware River subbasin subsided less than the New Jersey subbasin and, therefore, contains a much smaller preserved section of Jurassic strata. In the Pennsylvania subbasin, extension was partly taken up on the very shallow-dipping border fault system and partly along a dense network or minor normal faults (**z** in Figure 1B). A particularly stunning example of this type of faulting is exposed in a railroad cut south of Gwynedd, Pennsylvania (Figure 6). The faults strike NE, and the stratigraphic separation generally is less than a meter or two. Nevertheless, the faults produced an

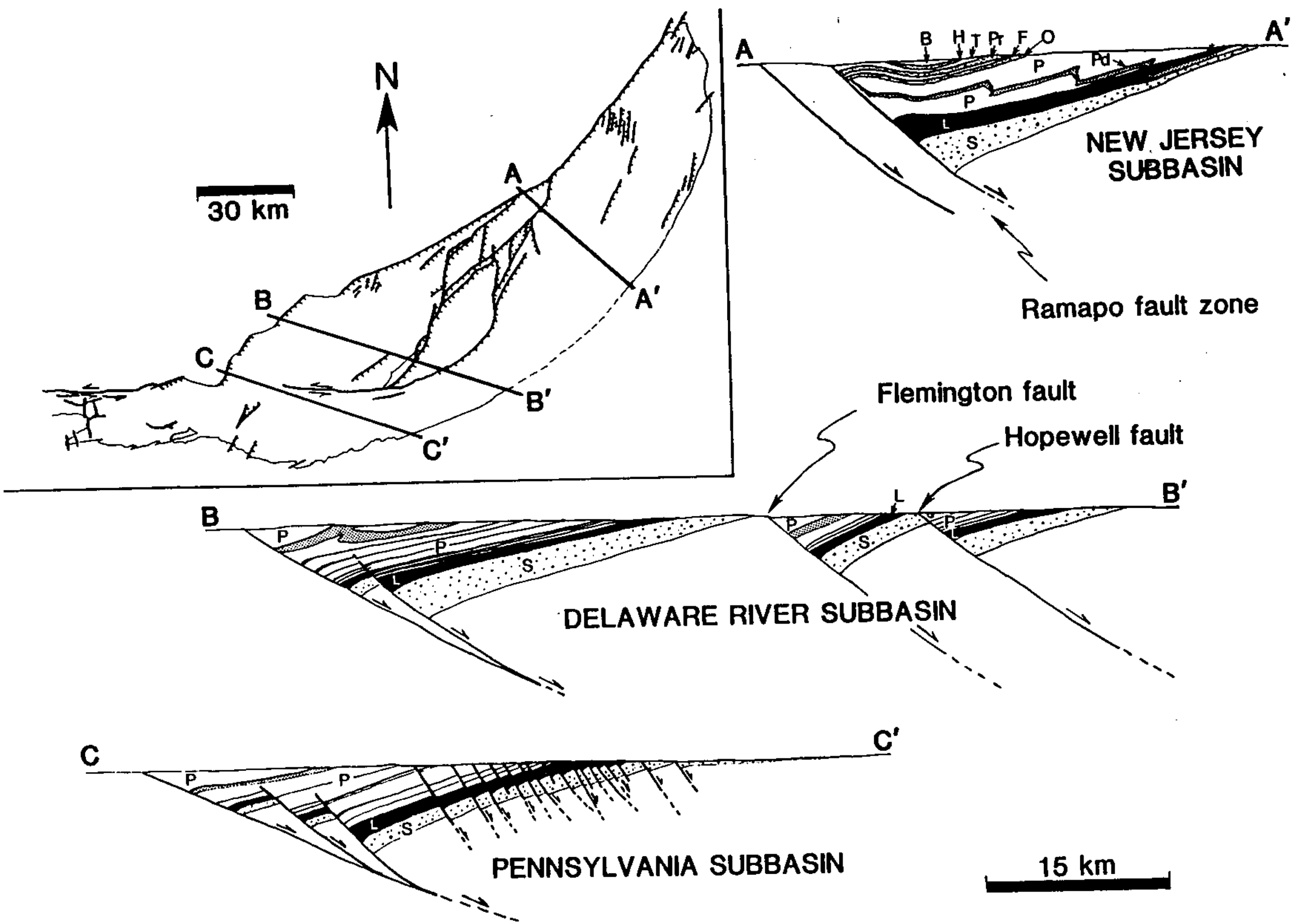

Figure 5: Representative cross-sections through the three subbasins of the Newark basin, illustrating the major structural differences. Abbreviations as in Figure 1.

apparent thickening of the section of 35%, and the amount of extension is 3.35% (Watson, 1958). Again, as a result of this distributed extension observed at Gwynedd and elsewhere in this subbasin, and in part because of the very shallow dip of the border fault in this region, the Pennsylvania subbasin subsided less than the New Jersey subbasin, and Jurassic sedimentary rocks are only preserved in the structural core of the Jacksonwald syncline.

The ESE-striking Chalfont fault has long been regarded as a down-to-the-south normal fault. The stratigraphic separation of the mapped contact between the Lockatong and Passaic formations is consistent with either down-to-the south normal faulting or left-lateral strike-slip. The offset of an Early Jurassic diabase intrusion (**db** in Figure 1B) holds promise in establishing the true nature of the type of fault slip. The intrusions on either side of the Chalfont fault belong to the same geochemical family (Smith *et al.*, 1975), suggesting that they may once have been a continuous feature. If the intrusion is a sill, then the exact nature of the slip cannot be determined. If, however, the intrusion is a dike intruded originally perpendicular to bedding, then the offset of the Lockatong-Passaic contact and of the dike require left-lateral strike-slip. Existing maps show the intrusion to be discordant with bedding over much of its length. Recent field work has established that the intrusion is steeply dipping south of the Chalfont fault and consists of a number of subparallel, perhaps *en echelon*, segments north of the fault, suggesting that the intrusion is a dike. Minor structures, consisting of steeply dipping, ESE-striking faults with subhorizontal slickenlines and a similarly oriented shear zone with a left-lateral sense of shear, observed in the town of Chalfont, Pennsylvania, immediately adjacent to the fault suggest that the Chalfont fault is a strike-slip fault.

Perhaps the most compelling reason for the strike-slip interpretation for the Chalfont fault stems from kinematic arguments. The Chalfont fault separates the Pennsylvania and Delaware River subbasins. South of the Chalfont fault, the basin fill was extended along a series of closely spaced normal faults (*e.g.*, Figure 6). Immediately north of the Chalfont fault, the

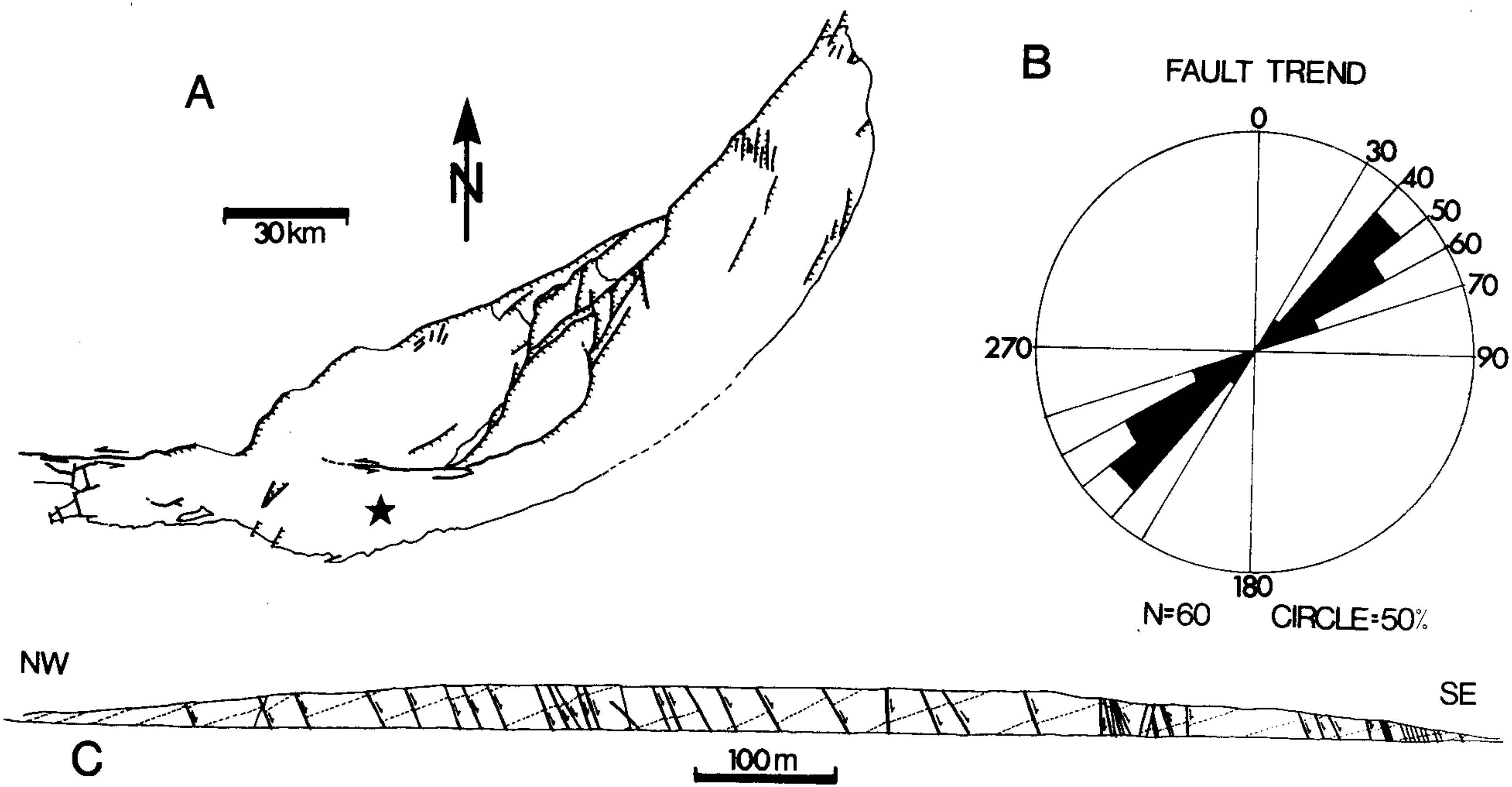

*Figure 6: High-density normal faulting in Reading Railroad cut south of Gwynedd, Pennsylvania. (**A**) Location of railroad cut marked by a star. (**B**) Rose diagram of fault trends. (**C**) Sketch of railroad cut. Dashed lines represent bedding. Data and sketch from Watson (1958).*

rocks are relatively unextended, although the whole subbasin was extended along the Flemington and Hopewell faults. The Chalfont fault dies out to the west because the spacing and intensity of the normal faulting similarly die out. At the fault's termination, the rocks on either side are relatively unextended. The differential extension across the Chalfont fault suggests that it is a left-lateral transfer fault, kinematically required to take up the variations in strain in an extending region (Bally, 1981; Gibbs, 1984; Lister *et al.*, 1986). In the same vein, a broad zone consisting of anastomosing and bifurcating NNE-striking normal faults accommodated the variations in strain between the New Jersey and Delaware River subbasins.

The structures and strata preserved within the fault blocks of the Delaware River subbasin allow us to constrain the timing and origin of the intrabasinal faulting. Strata within the hanging walls of the Flemington and Hopewell faults dip more steeply than those of the border fault (Figure 5), indicating that the hanging walls of the two intrabasinal faults experienced an added component of rotation over that of the border fault, suggesting that these two faults were at some point in the basin's history more active than the border fault. In addition, the Flemington and Hopewell faults dip somewhat more steeply than the border fault (Ratcliffe and Burton, 1988).

None of the formations (Stockton, Lockatong, Passaic, and Feltville) preserved within the hanging walls of the Flemington and Hopewell faults shows any evidence of syndepositional faulting (Olsen, 1980), strongly suggesting that the faulting post-dates the Early Jurassic Feltville Formation. In fact, the timing of faulting may coincide with a period of extensive hydrothermal alteration hypothesized to have reset the radiometric clocks of Newark igneous rocks to 175 Ma (Sutter, 1988). Hence, the Flemington and Hopewell faults appear to have developed late in the history of the basin, possibly to more easily accommodate the extension than was possible on the shallow-dipping, possibly shallowing, border fault (Schlische and Olsen, 1988). As extension progressed and the crust was tectonically denuded, the already shallow-dipping border fault system may have been isostatically rotated to an even shallower dip, which Sibson (1985) has shown makes slip more difficult. Extension may then have been transferred to two new, more steeply dipping faults whose hanging walls suffered an added component of rotation over that which had been imparted to the previously unfaulted hanging wall of the border fault. A displacement field and folding model similar to that postulated to have occurred along the border fault may have been responsible for the folds developed in the hanging wall of the Flemington fault.

In the New Jersey subbasin, the steeper dip of the border fault system prevented it from locking during the course of extension. Hence, the New Jersey subbasin is relatively unextended. In contrast, numerous small faults took up the extension when the border fault locked in the Pennsylvania subbasin. The differences in the style of faulting among the subbasins may reflect the initial dips and, therefore, the thicknesses of the hanging wall blocks (Figure 5). In the Pennsylvania subbasin, in which the utilized border faults were located furthest toward the hinterland and therefore had the shallowest dips, the hanging wall block was thinner, consequently weaker, and therefore prone to the high-density normal faulting. Note that the thickest hanging wall block--that of the New Jersey subbasin--is relatively unfractured. The Chalfont fault and the accommodation zone between the New Jersey and Delaware River subbasins may delineate significant subsurface changes in the dip of the border faults.

The Chalfont fault is additionally significant for it allows us to further constrain the early Mesozoic extension direction. Transfer faults are kinematically analogous to transform faults in the oceanic crust and hence are parallel to the extension direction. The Chalfont fault, therefore, gives an ESE extension direction, the same provided by the average strike of Early Jurassic-aged diabase dikes. This extension direction is more east-directed than that given by Ratcliffe and Burton (1985) and therefore fails to account for their observed right-oblique slip along the Ramapo fault, according to their fault reactivation model. These differences may be resolved if the extension direction varied after the Early Jurassic or if the Ramapo fault were reactivated in a stress field unrelated to Newark basin extension. A third possibility is that the right-oblique slip resulted from a combination of dip-slip (predicted by our extension direction) and the fold-forming shortening of the hanging wall in that region.

SUMMARY

The Newark basin is characterized by transverse folds, many of which were growing during sedimentation. These folds apparently formed as accommodation structures within the upper surface of the synformal basin induced by a variable displacement field for the border fault system. A similar mechanism could account for the folds developed in the hanging walls of the intrabasinal faults, which formed late in the history of rifting to more easily accommodate the extension than was possible on the shallow-dipping border fault. Transverse folds are also well developed in the Hartford-Deerfield, Gettysburg, Culpeper, and Dan River basins of the Newark Supergroup. If our model of fold formation is correct, then such folds should be an important component of other rift basins, where such folds have not been reported. Is this a function of the poor exposure of these basins with respect to the Newark basin? Or are these folds simply not present? If the latter is true, we must ask: What makes the basins of the Newark Supergroup so special?

ACKNOWLEDGMENTS

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A CORED STRATIGRAPHIC SECTION THROUGH THE NORTHERN NEWARK BASIN, NEW JERSEY*

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INTRODUCTION

The U.S. Army Corps of Engineers has proposed a flood diversion tunnel, about 22 km long, for the Passaic River basin in northern New Jersey. This tunnel is to be excavated entirely within the nonmarine Mesozoic rocks of the Newark Basin, trending across their regional northeast strike (Fig. 1). Continuous rock cores, ranging from 18 to 141 m long, were recovered from 27 vertical holes drilled in 1985 and 1986 and from five older holes scattered along the proposed tunnel alignment. These holes were drilled primarily to examine the engineering qualities of the rock. The close spacing of the holes provides stratigraphic overlap of the gently northwest dipping beds, enabling precise correlation of strata in adjacent holes in most instances. Geophysical logs provided by the U.S. Geological Survey support the stratigraphic correlation and were especially helpful where visual correlation was difficult. This correlation provides the first relatively complete stratigraphic section of the Jurassic rocks of the northern Newark Basin.

The thickness of the Mesozoic succession in the northern Newark Basin is approximately 5000 m from the base of the Stockton Formation to the eroded top of the Boonton Formation (Fig. 2). The proposed tunnel would cut through about 2400 m of strata from the upper part of the Passaic Formation through the lower half of the Boonton Formation (Fig. 2). A total of 2571 m of rock core has been recovered representing about 1650 m of stratigraphic section (Table 1), most of which is Early Jurassic in age. Several cores that were taken from holes drilled along strike in the Boonton Formation show lateral facies changes within the same stratigraphic units. Older Passaic strata along the projected tunnel route will be cored in future programs.

Most of the sedimentary portions of the cores are red sandstone with a smaller component of red siltstone and mudstone, and minor amounts of black shale, gray mudstone, and red conglomerate. The portions of the Passaic Formation that were cored differ from the overlying strata by being more coarse grained, by having poorly preserved internal stratification, and by the absence of black shale and gray mudstone. The Feltville, Towaco, and Boonton Formations are dominated by ripple cross-laminated sandstone and siltstone with distinct black shale and gray mudstone units at intervals of 12 to 60 m. Although these three formations are lithologically similar, each has a characteristic sedimentary package different from the others. Each of the three basalt formations (Fig. 1) consists of two or more extrusive flows whose tops are delineated by zones of vesicles and possible soil horizons. The basalts contain abundant zeolites along faults and minor fractures and within vesicles.

CORE DESCRIPTIONS BY FORMATION

Passaic Formation

The Passaic Formation (Olsen, 1980a) was drilled discontinuously to a stratigraphic level about 1100 m below the Orange Mountain Basalt (Fig. 2). Only 430 m of core was collected, including 190 m of continuous section at the top of

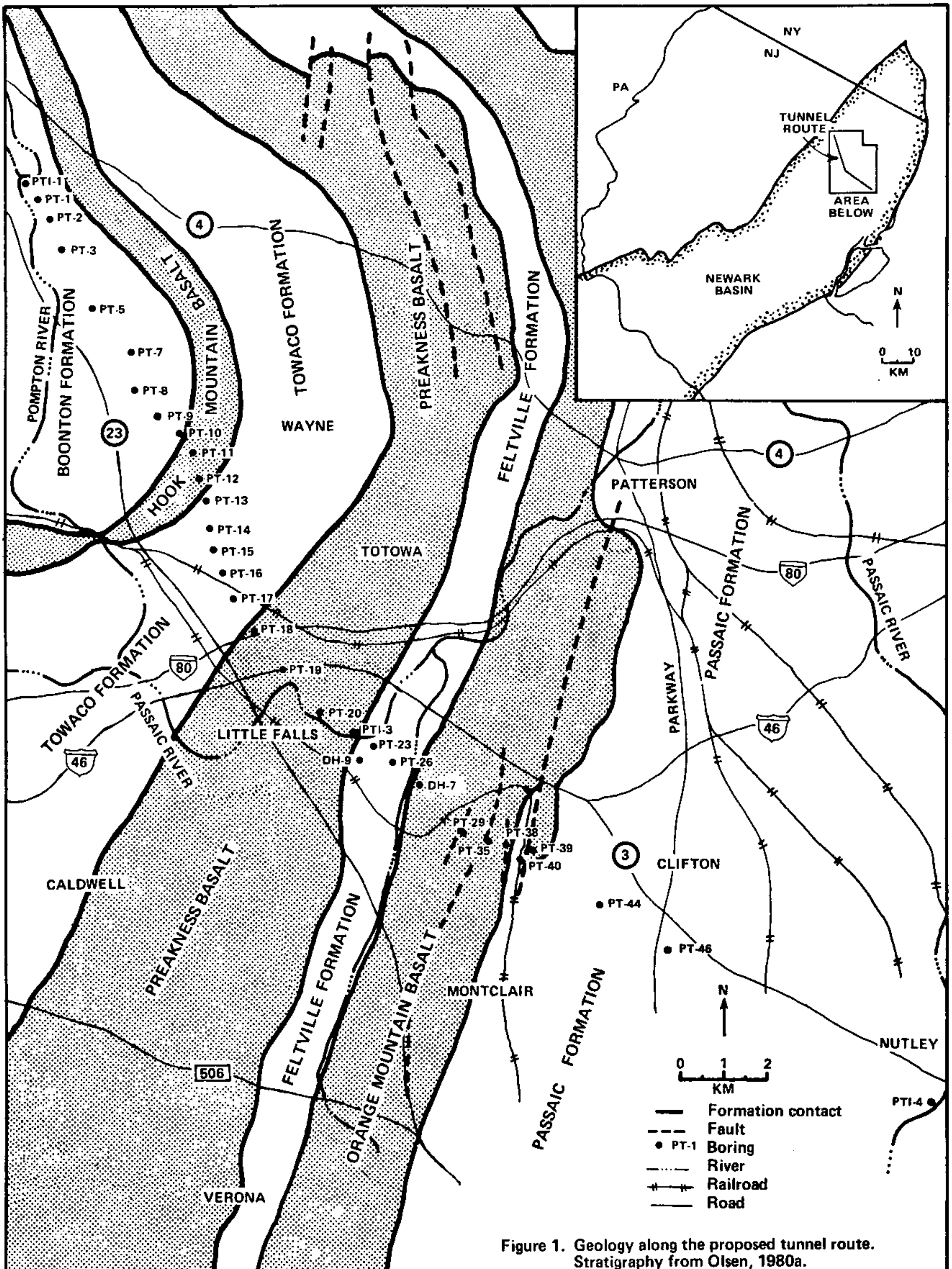
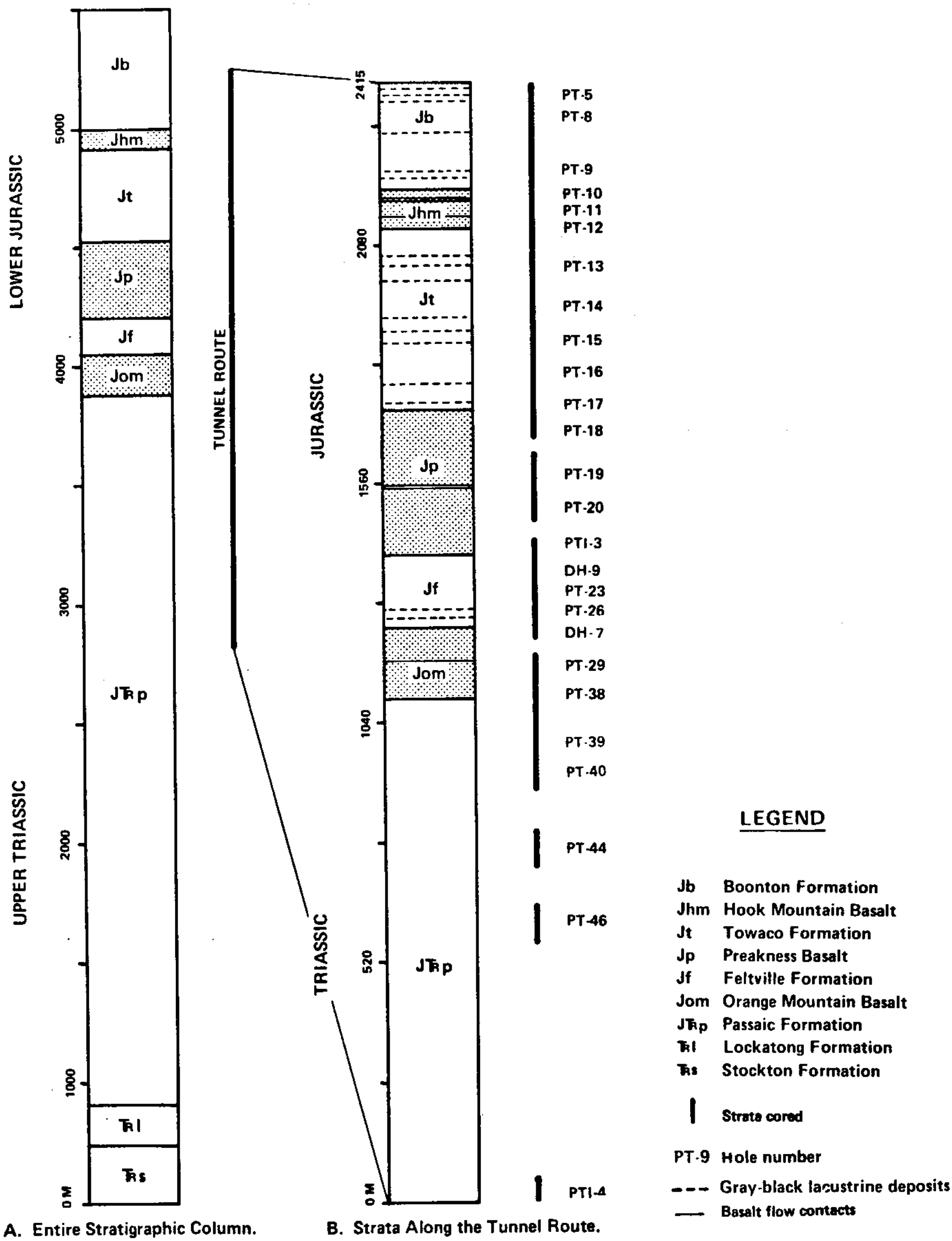


Figure 1. Geology along the proposed tunnel route. Stratigraphy from Olsen, 1980a.



A. Entire Stratigraphic Column. B. Strata Along the Tunnel Route.

Figure 2. Stratigraphic column of the northern Newark basin showing strata cored (modified from Olsen, 1980a).

TABLE 1: Length of core recovered from each drill hole.

Hole #	Length	Formation(s)	Hole #	Length	Formation(s)
PTI-1	40 m	Boonton	PT-18	43 m	Preakness
PT-1	65 m	Boonton	PT-19	69 m	Preakness
PT-2	81 m	Boonton	PT-20	84 m	Preakness
PT-3	74 m	Boonton	PTI-3	65 m	Preakness &
PT-5	86 m	Boonton			Feltville
PT-7	86 m	Boonton	DH-9	31 m	Feltville
PT-8	91 m	Boonton	PT-23	57 m	Feltville
PT-9	127 m	Boonton &	PT-26	61 m	Feltville &
		Hook Mtn.			Orange Mtn.
PT-10	36 m	Boonton &	DH-7	18 m	Orange Mtn.
		Hook Mtn.			
PT-11	77 m	Hook Mtn. &	PT-29	105 m	Orange Mtn. &
		Towaco			Passaic
PT-12	86 m	Hook Mtn. &	PT-35	124 m	Orange Mtn. &
		Towaco			Passaic
PT-13	103 m	Towaco	PT-38	141 m	Orange Mtn. &
PT-14	118 m	Towaco			Passaic
PT-15	89 m	Towaco	PT-39	125 m	Orange Mtn. &
PT-16	105 m	Towaco &			Passaic
		Preakness	PT-40	128 m	Passaic
PT-17	55 m	Towaco &	PT-44	87 m	Passaic
		Preakness	PT-46	64 m	Passaic
			PTI-4	50 m	Passaic

the formation. The general sedimentary style consists of irregularly-spaced, 0.3-5.0 m thick, graded sequences. Each sequence ranges in grain size from pebbly sandstone that grades upward into medium-grained sandstone to muddy sandstone that grades upward into muddy siltstone. Vertical successions of several of these graded sequences appear to define 15 to 25 m thick fining-upward packages, consisting of sandstone beds with pebbles of limestone and quartz at the base, and muddy siltstones with carbonate nodules at the top. All of the strata are bioturbated and stratification generally is poorly preserved. Decimeter-scale crossbeds are visible in some sandstone beds, and a few thin beds of interlaminated mudstone and fine-grained sandstone occur near the tops of the large-scale, fining-upward packages. The bioturbation includes root casts and burrows, in particular Scovenia (Olsen, 1980b). In addition, mudcracks and small scours are present in the laminated portions of the cores.

The Passaic deposits probably are fluvial in origin. However, it is not clear whether the small-scale graded sequences represent channel deposits of small braided streams, or if they are sections through large trough crossbeds within 15-25 m thick channel-fill or point bar deposits of large rivers.

Orange Mountain Basalt

The lower 100 m and the upper 20 m of the 160 m thick Orange Mountain Basalt were cored (Fig. 2). At least two tholeiitic basalt flows are separated by a thick vesicular zone, with the lower flow being 82 m thick. The bottom meter of the lower flow exhibits pillow structures indicative of extrusion into standing water, perhaps a shallow lake. The top of the lower flow consists of 10 m of vesicular basalt capped by a brown, weathered layered that may be a paleosol.

Feltville Formation

The entire 155 m of the Feltville Formation was cored (Fig. 2). Two calcareous, black laminated shales, each about 2 m thick, separated by about 20 m of predominately gray sandstone, siltstone, and mudstone, occur approximately 20 m above the basal contact. This sequence is equivalent to the informal Washington Valley member of Olsen (1980b). The bulk of the Feltville consists of alternations of red sandstone and muddy siltstone to silty mudstone. The sandstones form 1 to 4 m thick fining-upward sequences. The thickest sequences have crossbedded (10-20 cm thick sets), coarse-grained sandstone overlying a sharp basal contact. The coarse-grained sandstone grades upward into medium- to fine-grained sandstone with low-angle, climbing-ripple cross-lamination, which grades into rippled to flat-laminated siltstone. This, in turn, grades upward into bioturbated muddy siltstone, which may have claystone partings or, more typically, is massive. Thinner fining-upward sequences are composed of medium-to fine-grained sandstone with abundant low-angle, climbing-ripple cross-laminae. The muddy siltstone and silty mudstone layers typically are massive to poorly bedded with abundant root casts, burrows (although Scovenia-like burrows occur only in a few zones), and carbonate nodules. Mudcracks are common in the upper portions of mudstone and siltstone to increasingly thicker and more closely spaced fining-upward sandstone sequences.

The sedimentary structures in the Feltville indicate primarily fluvial deposition in low-gradient, meandering or anastomosing streams with crevasse-splay overbank deposits. The black shales of the Washington Valley member indicate relatively deep lake conditions, whereas the gray siltstones were deposited in shallower water. The overlying rippled sandstones are probably deltaic. The lower portions of the coarsening-upward sequences may represent delta front or distributary deposits in shallow lakes.

Preakness Basalt

The Preakness Basalt was cored discontinuously, recovering 259 m of its 310 m thickness (Fig. 2). At least two tholeiitic basalt flows are separated by 3 m of fine-grained sandstone and siltstone, the latter apparently defining a fluvial, fining-upward sequence. The 145 m thick basalt below the sedimentary layer has coarsely-crystalline gabbroic layers as much as 11 m thick. The upper basalt unit also contains gabbroic layers, but they are less than 8 cm thick. The coarsely crystalline layers may represent pockets of slowly cooled lava, segregation veins, or intrusive layers.

Towaco Formation

The entire 380 m of the Towaco Formation was cored (Fig. 2). The Towaco resembles the Feltville, except that it has eight calcareous, black laminated shales, which generally are thicker (3-5 m) than those of the Feltville, and has sandstones that typically are more fine grained (medium- to fine-grained sandstone). The black laminites commonly grade both upward and downward into gray bioturbated silty mudstone and sandstone. Red sandstones form 1-3 m thick fining-upward sequences, typically consisting of a basal medium-grained sandstone with low-angle, climbing-ripple cross-lamination, grading to fine-grained sandstone with high-angle, climbing-ripple cross-lamination. The top of each graded sequence consists of flat-laminated to massive bioturbated siltstone. Soft-sediment deformation structures are common in the sandstones, including load casts, pseudonodules, oversteepened cross-laminae, and small faults. Scovenia-like burrows are seen, but root casts and other small burrows are more abundant. Mudcracks occur irregularly throughout the section in red mudstones, but are less common than the soft-sediment deformation. Sandstones overlying mudcracked mudstones typically are rich in mud clasts and fill irregular scour pockets.

The black laminites are interpreted as deep lake deposits which grade into bioturbated siltstones that were deposited in shallow water. The rise and fall of lake levels apparently was in response to cyclic climatic variations (Olsen, 1986). The climbing ripples in graded sandstone sequences suggest pulses of deposition in standing water. These features and the abundance of soft-sediment deformation structures are similar to shallow water delta deposits described by Smoot and others (1985). In addition, some of the sandstone packages represent small stream deposits while the mudcracks indicate sporadic subaerial exposure.

Hook Mountain Basalt

The entire 85 m of the Hook Mountain Basalt was cored (Fig. 2) It consists of at least four tholeiitic basalt flow units, which are from base to top, 29, 35, 3, and 18 m thick. The top of each flow unit has brown brecciated basalt grading down into a thin vesicular layer. The brown brecciated layers are interpreted as paleosols developed on upper flow surfaces.

Boonton Formation

Approximately 225 m of continuous section was cored in the lower part of the 500 m thick Boonton (Fig. 2). Six, 3 to 5 m thick, calcareous, black laminated shales which grade into gray siltstones are distributed rhythmically through the dominant red sandstones and mudstones. The shale sequences are very similar to those of the Towaco Formation. The sandstone-mudstone red bed units are more similar to those of the Feltville, although the bases of some fining-upward sequences contain limestone pebble conglomerates. Evaporite crystal molds were observed in one shale unit near the base of the Boonton, and several red silty mudstone layers have possible sand-patch fabric (Smoot and Olsen, 1985).

The similarities to the Towaco Formation suggest that the Boonton had similar, climatically induced, lake level fluctuations. The fluvial deposits, similar to those in the Feltville, may differ from the predominately deltaic Towaco sandstones because the area cored in the Boonton is closer to the basin margin or because channeled areas were not crossed in the

Towaco core transect. It also could indicate that longer periods of dryness followed lake regressions in the Boonton, allowing stream channel systems to develop over the old lake bottom. Arid evaporitic conditions for at least part of the lower Boonton accumulation are suggested by the evaporite pseudomorphs and is supported by the possible sand-patch fabric, which is produced by sediment accumulation on an efflorescent, salt-encrusted surface.

CONCLUSIONS

The cored section of the Passaic Formation indicates depositional conditions considerably different from the sections cored through the younger sedimentary deposits. The coarser grain size of the Passaic suggests that stream gradients were steeper than those of the Feltville, Towaco, and Boonton Formations. The Passaic was deposited either by a complex of small braided streams or by larger rivers than those that deposited the younger formations. The black shale and gray siltstone sequences in the Feltville, Towaco, and Boonton Formations were deposited in lakes whose levels fluctuated, apparently in response to cyclic climatic variations (Olsen, 1986). The cyclic nature of these lake sequences is not obvious, probably due to superimposed fluvial and deltaic aggradational sequences. The Feltville Formation may have fewer black shale sequences than the younger formations, because its period of accumulation was shorter, spanning only two climatic periods of high inflow. The red sandstones and siltstones in the Feltville, Towaco, and Boonton Formations were deposited by low velocity streams or in deltas, as indicated by the abundance of ripple cross-laminae. The Feltville and Boonton cores indicate predominately fluvial channel sequences in the coarser sediments, while the Towaco Formation, dominated by graded ripple sequences, probably is mostly deltaic. The Feltville cored intervals are coarser grained than those of the Towaco despite their location more basinward relative to the northern border fault. This suggests that the Feltville may have had steeper gradients or that the Towaco was more poorly drained. It also is possible that the Towaco cores failed to cross a channeled area and the regional depositional conditions were very similar. The presence of evaporite pseudomorphs in the lower Boonton Formation suggests more arid conditions than for the Feltville and Towaco. Overall, the cored sequence through the sedimentary formations in the northern Newark basin suggests a gradual increase in standing water from Passaic deposition to the Towaco and a decrease in depositional gradient. A trend toward drier conditions from the Towaco Formation to the Boonton Formation also is indicated. These changes may be due entirely to climatic changes and/or to changes in the basin drainage due to tectonic activity. Basalt flows largely were subaerial, followed by periods of volcanic inactivity long enough to allow for soil development.

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THE WATCHUNG BASALTS REVISITED

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ABSTRACT

Previously published conclusions relating to the Watchung Basalts (Puffer and others, 1981) have been re-examined in light of more recent geochemical data. The new data support previous observations that the Orange Mountain Basalt is a "hi-Ti" (HTQ) type basalt with very little chemical variation; it essentially is equivalent to other Early Jurassic HTQ rocks such as the Talcott Basalt of the Hartford Basin, the Mount Zion Church Basalt of the Culpeper Basin, and the chill zone of the Palisades sill. This very narrow chemical range supports the interpretation that HTQ basalt is a primary magma. The composition of the Preakness Basalt, however, is found to be more complex than described previously. Although the lower flows of the Preakness are iron- and titanium-enriched basalts, the uppermost flow is a low titanium (LTQ) basalt. Presumably, the lower Preakness flows extruded from the upper levels of a differentiated LTQ magma chamber. The major-element composition of the Hook Mountain Basalt, particularly magnesium and titanium concentrations, supports the interpretation that it fractionated from a parental Orange Mountain magma. However, although inconclusive, REE data suggest an alternative source, perhaps a LTQ magma.

Geologic Setting

The Newark Supergroup of the Newark Basin (New York, New Jersey, and Pennsylvania) is divided by Olsen (1980) into nine formations (from bottom to top): Stockton Formation (maximum

1800 m); Lockatong Formation (maximum 1150 m); Passaic Formation (maximum 6000 m); Orange Mountain Basalt (maximum 200 m); Feltville Formation (maximum 600 m); Preakness Basalt (maximum 300+ m); Towaco Formation (maximum 340 m); Hook Mountain Basalt (maximum 110 m); and Boonton Formation (maximum 500+ m).

The Newark Basin contains the thickest preserved section of the Newark Supergroup. The sedimentary rocks and volcanic flows of the basin dip 5°-25° to the northwest, typically dipping 15° throughout most of northeastern New Jersey. Red siltstones and tholeiitic basalts are the dominant constituents of the basin. Mesozoic sediment deposition probably began during middle Carnian (early Late Triassic) time while the uppermost exposed deposits are Sinemurian (middle Early Jurassic) in age (Cornet, 1977).

The three Watchung Basalt flows contained within the Newark Basin record an early rifting, prespreading stage of tectonic development. The Watchung igneous activity of the Newark Basin was both preceded and followed by continental red-bed sedimentation and is now part of a widespread province of Early Jurassic eastern North America (ENA) basaltic rock exposures extending from Nova Scotia to Georgia.

Paleontological, stratigraphic, and geochemical evidence (Olsen, 1988; Puffer and others, 1981) indicate that the three Watchung Basalts extruded in rapid succession during the Hettangian age of the early Jurassic period. Olsen and Fedosh (1988) suggest that the time interval separating the Watchung Basalts, and the stratigraphically equivalent flows of the Hartford Basin, Connecticut, was only 550,000 years.

Most exposures of the Watchung Basalts display closely spaced columnar joints, but a few, particularly of the uppermost flow of the Orange Mountain Basalt, are pillowed. One of the best exposures of columnar jointing is found at the type section of the Watchung Basalts (Olsen, 1980) along Route 280, West Orange, New Jersey (Fig. 1). The structural sequence is comparable to that of the chemically similar Giants Causeway of northern Ireland as described by Tomkeieff (1940), the Columbia River basalt flows (Long and Wood, 1986), and several other flows of the western United States (DeGraff and Aydin, 1987).

The basalt flows, together with the entire Newark Basin, have been subjected to some burial metamorphism. The vesicles of the basalt and the spaces between any pillows were partially filled with gypsum and glauberite, presumably precipitated by volcanic vapors and solutions released during the cooling of the basalt lava. The vesicules and vugs were filled further with carbonates and clay, as typically is the case with flows subjected to prolonged groundwater alteration. Then, during zeolite facies burial metamorphism, the clay and carbonates were recrystallized into a complex assemblage of zeolites (including stilbite, heulandite, chabazite, pectolite, analcime, and datolite) and prehnite (Puffer and Laskowich, 1984).

In addition, copper minerals probably were recrystallized, along with the mobilization of copper-bearing fluids, during zeolite facies mineralization. The base of the Orange Mountain Basalt coincides with over 20 copper deposits that were mined during the Civil War (Puffer, 1987).

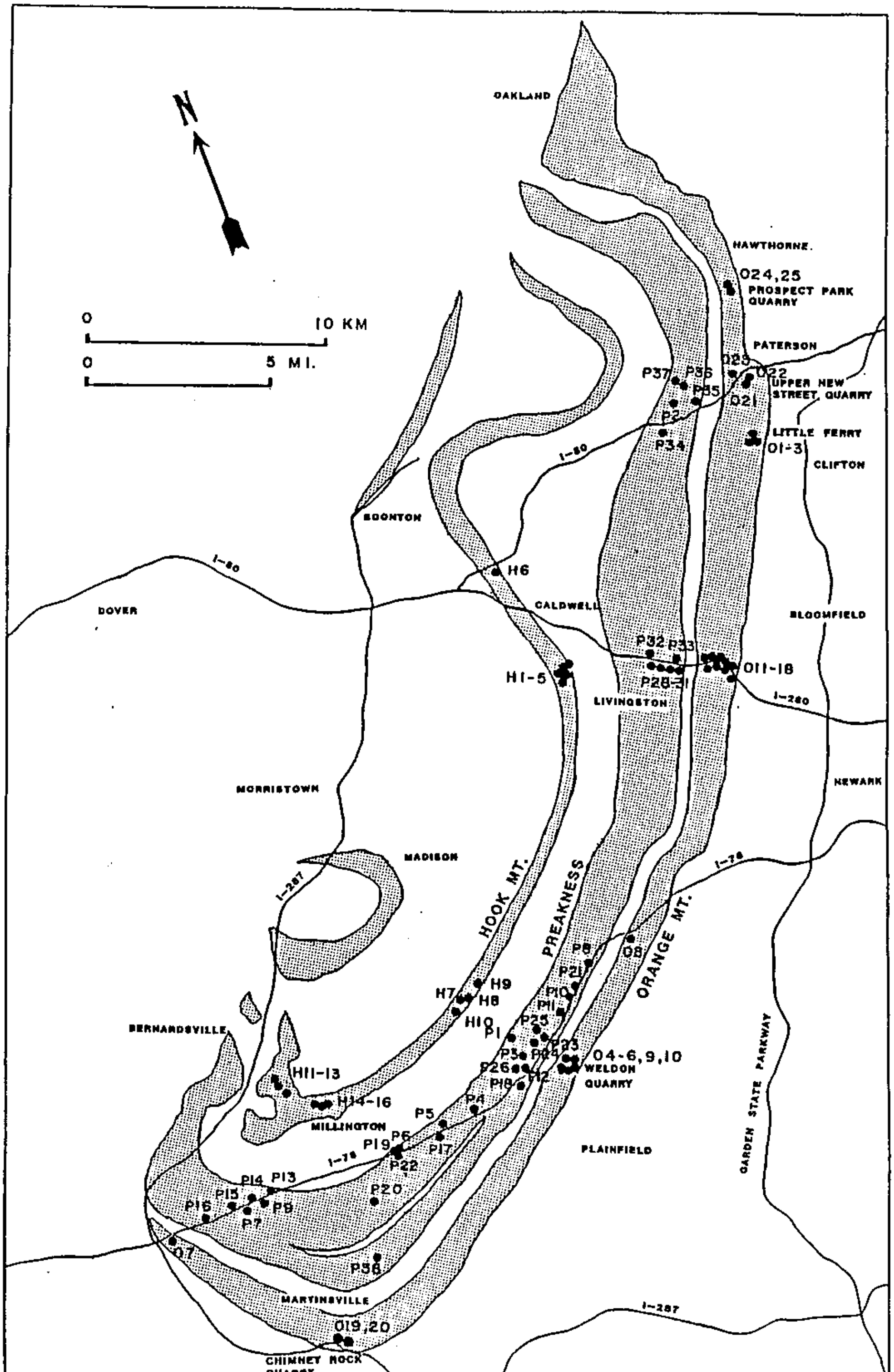

Fig. 1. Map of Watchung Basalts with locations of analyzed samples.

Petrologic and Geochemical Characteristics

Orange Mountain Basalt

The Orange Mountain Basalt averages 183 m thick (Faust, 1975) and consists of three flow units. The upper flow unit characteristically is fine grained, pillowed, and amygdaloidal, whereas the lower two flows typically display well-developed columnar joints (lower colonnade and entablature).

The Orange Mountain Basalt is a quartz normative tholeiite composed of plagioclase and augite with minor orthopyroxene and altered olivine in a glassy mesostasis containing quench dendrites of Fe-Ti oxides. Augite phenocrysts, glomeroporphyritic aggregates of augite, orthopyroxene, olivine, and a few plagioclase phenocrysts are characteristic of the basalt. Typical modes average 35 percent plagioclase (An_{65}), 35 percent pyroxene (augite ($Wo_{10}En_{55}Fs_{35}$), pigeonite, and minor hypersthene), 28 percent glassy mesostasis, and 3 percent opaque Fe-Ti oxides. Accessory and trace minerals include apatite, biotite, alkali feldspar, and pyrite.

The chemical composition of the Orange Mountain Basalt falls within the chemically diverse group of light REE-enriched "continental" tholeiites. The Orange Mountain Basalt is similar particularly to the Karroo Basalts of the Basutoland subprovince, South Africa (Cox and Hornung, 1967), and, to a slightly lesser extent, some other continental basalts such as the Keweenawan Basalts of the Lake Superior area and the Atrium Basalts of Northern Ireland. The Orange Mountain Basalt, however, does not plot on most tectonic discrimination diagrams within the compositional fields designated as characteristic of "within-plate" basalts, such as those developed by Pearce and Cann (1973). As noted by Marsh (1987) and Duncan (1987), this same geochemical characteristic also is exhibited by the Karroo Basalts. Marsh (1987) concludes: "Basalt geochemistry alone is unhelpful in determining the tectonic setting of continental flood basalts although the rift-related environments may be identified by the petrology and geochemistry of the whole igneous suite." Marsh (1987) also believes the source of the Karroo magma was an alkali-enriched subcontinental lithosphere, as supported by the work of Hawkesworth and Van Calsteren (1984) and Erlank and others (1982).

The Orange Mountain Basalt fits into the high-titanium group (HTQ) of ENA tholeiites as proposed by Weigand and Ragland (1970). The chemistry of the basalt is extremely uniform throughout (Fig. 2), averaging 51 (weight) percent SiO_2, 1.0 percent TiO_2, 14 percent Al_2O_3, 10 percent FeO (total), 8 percent MgO, 10.5 percent CaO, 2.1 percent Na_2O, 0.13 percent P_2O_5, 125 ppm Cu, 180 ppm Sr, 60 ppm Ni, and 115 ppm Zr, all within a very narrow range. The REE content of the Orange Mountain Basalt is within the "high-titanium" range of Ragland and others (1971) and plots close to and parallel with the REE distribution pattern of the lower chill margin of the Palisades sill (Fig. 3).

The chemical composition of the Orange Mountain Basalt (Table 1) virtually is equivalent in all respects to samples of Palisades chill analyzed by Walker (1969) and Shirley (1987).

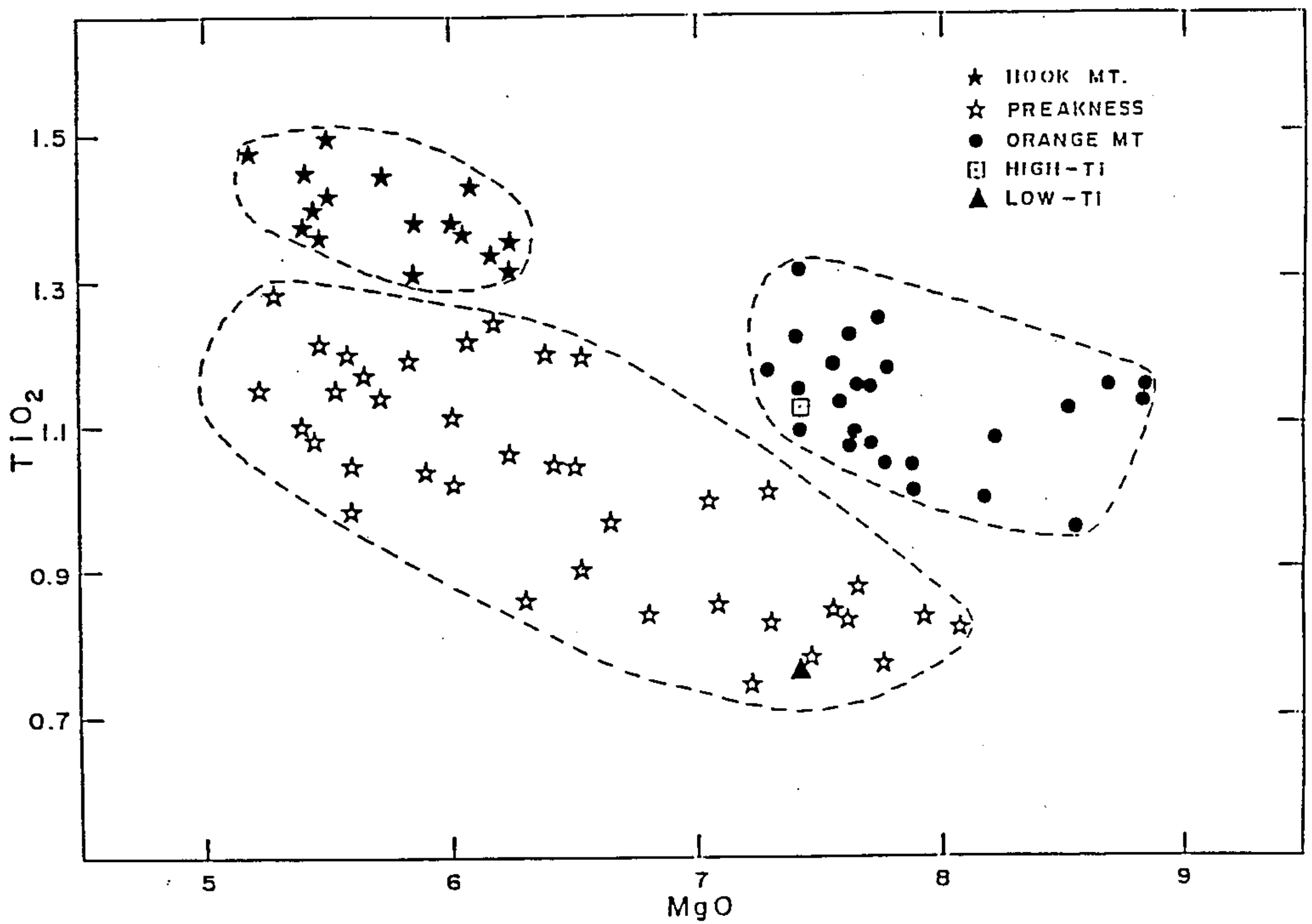

Fig. 2. MgO/TiO_2 distribution of Watchung Basalt samples located on Fig. 1. High-Ti and Low-Ti average values are after Weigand and Ragland (1970).

TABLE 1

Chemical Composition of Watchung Basalts
(typical samples)

	012 Orange Mt.	P5b Preakness first flow	P116 Preakness second flow	P146 Preakness third flow	H1 Hook Mt.
SiO_2	51.44	51.65	51.20	50.83	49.89
TiO_2	1.19	1.05	1.21	0.83	1.44
Al_2O_3	14.75	14.19	15.30	15.65	13.26
FeO	10.54	11.41	12.45	10.15	14.89
MnO	0.16	0.19	0.21	0.17	0.19
MgO	7.77	6.48	5.41	7.60	5.65
CaO	10.61	9.48	9.55	10.31	9.88
Na_2O	2.35	3.05	3.21	2.49	2.53
K_2O	0.32	0.63	0.59	0.36	0.32
P_2O_5	0.11	0.12	0.12	0.08	0.14
H_2O^+	0.42	0.76	0.48	0.61	0.69
Total	99.66	99.01	99.73	99.08	98.88
Ba	162	142	135	88	110
Co	37	38	39	38	43
Cr	374	76	53	251	122
Cu	120	78	80	70	167
Ni	84	38	25	55	50
Sr	206	136	151	141	118
V	250	315	340	248	324
Zn	72	84	95	72	115
Zr	78	88	91	56	98
La	10.1	9	11	7.4	8.1
Ce	22.5	19	26	15	19
Nd	13.5	14	16	9	12
Sm	3.8	2.8	3.0	2.5	2.9
Eu	1.11	1.02	1.25	0.90	1.1
Gd	3.6	3.25	3.8	4.0	4.0
Dy	3.8	5.0	5.0	4.2	5.5
Er	2.3	3.5	3.8	2.5	4.0
Yb	2.3	3.1	3.2	2.2	3.1

Although the Orange Mountain Basalt and the Palisades sill very probably are comagmatic, Shirley (1987) has found that some Palisades rocks contain such low titanium values that they may be misinterpreted as belonging to the quartz normative low-titanium basalt group (LTQ) of Weigand and Ragland (1970).

Of the 52 Palisades samples analyzed by Shirley (1987), 10 contain less than 0.9 percent TiO_2. Eight of the 10 samples were collected from the lower 71 m of the sill where MgO values exceed 8 percent and Cr and Ni concentrations exceed 600 and 100 ppm, respectively. This is in contrast to the 7.44 percent MgO, 218 ppm Cr, and 48 ppm Ni contents of average LTQ basalt (Weigand and Ragland, 1970). The low titanium content of these mafic Palisades samples very likely is due to the accumulation of olivine and/or pyroxene in a HTQ magma or, as suggested by Husch (pers. comm., 1988), may be the result of a pulse of olivine normative and titanium-depleted Quarryville type (Smith and others, 1975) magma. However, 2 of the 10 samples are from 327 and 329 m above the base of the sill and contain only 6.6 and 6.1 percent MgO, 78 and 90 ppm Cr, and 66 and 35 ppm Ni, respectively. Their respective TiO_2 contents of 0.9 and 0.8 percent also contrasts with the 1.1 to 1.9 percent TiO_2 content of the adjacent 18 samples within 20 m above or below them. Thus, the possibility exists that these 2 samples are derived from a LTQ-like magma.

The chemistry, mineralogy, and texture of the Orange Mountain Basalt also is equivalent to the first Early Jurassic basalts in the other basins of the Newark Supergroup, such as the Talcott Basalt of the Hartford Basin, Connecticut, and the Mount

Zion Church Basalt of the Culpeper Basin, Virginia (Puffer and others, 1981, Puffer, 1984).

The Preakness Basalt

The Preakness Basalt of the Newark Basin consists of at least three flows locally separated by thin layers of red siltstone. The combined thickness of the Preakness flows averages 215 m (Olsen, 1980).

The very coarse-grained appearance of the first or lowermost flow of the Preakness, resembling a diabase, may be related to the unusual flow thickness that may have included intrusive pulses similar to those proposed by Philpotts and Burkett (1988). The lower flow is well exposed at several locations along the length of the Second Watchung ridge, including the type locality of the Preakness Basalt on Interstate 280 (Fig. 1).

Thin sections of basalt sampled at the base and throughout the lower colonnade of the lowermost Preakness flow are fine to medium grained and porphyritic. Typical unaltered lower colonnade samples consist of about 50 percent pyroxene and 43 percent plagioclase as an intergranular mixture with about 3 percent plagioclase phenocrysts and 5 percent dark, fine-grained glassy mesostasis enriched in quench oxides. Typical unaltered entablature samples consist of a coarse intergranular mixture of about 45 percent pyroxene, 50 percent plagioclase, 3 percent opaque oxides, with only about 2 percent brown devitrified glass. Samples affected by only slight degrees of apparent alteration, however, are found to contain considerable albite. Despite this, the plagioclase composition of the least altered samples averages about An_{57}, on the basis of 40 microprobe analyses, with a range from An_{50} to An_{60}, excluding albite determinations. On the basis of 35 microprobe analyses, about two-thirds of the pyroxene of the entablature samples is pigeonite with about one-third augite ($WO_{20}En_{45}FS_{35}$). On the basis of 20 microprobe analyses, the magnetite composition typically is USP_{60}, and, on the basis of 12 microprobe analyses, the ilmenite composition is Hem_8.

The chemical composition of the first flow of the Preakness is characterized by 1.2 percent TiO_2, 6.1 percent MgO, 40 ppm Cr, and 95 ppm Zr. The chemical composition and texture virtually is identical to the Holyoke Basalt of Connecticut exposed along Rt. 66 and to typical Sander Basalt of the Culpeper Basin, Virginia.

Mapping of the middle flow, or flows, of the Preakness Basalt still is incomplete and has been difficult. Flow contacts rarely are exposed and are difficult to recognize. Although abandoned copper mines and several very large trap-rock quarries are found along the length of the Orange Mountain Basalt ridge, there are no such quarries exposing Preakness Basalt. The existence of a middle Preakness flow is reasonably well established on the basis of exposures near Totawa, New Jersey, and along Interstate 78, but it has not been demonstrated that the flow at both of these locations is continuous.

A high degree of textural and chemical variation further complicates the mapping of what probably is a middle flow, but some preliminary generalizations may be warranted. Most samples from the middle flow are medium-grained, intergranular mixtures of 35 percent pyroxene (both pigeonite and augite), 59 percent plagioclase (An_{55}), 4 percent opaque oxides, and 2 percent brown devitrified glass. Some samples are aphyric, but large plagioclase phenocrysts make up as much as 7 percent of some other middle flow samples. The low pyroxene content and high plagioclase content is responsible for the low magnesium and high aluminum content of the rock (Table 1). The high modal content of Fe-Ti oxides is responsible for the high TiO_2 content. Loss of pyroxene and accumulation of plagioclase also seems to be responsible for a slightly positive Eu anomaly (Fig. 3). The plagioclase-rich samples plot at the fractionated end of the Preakness field on a TiO_2 versus MgO diagram (Fig. 2).

The uppermost flow of the Preakness Basalt is exposed best along Interstate 78 (Fig. 1); elsewhere exposures are altered extensively. The somewhat vuggy nature of the flow seems to have accelerated alteration. The flow typically is medium grained and slightly porphyritic, consisting of glomeroporphyritic clots of pyroxene (largely augite) and plagioclase (An_{60}) in a dark mesostasis enriched in quench oxides. The mesostasis makes up 30 to 50 percent of the rock. The pyroxene and plagioclase content of the rock is approximately equal. Large euhedral phenocrysts of plagioclase and pyroxene are not uncommon.

The chemical composition of the uppermost flow (Table 1) is characterized by a consistently low TiO_2 content ranging from 0.7 to 0.9 and averaging 0.8 percent (Fig. 2). The basalt qualifies in all respects as a typical ENA LTQ basalt. The occurrence of LTQ basalt within the Newark Basin is not surprising in light of

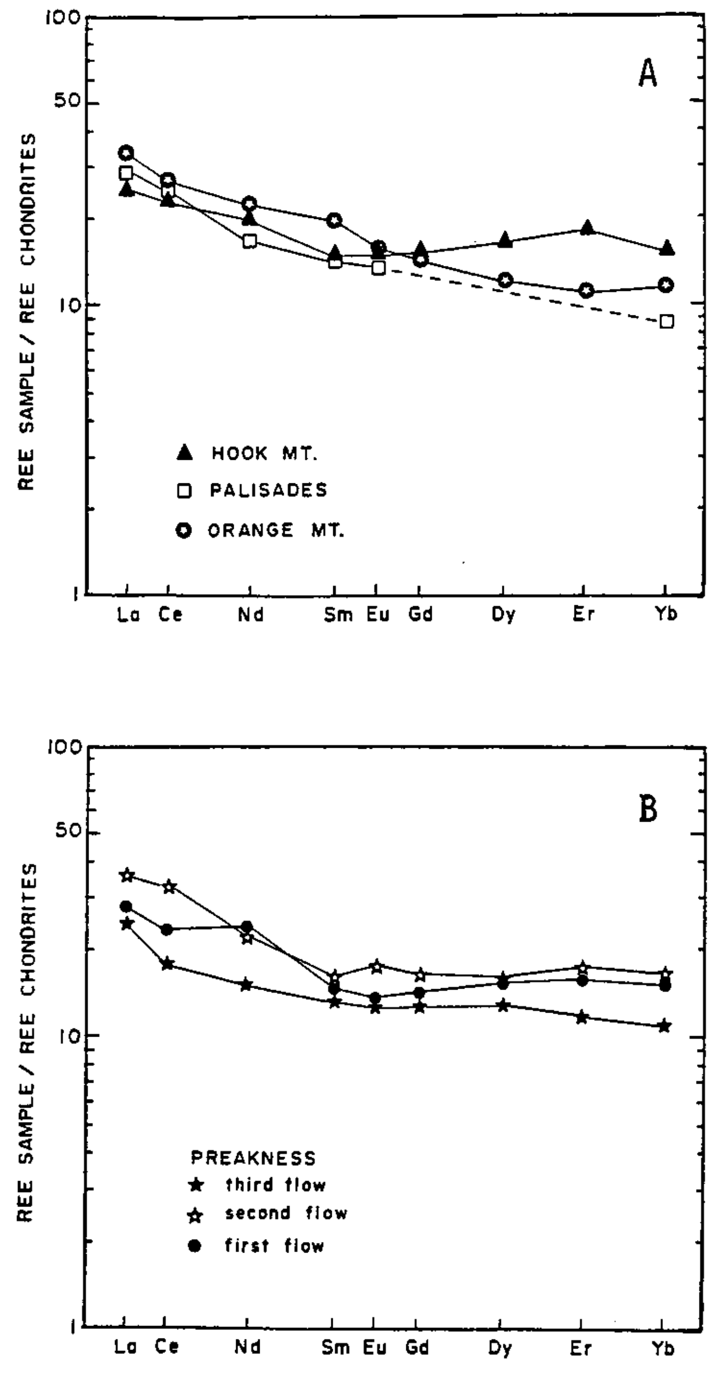

Figure 3. Chondrite normalized (after Masuda and others, 1973) REE distribution of the five typical Watchung Basalt samples of Table 1 and the Palisades sill lower chilled margin (sample 80B1) analyzed by Shirley (1987).

its occurrence in the Gettysburg Basin to the south (Smith and others, 1975) and in the Hartford Basin to the north (Philpotts and Martello, 1986).

The overlapping compositions of the three Preakness Basalt flows in Figure 2 and other variation diagrams support interpretations that all three Preakness flows are genetically related to each other. Apparently, the two lower Preakness flows fractionated out of a LTQ magma, as also seems to have been the case in the Buttress-Ware dike to Holyoke Basalt fractionation trend found in the Hartford Basin (Philpotts and Martello, 1986). The fractionated, more felsic upper portion of a differentiated LTQ magma chamber apparently extruded as the first two Preakness flows before a more mafic, parental pulse of LTQ magma finally was extruded as the uppermost flow.

The REE content of the third flow of the Preakness Basalt is within the "low-Ti" range of Ragland and others (1971) but the first and second flows contain distinctly higher REE concentrations. The distribution patterns (Fig. 3), however, are close to each other and are reasonably parallel, supporting a genetic relationship controlled by fractionation. The slightly positive Eu anomaly displayed by the second flow (Fig. 3) is consistent with a plagioclase enrichment that approximately is balanced by the relatively low plagioclase content of the first Preakness flow.

Hook Mountain Basalt

The Hook Mountain Basalt extruded about 550,000 years after the extrusion of the Orange Mountain Basalt and the intrusion of the Palisades sill (Olsen and Fedosh, 1988). The Hook Mountain Basalt, which chemically is equivalent to the Hampden Basalt of the Hartford Basin, extruded onto 340 m of underlying Towaco Formation. The basalt flowed toward the southwest, unlike the northeast flow directions of the Orange Mountain and Preakness Basalts (Manspeizer, 1969). The Hook Mountain Basalt averages 91 m in thickness (Faust, 1975) and consists of at least two amygdaloidal and deeply altered flows.

The basalt is composed of plagioclase, clinopyroxene, and Fe-Ti oxides in a fine-grained to glassy and typically vesicular mesostasis. Phenocrysts of plagioclase and pyroxene are common.

The SiO_2, Na_2O, Cr, Ni, Rb, and Sr contents of the Hook Mountain Basalt are intermediate between those of the Orange Mountain and Preakness Basalts. Puffer and others (1981), therefore, point out that the Hook Mountain Basalt could not have developed as the result of continuing the same fractionation trend (from an initial Orange Mountain magma) that yielded the Preakness Basalt. Philpotts and Reichenbach (1985) have suggested that the chemically equivalent Hampden Basalt, the uppermost flow unit of the Hartford Basin, may have developed out of HTQ magma via an independent parallel fractionation route. They suggest that a 31 percent fractional crystallization of Talcott magma (a HTQ type chemically equivalent to the Orange Mountain Basalt), forming plagioclase (An_{72}), augite ($Ca_{37}Mg_{58}Fe_5$), orthopyroxene (En_{87}), and olivine (Fo_{90}), would produce a residual liquid of Hampden composition. The fractionating magma was ponded, at least temporarily, in a shallow crustal environment.

The REE distribution pattern of the Hook Mountain Basalt (Fig. 3) plots close to that of the Orange Mountain Basalt and also is within the "high-Ti" range of Ragland and others (1971). The Hook Mountain Basalt, however, contains less light REEs than the Orange Mountain Basalt despite its more highly evolved major-element concentrations (including iron and titanium). These new REE data (Table 1) are consistent with the previously published data of Kay and Hubbard (1978) and Ragland and others (1971). For all three data sets, the fractionated "high-Fe" samples contain less light REE than the more mafic "high-Ti" samples. Further, there also is a distinct "cross-over" in the distribution patterns (Fig. 3), resulting in a higher heavy REE content for the "high-Fe" (Hook Mountain) samples than for the "high-Ti" (Orange Mountain) samples. This cross-over pattern, however, need not be inconsistent with interpretations which include the fractionation of Hook Mountain Basalt from a parental Orange Mountain magma. If assimilation of a light REE-depleted phase or removal of a light REE-enriched phase occurred during fractionation, the observed REE distribution patterns could have been produced. Assimilation of OPX has been proposed by Philpotts and Martello (1986) and by Smith and others (1975), but the light REE versus heavy REE content of the assimilated OPX has not been determined. Alternatively, the Hook Mountain Basalt may have fractionated from a parental LTQ magma along a fractionation path that differs from the one that led to the first flow of the Preakness Basalt.

Conclusions

1. Earlier published conclusions pertaining to the petrology of the Orange Mountain Basalt (Puffer and others, 1981) have held up to rigorous testing. The Orange Mountain Basalt is a "high-Ti" or ENA HTQ basalt, displaying very little compositional or textural variation. Although its chemical composition does not plot within the "continental" or within-plate fields of most tectonic discriminant diagrams, the geochemistry and tectonic setting of the Orange Mountain Basalt and other ENA HTQ rocks are virtually identical to the southern Karroo Basalts of South Africa. Both basalt provinces result from widespread extrusion in the center of Pangaea through a presumably attenuated, uplifted, and "hot" continental crust well before the onset of actual sea-floor spreading (Anderson, 1984). Just as was the case with the Karroo Province, the ENA HTQ magma was probably either a primary melt from a lithospheric source (Hawkesworth and Van Calsteren, 1984; Erlank and others, 1982) or a magma that fractionated from a more mafic parent before breaking through the crust once its density dropped below a threshold value (Cox, 1980).

2. The petrology and geochemistry of the Preakness Basalt is much more diverse than previously described (Puffer and Lechler, 1980). Although most exposures, including the type section, are within the basal flow (a coarse-grained flow averaging 1.2 weight percent TiO_2), the uppermost of probably three flows averages only 0.8 percent TiO_2 and is porphyritic. Mapping of the middle flow is incomplete but preliminary data indicate a highly varied

composition including a TiO_2 content that reaches as high as 1.4 percent. The uppermost Preakness flow is a "low-Ti" or LTQ type ENA basalt. The lower flows presumably were extruded from the fractionated upper levels of a LTQ magma chamber ponded within the crust.

3. The Hook Mountain Basalt is a highly evolved high-Fe type ENA basalt that Puffer and Lechler (1980) conclude cannot be produced by a continuation of the Preakness Basalt fractionation trend. The composition of Hook Mountain Basalt plots on a TiO_2 versus MgO diagram at a position that suggests fractionation out of HTQ magma, such as the Orange Mountain Basalt, along a line distinct from, but parallel to, a line through the Preakness Basalt population. However, although inconclusive, REE data suggest the Hook Mountain Basalt fractionated from a LTQ magma along a fractionation path that differs from one that led to the first flow of the Preakness Basalt.

4. The correlations of Newark Basin volcanics with Hartford Basin volcanics suggested by Puffer and others (1981) are supported by the paleontological work of Olsen (1988), the radiometric work of Sutter (1988), and the petrologic work of Philpotts and Martello (1986). There is little doubt that the same HTQ to LTQ to high-Fe volcanic sequence of the Newark Basin also is found in the Hartford Basin. There also is little doubt that the Orange Mountain and Preakness Basalts of the Newark Basin correlate with the Mount Zion Church and Sander Basalts, respectively, of the Culpeper Basin, Virginia, and perhaps with some widespread African basalt occurrances.

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PROGRESS IN UNDERSTANDING THE PALEOMAGNETISM OF THE PREAKNESS BASALT

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ABSTRACT

The analysis of data from 35 samples, from three flow units in the Preakness Basalt, demonstrates considerable variation in magnetic declination and inclination both within individual flow units and from one flow unit to another. In the lowest flow unit, there appears to be a systematic change in inclination from the base upward, which may correlate with a similar change observed at another locality by McIntosh and others (1985). Variations in inclination and declination within the second flow unit are large and unexplained. Except near its base, the direction of magnetic remanence is uniform in the uppermost flow unit. The magnetic inclination in the uppermost flow unit is much steeper than in the rest of the Preakness Basalt. Similar measurements at 3 sites near the top of the Preakness by McIntosh and others (1985) are interpreted to indicate that paleomagnetic correlation of the uppermost flow unit may be possible.

INTRODUCTION

The major lava flows in the Newark Basin are located at the northwest margin of the basin in northern New Jersey. Collectively, they are referred to as the Watchung Basalts (See figure 1) and are believed to have been extruded between 181 and 199 million years ago (Seidemann and others (1984). The Watchung Basalts are made up of three lava flows, each with multiple flow units. From oldest to youngest, they are designated: Orange Mountain Basalt; Preakness Basalt; and Hook Mountain Basalt. For

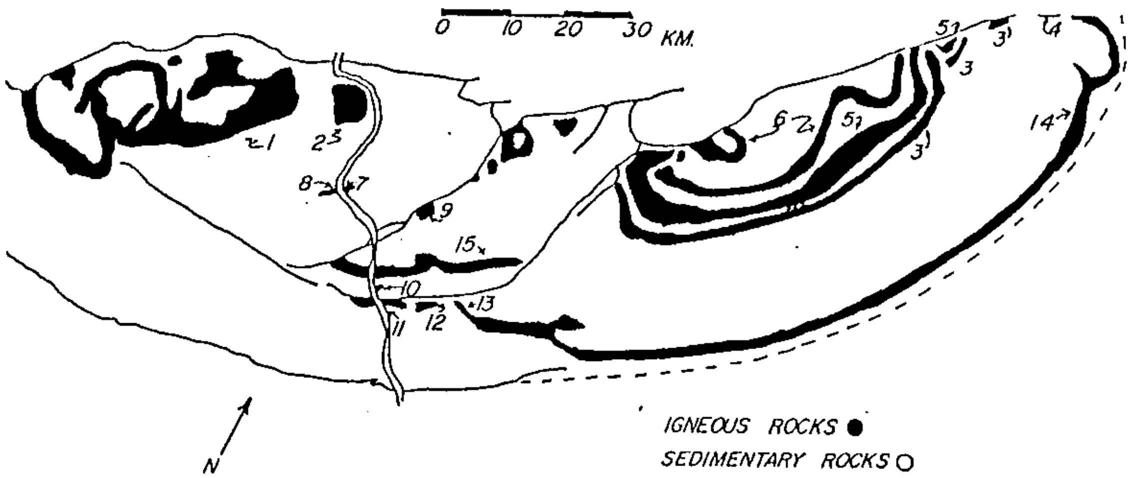


Figure 1: Locations of intrusive and extrusive bodies in the Newark Basin. Key to numbers: 1-Quakertown diabase; 2-Coffman Hill diabase; 3-Orange Mountain Basalt; 4-Ladentown flow; 5-PreaknessBasalt; 6-Hook Mountain Basalt; 7-Byram diabase (NJ localities); 8-Byram diabase (PA localities); 9-Sand Brook flow; 10-Belle Mountain diabase; 11-Baldpate diabase; 12-Pennington Mountain diabase; 13-Rocky Hill diabase; 14-Palisades Sill; 15-Lambertville (Sourland Mountain) diabase (modified from Olsen, 1980).

details of the petrology and geochemistry of these units, see Puffer (1988).

Hozik and Colombo (1984) published a figure showing the virtual geomagnetic poles determined from paleomagnetic studies of Newark Basin rocks (including an error in plotting a pole based on Opdyke's (1961) work). That figure was similar to figure 2, which shows additional data collected for this study. Both versions show that the poles related to the Preakness Basalt are located at longitudes significantly west of those for the other two lava flows and most of the intrusions in the Newark Basin. That observation led to the question of why the Preakness data should show such a departure.

PREVIOUS WORK

The Watchung lavas are such an obvious feature in the Newark Basin that one would expect that they would have been extensively studied. In fact, prior to this study, there were only two published paleomagnetic studies which sampled the Watchung lavas: Opdyke's (1961) study of both igneous and sedimentary rocks, and the more extensive work of McIntosh, Hargraves, and West (1985).

In the following discussion, only samples collected in the Preakness Basalt are considered. Opdyke (1961) collected 18 samples from two sites in the Preakness Basalt as part of his study of paleomagnetism in the Newark Basin. In 1975, McIntosh collected 27 samples from 4 sites in the Preakness Basalt (data from one site were disregarded because of lack of stratigraphic control). Data from the remaining three sites were formally

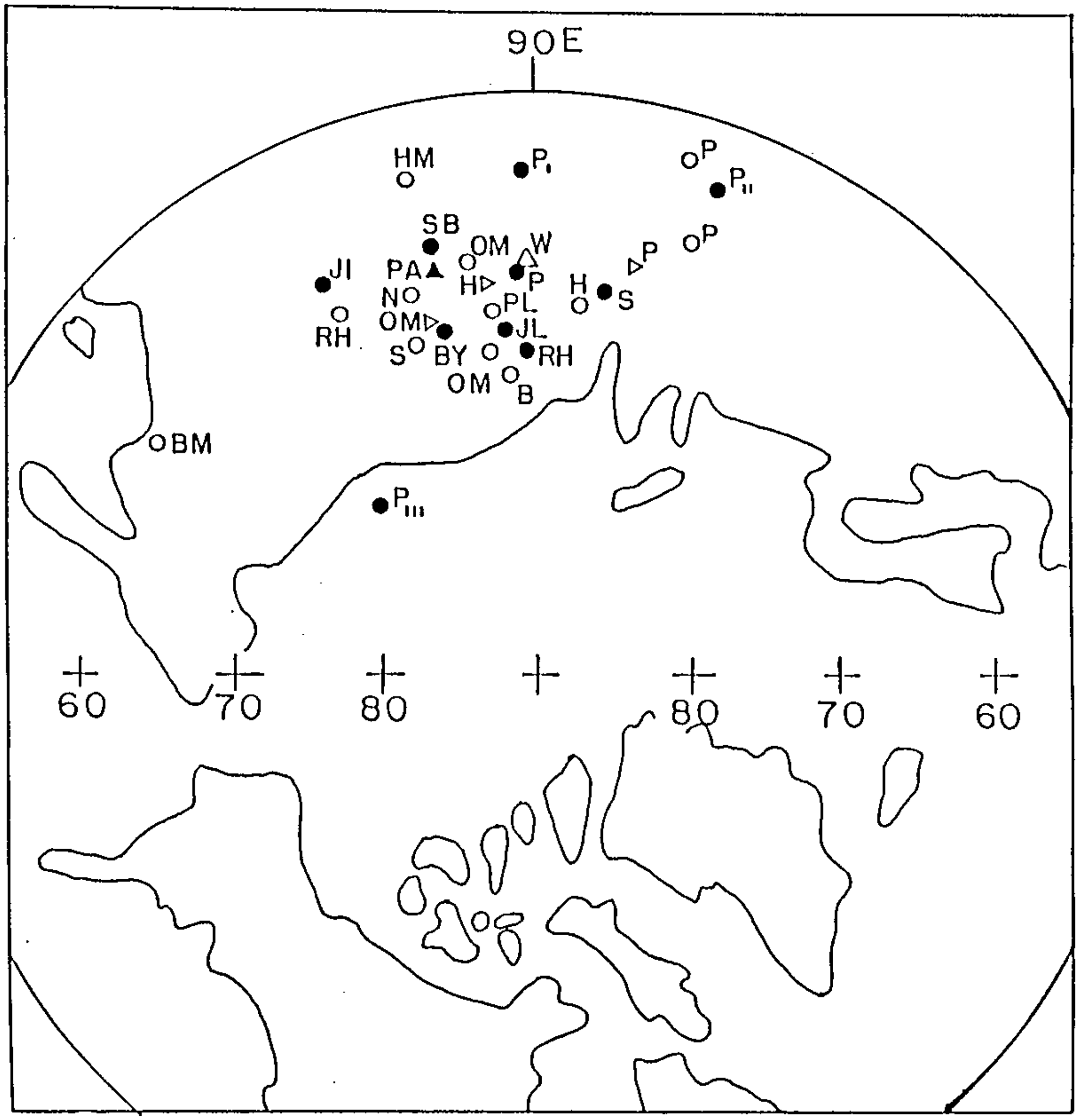


Figure 2. Virtual geomagnetic poles for rocks of the Newark Basin. Solid circles indicate data from Hozik and Colombo (1984), Hozik and Opitz (1988), or Stuck and others (1988). Open circles are from Opdyke (1961). Open triangles are from McIntosh and others (1985). Solid triangle is from Beck (1972). Key to letters: SB-Sand Brook basalt; RH-Rocky Hill diabase; BY-Byram diabase; S-Lambertville (Sourland Mountain) sill; HM-Haycock Mountain diabase; BM-Belle Mountain Diabase; PA-Pennsylvania intgrusives; PL-Palisades sill; OM-Orange Mountain Basalt; P-Preakness Basalt; PI, PII, PIII-first, second, and third flow units of the Preakness Basalt, respectively; JI-Jacksonwald intrusion; JL-Jacksonwald lava flows; H-Hook Mountain Basalt; N-Newark Basin sediments; W-Watchung basalts.

published in 1985 (McIntosh, Hargraves, and West) along with the results of analyses of 32 additional samples from 6 more sites. Virtual geomagnetic poles calculated for each of these 11 sites are shown in Table 1 and plotted on figure 3. The important point is that there is a great deal of scatter in the data from the Preakness Basalt. Because of the extensive scatter, it seemed possible that the divergent virtual geomagnetic poles from the Preakness Basalt may have been due to inadvertent sampling bias, but that explanation begs the question of why there is so much variation in the Preakness data.

Because it is well known that there are multiple flow units within the Preakness Basalt (Manspeizer, 1980; Puffer, 1988), I decided to try to determine how much of the variation in the Preakness results could be due to secular variation from flow unit to flow unit, and how much was variation within individual flow units. John Puffer and Chris Laskowich kindly guided me to localities where individual flow units could be identified.

METHODOLOGY

Samples were collected in each flow unit. Orientation was by means of a Brunton compass, and is probably accurate to within 3 degrees. Each sample was carefully located in terms of stratigraphic height above the base of the flow unit using a measuring tape and a Brunton compass. Samples were spun in a Molespin Mini-spin 1 spinner magnetometer at the University of Massachusetts at Amherst. Pilot samples were step-wise demagnetized (at 2.5, 5, 10, 20, 30, 40, 60 and 80 mT) in a

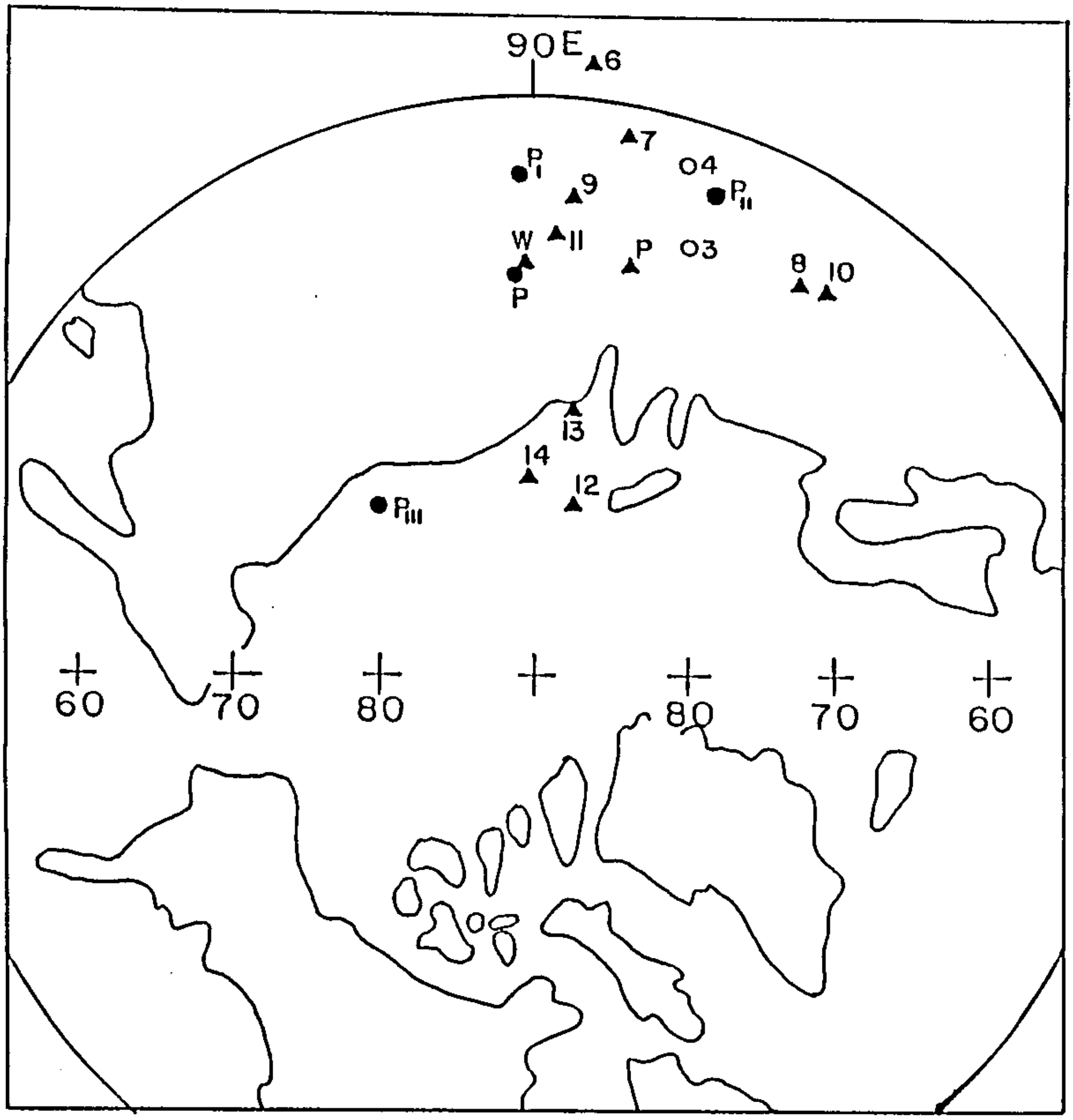


Figure 3. Virtual geomagnetic pole positions for individual sites in the Preakness Basalt. Solid circles indicate data from Hozik and Opitz (1988). Open circles are from Opdyke (1961). Solid triangles are from McIntosh and others (1985). Numbers refer to site numbers. PI, PII, PIII are from the first, second, and third flow units respectively. P is Preakness average. W is Watchung basalts.

TABLE 1: Magnetic data of Opdyke (1961) and McIntosh and others (1985)

SITE	S	N	K	A	D	I	VGP LAT	VGP E LONG	SOURCE
3	3	10	157		18	25	58.3	70.5	Opdyke (1961)*
4	2	8	350		19	14	52.5	73.4	Opdyke (1961)*
6	4	8	383	6.3	14.4	0.8	47.5	84.1	M,H,W (1985)*
7	13	13	40	6.7	15.8	9.9	51.5	79.9	M,H,W (1985)*
8	6	12	337	3.7	25.2	31.1	57.6	56.2	M,H,W (1985)*
9	10	10	273	2.9	11.2	17.3	56.7	84.9	M,H,W (1985)*
10	6	6	18	16.3	26.5	32.2	57.5	53.0	M,H,W (1985)*
11	4	4	402	4.6	9.4	22.6	59.9	86.8	M,H,W (1985)*
12	4	4	230	6.1	6.5	49.2	78.2	76.8	M,H,W (1985)*
13	4	4	357	4.9	8.4	40.3	70.9	81.1	M,H,W (1985)*
14	4	4	99	9.3	3.8	45.8	76.2	91.1	M,H,W (1985)*
MEAN	55	65	21	11.5	13.9	27.9	61.4	76.4	M,H,W (1985)*

Opdyke (1961)* --Pole calculated from data presented in Opdyke (1961)
M,H,W, (1985)* --Pole calculated from data presented in McIntosh and others (1985)

Table 2: Data from the present study

SITE	N	K	A	UNCORRECTED FOR STRUCTURE D	I	VGP LAT	VGP E LONG	CORRECTED FOR STRUCTURE D	I	VGP LAT	VGP E LONG
1	8	41	8.8	11.7	15.0	55.2	85.2	7.8	11.1	54.0	92.5
2	12	9	15.1	14.6	21.1	57.4	78.4	20.6	18.4	53.6	69.9
3	15	36	6.4	7.8	45.1	74.4	78.9	-7.9	45.3	74.5	133.1
MEAN	35	10	8.1					6.7	29.0	64.0	90.9

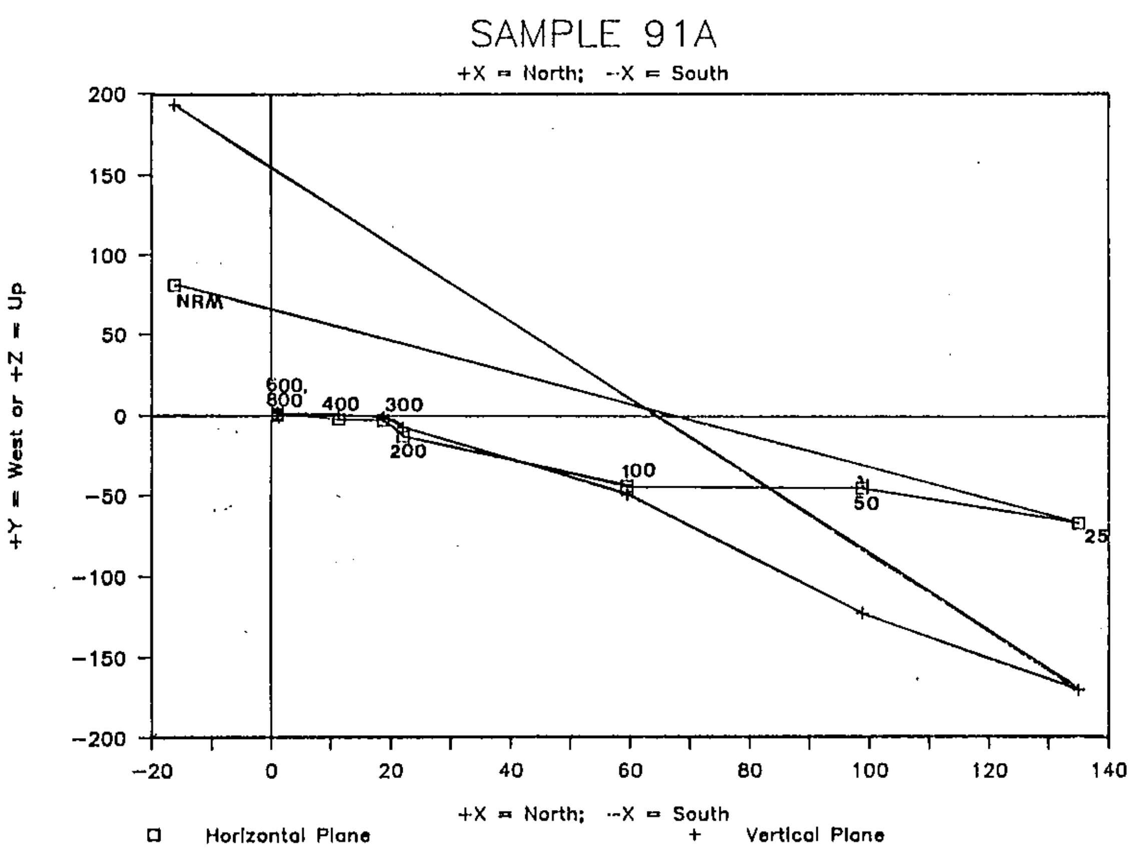


Figure 4. Zijderveld diagram for sample 91A, typical of lowest flow unit.

Schoenstedt alternating field demagnetizer. A few samples were step-wise demagnetized thermally (at 50, 100, 200, 300, 400, 500, 600, and 690 degrees Celsius). Unfortunately, logistical and instrumental problems prevented the determination of complete thermal demagnetization spectra on samples from each flow unit. Remanence remained stable beyond 20 mT in the pilot samples, and all remaining samples were demagnetized at that level.

RESULTS

The lowest flow unit was sampled in an abandoned quarry along Riverview Road in Little Falls. Eight samples yielded satisfactory results. A typical Zijderveld diagram is presented in figure 4. Declination and inclination before treatment, after magnetic cleaning at 20 mT, and after the application of the structural correction determined from the attitude of bedding in the sedimentary rocks below the flow are shown in figure 5. Mean data for the site are shown in Table 2. Changes in declination and inclination as a function of height above the base of the flow are shown in figure 6.

It is interesting to compare these results with those of McIntosh (1975). In a section of the Preakness basalt along I-280, McIntosh sampled a continuous section of more than 50 feet in stratigraphic thickness, and found a systematic change in inclination, and perhaps, declination (See figure 7). In that section, there is no evidence of multiple flow units. Although the Little Falls locality exposes less than 20 feet of section, there is a similar variation in inclination of approximately 25

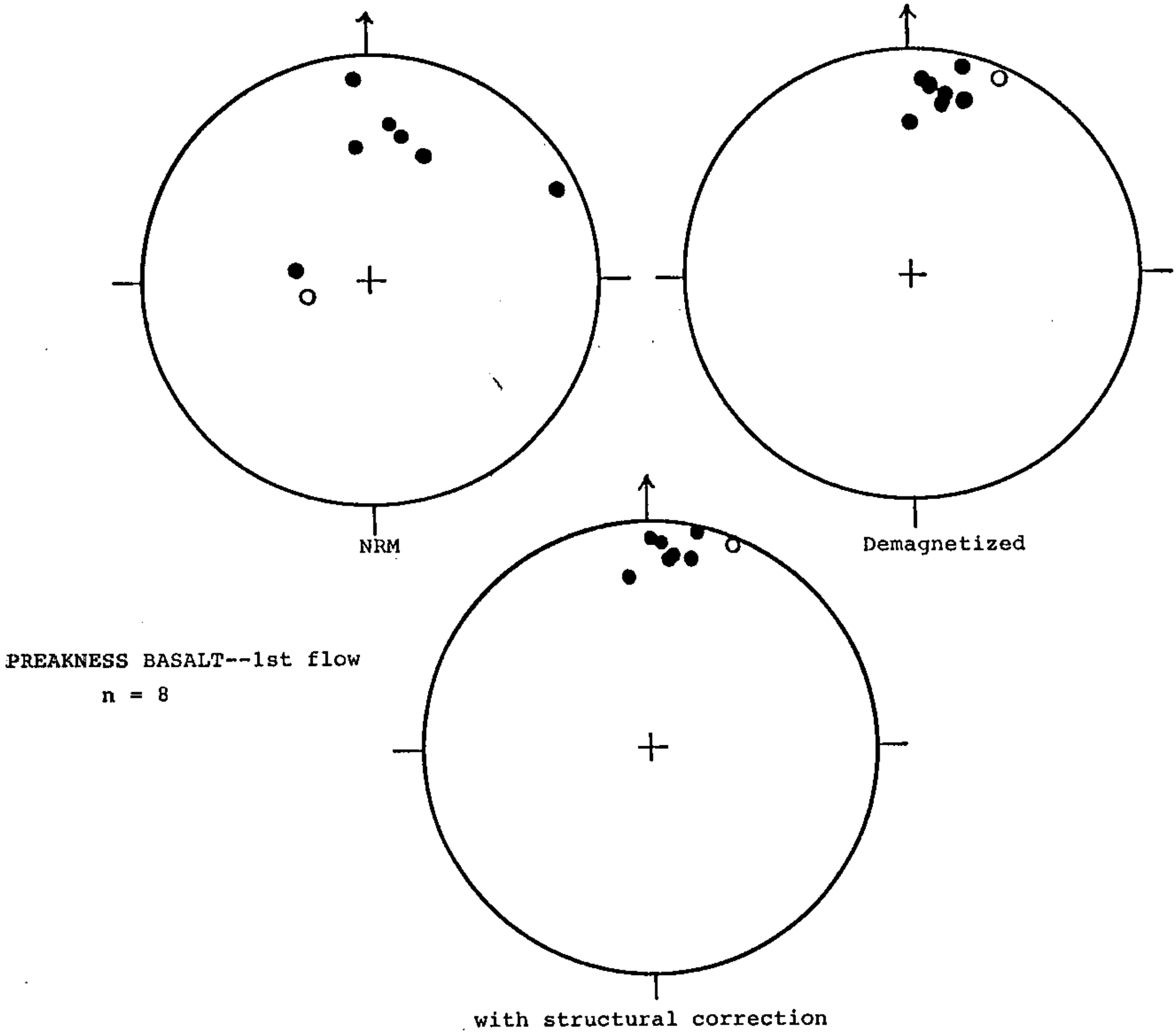


Figure 5. Plots of declination and inclination from individual samples from the first flow unit before treatment, after cleaning at 20 mT, and after the application of the structural correction.

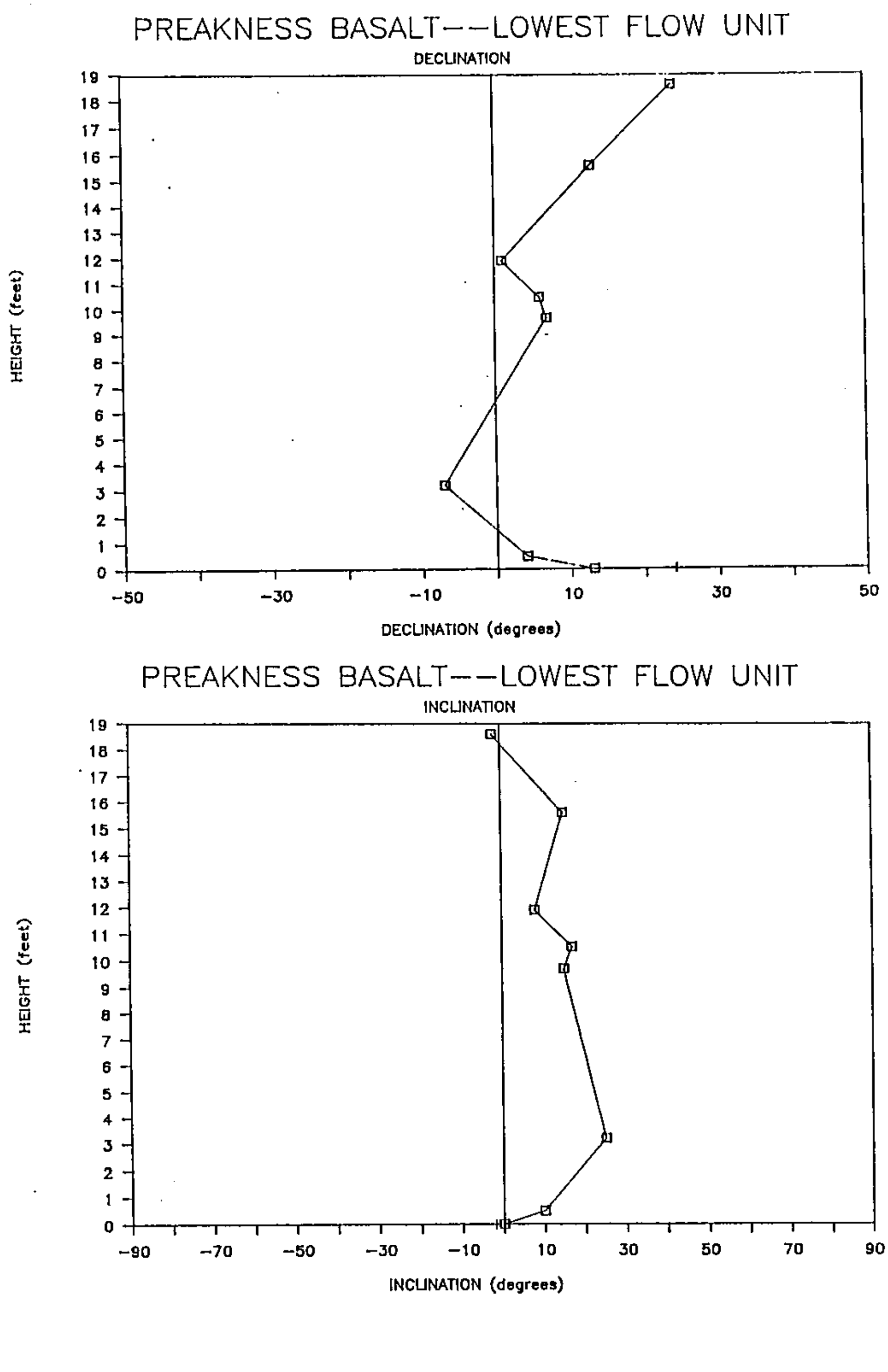


Figure 6. Changes of declination and inclination as a function of height above the base of the first flow unit. Note systematic change in inclination.

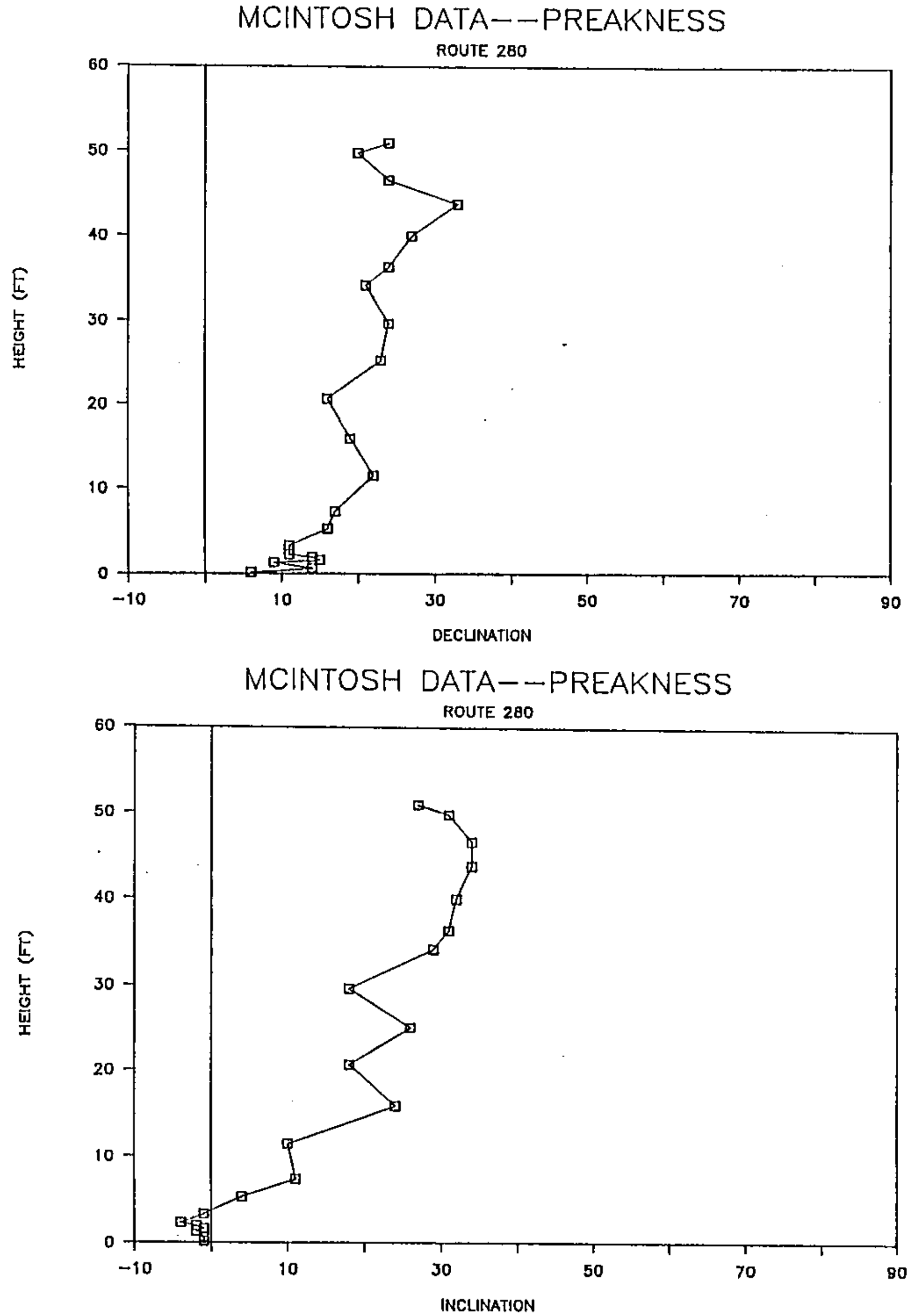


Figure 7. Changes in declination and inclination near the base of the Preakness Basalt found by McIntosh (1975) at sites along I-280.

degrees. This at least suggests that the variation might be continuous along the base of the Preakness. McIntosh and others (1985) suggest that this variation could be due to a secular change. Changes in declination appear to be less systematic, and the correlation between the two sites is not obvious.

The middle flow unit was sampled in an outcrop at the corner of Claremont and Columbus Avenues in Totowa. We were able to expose red sedimentary units at the base of the outcrop, precisely locating the contact. Twelve samples from a 100 foot section of basalt, yielded acceptable results. Zijderveld diagrams for alternating field and thermal demagnetization are shown in figure 8. Declination and inclination before treatment, after magnetic cleaning, and after the application of the structural correction determined from the attitude of bedding in the sedimentary rocks below the flow are shown in figure 9. Mean data for the site are shown in Table 2. Changes in declination and inclination as a function of height above the base of the flow are shown in figure 10.

The second flow unit exhibits more variation than any other flow unit (reflected in an alpha-95 nearly twice that of the other flow units). Neither the variation in declination nor the variation in inclination appears to be systematic. I have no clear explanation for this variation, although we are currently examining the possibility that it is due to changes in mineral chemistry.

The third flow unit was sampled in a road cut and adjacent quarry along Central Avenue in North Caldwell, New Jersey. The

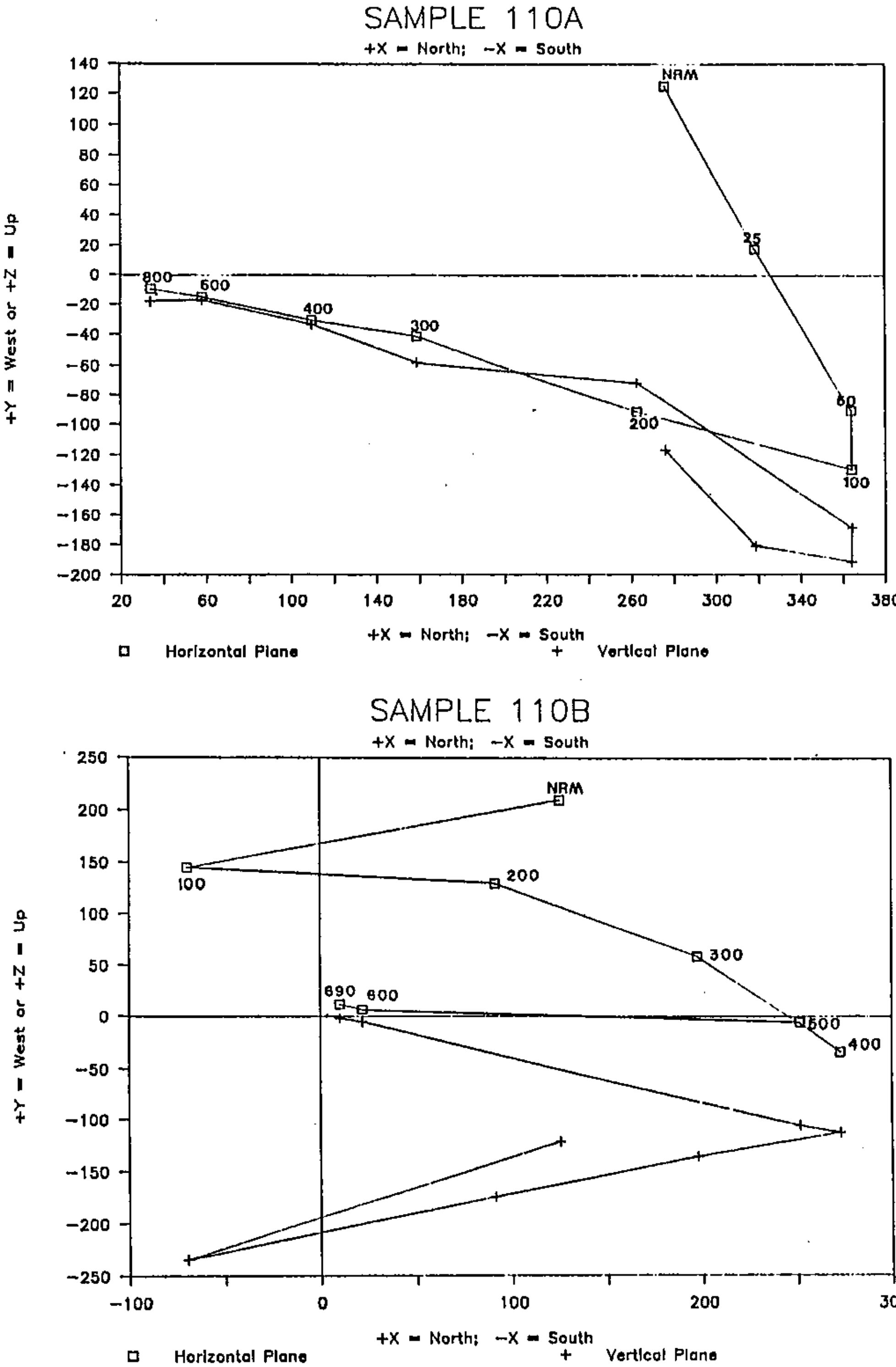


Figure 8. Zijderveld diagrams for samples 110A and 110B, typical of results from the second flow unit. Sample 110A was subjected to alternating field demagnetization. Sample 110B was thermally demagnetized.

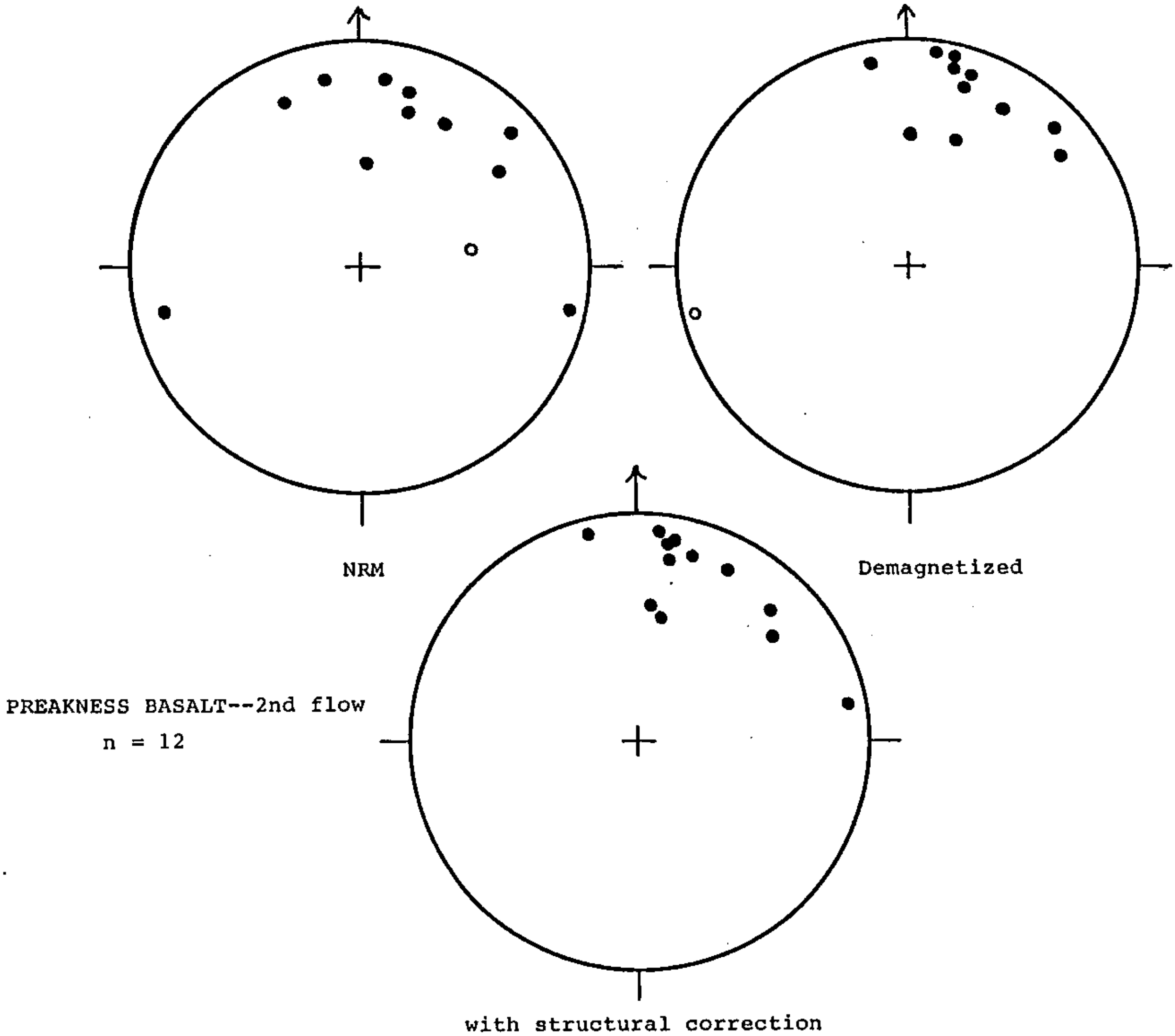


Figure 9. Plots of declination and inclination from individual samples from the second flow unit before treatment, after cleaning at 2Ø mT, and after the application of the structural correction.

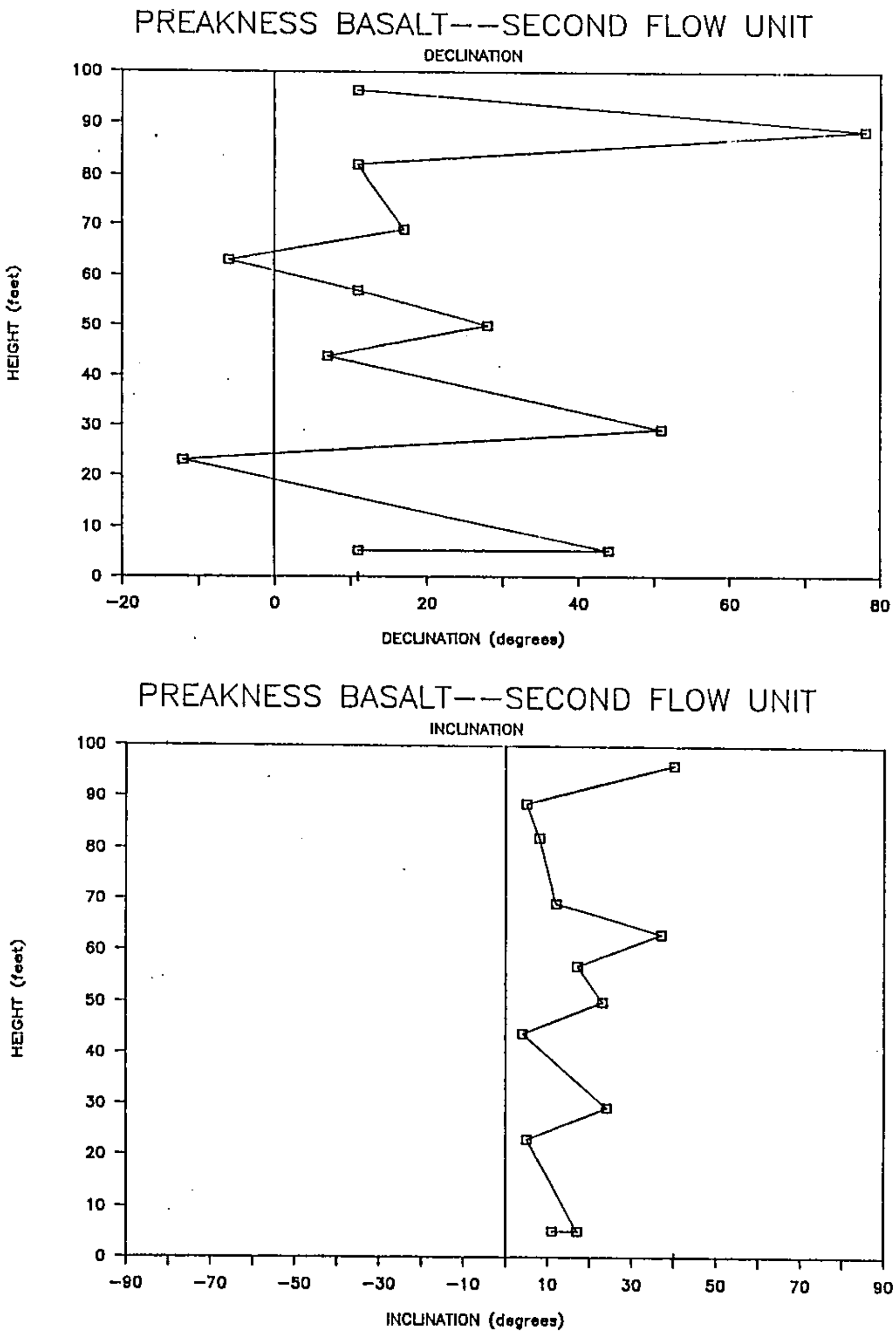


Figure 10. Changes in declination and inclination as a function of height above the base of the second flow unit.

base of the flow unit is exposed in outcrop along the road, enabling the accurate determination of stratigraphic height above the base. Just over 40 feet of section are exposed at this locality. Fifteen samples yielded acceptable results. A typical Zijderveld diagram for alternating field demagnetization is shown in figure 11. Declination and inclination before treatment, after magnetic cleaning, and after the application of the structural correction determined from the attitude of bedding in the sedimentary rocks below the flow are shown in figure 12. Mean data for the site are shown in Table 2. Changes in declination and inclination as a function of height above the base of the flow are shown in figure 13.

Except for the lowest 5 feet of the flow unit, both declination and inclination are relatively constant. The sample 39 feet above the base gives anomalous results for both declination an inclination, which distorts the visual appearance of the graphs. Alpha-95 of 6.4 degrees indicates the consistency of the data. It is important to note that the inclination determined at this locality is markedly steeper than in other flow units within the Preakness. If such an inclination could be found elsewhere at the same stratigraphic level in the Preakness Basalt, it might be possible to correlate the third flow unit paleomagnetically.

DISCUSSION

Again examining figure 1, it is clear that three of the sites sampled by McIntosh and others (1985) have inclinations

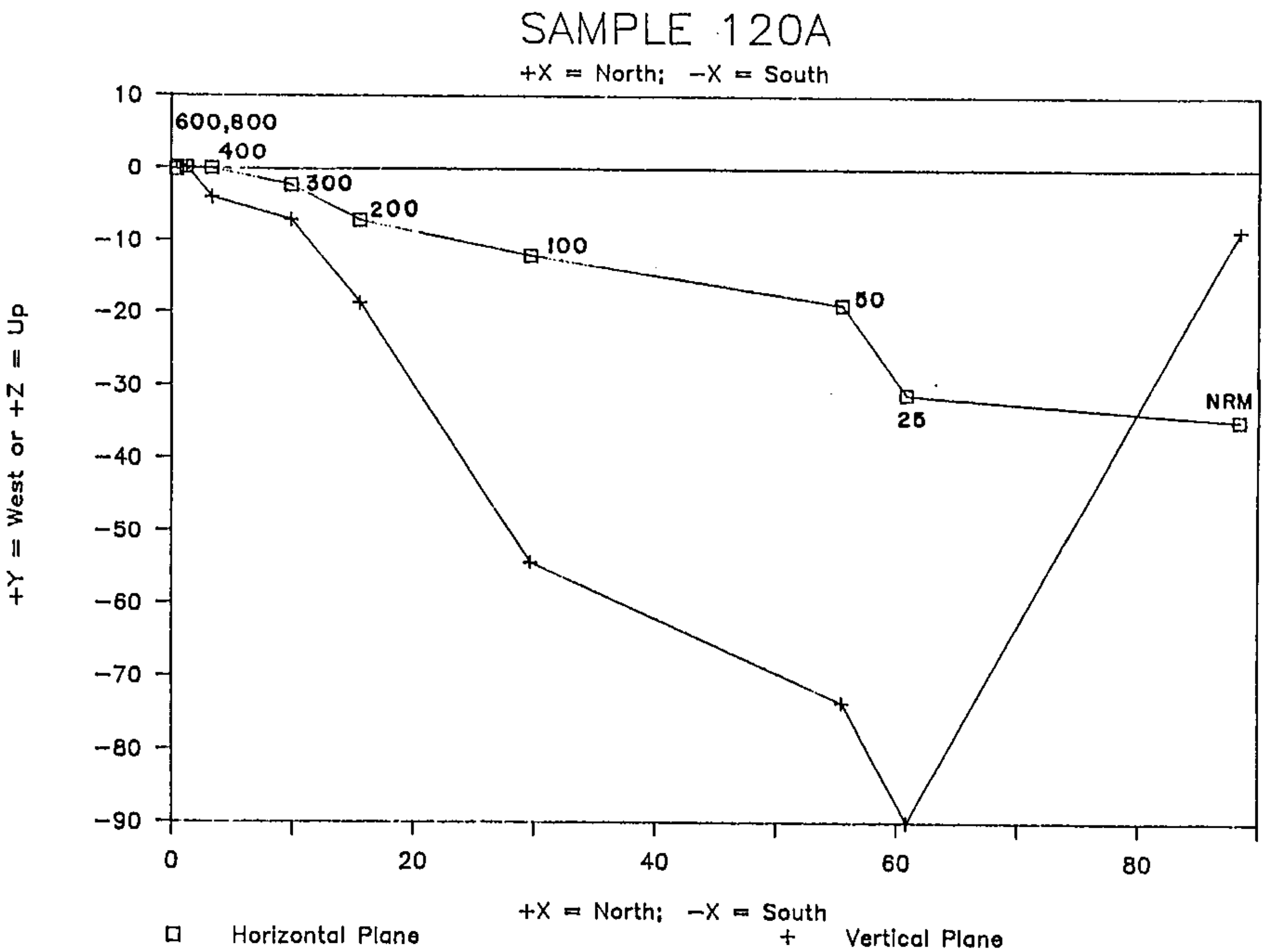


Figure 11. Zijderveld diagram for sample 12ØA, typical of results from the third flow unit.

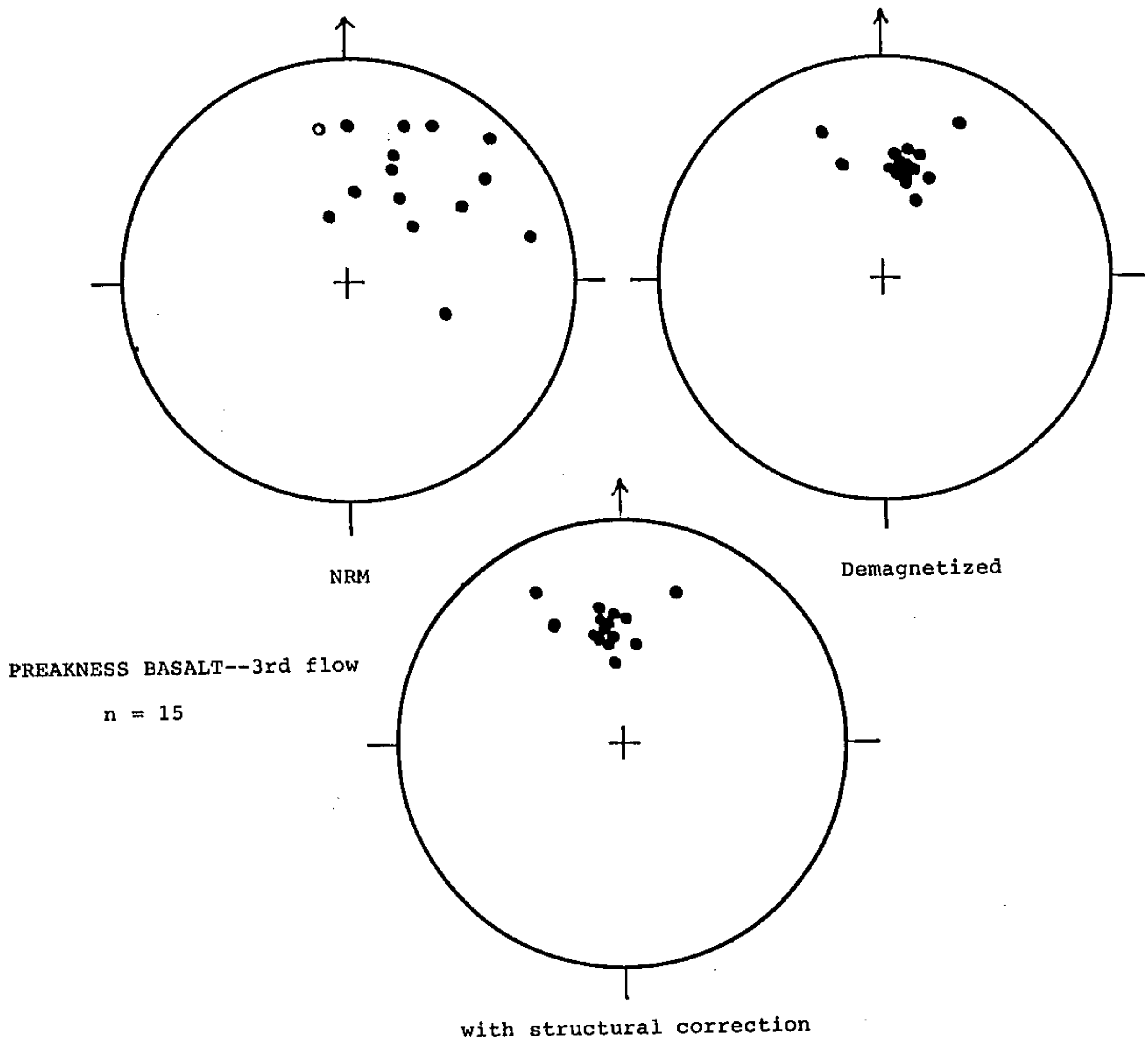
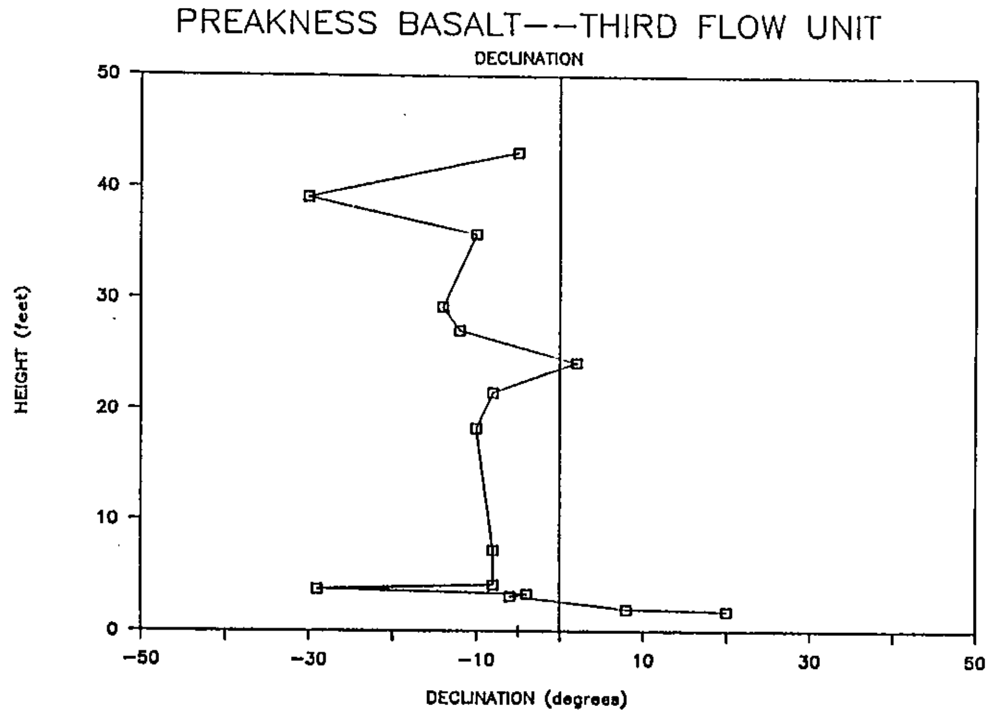


Figure 12. Plots of declination and inclination from individual samples from the third flow unit before treatment, after cleaning at 20 mT, and after the application of the structural correction.



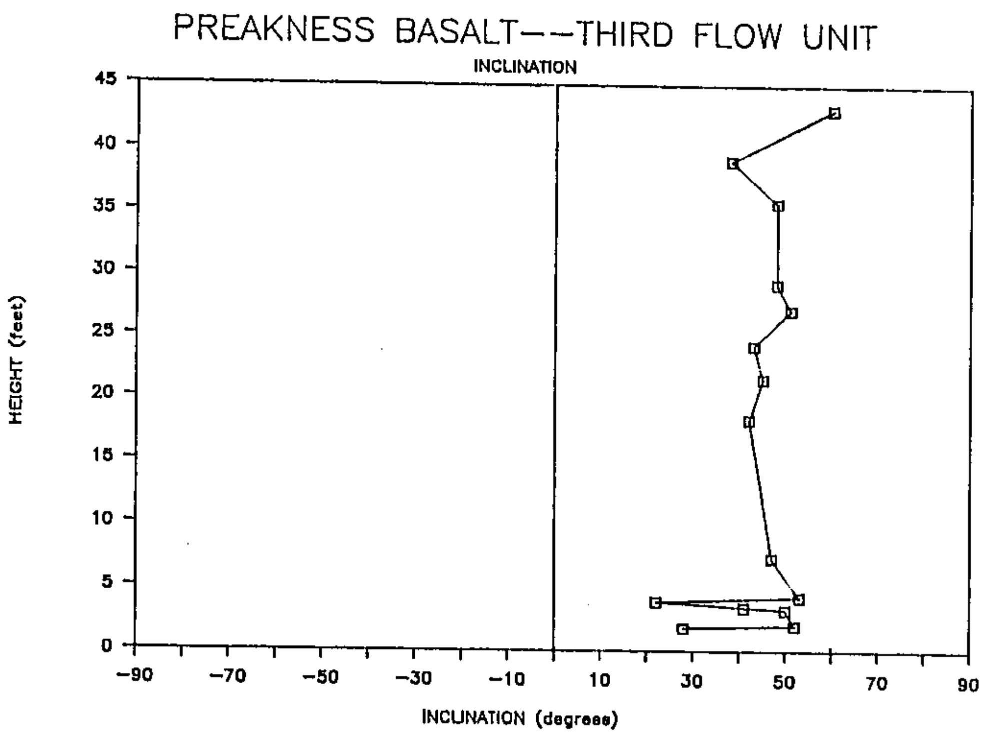


Figure 13. Changes in declination and inclination as a function of height above the base of the third flow unit.

markedly steeper than other Preakness data and consistent with inclinations determined from the third flow unit in the present study. Plotting the locations of McIntosh and others' (1985) localities and Opdyke's (1961) localities on a map (figure 14) shows that at least two, and possibly three of the McIntosh and others (1985) sites are located near the top of the Preakness Basalt. These are sites 12, 13, and 14. Sites 13 and 14 clearly could be in the uppermost flow unit, and depending on its precise location, so could site 12. That steep inclinations are measured near the top of the Preakness Basalt both in its central region (site 3) and near its western end strongly suggests that it might be possible to correlate the uppermost flow unit by means of paleomagnetism.

A comparison of virtual geomagnetic poles for Newark Basin igneous rocks is instructive. Although all Newark Basin igneous rocks plot as a cluster, there is a considerable amount of scatter. Data from the Preakness Basalt show that while the poles determined from data at individual sites show considerable scatter, a simple average of all Preakness samples collected in connection with this study plots in the middle of all the Newark Basin data (64.0 N, 90.9 E). This is almost identical to the average pole of "Watchung Basalts" (63.0 N and 90.1 E) reported by McIntosh and others (1985).

What the data probably indicate is that all of the Newark igneous activity was confined to a relatively short period of time. The lava flows, which cooled quickly, may have recorded short term, secular variations, the most divergent of which are

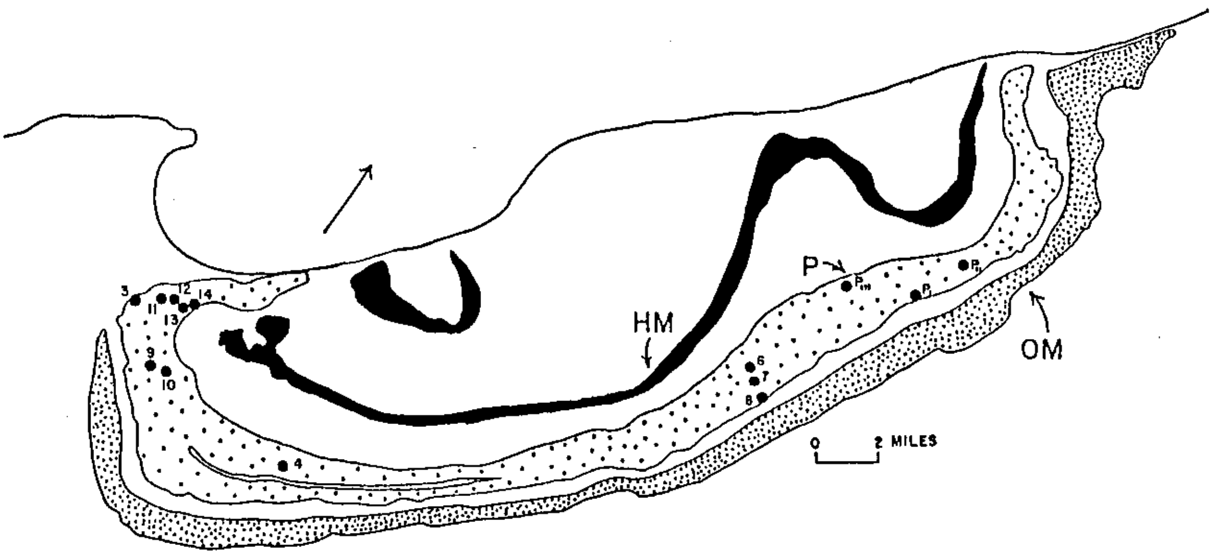


Figure 14. Sketch map of the Watchung basalts showing the locations of samples discussed in this paper. PI, PII, and PIII, are localities sampled for this report. Localities 3 and 4 are from Opdyke (1961). Localities 6, 7, 8, 9, 10, 11, 12, 13, and 14 are from McIntosh and others (1985).

probably preserved in the Preakness Basalt. Given the variability of results, even within one flow unit in the Preakness, it is also clear that care is required in attaching great significance to small variations in declination and inclination. Pole positions determined without regard to a careful sampling of all flow units to average out possible secular variation need to be interpreted with caution.

CONCLUSIONS

While it is clear that we still do not understand the reasons for all of the magnetic variation in the Preakness Basalt, we have made some progress:

1) Studies of variation of inclination and declination within individual flow units show considerable intra-flow variations.
    a) There seems to be a systematic change in inclination upward from the base of the lowest flow unit.
    b) There is a large amount of apparently random variation in the second flow unit, the cause of which is presently unknown.
    c) There is little variation in the uppermost flow unit except very near the base.

2) While the first and second flow units are indistinguishable paleomagnetically, the uppermost flow unit apparently is characterized by a much steeper magnetic inclination. Similar inclinations found by other workers near the top of the Preakness Basalt suggest that it may be possible to trace this flow unit using paleomagnetic methods.

3) When the data used to determine a virtual geomagnetic pole for the Preakness Basalt have enough samples to adequately represent the uppermost flow unit, the pole lies well within the cluster of those determined for other igneous intrusions in the Newark Basin. One may speculate that this means that there was a significant magnetic excursion during Preakness time. Given the short time period suggested by Olsen and Fedosh (1988) for the extrusion of the Watchung basalts (600,000 years), a magnetic excursion during Preakness time must have been very rapid indeed.

4) Given the large amount of variation in the orientation of the remanent magnetism both within flows and between them, it is clear that care must be exercised in interpreting small changes in declination and inclination in these rocks.

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THE ROLE OF SEDIMENTARY CONNATE BRINES IN THE MINERALIZATION OF THE WATCHUNG BASALTS

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ABSTRACT

The distribution of specific minerals and groups of minerals provides evidence about the processes that generated the secondary copper-prehnite-zeolite mineralization in the Watchung basalts. These secondary minerals resulted from reaction between sediment-derived brines and glassy basalt in porous, permeable zones. This was a hydrothermal process that took place after significant burial of the basalts.

INTRODUCTION

In the beginning of this century, the widespread and locally spectacular secondary mineralization in the Watchung basalts was the subject of substantial scientific investigation. However, since the publication of W. T. Schaller's paper in 1932, the study of these minerals has stagnated and been left to amateurs who tend to be good at collecting and observing, but not at interpreting and molding information into a systematic study. Scientific works published in the last fifty-five years which touched on the secondary minerals was adjunct to petrologic or structural studies and was based largely on the work of Schaller.

It is hoped that this paper will demonstrate that the sediments of the Newark basin played a major role in a mineralizing system quite different from the one envisioned by Schaller. It also suggests areas where detailed investigations are needed. A number of papers have been published in recent years, such as Kile and Modreski (1988), and Wilton and Sinclair (1988), concerning both prehnite-zeolite mineralization in basalts and red bed type copper deposits. These works will provide a foundation for future studies.

During the last eighteen years, the author has had frequent opportunities to examine all the operating quarries in the

Watchung basalts and share information with many observant amateur mineral collectors. During this time, observations on the distribution of specific minerals and groups of minerals has provided evidence that the secondary mineralization, for which the Watchung basalts are famous, resulted from the interaction of sediment-derived connate brines with the basalts. The evidence includes the distributions of copper minerals, hydrocarbons and iron sulfides, calcium and sodium zeolites, anhydrite and glauberite, as well as the general abundance of datolite.

COPPER MINERALS

Copper minerals are widespread in the First Watchung (Orange Mountain) basalt along its basal contact with the Passaic formation, in fissures, breccias, and in zones of amygdules and pillows (Woodward, 1944; Sassen, 1978; Peters, 1984; Cummings, 1987). Their mineralogy and mode of occurrence are consistent with the red bed type copper deposit described by Lindgren (1933), Brown (1971), Rose (1976), Lincoln (1981), and Wilton and Sinclair (1987). The copper minerals are an integral part of the general prehnite-zeolite assemblage characteristic of the Watchungs and were not superimposed on it by a separate event. At all localities where copper minerals are present, from Chimney Rock to Prospect Park, distribution follows a similar pattern (Fig. #1).

Native copper is limited to a zone extending a few feet on either side of the First Watchung basalt-Passaic formation contact. It is most abundant as fracture filling and replacement

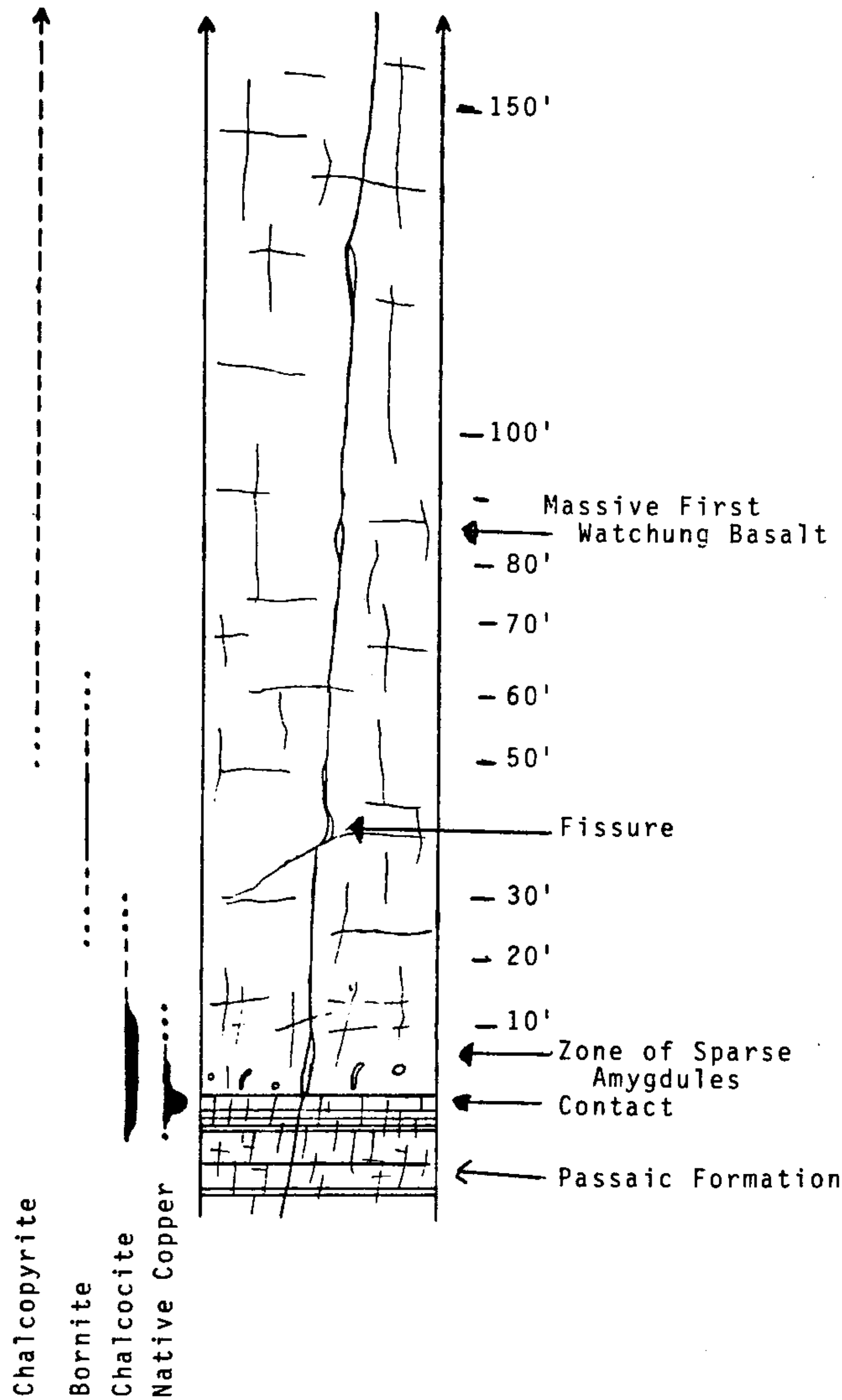

Figure #1. Sketch of copper mineral distribution at the Chimney Rock quarry, Bound Brook, New Jersey.

in the Passaic red beds, but occurs sporadically in fissures, breccias, and amygdules near the base of the basalt.

Chalcocite paragenetically follows native copper. Its zone of deposition completely overlaps that of copper and extends a short distance farther upward into the basalt. Chalcocite was deposited almost exclusively as an open space filling.

As the fissures are followed upward, away from the basalt-mudrock contact, there is a transition from chalcocite to bornite and then from bornite to chalcopyrite. The complete sequence can be seen only at the Chimney Rock quarry, near Bound Brook, but partial sequences are discernable at many localities.

Exposures at the Fanwood and Summit quarries (Cummings, 1987) exhibit the lower two of three amygdaloidal horizons present in the central Watchung syncline. Bornite, often rimmed by chalcopyrite, is characteristic of the lower horizon, while chalcopyrite alone, locally followed by galena and greenockite, is found in the middle horizon.

In the Paterson area (Peters and Peters, 1978), chalcopyrite, sometimes followed by galena and greenockite, is the only copper sulfide found in the amygdaloidal and pillow zones. Rarely, bornite and, even more rarely, native silver, chalcocite, and covellite are found in fissures and crosscutting breccias that probably represent fluid feeders.

The data of Brown (1971), Barnes (1975), Rose (1976), Lincoln (1981), and Wilton and Sinclair (1987) show that in a red bed type system, the order of metal sulfide deposition is a function of decreasing oxidation potential (Eh). In this type of metal-depositing system, oxidized fluids carrying metals as chloride complexes encounter a reduced environment where S^{--} is stable. The addition of increasing amounts of reduced sulfur to the fluids, from sulfate reduction or pyrite replacement, causes deposition of copper (and other metal) sulfides. Native copper is stable in the hematite field and deposits nearest the source of the fluids where Eh is highest. As these fluids penetrate the reduced environment and their Eh shifts toward, and then into, the magnetite field, increases in the availability of reduced sulfur cause chalcocite, then bornite, and finally chalcopyrite to reach saturation and crystallize sequentially at progressively more remote distances from the fluid source. Galena, greennockite, and sphalerite are more soluble than the copper sulfides and deposit at the outer fringe of copper mineralization.

This "textbook" distribution of copper minerals, observed in the First Watchung basalt, suggests that the fluids migrated out of the highly oxidized red beds into fractures in the basalt and then into the permeable, glassy, and reactive amygdaloidal and pillow horizons. The chemical changes in the oxidized fluid, caused by its progressive reaction with basalt, resulted in the zoned distribution of copper minerals and associated sulfides.

HYDROCARBONS, IRON SULFIDES, AND ZEOLITES

At the Millington quarry (Cummings, 1985), the largest exposure of the Third Watchung (Hook Mountain) basalt, localized veins are rich in semisolid petroleum hydrocarbons. Amygdules in the areas surrounding concentrations of hydrocarbon-bearing veins contain abundant pyrite, pyrrhotite, natrolite, and analcime (Fig. #2). In other parts of the quarry, where hydrocarbons have not been observed, iron sulfides are sparse or absent, and the calcium zeolites, stilbite and heulandite, occur in place of sodium zeolites. The transition between areas dominated by natrolite and analcime and those equally rich in stilbite and heulandite is very abrupt and takes place over no more than a few feet.

Although the distribution of hydrocarbons, iron sulfides, and the various zeolites has not been mapped in detail, the broad pattern is quite striking and is known well enough to allow some inferences. The presence of petroleum hydrocarbons is the most convincing evidence seen to date that fluids migrated from the sediments into the basalts. The halo of iron sulfides and sodium zeolites, in those parts of the amygdaloidal horizon surrounding its intersection with hydrocarbon-bearing veins, is thought to be indicative of the migration of fluids rich in H_2S and Na_2SO_4 outward from major feeder conduits through permeable, reactive horizons in the iron and calcium-rich basalt.

ANHYDRITE AND GLAUBERITE

Evaporites were widespread in small amounts in the Newark sediments. Cavities, once occupied by evaporite mineral crystals, are abundant in thin layers--usually less than two feet thick--at many outcrops. These layers probably represent sabkhas marginal to saline playas. The shape of the cavities indicates that glauberite was the dominant mineral in many of these evaporite deposits.

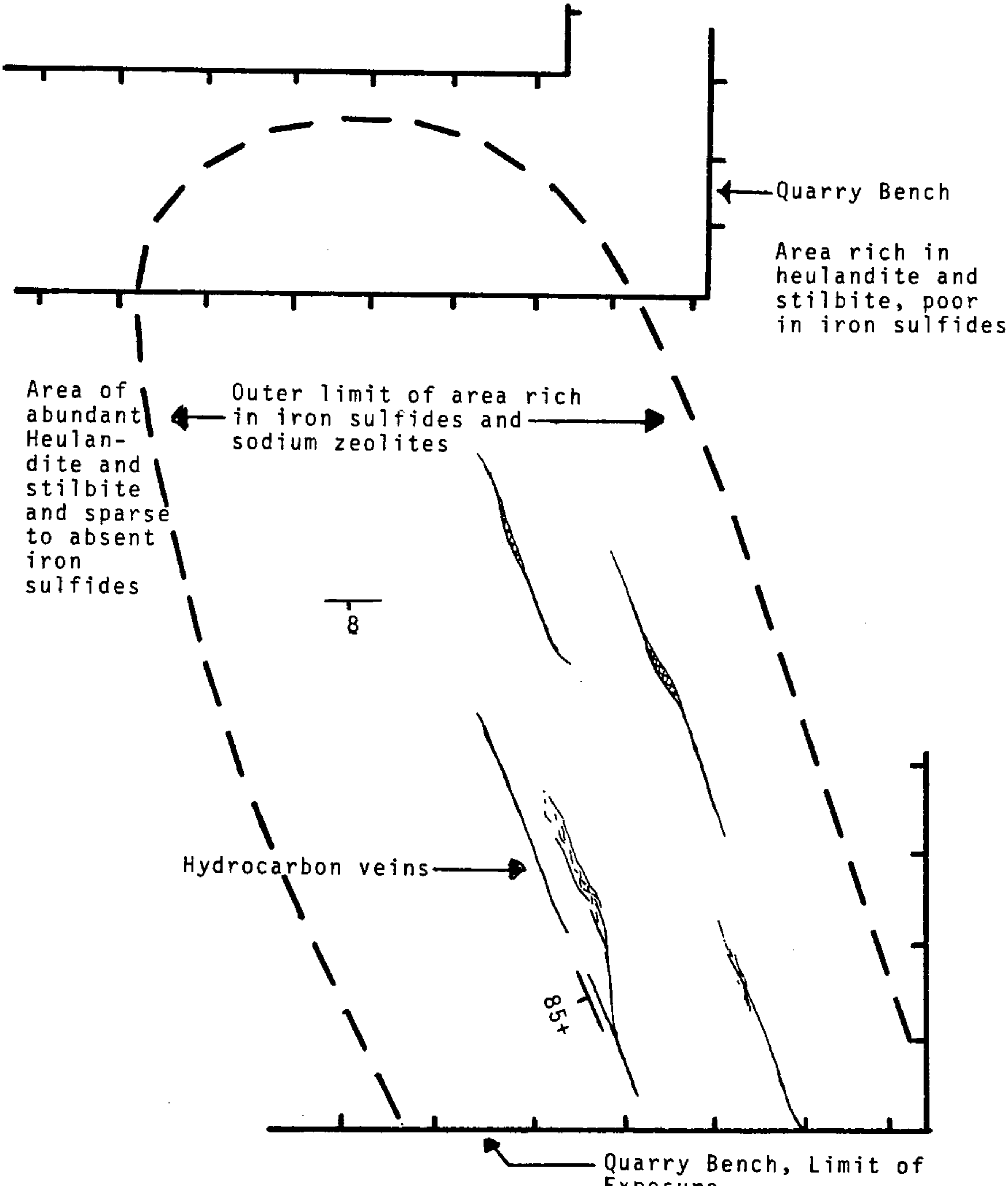

Figure #2. Sketch illustrating the relationship between hydrocarbon-bearing veins and amygdule minerals. For stratigraphic column see Cummings (1985). Based on exposures at the Millington ̂quarru, Bernards Township, Somerset County, New Jersey.

The connate brines, trapped in the Newark sediments, originated in the sulfate-rich playas. If glauberite was the dominant evaporite mineral, then the connate brines probably had a high sodium content. A solution at 35° C in equilibrium with glauberite contains 22.65% Na_2SO_4 and 0.206% $CaSO_4$ (Hill and Wills, 1938). (This dominant solubility of Na_2SO_4 will increase with temperature due to the retrograde solubility of gypsum and anhydrite.)

In contrast to the sediments, the overwhelmingly dominant saline mineral in the basalts is anhydrite. Anhydrite formerly occurred at nearly every locality in the basalts, exhibiting significant secondary mineralization. The characteristic rectangular cavities left by anhydrite are seen in every type of mineralized open space. Glauberite is common in the basalts only in the pillow lavas and an occasional fissure or breccia, sites that apparently had a higher water/rock ratio than normally prevailed.

It is most likely that SO_4^{--} was supplied to the basalt largely as Na_2SO_4. At some point of burial, perhaps 1-1/2 to 2 km, the temperature increased enough for significant reaction to begin between the migrating brines and the glassy basalt. Calcium, released by the basalt, deposited in anhydrite. Glauberite was stable only where the water/rock ratio was relatively high, limiting the amount of calcium available.

The concentration of glauberite in the sediments and anhydrite in the basalts, in addition to the nature of the sites where glauberite is found in the basalts, is best explained by fluid movement from the sediments into the basalts. The occurrence of sulfates in the basalts is not related to the movement of lava into saline lakes as was suggested long ago by Schaller.

DATOLITE

A final possible indicator that the sediments were the source of the mineralizing fluids is the amount of datolite found at many localities. Datolite, $CaBSiO_4OH$, is so common that it is questionable whether the basalt could supply the necessary boron. Datolite contains more than 4,000 times the boron found in normal basalt. The sediments are an attractive alternative source, although their boron geochemistry has never been studied.

Clay-rich mudrocks, graywackes, and shales deposited in marine and some saline lacustrine environments contain far more boron, up to 100-125 ppm, than any other common rock type. Boron is adsorbed on poorly crystallized clays. When the clays recrystallize during the later stages of diagenesis, between about 90^0 and 115^0 C, they release water and boron which become part of the connate brine. It also is possible that the evaporites in the Newark sediments contained soluble borates and their dissolution added boron to the connate water.

In either case, the temperature in the sediments was not sufficient to fix boron in tourmaline, so any boron present would migrate in solution until it encountered the more reactive basalts where large amounts of boron now reside in datolite.

SUMMARY

Since Walker (1951) demonstrated that the zeolites in the Antrim basalts postdated cooling of the plateau lava pile, and Coombs et al. (1959) pointed out that cavity mineralization in fresh basalts is very rare, virtually all studies of zeolite facies mineralization in basalts have concluded that it is a postburial feature resulting from reaction between the basalt and the available ground water. The mineralogy depends on the physical and chemical details of the plumbing system. Although there is a great deal of variation in these details, the general nature of the system is constant. The secondary assemblage in the Watchung basalts reflects its particular circumstances, massive flows with well-developed amygdaloids sandwiched between thick red beds containing saline brines. The combination of red bed copper, zeolite, and evaporite (saline) mineralization is one of the world's best examples of this type of depositional system.

ACKNOWLEDGEMENTS

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A REVIEW OF THE PETROLOGY AND GEOCHEMISTRY OF MESOZOIC DIABASE FROM THE CENTRAL NEWARK BASIN: NEW PETROGENETIC INSIGHTS

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INTRODUCTION

Mesozoic diabase from the central Newark Basin in west-central New Jersey and eastern Pennsylvania occurs in at least seven sheet-like intrusive bodies (Fig. 1). These intrusions, previously studied by Husch and others (1984), Trione and Husch (1985), Eliason and Husch (1986), and Husch and Roth (1988) are, from west to east, the Point Pleasant or Byram diabase, the Stockton diabase, the Lambertville sill, the Belle Mountain diabase, the Baldpate Mountain diabase, the Pennington Mountain diabase, and the Rocky Hill diabase (Fig. 1). Numerous smaller, fine-grained dikes also are present, such as the Solebury dike and the Quarry dike (Fig. 1), the latter studied by Husch and Schwimmer (1985).

With obvious exceptions at the western end of the Rocky Hill diabase and the western and central portions of the Lambertville sill (Fig. 1), all seven intrusions are roughly conformable with enclosing Late Triassic sedimentary rocks (Olsen, 1980) of the Lockatong and Passaic Formations. Although none of the sheets or dikes from the central Newark Basin have been dated isotopically, other evidence indicates they were intruded at essentially the same time (approximately 200 my: Dallmeyer, 1975; Seidemann and others, 1984; Sutter, 1988) as numerous radiometrically dated basalts and diabases from the eastern North America (ENA) Mesozoic basalt province, including the Palisades sill. Based upon sedimentological evidence, the three Watchung Basalts, which, from bottom to top, are the Orange Mountain Basalt,

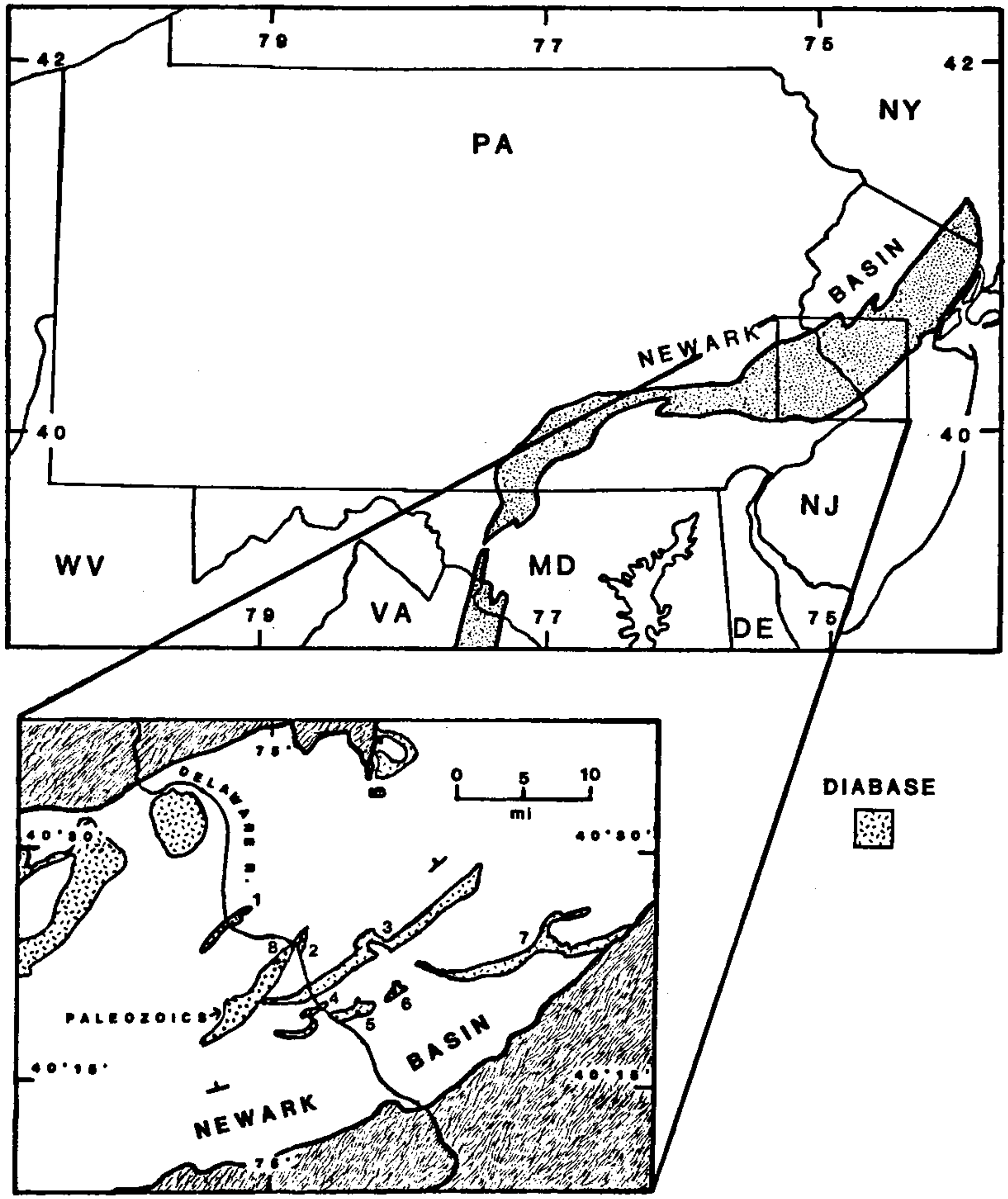


Figure 1. Location of Newark Basin and geologic sketch map of Mesozoic diabase in the Delaware River region. Individually numbered intrusions shown are: 1) Byram (Point Pleasant) diabase; 2) Stockton diabase; 3) Lambertville sill; 4) Belle Mountain diabase; 5) Baldpate Mountain diabase; 6) Pennington Mountain diabase; 7) Rocky Hill diabase; 8) Quarry dike locality.

Preakness Basalt, and Hook Mountain Basalt (Olsen, 1980), were extruded over a span of approximately 500,000 years during Early Jurassic time (Olsen and Fedosh, 1988). Paleomagnetic data from Hozik and Colombo (1984) show that nearly all of the central Newark Basin diabases give an Early Jurassic pole position essentially identical to the one found for the Watchung Basalts and Palisades sill. Thus, the diabases of the central Newark Basin represent a contemporaneous extension of the northern Newark Basin magmatic province, providing a "link" between it and the Mesozoic diabases of the southern Newark Basin and Gettysburg Basin (Smith and others, 1975).

In many respects then, including age, structure, and geochemistry (see discussion below), the central Newark Basin diabase intrusions are very similar to the intensively studied Palisades sill (Walker, 1940; Walker, 1969; Shirley, 1987), located some 20 miles to the northeast. Indeed, the Rocky Hill diabase most probably is a lateral extension of the Palisades sill, while others, such as the Lambertville sill, also may be distal continuations of the Palisades sill re-exposed by subsequent, post-intrusion normal faulting (Lewis, 1907; Bascom and others, 1909; Van Houten, 1969).

PETROGRAPHY

The hypabyssal, largely diabasic nature of the Mesozoic igneous rocks of the central Newark Basin has been known since the early studies of Phillips (1899) and Bascom and others (1909). Whole-rock compositions range from mafic-rich pyroxene cumulates to iron-rich residual fractionates to highly evolved granophyric differentiates. Texturally, the rocks vary from nearly glassy, fine-grained basalts to coarsely crystalline, granular gabbros. Diabasic (subophitic) textures are common in almost all intrusions.

Petrographic observations show that most diabase samples from the central Newark Basin region are composed predominantly of plagioclase (An 50-70), clinopyroxene (augite and pigeonite), orthopyroxene (En 70-80), and Fe-Ti oxides. Accessory minerals include biotite, apatite, quartz, alkali feldspar, sphene, and various opaque sulfides. In highly fractionated granophyric samples, clinopyroxene often is replaced by deuteric hornblende and biotite, plagioclase compositions may be much more sodic (An 20-50) than normal, particularly near grain boundaries, and quartz and alkali feldspar usually occur in graphic micro-pegmatitic intergrowths. Olivine rarely is found in diabase from the region; there are no known olivine accumulation zones as in the Palisades sill. In those few samples from chill margins where olivine is seen, it is present in only small amounts (less than 2 modal percent) and commonly is embayed and anhedral in form, suggesting extensive peritectic reaction with the host basaltic liquid.

On the other hand, accumulations of large (up to 5 mm), euhedral clinopyroxene and/or orthopyroxene are found in porphyritic and/or seriate samples from the Lambertville sill and Point Pleasant diabase. In addition, glomeroporphyritic masses of clinopyroxene occur in almost every chill margin sampled to date;

glomeroporphyritic clinopyroxene is a feature typical of many ENA basalts such as the Orange Mountain Basalt (Puffer, 1988) of the northern Newark Basin and the Talcott Basalt (Philpotts and Martello, 1986) of the Hartford Basin. Microphenocrysts (less than 0.7 mm in length) of plagioclase are seen only in the chill margin from the Point Pleasant diabase and within the nearby Quarry dike. For all chill margin samples, phenocrysts are set in a fine-grained (averages 0.1 mm) intergranular or subophitic groundmass of augite and plagioclase. With few exceptions, groundmass glass is found in only trace amounts. Textures for coarser grained diabase typically are holocrystalline and subophitic or granular with grain sizes ranging from 1 to 6 mm. A preferred alignment of crystals in the coarse-grained rocks is seen only in a few isolated localities and, at best, is poorly developed.

Almost every sample collected exhibits at least some trace amount of alteration, as evidenced by the saussuritization of plagioclase and the uralitization, epidotization, and/or chloritization of pyroxene and olivine. However, with few exceptions, samples were not analysed geochemically if observed alteration effects were extensive (major replacement of orignal minerals). More importantly, no statistically meaningful correlation between whole-rock geochemistry and extent of alteration was found for the samples analysed (see also Husch and Schwimmer, 1985).

TABLE 1

	LAMBERTVILLE SILL												
	ELS11	ELS6	ELS24	ELS5	ELS25	ELS23	ELS10	ELS13	ELS14	ELS15	ELS18	ELS19	ELS21
SiO_2	52.37	52.52	52.62	52.60	52.36	53.39	52.68	52.66	52.88	52.23	53.06	52.92	52.68
TiO_2	0.94	1.14	1.19	1.16	1.13	1.41	1.04	0.95	1.03	1.13	1.01	1.03	1.43
Al_2O_3	11.47	15.20	15.85	14.11	14.15	15.27	12.25	15.23	14.09	13.37	15.07	15.54	15.11
Fe_2O_3	1.92	1.92	1.96	1.97	1.97	2.05	2.23	1.83	2.04	2.01	1.93	1.99	2.27
FeO	8.14	8.14	8.34	8.36	8.36	8.71	9.51	7.79	8.64	8.52	8.19	8.46	9.64
MnO	0.18	0.17	0.17	0.17	0.17	0.17	0.20	0.16	0.18	0.18	0.17	0.17	0.21
MgO	10.66	6.68	5.03	7.81	7.81	5.14	8.27	7.19	7.35	8.42	6.14	5.70	4.72
CaO	11.87	9.03	7.98	10.65	10.63	9.90	11.21	11.30	11.04	11.33	10.77	10.56	8.86
Na_2O	1.80	2.69	4.16	2.13	2.44	2.86	1.83	2.23	2.03	1.95	2.87	2.53	3.60
K_2O	0.53	2.33	2.53	0.86	0.81	0.89	0.64	0.53	0.58	0.73	0.63	0.61	1.31
P_2O_5	0.12	0.18	0.17	0.18	0.17	0.21	0.14	0.13	0.14	0.13	0.16	0.17	0.17
Ba	88	272	259	148	149	178	142	142	146	128	132	147	175
Cr	590	173	22	278	280	49	322	186	104	359	31	22	9
Cu	70	111	128	115	110	131	147	96	118	107	103	109	51
Ni	102	72	56	96	96	56	95	86	87	98	63	60	44
Rb	15	93	95	33	32	34	22	19	22	29	21	19	48
Sc	29	36	29	37	39	33	42	35	39	38	36	35	34
Sr	109	162	300	235	222	227	140	191	177	165	246	201	250
V	167	281	278	257	252	290	214	249	297	267	241	268	331
Zr	34	95	102	107	97	113	98	78	99	85	102	88	82
Q	1.23	0.00	0.00	2.90	1.28	4.93	3.91	3.30	4.46	2.39	2.28	5.03	0.00
OR	3.13	13.77	14.95	5.08	4.79	5.26	3.78	3.13	3.43	4.31	3.72	3.60	7.74
AB	15.23	22.76	30.69	18.02	20.65	24.20	15.49	18.87	17.18	16.50	24.29	21.41	30.46
AN	21.65	22.52	17.10	26.40	25.27	26.20	23.32	29.98	27.62	25.57	26.38	29.25	21.20
NE	0.00	0.00	2.44	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
DI	29.44	17.27	17.66	20.58	21.48	16.18	25.65	20.58	21.45	24.24	21.35	17.29	18.01
HY	24.47	14.07	0.00	21.54	21.15	17.08	22.32	19.38	20.62	21.62	16.89	18.17	14.90
OL	0.00	4.43	11.65	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
MT	2.78	2.78	2.84	2.86	2.86	2.97	3.23	2.65	2.96	2.91	2.80	2.89	3.29
IL	1.79	1.98	2.26	2.20	2.15	2.68	1.98	1.80	1.96	2.15	1.92	1.96	2.72
AP	0.28	0.42	0.39	0.42	0.39	0.49	0.32	0.30	0.32	0.30	0.37	0.40	0.39

	LAMBERTVILLE SILL						POINT PLEASANT				STOCKTON	
	LS2	LS3	LS8	LS9	LS11	LSGRAB	PP1	PP2	PP5	PP6	LQ1	LQ4
SiO_2	52.19	52.63	53.08	52.78	52.76	52.55	52.58	52.42	52.51	52.58	52.82	52.81
TiO_2	0.90	1.11	1.03	0.97	0.90	1.06	1.13	1.11	0.76	1.10	1.00	1.00
Al_2O_3	11.50	14.23	18.09	15.47	14.99	13.44	14.63	14.19	10.29	14.26	14.50	14.27
Fe_2O_3	1.93	1.94	1.69	1.88	1.81	1.93	1.94	1.93	1.87	1.95	2.06	2.11
FeO	8.23	8.26	7.20	7.99	7.70	8.20	8.26	8.22	7.95	8.30	8.77	8.98
MnO	0.17	0.16	0.15	0.17	0.16	0.17	0.16	0.16	0.18	0.16	0.19	0.20
MgO	11.97	8.04	4.65	6.92	7.75	9.05	7.52	8.14	15.39	8.03	6.98	7.15
CaO	10.94	10.93	10.55	10.83	11.22	11.17	10.69	11.00	9.03	10.95	10.64	10.52
Na_2O	1.67	1.99	2.84	2.35	2.25	1.86	2.34	2.08	1.56	1.98	2.35	2.23
K_2O	0.47	0.59	0.58	0.55	0.35	0.54	0.61	0.61	0.35	0.55	0.58	0.60
P_2O_5	0.03	0.12	0.14	0.09	0.11	0.03	0.14	0.14	0.11	0.14	0:11	0.13
Ba	.125	163	152	141	121	157	145	147	90	140	139	146
Cr	470	274	35	91	174	411	232	281	482	278	105	106
Cu	106	104	94	91	98	97	98	105	101	100	109	90
Ni	110	90	55	76	86	98	82	89	122	89	69	70
Rb	13	19	17	19	13	18	21	16	10	18	18	15
Sc	40	36	30	35	36	37	35	36	37	36	40	40
Sr	121	162	218	173	142	149	165	169	109	161	183	171
V	241	260	232	261	245	235	256	264	229	258	283	300
Zr	71	93	96	84	80	87	96	94	65	96	82	82
Q	0.68	3.71	4.03	3.34	3.27	3.00	2.74	2.86	0.00	3.81	3.42	3.80
OR	2.78	3.49	3.43	3.25	2.07	3.19	3.60	3.60	2.07	3.25	3.43	3.55
AB	14.13	16.84	24.03	19.89	19.04	15.74	19.80	17.60	13.20	16.75	19.89	18.87
AN	22.50	28.15	34.90	30.04	29.77	26.73	27.62	27.58	20.04	28.40	27.30	27.16
NE	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
DI	25.44	20.58	13.62	18.88	20.49	23.11	19.99	21.21	19.24	20.36	20.30	19.83
HY	29.90	22.03	15.26	19.83	20.78	23.35	20.97	21.91	40.49	22.19	20.53	21.54
OL	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.55	0.00	0.00	0.00
MT	2.80	2.81	2.45	2.73	2.62	2.80	2.81	2.80	2.71	2.83	2.99	3.06
IL	1.71	2.11	1.96	1.84	1.71	2.01	2.15	2.11	1.44	2.09	1.90	1.90
AP	0.07	0.28	0.32	0.21	0.25	0.07	0.32	0.32	0.25	0.32	0.25	0.30

TABLE 1-CONT.

	BALDPATE MOUNTAIN											
	MS1	MS2	MS3	TMS1	T1	T2	T3	T4	T5	T6	T7	T8
SiO_2	61.29	61.64	61.59	60.79	57.61	52.94	52.93	52.77	53.38	53.27	52.52	51.73
TiO_2	1.63	1.59	1.51	1.70	2.59	1.33	1.00	1.10	1.15	1.03	0.83	3.06
Al_2O_3	12.05	11.85	12.11	11.75	11.72	15.60	15.33	13.67	13.73	15.09	17.13	12.57
Fe_2O_3	2.46	2.41	2.44	2.44	2.81	2.04	1.93	2.00	1.94	1.73	1.61	3.25
FeO	10.46	10.24	10.38	10.38	11.97	8.66	8.19	8.49	8.25	7.37	6.87	13.81
MnO	0.18	0.22	0.22	0.19	0.19	0.19	0.17	0.18	0.19	0.16	0.15	0.24
MgO	1.09	0.96	1.00	1.23	1.75	5.32	6.51	8.20	7.75	7.13	6.08	3.51
CaO	4.53	4.80	4.50	5.13	5.51	10.46	10.93	10.46	10.46	11.20	11.18	6.70
Na_2O	3.57	3.38	3.34	3.76	3.21	2.48	2.26	2.28	2.33	2.33	2.87	3.19
K_2O	2.29	2.41	2.42	2.14	2.11	0.80	0.62	0.71	0.69	0.55	0.65	1.72
P_2O_5	0.45	0.50	0.49	0.49	0.53	0.18	0.13	0.14	0.13	0.14	0.11	0.22
Ba	451	478	508	454	450	231	180	192	140	163	160	353
Cr	9	9	12	9	12	59	74	306	317	276	76	4
Cu	325	350	258	325	271	163	130	103	142	111	91	221
Ni	18	18	17	10	14	59	80	101	96	86	66	28
Rb	66	69	73	70	80	28	25	32	32	24	17	55
Sc	21	19	20	25	32	36	36	36	36	33	24	38
Sr	142	142	150	160	165	219	194	171	168	171	224	227
V	23	22	25	38	139	293	283	273	298	261	200	570
Zr	306	279	296	312	301	124	124	95	110	105	82	203
Q	16.99	18.07	18.11	15.53	13.20	4.65	4.22	2.49	3.75	4.03	0.91	3.04
OR	13.53	14.24	14.30	12.65	12.47	4.73	3.66	4.20	4.08	3.25	3.84	10.16
AB	30.21	28.60	28.26	31.82	27.16	20.99	19.12	19.29	19.72	19.72	24.29	26.99
AN	10.09	10.05	10.90	8.86	11.34	29.07	29.86	24.97	24.97	29.09	31.94	14.90
NE	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
DI	8.21	9.15	7.16	11.61	11.47	17.87	19.28	21.18	21.25	20.81	18.63	14.33
HY	13.26	12.22	13.72	11.63	14.18	16.79	18.87	22.56	20.94	18.32	16.23	19.54
OL	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
MT	3.57	3.49	3.54	3.54	4.07	2.96	2.80	2.90	2.81	2.51	2.33	4.71
IL	3.10	3.02	2.87	3.23	4.92	2.53	1.90	2.09	2.18	1.96	1.58	5.81
AP	1.04	1.16	1.14	1.14	1.19	0.42	0.30	0.32	0.30	0.32	0.25	0.51

	BALDPATE MOUNTAIN		ROCKY HILL			PENNINGTON MOUNTAIN				BELLE MOUNTAIN		
	T9	T10	KQ6	KQ7	KQ8	PQ2	PQ3	PQ4	PQ6	WH1	WH2	WH3
SiO_2	52.77	52.95	53.08	52.90	53.43	53.89	53.76	61.13	54.86	52.99	53.44	52.90
TiO_2	1.10	1.12	1.06	0.97	1.09	1.84	1.65	1.52	2.40	1.37	1.32	1.30
Al_2O_3	16.09	16.21	17.01	15.82	16.70	14.55	15.73	12.11	12.10	16.40	16.85	16.32
Fe_2O_3	1.84	1.92	1.85	1.88	1.82	2.50	2.25	2.47	3.13	2.13	1.82	1.98
FeO	7.84	8.16	7.86	8.01	7.75	10.64	9.59	10.53	13.33	9.08	7.73	8.43
MnO	0.16	0.18	0.16	0.17	0.15	0.18	0.18	0.26	0.27	0.17	0.16	0.19
MgO	5.72	5.40	5.12	6.54	5.29	3.56	3.67	1.05	2.04	4.07	5.19	5.63
CaO	9.25	10.32	10.37	10.43	10.37	8.51	9.13	4.74	5.68	8.37	9.92	7.33
Na_2O	2.83	2.89	2.65	2.33	2.58	2.84	2.88	3.23	3.77	3.57	2.86	4.46
K_2O	2.25	0.71	0.72	0.83	0.70	1.28	0.96	2.49	1.87	1.74	0.60	1.33
P_2O_5	0.15	0.14	0.12	0.12	0.12	0.21	0.20	0.47	0.55	0.11	0.11	0.13
Ba	248	142	169	164	175	280	236	520	440	281	174	303
Cr	44	26	43	87	48	22	22	24	20	27	62	94
Cu	134	123	103	91	99	200	195	178	210	168	117	119
Ni	64	61	57	71	57	39	43	19	21	45	64	66
Rb	99	30	20	27	20	41	25	79	54	53	18	36
Sc	32	33	31	34	31	34	33	21	30	31	35	34
Sr	316	205	192	187	191	195	198	147	173	225	221	204
V	269	314	233	224	231	331	295	29	68	271	257	266
Zr	108	107	89	83	104	161	141	323	208	109	115	108
Q	0.00	2.80	4.12	3.48	4.92	6.61	6.41	17.46	6.75	0.00	4.65	0.00
OR	13.30	4.20	4.26	4.91	4.14	7.56	5.67	14.72	11.05	10.28	3.55	7.86
AB	23.95	24.45	22.42	19.72	21.83	24.03	24.37	27.33	31.90	30.21	24.20	37.74
AN	24.56	29.16	32.39	30.26	31.92	23.17	27.16	11.19	10.57	23.59	31.37	20.58
NE	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
DI	16.72	17.40	15.06	16.95	15.42	14.88	14.10	8.06	12.08	14.37	14.10	12.29
HY	13.63	16.75	16.77	19.85	16.78	16.14	15.42	13.69	17.29	14.71	17.11	4.50
OL	2.75	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.90	0.00	11.38
MT	2.67	2.78	2.68	2.73	2.64	3.62	3.26	3.58	4.54	3.09	2.64	2.87
IL	2.09	2.13	2.01	1.84	2.07	3.49	3.13	2.89	4.56	2.60	2.13	2.47
AP	0.35	0.32	0.28	0.28	0.28	0.49	0.46	1.09	1.27	0.25	0.25	0.30

TABLE 1-CONT.

LAMBERTVILLE SILL-NORTHEAST REGION

	LSR1	LSR2	LSR3	LSR4	LSR5	LSR6	LSR7	LSR12	LSR13	LSR14	*LSR8	*LSR9	*LSR10	*LSR11
SiO_2	52.03	51.76	52.42	51.42	52.09	51.48	52.46	52.39	52.26	52.37	51.74	51.93	51.58	52.47
TiO_2	1.07	1.00	1.10	0.98	0.95	0.90	0.98	0.98	1.52	1.17	0.80	0.77	0.80	0.88
Al_2O_3	13.73	13.19	14.08	12.93	11.61	11.43	13.42	13.38	14.47	14.66	14.48	14.55	15.26	16.04
Fe_2O_3	1.95	1.90	1.93	1.93	2.00	2.00	1.87	1.96	2.36	1.93	1.84	1.87	1.83	1.75
FeO	8.28	8.07	8.22	8.22	8.50	8.51	7.97	8.35	10.05	8.19	7.84	7.93	7.78	7.46
MnO	0.19	0.19	0.18	0.19	0.19	0.20	0.19	0.20	0.20	0.18	0.19	0.18	0.19	0.17
MgO	9.13	9.90	8.41	10.05	10.66	12.41	9.66	9.31	4.91	7.68	8.82	8.32	8.04	6.04
CaO	10.97	11.42	11.03	11.41	11.47	10.88	11.09	11.01	10.35	10.91	11.46	11.61	11.40	11.81
Na_2O	2.02	1.93	1.83	2.20	1.91	1.63	1.90	1.86	2.66	2.15	2.23	2.25	2.50	2.74
K_2O	0.52	0.53	0.66	0.56	0.50	0.45	0.55	0.44	1.05	0.62	0.50	0.49	0.54	0.55
P_2O_5	0.11	0.11	0.14	0.10	0.08	0.06	0.11	0.12	0.17	0.14	0.10	0.10	0.08	0.09
Ba	125	128	138	120	103	123	122	126	223	154	122	112	133	133
Cr	442	578	378	588	602	670	431	374	8	258	234	209	179	22
Cu	96	87	100	85	89	83	89	97	128	107	87	85	82	80
Ni	103	107	93	111	119	132	104	103	36	87	85	84	78	53
Rb	11	16	12	14	21	3	19	18	60	15	16	20	16	23
Sc	37	40	35	39	41	42	41	39	36	37	37	39	36	32
Sr	184	143	189	135	106	111	179	186	254	208	132	143	142	156
V	256	267	264	257	254	238	266	276	292	266	279	290	252	255
Zr	90	82	88	80	63	75	85	90	139	104	79	79	81	61
Q	1.73	0.66	3.52	0.00	0.53	0.00	2.08	2.90	2.78	3.00	0.10	0.69	0.00	1.39
OR	3.07	3.13	3.90	3.31	2.95	2.66	3.25	2.60	6.21	3.66	2.95	2.90	3.19	3.25
AB	17.09	16.33	15.49	18.62	16.16	13.79	16.08	15.74	22.51	18.19	18.87	19.04	21.15	23.19
AN	26.86	25.76	28.26	23.75	21.63	22.54	26.47	26.86	24.44	28.52	28.03	28.16	28.82	29.85
NE	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
DI	21.78	24.41	20.77	26.06	28.11	25.01	22.53	21.89	21.46	20.12	22.87	23.41	22.14	23.07
HY	24.35	24.79	22.86	19.98	25.69	29.85	24.98	25.03	15.91	21.17	22.77	21.41	18.43	14.85
OL	0.00	0.00	0.00	3.40	0.00	1.34	0.00	0.00	0.00	0.00	0.00	0.00	1.91	0.00
MT	2.83	2.75	2.80	2.80	2.90	2.90	2.71	2.84	3.42	2.80	2.67	2.71	2.65	2.54
IL	2.03	1.90	2.09	1.86	1.80	1.71	1.86	1.86	2.89	2.22	1.52	1.46	1.52	1.67
AP	0.25	0.25	0.32	0.23	0.19	0.14	0.25	0.28	0.39	0.32	0.23	0.23	0.19	0.21

TABLE 1. Major- and trace-element and CIPW normative compositions of diabase from the central Newark Basin. Major elements normalized to 100 percent anhydrous. Fe2O3/FeO ratio set at .235. First set of Lambertville sill data is from the Delaware River area and the second set is from the sill's northeast end in the area of Sourland Mountain (Husch and Roth, 1988). *LTQ-like samples.

WHOLE-ROCK GEOCHEMISTRY

Major- and trace-element data

Major- and trace-element whole-rock compositions, determined by direct current plasma atomic emission spectroscopy (Feigenson and Carr, 1985), and calculated CIPW norms for all analysed diabase samples collected from the seven sheet-like intrusions of the central Newark Basin are presented in Table 1. These data represent an updated compilation of analyses presented by Husch (1988), and in some instances include new analytical results. Quarry dike analyses (Husch and Schwimmer, 1985) are presented in Table 2. All major-element values in Tables 1 and 2 have been normalized to 100 percent anhydrous with the Fe_2O_3/FeO ratio set at 0.235. This value represents the average for Early Jurassic igneous rocks of similar composition and occurrence from the Newark and Gettysburg Basins (Smith and others, 1975; Puffer and Lechler, 1979; Puffer and others, 1981). Original major-element (all Fe as FeO) plus loss-on-ignition (LOI) totals range between 97.2 and 102.6 weight percent with LOI values between 0.82 and 2.17 weight percent. No correlation between LOI values and alkali content is found.

In general, diabase compositions (Table 1) are quite variable. They include: 1) high magnesium (>9% MgO) diabase; 2) less mafic (<9% MgO) diabase close in bulk chemistry to the quartz normative, high-titanium tholeiite (HTQ) of Weigand and Ragland (1970) and the York Haven tholeiite of Smith and others (1975); 3) a few less mafic (<9% MgO) diabase samples possibly

TABLE 2
QUARRY DIKE ANALYSES

Sample	1	2	3	4	5	6	7	8	9	10
SiO_2	54.46	52.84	52.67	52.77	52.97	52.75	52.91	52.96	52.63	52.85
TiO_2	1.27	1.15	1.11	1.12	1.13	1.13	1.09	1.07	1.08	1.06
Al_2O_3	16.37	14.01	14.10	14.03	14.41	14.45	14.34	14.23	14.30	14.24
Fe_2O_3	1.62	1.97	1.95	1.98	1.98	1.98	1.97	1.97	1.98	1.94
FeO	6.88	8.40	8.28	8.43	8.43	8.39	8.41	8.37	8.42	8.31
MnO	0.02	0.17	0.16	0.17	0.16	0.17	0.17	0.17	0.18	0.17
MgO	12.16	7.97	7.94	7.99	7.88	7.90	7.97	8.08	8.36	8.27
CaO	3.07	10.50	10.54	10.28	9.57	8.30	8.85	8.93	7.84	7.89
Na_2O	2.58	1.90	2.02	1.99	1.96	2.45	2.10	2.05	2.61	2.73
K_2O	1.41	0.96	1.11	1.11	1.38	2.34	2.05	2.05	2.47	2.41
P_2O_5	0.16	0.13	0.12	0.13	0.13	0.14	0.14	0.12	0.13	0.13
Ba	133	145	149	153	176	197	187	174	201	227
Cr	311	282	291	275	279	270	304	264	334	374
Cu	72	90	78	69	71	78	74	56	64	82
Ni	96	86	88	83	83	84	89	77	89	105
Rb	35	28	36	34	40	63	55	48	68	76
Sc	32	35	34	36	36	35	36	34	34	40
Sr	155	190	170	186	218	258	200	202	201	229
V	274	241	254	229	241	238	257	209	244	281
Zr	96	101	98	100	102	96	98	88	89	106

Sample	11	12	13	20	19	18	17	16	15	14
SiO_2	52.79	52.71	52.00	52.80	52.93	52.53	52.50	52.80	52.90	52.79
TiO_2	1.10	1.13	1.23	1.08	1.05	1.15	1.13	1.13	1.16	1.13
Al_2O_3	14.46	14.39	14.96	14.41	14.61	14.51	13.94	14.40	15.00	14.58
Fe_2O_3	1.95	1.97	1.97	1.86	1.96	1.96	2.00	1.94	1.89	1.96
FeO	8.33	8.40	8.39	7.91	8.36	8.34	8.51	8.27	8.05	8.38
MnO	0.17	0.18	0.12	0.17	0.17	0.17	0.18	0.18	0.15	0.16
MgO	8.22	8.03	9.30	8.25	7.94	7.92	8.10	7.88	8.32	7.76
CaO	8.06	8.50	6.82	8.34	7.88	8.00	8.26	8.28	8.27	9.68
Na_2O	2.31	2.37	3.09	2.81	2.20	2.62	2.76	2.44	2.40	2.10
K_2O	2.46	2.18	1.98	2.24	2.75	2.65	2.47	2.55	1.71	1.30
P_2O_5	0.14	0.15	0.15	0.13	0.14	0.15	0.15	0.14	0.15	0.15
Ba	207	184	152	195	221	221	233	225	175	173
Cr	314	293	360	333	307	301	296	303	311	321
Cu	106	102	142	105	112	118	95	116	117	128
Ni	97	93	101	99	92	93	92	95	99	101
Rb	70	63	51	67	80	82	74	80	52	47
Sc	35	37	40	36	37	36	37	37	37	40
Sr	211	165	266	160	182	171	189	181	188	208
V	270	266	307	270	273	286	271	276	288	299
Zr	90	97	106	93	99	94	99	101	103	103

Major elements normalized to 100 percent anhydrous; Fe_2O_3/FeO set at .235. Samples are presented in order from west to east across the dike.

related to the quartz normative, low-titanium tholeiite (LTQ) of Weigand and Ragland (1970) and the Rossville tholeiite of Smith and others (1975); 4) high-iron diabase derived most probably from an HTQ parent; 5) even more highly fractionated, residual granophyres. Although there is some variability in whole-rock compositions from the Quarry dike (particularly for the alkalis and calcium), all samples can be classified as HTQ basalt, particularly on the basis of their constant titanium content.

Table 3 presents the major- (normalized) and trace-element whole-rock compositions for samples collected from fine-grained chill zones in the central Newark Basin, including the chill zone of the Quarry dike. Also presented in Table 3 are average compositions for various ENA HTQ basalts and diabases. The analysed fine-grained chill zone samples are thought to give the best available approximation of magma compositions parental to the Early Jurassic intrusions of the central Newark Basin. This assumption is supported by their remarkable compositional and petrographic uniformity throughout the region and their close similarity to numerous other ENA HTQ basalts and diabases (Table 3); chill margin samples also contain the lowest initial strontium isotopic ratios ($\approx$.706). No other ENA magma type (olivine normative, high-iron, or LTQ) is represented by the chill margin samples analysed to date. This is in contrast to the multiple ENA magma types represented by fine-grained samples from the Gettysburg, Hartford, and Culpeper Basins (Smith and others, 1975; Phillpots and Martello, 1986, Froelich and Gottfried, 1988) and the northern Newark Basin (Puffer, 1988).

TABLE 3
MESOZOIC DIABASE CHILL AND BASALT COMPOSITIONS

	1	2	3	4	5	6	7	8	9	10
SiO_2	52.88	52.52	52.47	52.57	53.18	52.55	51.61	51.86	52.69	52.65
TiO_2	1.13	1.11	1.17	1.15	1.13	1.22	1.09	1.07	1.11	1.13
Al_2O_3	14.26	14.22	14.69	14.16	13.73	14.64	14.34	14.27	14.57	14.37
FeO*	10.13	9.98	9.95	10.15	10.16	10.40	10.20	10.86	9.97	10.58
MnO	0.16	0.16	0.18	0.17	0.19	0.16	0.15	0.16	0.20	0.19
MgO	7.91	8.16	7.69	7.82	7.99	7.67	8.32	7.98	7.83	7.49
CaO	10.26	11.02	10.93	10.66	10.48	10.44	11.41	11.24	10.90	10.78
Na_2O	2.01	2.08	2.16	2.29	2.30	2.06	2.15	2.06	1.99	2.15
K_2O	1.12	0.61	0.62	0.84	0.70	0.89	0.59	0.50	0.61	0.67
P_2O_5	0.13	0.14	0.14	0.18	0.14	0.14	0.13	0.12	0.12	----
Ba	156	147	154	149	166	195	182	174	160	---
Cr	298	281	258	279	311	315	260	322	302	277
Cu	99	105	107	113	122	110	127	123	121	111
Ni	92	89	87	96	98	95	61	72	89	81
Rb	37	16	15	33	32	---	37	22	25	21
Sc	36	36	37	38	36	37	---	---	---	---
Sr	189	169	208	229	170	175	183	186	187	186
V	265	264	266	255	285	235	272	270	310	---
Zr	101	94	104	102	102	120	116	87	115	92

Major elements normalized to 100 percent anhydrous.

1) Quarry dike chill (average of samples QD2, QD3, and QD14 of Husch and Schwimmer (1985)).
2) Byram (Point Pleasant) diabase chill (sample PP2 of Husch and others (1984)).
3) Lambertville Sill chill-Sourland Mountain region (sample LSR14 of Husch and Roth (1988)).
4) Lambertville sill chill-Delaware River area (average of samples ELS5 and ELS25 of Eliason (1986)).
5) Baldpate Mountain chill (average of samples T4 and T5 of Trione (1985)).
6) Palisades sill chill (sample W-889LC-60 of Walker (1969)).
7) Average Orange Mountain Basalt (Puffer and others, 1981).
8) Average Talcott Basalt (Puffer and others, 1981).
9) Average York Haven type basalt (Smith and others, 1975).
10) Average HTQ diabase chill and basalt (Weigand and Ragland, 1970).

The very limited compositional range of the chill margin samples also is not consistent with the suggestion of Philpotts and Burkett (1985) that two distinct, though similar, HTQ magmas were intruded in the central Newark Basin. Similarly, the conclusion of Husch and others (1984) that extensive deep crustal fractionation of a ponded HTQ basalt produced a variety of derived high-iron type parental magmas also is not supported.

Although no fine-grained equivalents are known from the region, there are four coarse-grained samples collected as part of a stratigraphic section through the northeast end of the Lambertville sill (Husch and Roth, 1988) that do have whole-rock compositions with some LTQ-like characteristics. These include generally lower concentrations of titanium, barium, chromium, copper, nickel, strontium, and zirconium and generally higher concentrations of aluminum, calcium, and sodium than is typical for HTQ-related rocks with similar MgO contents. Because the LTQ-like rocks from the Lambertville sill are coarse grained, they probably do not represent actual magma compositions. Indeed, three of the four samples in question have MgO contents higher than any reported for ENA LTQ related magmas; petrographically, these same samples appear to contain cumulate pyroxenes. However, when the average composition for the four samples is compared, its similarity with various ENA LTQ magma types is apparent, as are its differences with typical ENA HTQ and York Haven basalts (Table 4). Although other coarse-grained rocks with relatively low titanium contents are found in the Lambertville sill and other central Newark Basin intrusives, none can be typed

TABLE 4
LTQ AND HTQ MAGMA COMPOSITIONS

	1	2	3	4	5	6
SiO_2	51.93	51.32	51.51	51.27	52.69	52.65
TiO_2	0.81	0.83	0.76	0.75	1.11	1.13
Al_2O_3	15.08	14.94	14.91	16.79	14.57	14.37
Fe_2O_3	1.82	----	11.74	1.08	----	----
FeO	7.75	11.12	----	9.15	9.97	10.58
MnO	0.18	0.16	0.20	0.18	0.20	0.19
MgO	7.81	8.08	7.42	6.89	7.83	7.49
CaO	11.57	10.49	10.77	10.96	10.90	10.78
Na_2O	2.43	2.52	2.22	1.98	1.99	2.15
K_2O	0.52	0.36	0.48	0.40	0.61	0.67
P_2O_5	0.09	0.17	----	0.09	0.12	----
Ba	125	110	---	115	160	---
Cr	186	250	218	205	302	277
Cu	84	71	68	66	121	111
Ni	75	56	48	63	89	81
Rb	19	---	15	21	25	21
Sc	36	---	---	---	---	---
Sr	143	149	127	137	187	186
V	269	225	---	---	310	---
Zr	75	68	60	66	115	92

Major elements normalized to 100 percent anhydrous.

1) Average of four LTQ-like whole-rock compositions (samples LSR8, LSR9, LSR10, and LSR11 of Husch and Roth (1988)).
2) LTQ-related third flow of Preakness Basalt (Puffer, 1988).
3) Average LTQ basalt (Weigand and Ragland, 1970).
4) Average Rossville type basalt (Smith and others, 1975)
5) Average York Haven type basalt (Smith and others, 1975).
6) Average HTQ basalt (Weigand and Ragland, 1970).

unequivocally as a LTQ magma. In fact, many most likely are mafic-rich and titanium-poor because of pyroxene accumulation in HTQ liquids. Other titanium-poor samples, that are not mafic-rich, do not consistently exhibit LTQ-like characteristics.

Variation trends

Selected major-element oxide and trace-element versus MgO diagrams are presented in Figures 2A-H to illustrate important variation trends for the seven sheet-like diabases combined. As seen in Figures 2A-H, these trends, with few exceptions, essentially are identical to those found for a vertical section through the Palisades sill (Walker, 1969). Quarry dike samples are not plotted in Figures 2A-H since their MgO contents remain almost constant at about 8 weight percent and, therefore, would form tightly clustered to nearly vertical trends. As shown by Husch and Schwimmer (1985), this is because crystal fractionation was not a significant cause of the limited compositional variations found within the Quarry dike. Besides the Quarry dike samples, whole-rock samples with MgO contents greater than 9 weight percent also are not plotted. These samples are believed to be mafic enriched because of the accumulation of orthopyroxene and/or clinopyroxene and are discussed separately.

Figures 2A-C show that SiO2, TiO2, and P2O5 contents initially remain nearly constant or rise only slightly until MgO values fall to approximately 3-4 weight percent. At lower MgO values, concentrations increase dramatically for all three oxides, and then fall at the lowest MgO values (<2 wt %) for TiO2 and P_2O_5; FeO* (total) and V trends (not shown) exhibit similar behavior. Al_2O_3 (Fig. 2D) and Sr (not shown) concentrations, on the other hand, increase until MgO contents fall to approximately 4-5 weight percent. At lower MgO contents there is a clear reversal in trend and Al_2O_3 and Sr contents decrease markedly. In addition, it is noted that one sample is quite depleted in SiO_2 (and Al_2O_3). This somewhat anomalous sample also is higher in TiO_2, FeO* (not shown), V (not shown) and Sc (not shown) contents than samples with similar MgO contents that plot along the main variation trend. Petrographic studies show this sample to be enriched in Fe-Ti oxides (approximately eight modal percent) and the geochemical data are consistent with this observation. Comparable samples are found in the Palisades sill (Fig. 2). Finally, three of the four samples from the Lambertville sill with LTQ affinities plot just off the main HTQ-related variation trends for SiO_2, TiO_2, P_2O_5, and Al_2O_3.

CaO concentrations (Fig. 2E) exhibit a third pattern in that they decrease slightly until MgO contents fall to 4 or 5 weight percent. At this point there is a distinct increase in slope and CaO is depleted more rapidly. Sc concentrations (not shown) produce a similar pattern with the difference that the rapid depletion occurs at lower MgO values of approximately 2 weight percent or less. Significantly, at least four samples, and perhaps as many as six, are notably depleted in CaO and plot well below the main variation trend. As before, the LTQ-like samples plot just off the main variation trend.

In contrast to the behavior for CaO, all alkali elements and

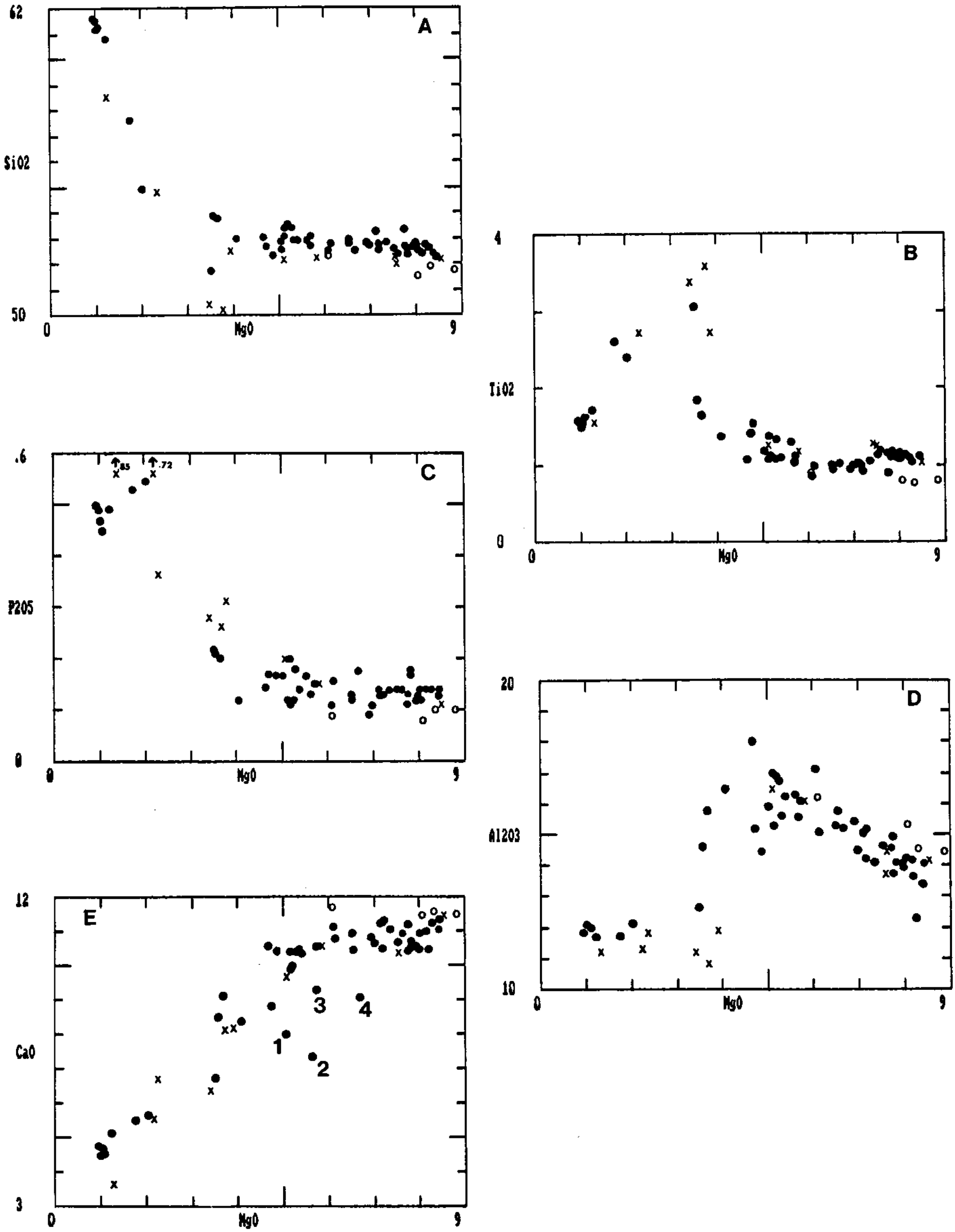
Zr act incompatibly throughout the entire range of MgO values shown. This behavior is exemplified by the variation trend for total alkali (Na_2O+K_2O; Fig. 2F). Not coincidently, all samples exhibiting a relative depletion in CaO are correspondingly enriched in alkalis (although not Zr). However, unlike the case for many other elements, there is no meaningful compositional distinction on the basis of total alkali content between the four possible LTQ-related samples and the numerous others clearly related to a HTQ parent magma.

Variation trends for Ni and Cr document their compatible behavior (Figs. 2G-H). However, Cr concentrations fall much more rapidly with decreasing MgO content than Ni. In fact, the variation trend for Ni is quite linear. This led Philpotts and Burkett (1985) to suggest that the entire range of rock compositions in the region was controlled almost entirely by orthopyroxene (of constant composition), either through its addition or removal. However, the obvious nonlinearity of the Cr variation trend over the entire MgO range is not compatible with this idea. Finally, the four samples with LTQ-like compositions plot below the main HTQ-related variation trends for Ni and Cr.

DISCUSSION

Crystal fractionation

Extensive olivine fractionation is not indicated by the variation trends presented in Figures 2A-H, particularly those for SiO_2, CaO, and Ni. If olivine fractionated to any meaningful degree early in the crystallization sequence, residual liquid


A
62
SiO2
50
0
MgO
9
B
4
TiO2
0
0
MgO
9
C
.6
.85
.72
P2O5
0
0
MgO
9
D
20
Al2O3
10
0
MgO
9
E
12
CaO
3
0
MgO
9
1
2
3
4

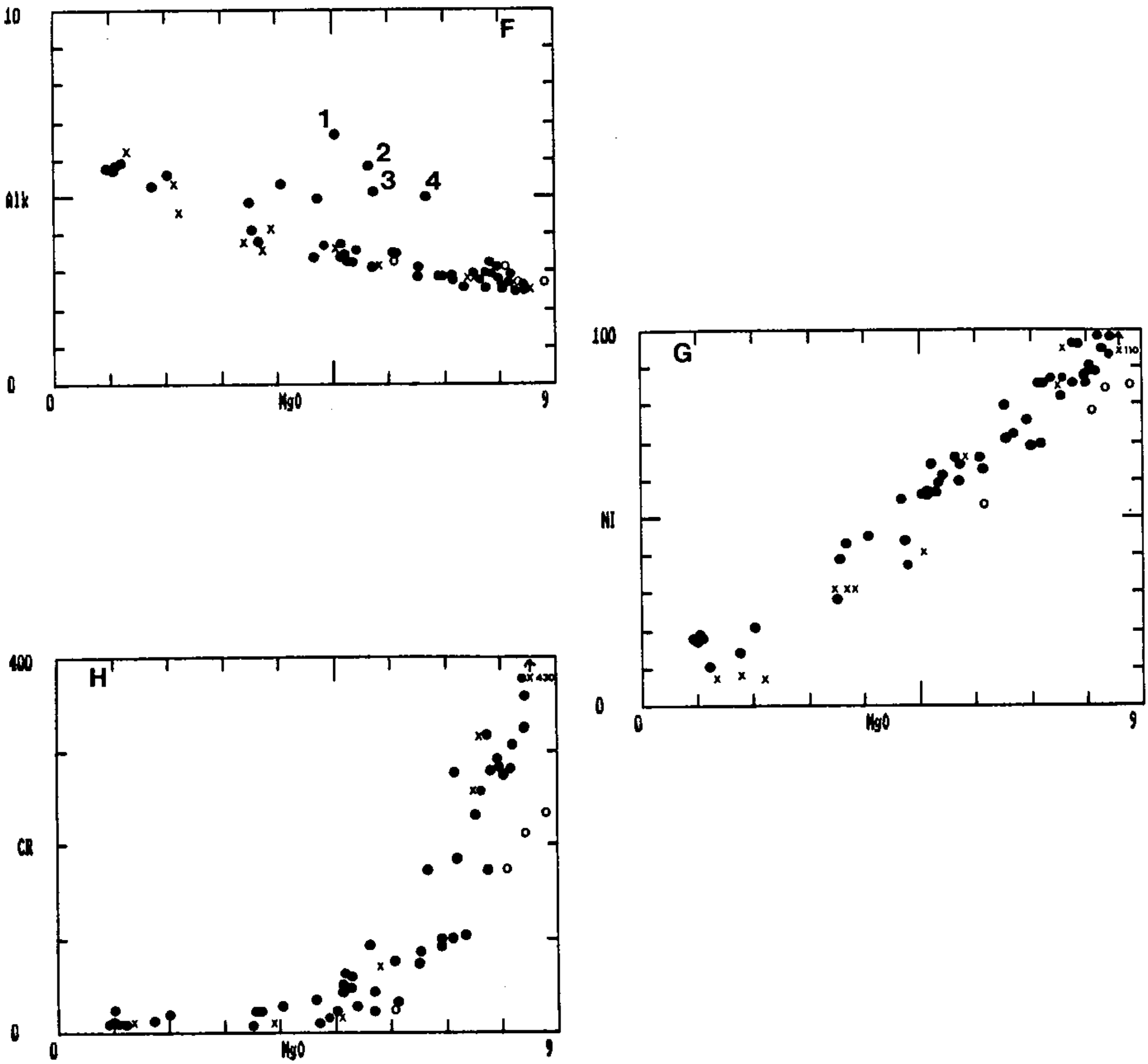


Figure 2. Selected whole-rock oxide and trace-element versus MgO variation trends for Mesozoic diabase from the central Newark Basin. HTQ-related samples are shown with filled circles; the four LTQ-related samples are shown with open circles. For comparison, whole-rock samples (Walker, 1969) from the Palisades sill (X) also are plotted. Oxide concentrations are given in weight percent and trace elements in parts per million. Samples with MgO contents greater than 9 weight percent and all samples from the Quarry dike are omitted. Alk is total alkali (Na2O+K2O). Samples with anomalously low calcium and high alkali contents labeled in E and F are: 1) ELS24; 2) WH3; 3) T9; 4) ELS6.

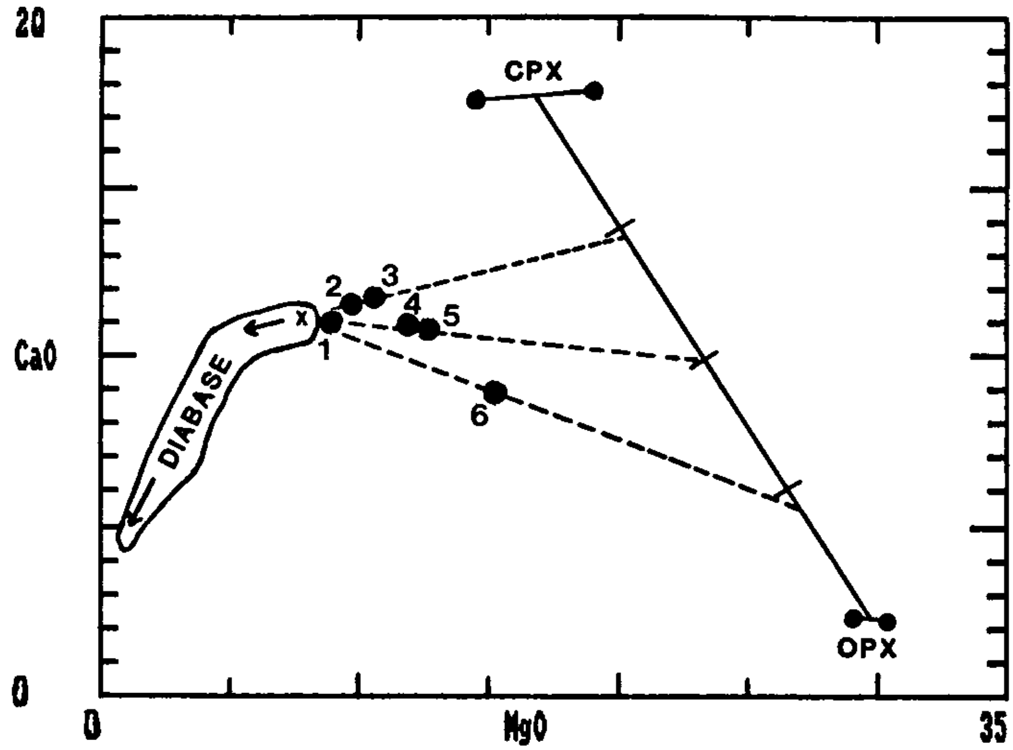
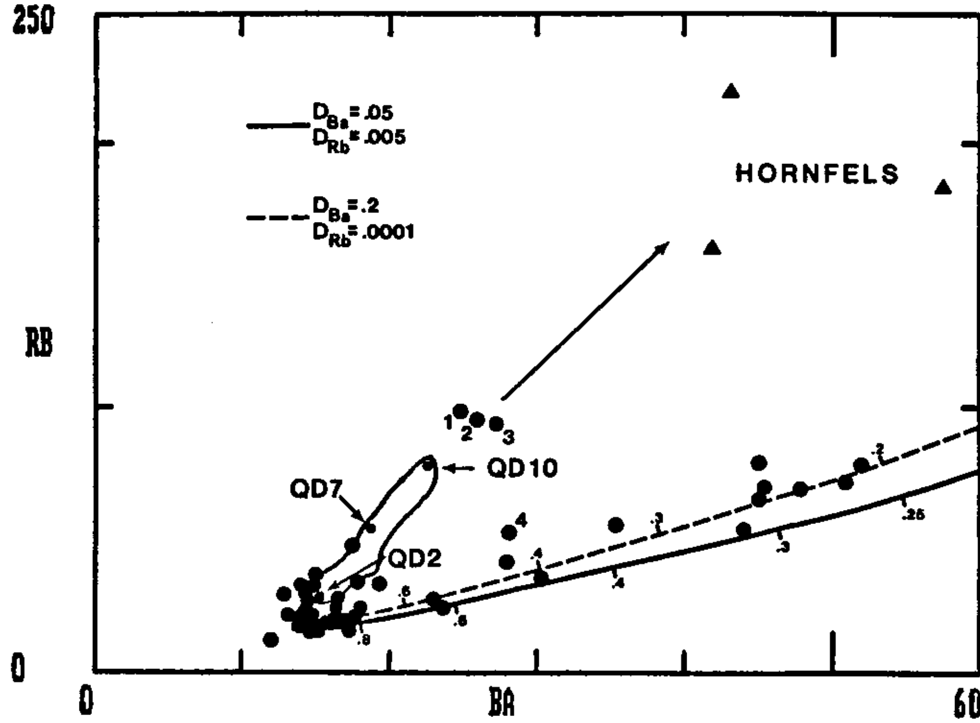


Figure 3. CaO versus MgO (weight percent) mixing diagram. Diabase field taken from Figure 2E. Selected samples with greater than 9 weight percent MgO are shown by filled circles. Three addition/removal lines with one end-member being the average regional HTQ chill margin composition (X) and the other end-member being a variable mixture of clinopyroxene (compositions from Walker (1969) and Philpotts and Reichenbach (1985)) and orthopyroxene (compositions from Smith (1973) and Philpotts and Reichenbach (1985)) are shown by the dashed lines. Six high-magnesium samples discussed in text are: 1) LSGRAB; 2) LSR2; 3) ELS11; 4) LS2; 5) LSR6; 6) PP5. Arrows in diabase field show direction of variation due to crystal fractionation.

concentrations for SiO_2 and CaO would have increased while those for Ni would have decreased much more dramatically than observed. Although it is true that these compositional variations might be produced by fractionating a SiO_2- and CaO-rich and Ni-poor phase, such as plagioclase, along with olivine, the low abundances of modal olivine in every intrusion (even those with high magnesium samples) and the complete lack of any olivine accumulation zones in the central Newark Basin argue against the possibility. Also, the early incompatible behavior for Al_2O_3 (Fig. 2D) and Sr is not consistent with the early extensive fractionation of plagioclase.

On the other hand, the early fractionation of high and low calcium pyroxenes is consistent with the observed petrography of the analysed samples and with the more highly compatible behavior of Cr as compared to Ni; Cr pyroxene/basalt distribution coefficients are two to four times higher than those for Ni (Henderson, 1982). Furthermore, the slight early depletion of CaO and Sc and the lack of early SiO_2 enrichment agree with the early crystallization of both pyroxene phases. A CaO-MgO two-oxide mixing diagram (Fig. 3) indicates that the early variation trend could have been produced by removing three times as much clinopyroxene as orthopyroxene from chill margin compositions, representing parental ENA HTQ basalts.

The change in Al_2O_3 and CaO trends seen in Figures 2D-E at a MgO concentration of approximately 5 weight percent is consistent with the onset of extensive plagioclase crystallization and fractionation, along with continuing pyroxene fractionation. This



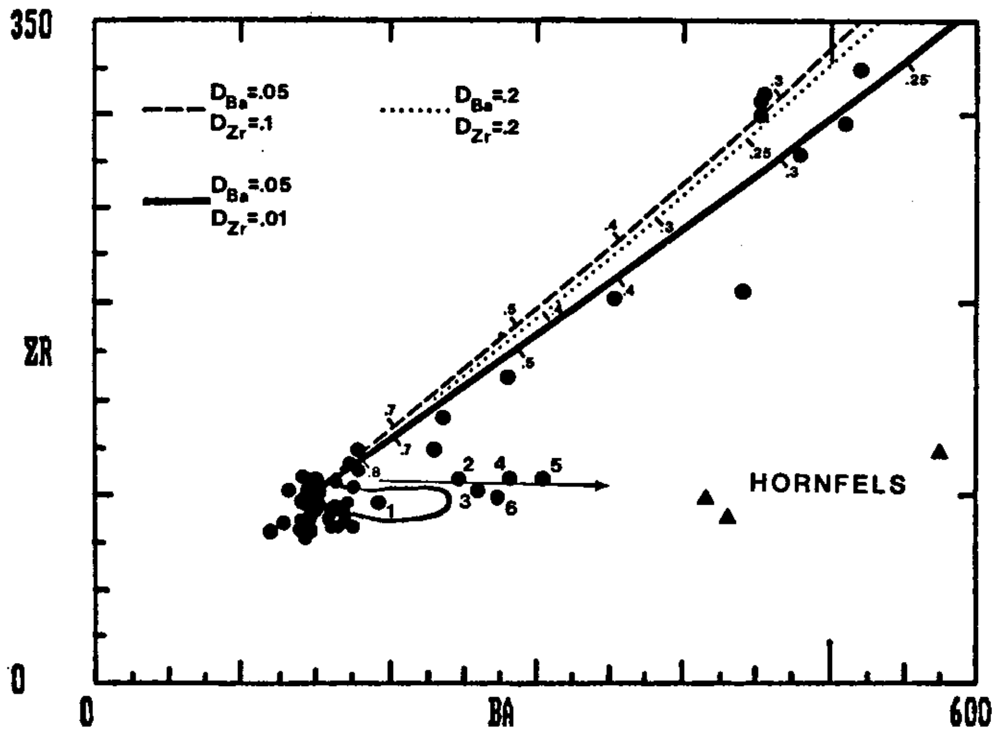


Figure 4. A) Rb versus Ba variation diagram. B) Zr versus Ba variation diagram. Fractionation curves shown are produced by applying the bulk distribution coefficients given and assuming a perfect fractional crystallization process. Values along curves give fraction of liquid remaining. Arrows indicate general trend of contaminating chill margin or slightly fractionated diabase compositions with analysed hornfels (from Elaison (1986) and Sturgis (1983)). Coarse-grained contaminated samples in A are: 1) T9; 2) ELS24; 3) ELS6; 4) WH1. Coarse-grained contaminated samples shown in B are: 1) sample D1 of Benimoff and Sclar (1984); 2) T9; 3) ELS24; 4) WH1; 5) WH3; 6) ELS6. Also shown are the Quarry dike fields with three samples labeled (A only; from Husch and Schwimmer, 1985).

would explain the rapid depletions of Al_2O_3, CaO, and Sr in rocks with lower MgO contents. Mass balance calculations, utilizing mineral compositions from Walker (1969), Smith (1973), and Philpotts and Reichenbach (1985), suggest that plagioclase fractionation began after approximately 20 percent pyroxene crystallization. Pyroxene and plagioclase fractionation would be followed at still lower MgO values by the addition of extensive Fe-Ti oxide and finally apatite fractionation. The former accounts for the enrichment of SiO_2 and the depletion of FeO*, TiO_2, Sc, and V in the late-stage granophyres, while the latter accounts for decreasing P_2O_5 values in the same rocks.

If a perfect fractional crystallization (or Rayleigh fractionation) model is applied, it is seen that even the most evolved, granophyric whole-rock compositions do not appear to require more than 75-80 percent crystallization (Figs. 4A-B). However, this value only is an estimate because the actual fractionation process may have been less effective (due to incomplete separation of crystal and residual liquid) than pure Rayleigh fractionation and the differentiated sheet-like intrusions probably were not completely closed systems during crystallization; new magma may have been injected repeatedly into the sheet-like bodies (Husch and Roth, 1988) while residual liquids migrated laterally (Husch, 1987) and/or vented to the surface (Puffer, 1988).

Finally, the close agreement between the variation trends for diabasic rocks of the central Newark Basin and those from the Palisades sill (Walker, 1969) suggests the possibility that similar differentiation processes were operating. The lack of olivine accumulation zones in the central Newark Basin diabases, or in any other documented HTQ-related intrusion from any other ENA basin, calls into question the necessity of having an early olivine fractionation event in order to produce very similar whole-rock compositions in the Palisades sill. This leads to the unexpected (if not heretical) conclusion that the formation of the famous olivine zone of the Palisades sill was not central to the overall differentiation of the sill! Thus, whatever process formed the olivine zone did not play a direct role in the *in situ* geochemical evolution of the entire intrusion. The restricted lateral extent of the olivine zone within the Palisades sill (Walker, 1940; Walker, 1969), the zone's abrupt terminations, both upward and downward, and the paucity of olivine in other coarse-grained rocks in the sill (Walker, 1969; Shirley, 1987) are consistent with this hypothesis. The formation of the olivine zone may have been a separate event, resulting perhaps from the intrusion of a distinct magma type, possibly olivine normative. Alternatively, the olivine zone could represent the intrusion of a HTQ-related magma in which there had been a prior accumulation of olivine at depth, leaving behind a residual melt parental to the Palisades sill. This would explain the similar slopes for chondrite normalized rare-earth element distribution patterns throughout the entire sill (Shirley, 1987).

Mafic-rich samples

Figure 3 also shows that high magnesium (>9% MgO) samples can be produced by the addition of variable amounts of clinopyroxene and orthopyroxene to chill margin composition. For example, simple mixing relationships suggest that in one instance (sample PP5) accumulation is dominated by orthopyroxene, while in others (samples LSR2 and ELS11) it is dominated by clinopyroxene; additional compositions (samples LSGRAB, LS2, and LSR6) can be produced by accumulating roughly equal proportions of the two pyroxenes. Modal determinations are in agreement with these results. Whether these accumulations took place _in situ_, at depth, or in a combination of the two circumstances has not been determined quantitatively. However, the accumulation of high pressure orthopyroxenes has been described for the Talcott Basalt by Philpotts and Riechenbach (1985) and sample PP5 is texturally very similar; large phenocrysts of euhedral orthopyroxene rimmed by clinopyroxene in a fine-grain groundmass are seen in both occurrences. On the other hand, _in situ_ accumulation for LSR2, LSR4, and ELS11 is more likely, considering that their compositions in Figure 3 plot directly along a mafic extension of the early variation trend, ultimately resulting in the granophyric compositions found in many of the central Newark Basin intrusions. In addition, they are distinct texturally from PP5 and lack large concentrations of orthopyroxene megacrysts.

Contamination

Whole-rock samples containing anomalously high alkali and

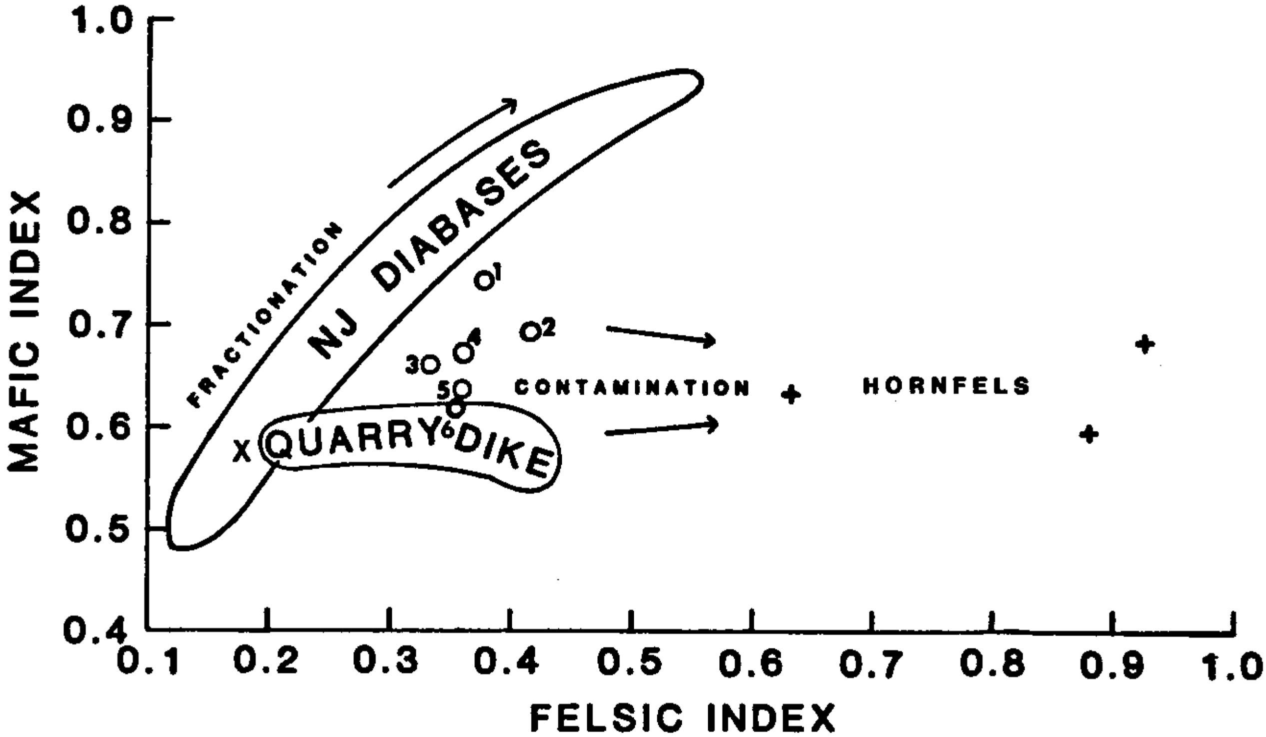


Figure 5. Mafic index (Fe2O3*/Fe2O3*+MgO) versus felsic index (Na2O+K2O/Na2O+K2O+CaO) variation diagram. Quarry dike and New Jersey diabase fields from Husch and Schwimmer (1985); New Jersey diabase field includes variation trend from the Palisades sill (Walker, 1969). Hornfels compositions from Sturgis (1983) and Benimoff and Sclar (1984; sample XB). Contaminated coarse-grained diabase samples shown by open circles are: 1) WH1; 2) WH3; 3) D1 (Benimoff and Sclar, 1984); 4) ELS24; 5) T9; 6) ELS6. Average composition of diabase chill margins shown by X.

low Ca concentrations have been discussed by Husch and Schwimmer (1985) and Benimoff and Sclar (1984; their sample D1 from an exposure of the Palisades sill) and are believed to result from the contamination of HTQ-related magma by local country rocks whose compositions are best represented by hornfels of Lockatong Formation (Figs. 4A-B, 5, and 6). As detailed by Benimoff and Sclar (1984), contamination apparently was controlled by the selective diffusion of alkalis from the contaminant into the magma and of calcium from the magma into an anatectic melt derived from the contaminating material. Thus, the presence of contamination is manifested most obviously by samples that contain distinct geochemical characteristics including: anomalously high values for the alkalis and low values for CaO, at a given value for MgO (Figs. 2E-F); plot off of reasonable fractional crystallization variation trends (Figs. 4A-B); plot to the right of the main diabase variation trend on a mafic index versus felsic index variation diagram (Fig. 5); and plot along one of the calculated mixing lines (Langmuir and others, 1978) on a Rb/Sr versus K/BA variation diagram (Fig. 6). The Quarry dike fields shown in Figures 4A-B, 5, and 6 also appear to be controlled by contamination from a very similar source.

Strong supporting evidence for this contamination mechanism, particularly for the Quarry dike, comes from isotopic data. On an initial 87Sr/86Sr versus 1/Sr diagram (Fig. 7), three Quarry dike samples plot directly on a mixing (or contamiantion) line between a typical HTQ chill (PP2) and a sample of Lockatong hornfels (KQ2). These two end-member compositions are the same as

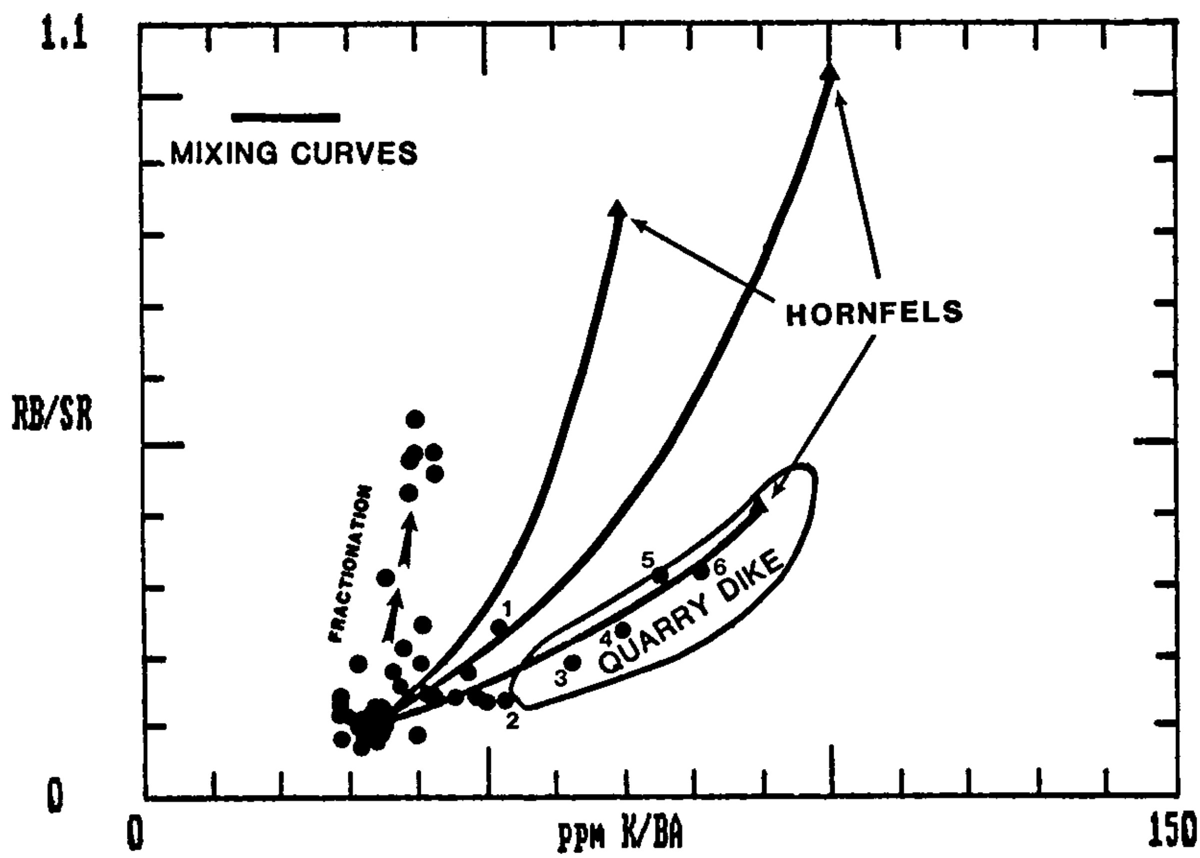


Figure 6. Rb/Sr versus K/Ba variation diagram. Three mixing curves extending from average chill composition to each of three analysed hornfels compositions (in descending order of Rb/Sr values they are samples ELS3 (Eliason, 1986), KQ10, and KQ2 (Sturgis, 1983), respectively). Mixing curves are generated using general equations of Langmuir and others (1978). Quarry dike field from Husch and Schwimmer (1985). Coarse-grained samples showing effects of contamination are: 1) WH1; 2) sample D1 of Benimoff and Sclar (1984); 3) ELS21; 4) ELS6; 5) T9; 6) ELS24. Note that all coarse-grained sample compositions (except WH1) and the Quarry dike field plot along the mixing curve produced by having sample KQ2 as the contaminant.

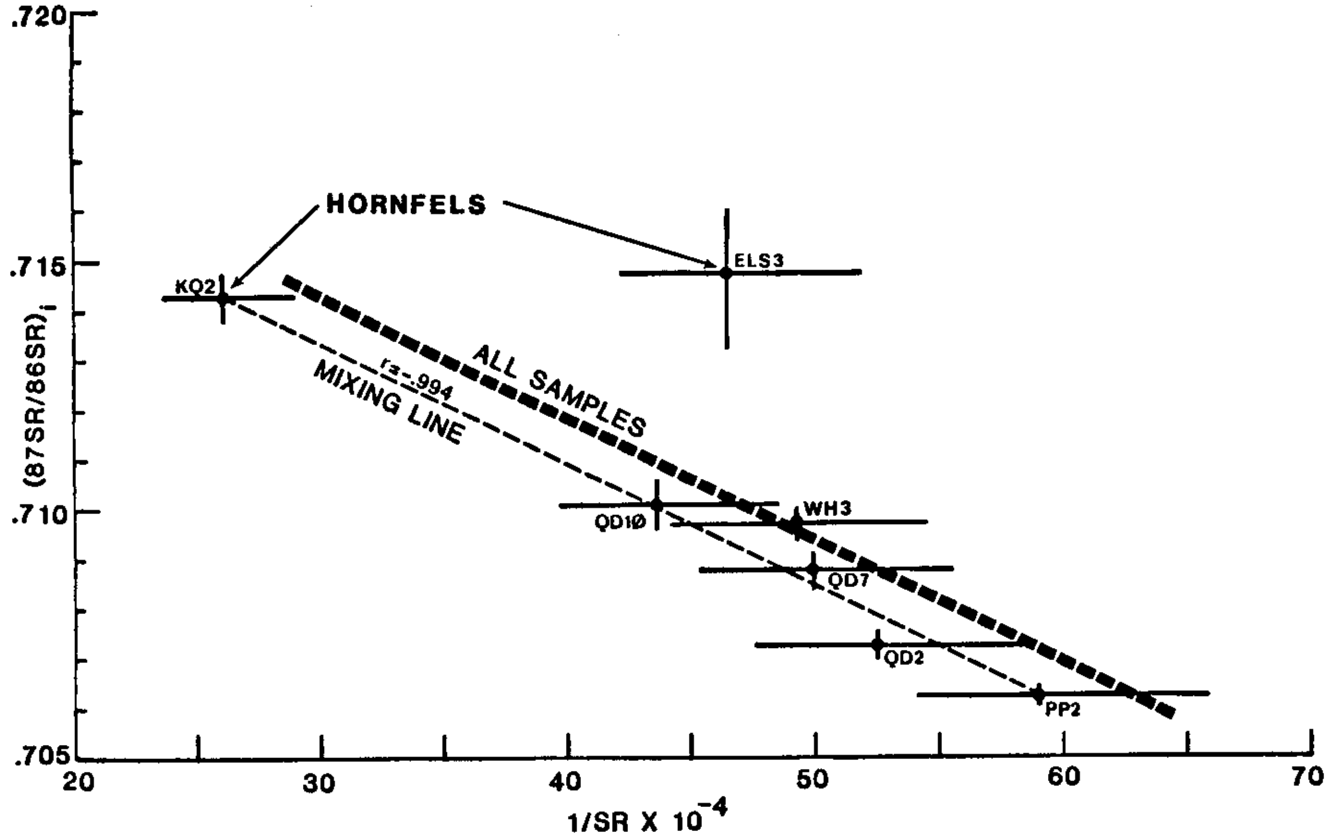


Figure 7. Calculated initial 87Sr/86Sr versus 1/Sr variation diagram. Initial strontium isotopic ratios are calculated for 200 m.y. (the best estimate for the actual age of the diabase intrusions). Sample PP2 represents typical HTQ basalt. Samples QD2, QD7, and QD10 represent Quarry dike chill, intermediately contaminated, and highly contaminated samples, respectively. Sample WH3 is a coarse-grained, contaminated (see preceding three figures) diabase. The mixing line connecting KQ2 with PP2 is indistinguishable from the best least-squares fit (correlation coefficient shown) for samples PP2, QD2, QD7, QD10, and KQ2. Error bars for calculated ratios are given for each sample.

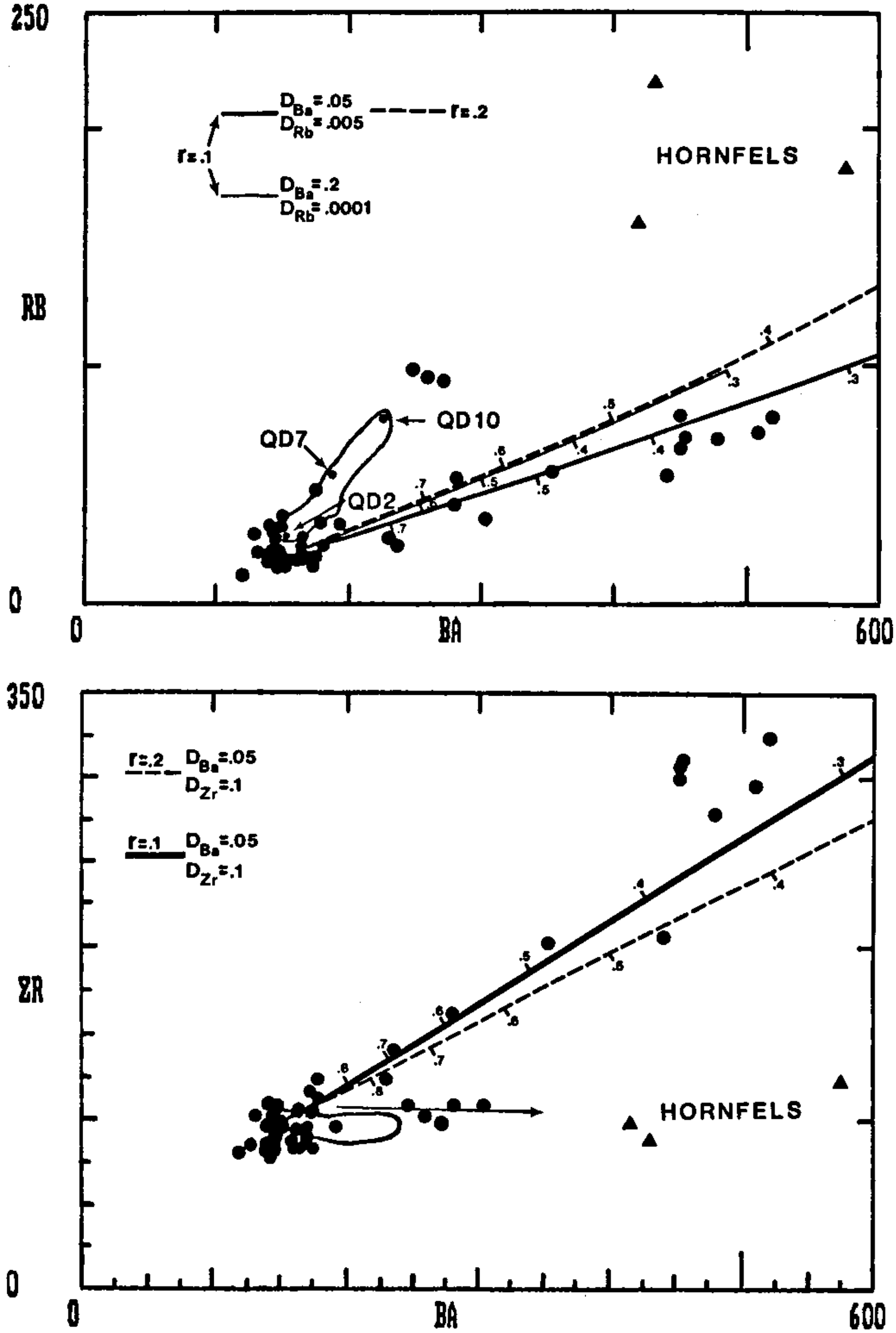


Figure 8. A) Rb versus Ba variation diagram. B) Zr versus Ba variation diagram. Plotted compositions and Quarry dike fields are identical with those shown in Figure 4. Variation trends shown are produced by applying the assimilation-fractional crystallization (AFC) model of DePaolo (1981). Bulk distribution coefficients and assimilation/crystallization (r) ratios are given for each curve shown. Values along curves give fraction of liquid remaining. Arrow in B indicates general trend of contaminating slightly fractionated diabase with hornfels.

those used to construct the preferred mixing curve in Figure 6. The apparent extent of contamination shown for the three Quarry dike samples is the same regardless of whether trace-element (Fig. 4A) or isotopic data (Fig. 7) is used; sample QD2 (chill) has the least amount of contamination, sample QD7 has an intermediate amount, and sample QD10 has the most. The latter sample also is from the center of the dike where total alkali content is greatest. In addition, many of the same coarse-grained samples from a number of the sheet-like intrusions plot as contaminated in Figures 2E-F, 4A-B, 5, and 6; one, WH3, plots along the mixing line in Figure 7. Thus, major-element, trace-element, and isotopic data are all consistent and indicate a contamination origin for the compositional variations found for the Quarry dike and the anomalous alkali-rich and calcium-poor coarse-grained samples. The contaminant in most instances appears to be the same material, hornfels of the Lockatong Formation.

A further refinement in the petrogenesis of the central Newark Basin diabases is introduced if contamination is combined with Rayleigh fractionation by utilizing the assimilation-fractional crystallization (AFC) model of DePaolo (1981). Figures 8A-B show that some of the more evolved coarse-grained sample compositions are reproduced more accurately if approximately 50-70 percent fractional crystallization occurs simultaneously with about 5-25 percent contamination. Note also that a number of other coarse-grained samples plot along a contamination line with a magmatic end-member plotting at approximately 15-20 percent crystallization (Fig. 8B). The possibility that these samples

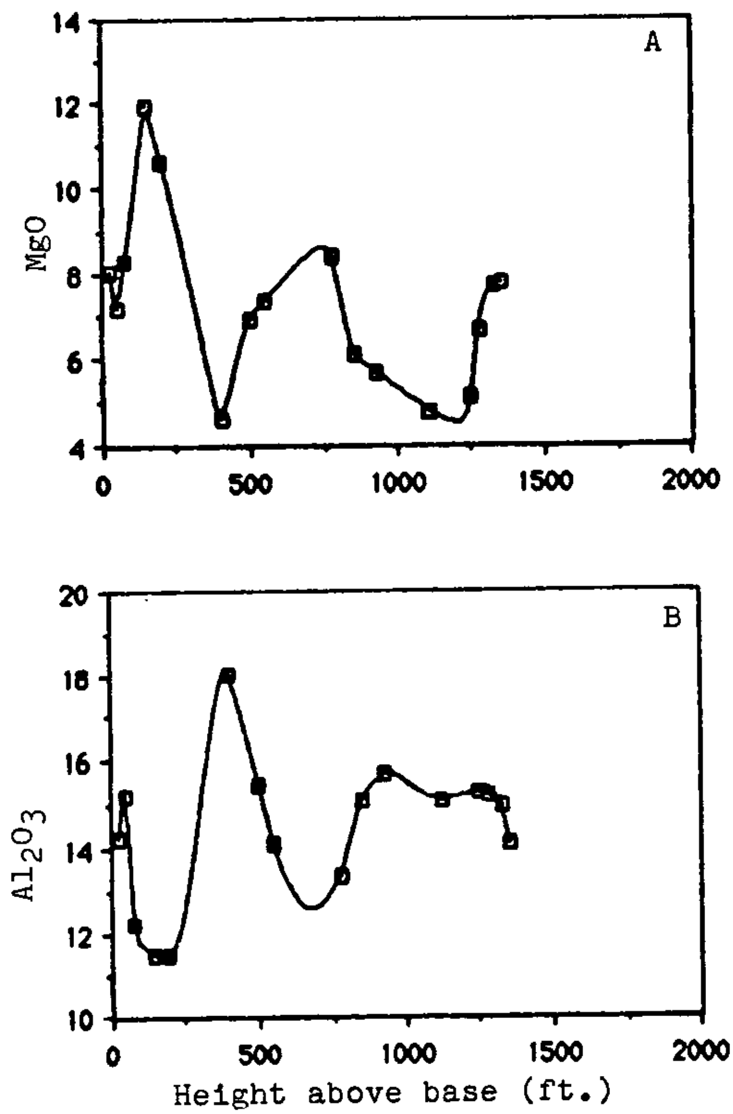


Figure 9. Variations in whole-rock compositions through the Lambertville sill along the Delaware River. A) MgO content (weight percent) versus height (feet) above the base of the sill. B) Al2O3 content versus height above the base of the sill. Note the near mirror-image nature of the variation patterns and the presence of an upper mafic zone at approximately 700 feet.

were affected by contamination after initial fractionation also is consistent with their MgO contents of 4-6 weight percent and their plotted positions in Figure 5.

Finally, none of the coarse-grained diabase samples shown to be contaminated on the basis of their geochemistry are located at a chilled contact with Triassic sedimentary rocks. This situation is analogous to the setting for the Quarry dike where measured contamination levels increase inward from the contact with Paleozoic sediments (Figures 4A and 7) Therefore, the source of contamination either is at depth, as is apparently the case for the Quarry dike (Husch and Schwimmer, 1985), or within the individual intrusions themselves, as documented by Benimoff and Sclar (1984) in the Palisades sill.

Multiple magma pulses

Two widely separated vertical sections through the Lambertville sill contain whole-rock compositions that suggest at least two distinct pulses of HTQ magma were intruded. The section across the northeast end of the sill also contains LTQ-related compositions which would necessitate the existence of a third separate pulse. Evidence for the two HTQ-related pulses is best seen in compositional variations for whole-rock samples collected along a section through the southwestern end of the sill at the Delaware River (Figs. 9A-B). MgO contents (Fig. 9A), along with Ni and Cr abundances, initially rise from the lower chilled contact zone and peak in a lower pyroxene accumulation zone. They then fall as an overlying differentiated diabase zone is traversed, rise again in a second, upper mafic-rich zone, fall again in a second, upper differentiated diabase zone, and rise a final time as the upper chilled contact zone is approached. Al_2O_3 (Fig. 9B), acting incompatibly along with TiO_2, Na_2O, K_2O, P_2O_5, Ba, Cu, Rb, Sr, and Zr, exhibits a variation pattern which is a near mirror image of that found for MgO. Not surprisingly, optically determined pyroxene and plagioclase compositions are consistent with the available whole-rock data. Orthopyroxenes are most magnesian ($\cong$En 80) and plagioclases are most calcic ($\cong$An 70) in both pyroxene-rich zones; orthopyroxenes are more iron-rich ($\cong$En 65) and plagioclases are more sodic ($\cong$An 55) in the two MgO-poor, differentiated zones.

The absolute timing of the pulses is not known. However, no internal chills have been observed, suggesting the time between pulses was of short duration. Indeed, it is tempting to correlate the injection of the HTQ pulses to the multiple magma injections documented for the Palisades sill (Walker, 1969; Shirley, 1987) and the multiple flows of the Orange Mountain Basalt (Puffer, 1988). The LTQ-related pulse in the northeastern end of the Lambertville sill then would correlate with the extrusion of the younger Preakness Basalt, a LTQ-derived flow (Puffer, 1988). This would place an upward limit of approximately 250,000 years on the multiple injection sequence (Olsen and Fedosh, 1988).

Lateral compositional variations

As reported by Husch (1987), most of the central Newark Basin sheet-like intrusions do not appear to be in mass balance with their HTQ parental magma composition. This is seen in Table 5 where, with the possible exception of the Delaware River section of the Lambertville sill, the calculated average composition for each intrusion is statistically different from the average regional chill margin composition. These compositional data support what easily is seen in the field; differentiated sheet-like intrusions with thick and extensive pyroxene accumulation zones, such as the Byram diabase and the Lambertville sill, lack complementary residual, granophyric compositions, and those diabase bodies that do have significant volumes of granophyric rock, such as the Baldpate Mountain and Pennington Mountain diabases, lack any significant amounts of mafic-rich (pyroxene-rich) compositions. However, when all intrusion averages are combined, the regional average for the coarse-grained diabases is in closer agreement with the average chill margin (parental magma) composition.

The regional mass balance on the one hand, coupled with the lack of mass balance locally on the other, could have been produced by having residual liqiuds migrate laterally from where pyroxenes were accumulating. Because there is a consistent tendency for the more granophyre-rich intrusions to be located higher up in the stratigraphy than the mafic-rich intrusions, the residual liquids appear to have migrated up-section as well. A sketch cross-section of the central Newark Basin constructed roughly along the Delaware River (Fig. 10) shows a single diabase intrusion cutting up-section prior to intrabasinal normal faulting. The portions of the intrusion that after faulting would

TABLE 5

INTRUSION AVERAGES[4] AND STANDARD DEVIATIONS

	1 CHILLS x̄	σ	2 LAMBERTVILLE x̄	σ	3 PT. PLEASANT x̄	σ	4 BALDPATE x̄	σ	5 ROCKY HILL x̄	σ	6 PENNINGTON x̄	σ	ALL[1] INTRUSIONS x̄	σ
N	5		14		3		9		3		4		5	
SiO_2	52.78	(0.27)	52.73	(0.33)	52.56	(0.04)	55.20	(3.91)	53.14	(0.27)	55.91	(3.51)	53.91	(1.54)
TiO_2	1.13	(0.02)	[3]1.06	(0.17)	1.00	(0.21)	1.58	(0.77)	1.04	(0.06)	1.85	(0.39)	1.31	(0.39)
Al_2O_3	14.13	(0.23)	14.35	(1.81)	13.06	(2.41)	14.17	(2.11)	16.51	(0.62)	13.62	(1.82)	14.34	(1.31)
FeO*	10.09	(0.08)	10.12	(0.79)	9.90	(0.23)	11.54	(2.77)	9.54	(0.16)	13.35	(1.95)	10.89	(1.57)
MnO	0.17	(0.01)	0.17	(0.02)	0.17	(0.01)	0.19	(0.03)	0.16	(0.01)	0.22	(0.05)	0.18	(0.02)
MgO	7.99	(0.13)	7.42	(2.16)	10.31	(4.40)	4.19	(2.41)	5.65	(0.78)	2.58	(1.26)	6.03	(2.99)
CaO	[2]10.78	(0.25)	10.84	(0.79)	10.22	(1.04)	8.40	(2.94)	10.39	(0.04)	7.02	(2.13)	[3]9.37	(1.61)
Na_2O	2.13	(0.15)	2.33	(0.55)	1.96	(0.39)	2.91	(0.47)	2.52	(0.17)	3.18	(0.43)	2.58	(0.48)
K_2O	[2]0.69	(0.11)	0.64	(0.23)	0.50	(0.14)	1.32	(0.80)	0.75	(0.07)	1.65	(0.68)	0.97	(0.49)
P_2O_5	0.14	(0.02)	0.13	(0.05)	0.13	(0.02)	0.27	(0.17)	0.12	(--)	0.36	(0.18)	0.20	(0.10)
Ba	156	(8.3)	141	(23)	125	(30)	[2]290	(142)	169	(5.5)	369	(133)	219	(106)
Cr	289	(15)	204	(192)	331	(133)	60	(86)	59	(24)	22	(1.6)	[3]135	(130)
Cu	109	(90)	101	(23)	100	(1.5)	198	(97)	98	(6.1)	196	(13)	139	(53)
Ni	93	(3.9)	80	(21)	98	(21)	48	(28)	62	(8.1)	31	(12)	[3]64	(26)
Rb	[2]25	(8.8)	22	(9.4)	16	(5.7)	44	(24)	22	(4.0)	50	(23)	31	(15)
Sc	36	(0.9)	36	(3.6)	36	(1.0)	30	(7.0)	32	(1.7)	30	(5.9)	[3]33	(3.0)
Sr	183	(26)	179	(45)	145	(31)	188	(34)	190	(2.6)	178	(24)	176	(18)
V	266	(11)	260	(41)	248	(16)	234	(168)	229	(4.7)	181	(154)	230	(30)
Zr	98	(4.5)	86	(19)	86	(18)	181	(92)	92	(11)	208	(81)	131	(59)

[1]Average of columns 2 - 6

[4]Chills and contaminated samples excluded

[2]N = 4

[3]Significantly different at the 90% confidence limit.

Underlined values are significantly different from the chill average (column 1) at the 95% confidence limit (unless noted.) Significance determined by the Mann-Whitney U (Rank) Test.

be exposed as the Byram diabase and Lambertville sill are low in the Triassic section, near a hypothetical feeder dike, and contain mafic-rich rocks. The more distal portions of the body, now re-exposed as the Baldpate Mountain and Pennington Mountain diabases are well up into the Passaic Formation and dominated by granophyric rock.

A possible mechanism for driving the lateral and upward migration of residual liquids is the multiple injection of new magma from below. As a new pulse was intruded, space would be made for it by draining the relatively low density residual liquids of an earlier pulse to structurally higher and more distal (to the feeder dike) portions of the body. The areal extent of the intrusion may have been increasing with time as magma injections and differentiation continued.

A similar situation may be the case for the Palisades sill where its mildly fractionated (averages about 5.3 weight percent MgO) northern extension in southeastern New York State is located relatively high up in the stratigraphy (Puffer and others, 1982). This contrasts with the New Jersey portion of the Palisades sill where it is found much lower in the Triassic section and mafic-rich compositions are common. Significant lateral variations in diabase composition and stratigraphic position also are common in the Culpeper Basin where pyroxene-rich bodies often are located miles from their granophyric complements (Froelich and Gottfried, 1988). Thus, the lateral and upward migration of residual liquids after pyroxene fractionation appears to be a common process in the petrogenesis of many Early Jurassic intrusive sheets from a

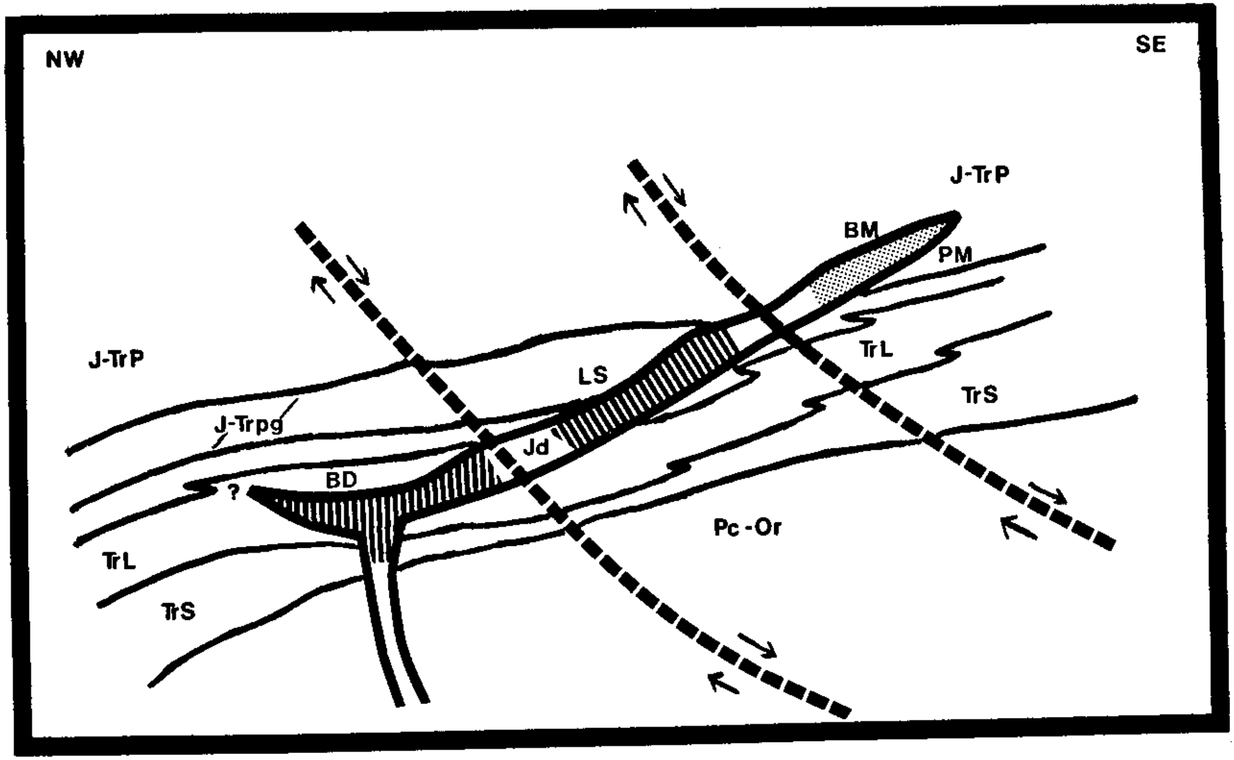


Figure 10. Sketch cross-section (not to scale) taken approximately along the Delaware River showing a single sheet-like intrusive diabase cutting up-section in its more distal regions. Dashed curves show approximate positions of post-intrusive faults which now result in the re-exposure of the sheet. Vertically ruled area shows where mafic-rich rocks are common and fractionated compositions are rare, diagonally ruled area shows where both mafic-rich and fractionated diabases are found, and dotted area shows where highly fractionated, granophyric compositions predominate. Locations of presently exposed diabases are: BD-Byram (Point Pleasant); LS-Lambertville sill; BM-Baldpate Mountain; PM-Pennington Mountain. Stratigraphic units are: Pc-Or, pre-Mesozoic basement; Trs, Stockton Formation; TrL, Lockatong Formation; J-TrP, Passaic Formation; J-Trpg, gray shale units of the Passaic Formation; Jd, Jurassic diabase.

number of separate basins. The classic view of these sills being differentiated only vertically directed, olivine dominated fractionation mechanisms (see Hyndman, 1985) is not supported by the available geochemical, petrographic, and structural data.

SUMMARY AND CONCLUSIONS

Petrographic, major- and trace-element, and Sr isotopic data for seven sheet-like diabase intrusions and one associated fine-grained dike located in the central Newark Basin indicate a petrogenesis involving: olivine-free fractional crystallization; selective contamination by surrounding country rock; multiple magma injections; and horizontal and vertical displacement of residual magmas. Chill margin compositions are homogeneous, have the lowest Sr isotopic ratios, and are typical quartz normative, high-titanium tholeiites. They are believed to be the best approximation of parental magma compositions. Differentiation initially was controlled by the fractionation of clinopyroxene with lesser amounts of orthopyroxne; accumulations of these phases in parental magmas produced mafic-rich compositions. Plagioclase removal was not significant until crystallization reached approximately 20-25 percent. Residual granophyric compositions were produced after approximately 75 percent solidification, assuming perfect fractional crystallization; lesser percentages are required if a combined assimilation-fractional crystallization model is applied. Locally, selective contamination from Triassic sedimentary wall rocks or xenoliths resulted in alkali enrichment and calcium depletion.

Calculated average compositions for the individual differentiated sheets often do not match their chill margin compositions. Typically, intrusions with mafic-rich rocks lack abundant granophyres; granophyre-rich intrusions lack any significant amounts of mafic-rich rocks. Generally, granophyre-rich sheets are found higher in the stratigraphy than those with mafic-rich units. This suggests that residual liquids were displaced both laterally and vertically from their complementary mafic fractions. The multiple injection of magma from below is seen in at least one differentiated sheet-like intrusion and this process could be the driving force for the migration of residual liquids.

Many of the petrogenetic processes proposed for the diabases of the central Newark Basin can be applied equally well to other compositionally similar diabase intrusions from other Mesozoic basins. This group includes the Palisades sill where an early _in situ_ olivine fractionation event is not required to produce the range of diabase compositions found.

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GEOLOGICAL CONTROLS OF RADON HAZARDS

IN THE NEWARK BASIN

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ABSTRACT

Elevated indoor radon in New Jersey has been evaluated using airborne radiometric data, a survey of indoor radon levels, and confirmatory radon testing. The radiometric data were reprocessed and portions were reflown. Data from ground sampling studies and in-home radon testing show that the airborne systems reliably detect areas of elevated uranium and its daughter radionuclide, radon. Areas with airborne anomalies of 6 ppm equivalent uranium or greater correlate well with clusters of homes where radon levels exceed 4 pCi/l and numerous homes exceed 200 pCi/l.

In the southwestern part of the Newark Basin where Triassic rocks underlie homes, more than 30% of the homes exceed the 4 pCi/l guideline. Elevated radon levels are less common in the northeastern part of the basin where glacial cover and restricted outcrop area of radioactive bedrock are mitigating factors.

Lacustrine cycles in the Lockatong and Stockton Formations are associated with the majority of elevated radon occurrences. Black, carbonaceous shale and siltstone lithologies deposited as lake-bottom sediments almost everywhere are enriched in uranium. Gray sandstone beds deposited in lake margin environments during both the filling and drying up of lakes also are commonly enriched in uranium. The distribution of uranium mineralization within lake-bed cycles appears to have been controlled both by organic matter content and original porosity of the sediments.

INTRODUCTION

Radon gas has been identified as a cause of lung cancer from studies of uranium mine workers (Lundin and others, 1971; Whittemore and McMillan, 1983) and has been recently found at elevated levels in many New Jersey homes. The natural occurrence of radioactive radon gas in homes became widely publicized in the eastern United States following the discovery of extremely high levels in Boyertown, Pennsylvania, in December of 1984. In the spring of 1986, high radon levels were found throughout a subdivision of homes in Clinton, New Jersey.

The Pennsylvania and New Jersey occurrences are associated with uranium-rich zones in the Precambrian crystalline rocks of the Reading Prong and in adjacent Paleozoic carbonate rocks, respectively. Radon is a product of radioactive decay of ^{238}U and ^{232}Th, the main naturally occurring isotopes of uranium and thorium. High radon levels in the Reading Prong are not surprising. The crystalline rocks of the Reading Prong have long been known to have radioactive mineral occurrences and the area had been identified as a prime area of uranium resources by the National Uranium Resources Evaluation Program (NURE) of the 1970s.

What has been surprising in developing a radiation protection program in New Jersey is that nearly half of the effort has been concentrated in the Piedmont Province. Again, geologic evidence should have led to this expectation. The NURE studies noted the potential for uranium mineralization in the Triassic rocks of the Newark Basin (Popper and Martin, 1982), which comprises most of the Piedmont Province in New Jersey. This potential was highlighted further in a paper by Turner-Peterson and others (1985) on the Lockatong and Stockton Formations.

A clearer picture is emerging of the geological controls of uranium distribution in the Newark Basin from data collected by the New Jersey Geological Survey over the past two years. Newark Basin data are discussed below in the context of statewide trends in radon. The regional evaluation of radon hazards includes: 1) aerial radiometry from the NURE program; 2) a 6000-home survey of radon throughout New Jersey; and 3) confirmatory measurements of indoor radon from the State radon program. In addition, preliminary information is presented on indoor radon levels at home sites and on radioactivity in measured geologic sections in the Newark basin.

REGIONAL TRENDS

Aerial Radiometric Data

As part of the NURE program, abundant aerial radiometric data were collected over New Jersey by the U.S. Department of Energy in the late 1970s. The original purpose of the program was to define potential uranium resources and identify areas for further exploration. Data collected during the geophysical surveys included gamma spectrometer counts of bismuth-214, thallium-208, and potassium-40. Bismuth-214 is a decay product within the uranium-238 decay series and can be used to estimate concentrations of uranium or radon, assuming radioactive equilibrium.

NURE flight-line data in New Jersey are spaced at three-mile intervals in the northern part of the State and six-mile intervals in the southern part of the State (Fig. 1). In addition, a 1/4-mile-spaced survey was conducted over a large portion of the Reading Prong. As part of a statewide radon study, these data sets were reprocessed and contoured. Areas of anomalous bismuth-214 gamma activity, defined as exceeding 3 standard deviations above the mean of 2.4 ppm equivalent uranium, were highlighted as having the greatest source potential for elevated radon (Fig. 2). Fifty of the 6 ppm anomalies were reflown to assure adequate delineation and to verify the reliability of the data; six of these were in Triassic rocks.

Indoor Radon Survey

New Jersey is divided broadly into six geologic regions or provinces for the purposes of assessing regional trends in radiometric signatures. Radon statistics for a 6000-home survey are summarized in Table 1 along with the NURE equivalent uranium averages associated with the sampled homes. Provinces with mean equivalent uranium (eU) exceeding the statewide mean (2.4 ppm eU) have associated average radon levels exceeding 4 pCi/l. These provinces also contain most of the NURE anomalies which exceed the mean by three times the standard deviation (Fig. 2). Conversely, provinces with average equivalent uranium values less than the statewide mean have associated radon averages less than 4 pCi/l and contain only a few NURE anomalies.

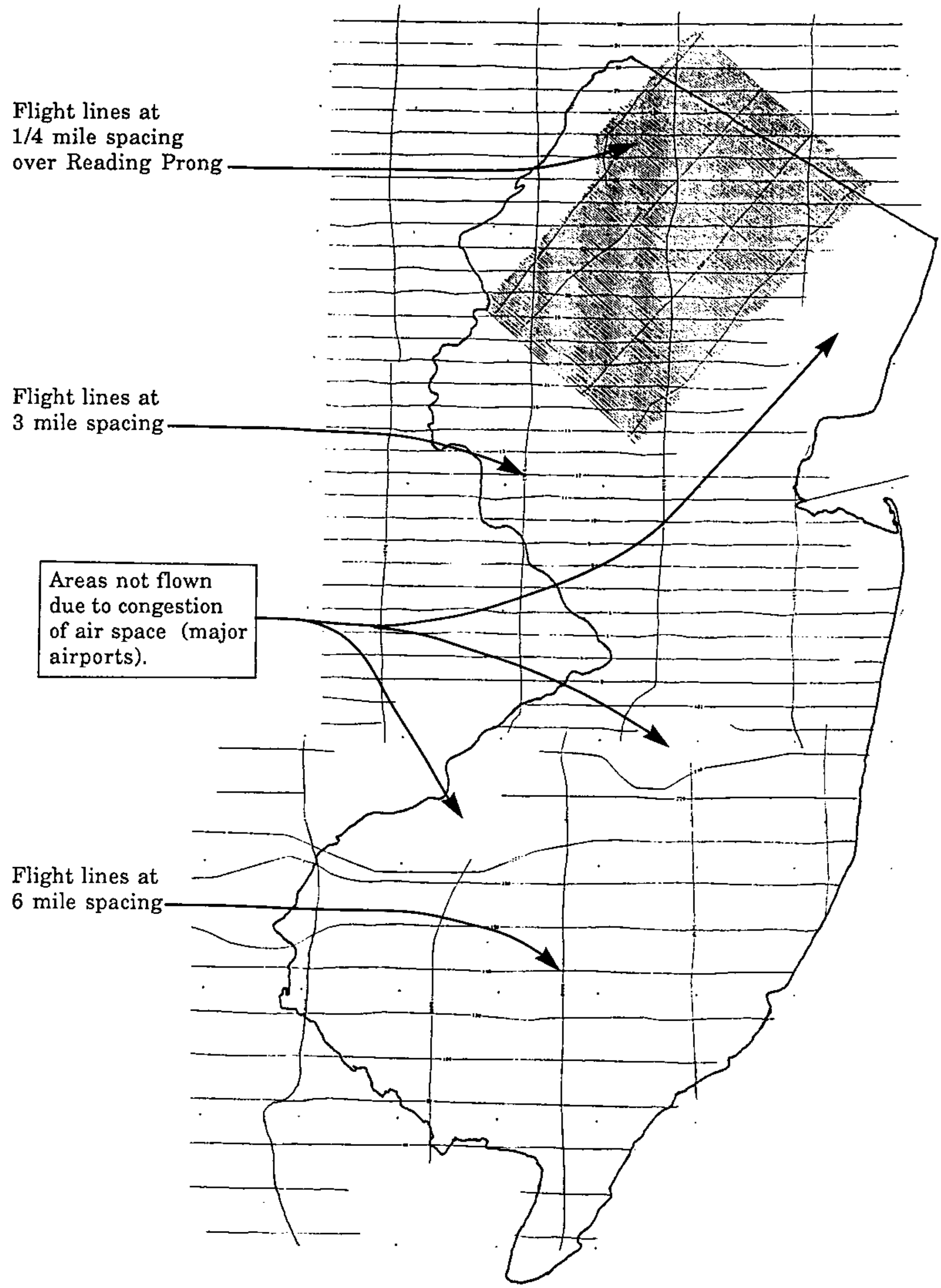

Figure 1. Flight line locations of NURE aerial radiometric data in New Jersey.

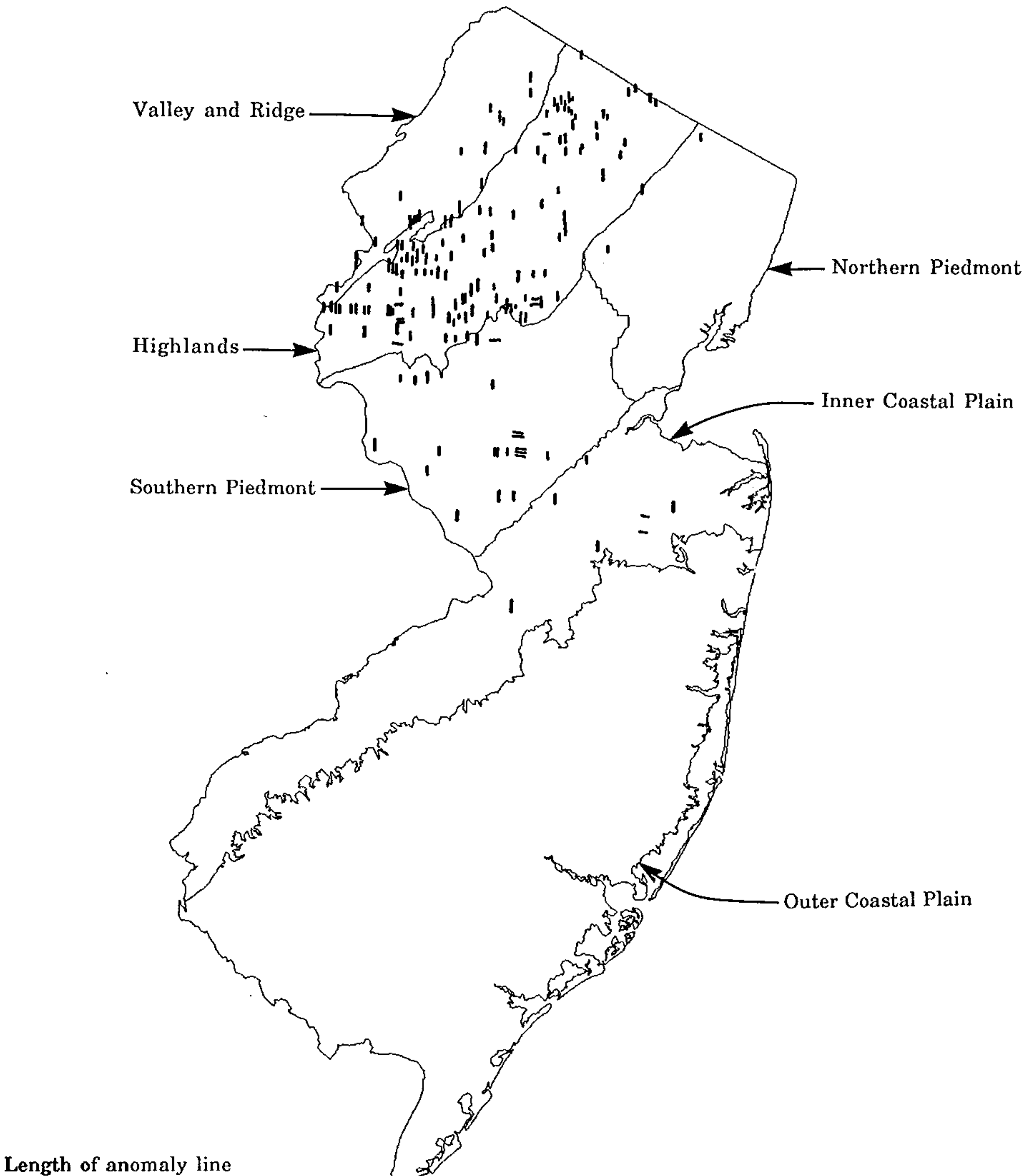

Figure 2. Locations of NURE gamma anomalies greater than 3 standard deviations above the mean (2.4 ppm equivalent uranium).

TABLE 1. INDOOR RADON STATISTICS AND NURE DATA FOR GEOLOGIC PROVINCES

Province	Arithmetic Average*	Geometric Mean*	Percent of Homes ≥4pCi/l	≥20pCi/l	NURE eU#
Valley & Ridge	7.6	4.5	57	8	2.5
Highlands (Reading Prong)	8.6	4.2	52	9	2.6
Southern Piedmont	4.9	2.5	32	4	2.7
Northern Piedmont	1.7	1.2	6	<1	2.1
Inner Coastal Plain	2.4	1.5	14	1	1.8
Outer Coastal Plain	1.4	1.0	5	<1	1.2
Statewide	5.2	2.4	33	5	2.4

* Average and mean indoor radon values in pCi/l are from Cahill and others (1988)
# Average equivalent uranium (eU) in ppm.

Cambro-Ordovican sedimentary rocks in the Valley and Ridge Province and Precambrian crystalline rocks in the Highlands or Reading Prong are the source of the most severe radon problems in the State. Radon levels in more than 50% of the homes in these provinces exceed 4 pCi/l.

The Piedmont Province is comprised of Triassic-Jurassic sedimentary rocks of the Newark Basin. There is a dramatic difference between the portions of the province north and south of the terminal glacial moraine. The southern Piedmont has an elevated radiometric signature and moderate radon levels, with 32% of the homes exceeding 4 pCi/l. In contrast, the northern

Piedmont has much lower radon and surficial radiometric levels due partly to a mitigating cover of glacial materials. Glacial cover is also a mitigating factor in the northern portions of the other high radon provinces discussed above.

The Coastal Plain Province, covering the southern half of New Jersey, is comprised of Cretaceous through Tertiary age sediments. Significant differences in radon and radioactivity are found between the Inner (northern) and Outer (southern) Coastal Plains. The Inner Coastal Plain consists dominantly of immature marine clastics containing abundant glauconite and enriched in potassium and phosphorous. This area exhibits a more elevated radon and aeroradiometric signature (Table 1) than the Outer Coastal Plain where clean quartz sands dominate.

Confirmatory Data

The correlation of indoor radon levels with airborne radiometric data for nine anomalous areas is summarized in Table 2. These areas have been ground checked by detailed geologic mapping, soil sampling, and ground radiometric traverses. The indoor radon data was collected under a State program of confirmatory monitoring in areas of elevated radon.

TABLE 2: NURE VALUES FOR INDOOR RADON CLUSTER AREAS

LOCATION	NURE eU ppm*	Percent of Homes Monitored >4pCi/l	>20pCi/l	>200pCi/l
Clinton	10	96	80	35
Montgomery	9	67	27	7
Ewing	8	78	40	5
Princeton	9	75	28	2
Bethlehem	7	75	35	11
Hampton	11	100	64	14
Bernardsville	9	87	45	9
Mansfield	10	88	65	7
Washington	10	87	57	13

* Peak value for the anomaly in the NURE data.

All studied localities are located within or immediately adjacent to airborne radiometric anomalies which exceed 6 ppm eU. As might be expected, the spatial correlation of anomalies with elevated indoor radon is improved with more closely spaced data. Anomalies in the 1/4-mile data are well defined, detected on multiple flight lines, and coincident with clusters of elevated radon homes (Bernardsville, Mansfield, and Washington). Anomalies detected in the three-mile data occur on one flight line, frequently are adjacent to areas where elevated radon has been detected, and occur along continuations of a geologic unit which causes home radon problems.

GEOLOGIC SOURCES OF RADON

Newark Basin Radon Clusters

The Montgomery, Ewing, and Princeton clusters of elevated indoor radon are situated on Triassic rocks of the Newark basin. All of these sites are located on uranium-enriched lithologies of the Lockatong and upper Stockton Formations. A sequence of strata near the base of the Lockatong Formation is associated with consistently high indoor radon levels . In the Ewing area strata which are higher in the Lockatong Formation also cause radon problems. The NURE data in this region are based on lines flown three miles apart. Three sigma anomalies in equivalent uranium were detected along the Lockatong-Stockton contact zone about one mile along strike from homes at the Montgomery and Ewing sites and a half mile away at the Princeton site. Helicopter and ground radiometric surveys conducted in 1987 showed that this zone of uranium enrichment is continuous and stratigraphically controlled. Radon problems are predicted wherever homes are built along this zone. Recent testing in two other areas along this contact zone have confirmed the prediction of elevated indoor radon.

Occurrences of Uranium-Thorium

Uranium-thorium (U/Th) mineralization has been reported in two main modes of occurrence in the Newark basin:

1) Scattered patchy, or disseminated mineralization in sandstone of the Stockton Formation. Johnson and McLaughlin (1957) reported torbernite ($Cu(UO_2)_2(PO_4)_2$ $8\text{-}12H_2O$) in a limonitic sandstone 'vein' north of Stockton, New Jersey. McCauley (1961) reported disseminated metazeunerite ($Cu(UO_2)_2$-$(AsO_4)_2$ $8H_2O$) and meta-autunite ($Ca(UO_2)_2(PO_4)_2$ $2\text{-}6H_2O$) in gray arkosic sandstone with brown iron-oxide specks from the Stockton Formation at Prallsville and Raven Rock, New Jersey. Turner-Peterson (1980) reported patchy alteration of clay clasts in iron-oxide-stained gray Stockton sandstone from the Raven Rock area.

2) Disseminated, submicroscopic uranium-thorium minerals in beds of black to gray carbonaceous mudstone to claystone and gray fine sandstone to siltstone in lake-bed sequences of the Lockatong and Passaic Formations. Turner-Peterson and others (1985) discussed U/Th mineralization in lake beds of the upper Stockton and the Lockatong Formations and believe uranium is complexed with humic matter. Zapecza and Szabo (1987) reported radioactive gray to black mudstone in the lower Passaic Formation. No discrete mineral phases containing U/Th have been

identified in the fine-grained lake bed occurrences. Autoradiographs of black mudstone typically show evenly distributed radioactive specks spaced a few mm apart, and commonly show radioactive concentrations in coarse-grained laminae.

The first mode of occurrence is only reported from the Stockton Formation near the Delaware River. It is not known whether the Stockton sandstone occurrences are localized or widespread. No radon problems have been identified as being associated with arkosic sandstone facies, although some radon occurrences in Princeton may be localized along fault zones in the Stockton sandstone.

Uranium-thorium mineralization of the second type, in bedded lake-bottom, lake-margin, and playa deposits of the Lockatong and lower Passaic Formations, is abundant and widespread and has been identified as a source of regional radon anomalies.

Radioactivity Anomalies in Lacustrine Cycles

Basin-fill sediments of the uppermost Stockton, the entire Lockatong, and the lower half of the Passaic Formations record a repeating succession of dry basin flooding, lake filling, lake recession, and return to dry basin conditions. These climate-forced cycles, which are discussed at length by Van Houten (1962, 1964) and Olsen (1980, 1984) produced rhythmically-bedded sedimentary sequences mostly 3-10 m thick. The cycles can be divided into three divisions (Fig. 3). Division 1, which records the onset of more humid conditions, commonly consists of greenish-gray fine sandstone or laminated tan siltstone with desiccation features, disrupted bedding, or pedogenic structures. Division 2 records the lake high stand and consists of lake-bottom beds of finely laminated to medium bedded carbonaceous claystone to siltstone, commonly calcareous and pyritic in some places. Division 3 records the recession of the main lake stage and return to more arid conditions and usually consists of thick bedded to massive, bioturbated siltstone with some desiccation crack layers and evaporite-mineral casts.

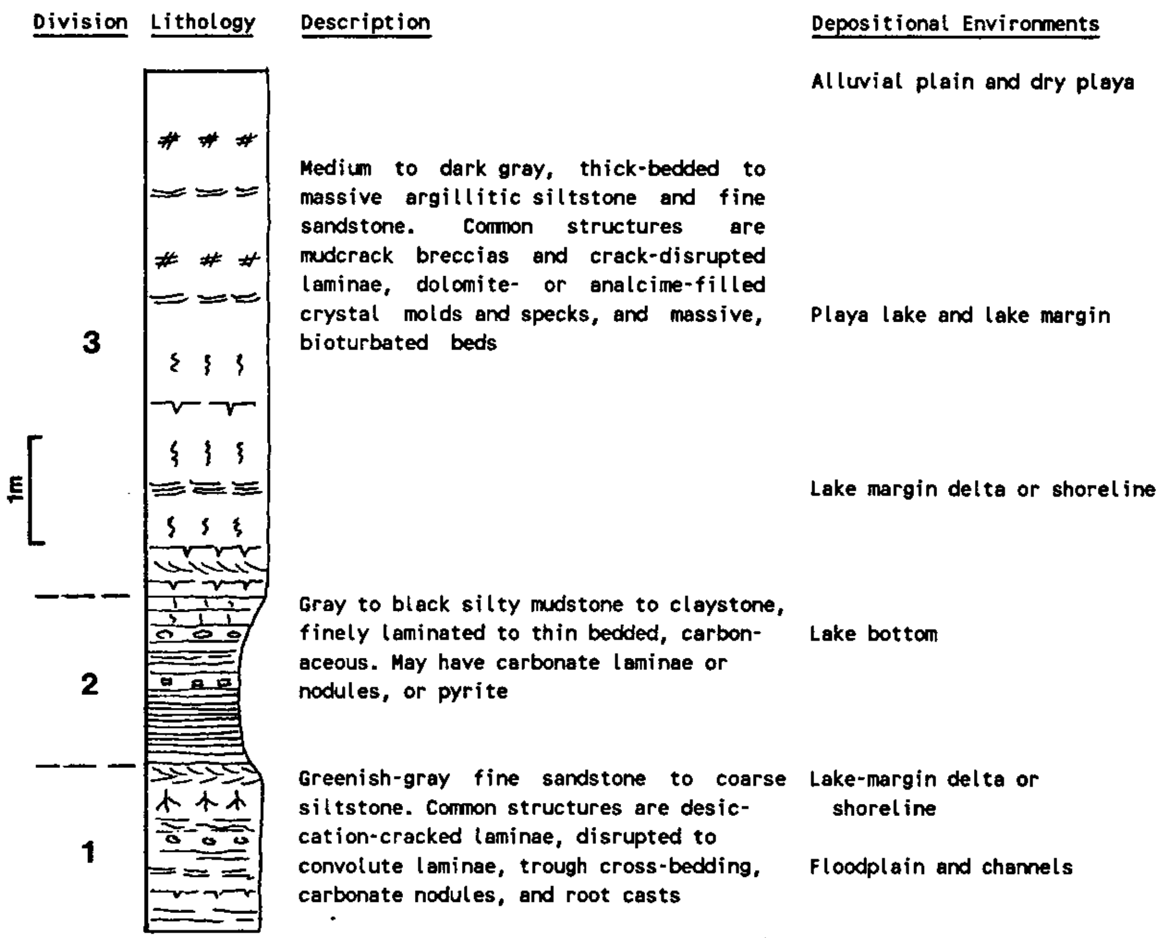

Figure 3. Diagrammatic section of a Lockatong lake cycle showing transgressive, lake high-stand, and regressive divisions.

Radioactivity anomalies occur in divisions 1 and 2, and in the lower part of division 3 with almost equal frequency (Table 3). Some anomalies have been followed along beds for more than 0.5 km and show little variation in intensity or thickness. Major anomalies can be traced over tens of kilometers with portable scintillometers.

TABLE 3. MAXIMUM RADIOACTIVITY IN INDIVIDUAL LAKE-BED CYCLES, LISTED BY CYCLE DIVISION.

Cycle number	Maximum radioactivity (CPS) Div. 1	Div. 2	Div. 3	Cycle number	Maximum radioactivity (CPS) Div. 1	Div. 2	Div. 3
B2	350	150	120	B15	225	200	150
B3	235	360	1175	B16	480	190	190
B4	160	170	200	B17	550	190	150
B5	130	150	450	B18	140	175	475
B7	130	145	180	B19	125	135	130
B8	325	800	380	B20	180	130	110
B9	300	650	130	B21	375	225	150
B10	150	250	160	WC1	900	500	200
B11	160	250	450	RR2	250	1100	250
B12	180	225	250	WV1	200	190	400
B13	190	160	280	WV2	550	250	100
B14	150	190	180	WV3	90	90	95

Note: Cycle number prefixes correspond to following localities: B = Byram cliffs on N.J. Route 29 (middle Lockatong Formation); WC = Warford Creek lake beds in lower Passaic Formation; RR = Raven Rock, outcrops in lowermost Lockatong Formation; WV = Woodsville Road, outcrops in lower Lockatong Formation. Radioactivity measured by hand-held scintillometer and given in units of counts per second (CPS).

Turner-Peterson and others (1985) reported uranium enrichment only in lake-bottom beds of division 2 and considered uranium occurrence to be controlled by the distribution of humic organic matter. However, several of the highest radioactivity anomalies are observed in lake-margin sandstones of divisions 1 and 3. Uranium/thorium mineralization appears to be controlled both by organic-matter content of lake-bottom mudstones and by porosity of sandstone beds adjacent to organic-rich layers. Additional petrographic and geochemical studies are needed to determine the exact mechanism and sequence of uranium mineralization.

CONCLUSIONS

The Early Mesozoic rocks of the Newark Basin are second only to the crystalline rocks of the Reading Prong in their geologic potential for radon hazards in New Jersey. Actually, more homes may be at risk in the Newark Basin than in the Reading Prong because of greater population density in the basin.

Clusters of homes where radon levels generally are greater than 4 pCi/l and exceed 200 pCi/l in several cases have been identified in association with lake-bed sequences of the lower and middle Lockatong Formation. Radon anomalies also occur in other areas underlain by lake-bed sequences of the upper Stockton, Lockatong, and lower Passaic Formations. Major anomalies persist within strata over distances of several kilometers to tens of kilometers. Elevated radon occurrences are common in the central Newark Basin where thick lake-bed strata are exposed in wide areas, and less common in the northern basin where lake-bed sequences are thinner and restricted to a zone above and below the Palisades sill. Surficial glacial sediments also reduce the radon hazard in the northern Newark Basin.

Radioactive anomalies occur both in dark gray to black mudstone and in gray sandstone beds of lake cycles. Uranium mineralization appears to have been controlled both by organic matter content and porosity of the sedimentary layers. Further study of the geochemistry and petrology of of the lacustrine sedimentary rocks is needed to determinme the controls of U/Th mineralization.

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EARTH SCIENCE TEACHERS WORKSHOP:

DINOSAURS OF THE NEWARK RIFT BASIN

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Introduction

Before *Hadrosaurus foulkii* was unearthed in Haddonfield, NJ, in 1858, and even before Gideon Mantell's discovery of *Iguanodon* in England, dinosaur fossils were found in New England but misidentified In 1802, a Yankee farmer named Pliny Moody plowed up a slab of red rock covered with curious impressions that resembled nothing so much as bird tracks. These were explained as the footprints of Noah's raven, in keeping with the prevalent opinion of the time which assigned all fossils to the catastrophe of the biblical Great Flood (Howard, 1975).

In the 1830s, Edward Hitchcock, a professor at Amherst College, organized an extensive collection of dinosaur footprints preserved in stone. He attributed the prints to birds, but some were so large they could not have been made by any bird known to science at that time. Hitchcock collected a number of different forms primarily from the Connecticut River valley; he gave them various scientific names based in many cases on minor differences in shape. His son, C. H. Hitchcock, first described footprints of the same sort from the Newark red beds of New Jersey (Lull, 1915). It wasn't until 1877 that the famed vertebrate paleontologist O. C. March of Yale demonstrated these tracks were made by dinosaurs.

Newark Basin Geology

Rocks of the Newark Basin underlie the Piedmont Lowland physiographic province of New Jersey, stretching diagonally across the state from the Hudson River on the northeast to Milford and Trenton on the Delaware River, and then continuing on across southeastern Pennsylvania to Gettysburg. In places, these strata attain a thickness of 10,000 meters (upwards of 32,000 ft.; Olsen and Galton, 1977). The Newark Basin is one of a series of basins extending along the east coast from Nova Scotia to South Carolina. These trough-like structures contain deposits of roughly the same age, composition, and fossil content; all their rock formations are included within the Newark Supergroup.

The world was a very different place when these red beds and lake deposits were deposited and widespread lavas flowed in the Newark Supergroup basins during late Triassic and early Jurassic times. The end of the previous era of geologic history, the Paleozoic Era, was marked by the accretion of most of the world's landmasses into the great supercontinent of Pangaea. Africa and Europe were connected to the east coast of North America and the bulge of South America was tucked into the West African bight. Thus, at the end of the Triassic, the first period of the Mesozoic Era, New Jersey lay completely land-locked deep within the interior of the giant Pangaean landmass. Continental climate conditions prevailed, and dry spells alternated with rainy seasons, producing periods of arid weathering followed by periods of widespread flooding and deposition.

By late Triassic time, Pangaea was beginning to be affected by internal stresses that eventually led to the complete breakup of the supercontinent. Huge rift valleys formed where blocks of the earth's crust dropped down along faults. The highlands surrounding these linear valleys were the source regions for huge quantities of eroded sediment which poured into the basins during rainy seasons. During subsequent dry spells, the widespread floodplain sediments would dry out, crack, and weather to a rich reddish color as a result of the oxidation of the iron contained in the silts and mud. The early dinosaurs walked across these mudflats, leaving their footprints in the soft, drying sediments to be baked and hardened and, eventually, covered by the next flood's deposits.

Associated with the red beds are lake deposits and basaltic lava flows. The cyclical extremes of climate produced fluctuating bodies of water, and the stresses associated with rifting were accompanied by volcanic activity. A modern analog to this geological environment is the East African Rift Valley, where tensional stresses are tearing apart the African continent today.

Triassic Fossils

The earliest sediments deposited in the Newark Rift Basin do not yield many dinosaur footprints. Instead the deposits of the Stockton Formation, with surface exposures largely along the southeastern margin of the Newark Basin, contain the remains of large amphibians called metoposaurs, coelocanth fish, and phytosaurs (relatives of the ancestral dinosaurs that resembled crocodiles). Plant fossils also are found in the Stockton Formation.

The next youngest rock unit is the Lockatong Formation. This Formation consists largely of gray siltstones and argillites deposited in a playa lake, a body of water whose areal extent expanded and contracted in response to the wet-dry climate cycles. In some layers, tremendous numbers of coelocanth fish (Diplurus) and other primitive bony fish (Semionotus) are preserved, suggesting mass mortality as the Lockatong lake waters evaporated and receded. Phytosaurs also are found in the lakebed deposits, sometimes in proximity to the fish (Baird, 1986). Another interesting find is the remains of a gliding lizard, Icarosaurus sieferi (Colbert, 1966), from the Granton Quarry in North Bergen, Hudson County, New Jersey. The plant fossils found in the dark gray argillites represent a flora of ferns, cycads, and conifers.

Dinosaur footprints are found at various stratigraphic levels within the Lockatong Formation. The animals which produced them probably were attracted by water and food resources and left their tracks in the mud around the lake margins. These early dinosaurs were small compared to later Cretaceous types with which most people are familiar. Typical of these early, small dinosaur footprints is the form known as Grallator (figure 1). Grallator probably represents a primitive, generalized bird-like carnivorous dinosaur of the type called coelurosaurs. Its diminutive footprints (3-4 inches on average) display three slender toes, with the two side toes unequal in size and smaller than the central digit. A number of species of this footprint form have been named, although factors such as growth stages, sexual dimorphism (male vs. female forms), varying gait, and various preserving sediments (sand, silt, wet mud, dry mud) all

Figure 1

<u>Anomoepus</u> footprint (X 0.7)
(primitive ornithopod)

<u>Anchisauripus</u> footprint (X 0.7)
(ancestral saurischian dinosaur)

<u>Grallator</u> footprints (X 0.7)
(small theropod dinosaur)

played a part in producing the range of types, so that there probably were fewer real species present than indicated by the many names given to the various footprint forms. This always is a problem with ichnofossils (trace fossils) which usually display a bewildering variety of forms. The *Grallator* footprints are thought to correspond to the skeletal remains of *Coelophysis*, a light, agile predator known from body fossils found in the Upper Triassic of New Mexico. These coelurosaur types were an early evolutionary success and persisted throughout the Mesozoic.

Another dinosaur footprint form, or ichnogenus, belongs to a second group of dinosaurs, the ornithischians. *Atreipus* tracks consist of hand and foot impressions, indicating an animal that walked on all fours, unlike the bipedal *Grallator* (Olsen and Baird, 1986). The hindprints are larger and three-toed, while the forelimb prints are much smaller. The animal that made these tracks probably was a very primitive form on the line to the hadrosaurs, ceratopsians, and other ornithischian dinosaurs. *Atreipus* prints range stratigraphically from the Lockatong Formation to the overlying rock unit, the Passaic Formation.

Expansion and contraction of the Lockatong lakes has been linked to variations in the earth's orbit and inclination (Olsen, 1986). Climatic conditions apparently grew more arid toward the end of the Triassic, as the dark gray lacustrine sediments became more limited in their extent and deposition of red beds predominated. The red mudstones, siltstones, and sandstones of the Passaic Formation reflect this increasing dryness (Van Houten, 1964, 1969).

The lower Passaic Formation has yielded some typical late Triassic fossils. Non-dinosaurian tracks include footprints of phytosaurs called Apatopus and the footprints known as Rhynchosauroides, representative of the rhynchosaurs. These were clawed, sheep-sized, parrot-headed reptiles that became extinct at the end of the Triassic.

Still other trackways present in the lower part of the Passaic are the footprints known as Cheirotheirium and Brachycheirotheirium, probably made by rauisuchid thecodonts, or pseudosuchians (literally, "false crocodiles"). These were animals close to the ancestral lineage of both dinosaurs and crocodiles.

A number of non-dinosaurian skeletal remains also have been found in the Passaic deposits. These include: phytosaur jaws and teeth; a skull and footprints of a horned-stem reptile, the procolophonid Hypsognathus fenneri; and an armored aetosaurian thecodont, Stegomus arctuatus, known from impressions of its armadillo-like body (Baird, 1986).

Jurassic Dinosaurs

Dinosaur footprints of the small grallatorid type and Atreipus are known throughout most of the Passaic Formation. At the very top of the Formation, a dramatic change occurs: Dinosaur footprints increase in number within the top meter of the Formation just below the Orange Mountain Basalt, the first lava flow of the Watchung Mountains. Several footprint forms suddenly become abundant at this level. The four-toed hand-and-foot impressions known as _Batrachopus_ probably were made by a crocodylomorph, a small, primitive ancestral crocodile. Besides the long-ranging _Grallator_, dinosaur footprints include _Anchisauripus minusculus_ and _Anchisauripus sillimani_. _A. sillimani_ (figure 1) is a medium-sized print, averaging 5 to 7 inches (13 to 17 cm), while _A. minusculus_ is a larger print, usually 12 to 14 inches (30 to 36 cm). There has been some speculation that all three dinosaur footprint forms represent growth stages of the same animal; to express this relationship, _Anchisauripus_ now is referred to as _Grallator_ (_Anchisauripus_) (see, for example, Olsen and Baird, 1986). A principle feature of _G._ (_Anchisauripus_) is the impression of the backward-pointing hallux claw, although the imprint of this fourth toe is not always apparent in all specimens.

The uppermost Passaic, just below the first of the Watchung lava flows, now is considered early Jurassic in age (Olsen, 1980). This uppermost meter of sediment reflects the change change in the fauna associated with the Triassic-Jurassic _Hypsognathus_, aetosaurs such as _Stegomus_, and various mammal-like reptiles disappeared. They were replaced in the Early Jurassic by diverse species of dinosaurs, crocodiles, and mammals (Olsen and others, 1987).

The three major lava flows of the Watchung basalts reflect the increasing tensional stress of Pangaean rifting. Igneous activity was widespread in the Eastern North American rift basins as the long process of supercontinental breakup began in the Early Jurassic, ultimately leading to the formation of the

Atlantic Ocean. It was within this tectonic setting that dinosaurs became widespread and, at least, locally abundant.

In the Feltville Formation, above the Orange Mountain Basalt, the footprint form called Anomoepus first appears in the fossil record. This track type is distinguished by its gracile toes (see figure 1). Anomoepus is an early ornithischian footprint, more typically ornithopod than the more primitive Atreipus prints. As with those Triassic prints, Anomoepus sometimes may be recovered in hand-foot sets, the hand showing five toes and being distinctly smaller. The Feltville also contains Grallator and Anchisauripus footprints.

Above the Feltville is the Preakness Basalt (Second Watchung Mountain) and then another series of red beds with some gray lake deposits interbedded. This is the Towaco Formation, which has produced numerous dinosaur footprints at Riker Hill in Roseland, New Jersey, and elsewhere. In addition to Grallator, Anchisauripus, and Anomoepus, another footprint form becomes common at this level, the gigantic Eubrontes. Measuring up to 18 inches (46 cm) in length from heel to middle claw, this three-toed form represents an early large dinosaur, probably one of the first of the great meat-eaters.

The Towaco Formation is capped by the Hook Mountain Basalt, the third and youngest of the Watchung basalt flows. The uppermost formation of the Newark supergroup in New Jersey is the Boonton Formation, located along the northwest margin of the rift basin. In addition to dinosaur footprints found in the red beds, a large number of fish (Semionotus) were uncovered during excavations for a reservoir at Boonton in Morris County. Faults cut off the northwestern boundary of the Newark Basin and the crystalline rocks of the New Jersey Highlands define the edge of the old rift basin. Very coarse-grained tongues of conglomerate are found along the northwestern margin. These beds often are referred to as the Hammer Creek Conglomerate, and seem to represent alluvial fans of the type found on the fringes of arid mountains where flash floods deposit large pebbles and cobbles.

Conclusions

Within the extensive thickness of the Newark Basin deposits, we can see the story of the early dinosaurs rise to a dominant position among land animals. While initially the dinosaurs do not appear to have been an overwhelming and immediate success, after the Triassic-Jurassic extinction they apparently took over most of the environmental niches available to large land animals.

It is interesting to speculate on what led to the dinosaurs dominance of the land during most of the Mesozoic Era. Virtually identical footprint forms are widespread in rocks of Newark Supergroup age; they are found in the American Southwest (Lockley, 1986), Wales (Tucker and Burchette, 1977), continental Europe, South Africa (Haubold, 1986), and Iran (DeLapparent, 1973). As a group, dinosaurs were characterized by their ability to rear back on their hind limbs and go about in a more or less two-legged gait; this type of stance is called "facultative bipedality." Changes in limb proportions and hip structure allowed them to adopt this semi-erect posture. It is tempting to suppose that this mode of locomotion gave the early dinosaurs an advantage in the broken landscape of Pangaea, where rugged mountains alternated with deep rift valleys. Certainly, bipedality would aid a group's dispersion across such rough terrain, and the continental climate may have modified the metabolic processes of the dinosaurs as well. The origins of warm-bloodedness may lie in the necessity of adapting to fluctuations of extreme cold and heat, leading perhaps to size increase as a way of retaining body heat. As a whole, dinosaurs may be characterized as Pangaean animals, a large part of whose development was recorded in the interior rift valleys of the rupturing supercontinent.

Methods For Studying Dinosaur Tracks

Collecting

Obviously, the most desirable technique for the study of dinosaur tracks is to quarry them out of the rock and take them back to the laboratory, museum, or university for further scrutiny. This is not always practical for several reasons. Sometimes the material is on private property or publicly protected land where collecting is not permitted. One should obtain permission from the appropriate authorities before attempting to extricate tracks. The stratum that contains the actual mold of the footprint (negative) may be very weak incompetent mudstone or siltstone that easily shatters upon striking. In thin-bedded layers, the footprints will be the thinest and weakest section of the bed, and hence will be the most likely place for the rock to break. Usually the overlying stratum is a coarser grained, stronger sandstone that slabs out very nicely, preserving intact the natural cast (or positive) of the footprint. If footprints are cracked upon extraction, the pieces may be reassembled more easily in the lab with the aid of "witness marks," arrows written in indelible ink on the fragments indicating how the pieces should be reassembled.

Finally, transporting and storing lots of large and heavy slabs becomes a major logistical challenge, and one may wish to consider other ways to "collect" dinosaur footprints.

Casting

Plaster or latex casts may be made in the field or lab. A simple casting frame may be made out of firring strips or other lumber. An adjustable frame may be constructed by nailing together two sections to form an "L" shape, and then accommodating this frame to any size footprint by sliding and adjusting the two sides past each other to the appropriate length, securing them with thick rubberbands. This will conserve plaster and thus decrease drying time.

While plaster is readily available, latex enjoys several advantages: it is lighter in weight, less brittle, sets more quickly, produces a more authentic cast, and is easier to store and transport.

Photography

Frequently the best way to "collect" footprints is to take pictures of the tracks. This can provide a record of their field associations to other tracks and to the rocks in which they are found. Photographs of faint impressions may be enhanced by wetting or chalk outlining. Use of lower angle lighting to increase shadowing may be helpful.

Mapping

Drawing a detailed field map of a bedding plane surface can aid in preserving useful data about dinosaur tracks such as relationships of trackways to each other and to sedimentary features, compass orientations (including preferred directions, if any), footprint length, stride length, and diversity and density of footprints.

Rubbing

A quick method of preserving a footprint image is to make a rubbing of the track. When I first saw this technique applied to gravestones, I decided to experiment with rubbings of dinosaur footprints. The results were accurate, and the method is simple and inexpensive. Simply take a sheet of unlined white paper (typing paper will do), place it over the print (positive or negative), and run the side of the lead of a soft pencil (#1 or#2) over the surface of the paper as it is pressed on the track. The ichnofossil will be reproduced accurately as a dark shading on the paper (see figure 1). This is an easy and economical way to mass-produce dinosaur footprint images.

Localities

While the potential for discovering dinosaur footprints exists in most parts of the Newark Basin, tracks have been found in the past at Milford, Whitehall, Avondale, Pompton Falls, New Vernon, Roseland, Clifton, and Newark (Gallagher, 1983). No dinosaur bones of Triassic-Jurassic age have ever been found in New Jersey, although dinosaur bones have been found in the Newark Supergroups strata of the Hartford Basin, Connecticut, as well as in the Bay of Fundy area of Nova Scotia (Olsen and others, 1987).

Acknowledgements

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FIELD GUIDES

FIELD GUIDE TO THE LATE TRIASSIC PORTION OF THE NEWARK BASIN SECTION IN THE DELAWARE VALLEY, NEW JERSEY

By

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with contributions by

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INTRODUCTION

Eastern North America includes the classic Atlantic-type passive continental margin formed by the breakup of the supercontinent of Pangaea. The Triassic initiation of the breakup was marked by the formation of rifted crust all along the axis of the future Atlantic, from Greenland to Mexico. In eastern North America, nine major rift basins, mostly half-graben, and several minor basins are exposed from Nova Scotia to South Carolina, with many more buried below the coastal plain and on the continental shelf (Figure 1). The exposed rift basins, which closely follow the trend of the Appalachian orogen, filled with thousands of meters of continental sediments, minor mafic volcanic rocks, and diabase plutons and dikes over a period of 45 million years. The faulted, tilted, and eroded rift strata are termed the Newark Supergroup (Van Houten, 1977; Olsen, 1977; Froelich and Olsen, 1984). The purpose of this field trip and paper is to provide a review of the geology of the Newark Basin (Figures 1, 2) division of the Newark Supergroup in the Delaware Valley.

GEOLOGICAL CONTEXT

Newark Supergroup rifts lie entirely within the Appalachian Orogen which, as presently understood, is divisible into three broad structural zones: 1) a western thin-skinned zone of allochthonous sheets lying on the essentially undeformed North American craton; 2) a central thick-skinned zone of rooted thrust slices and accreted terranes; and 3) an eastern zone transitional

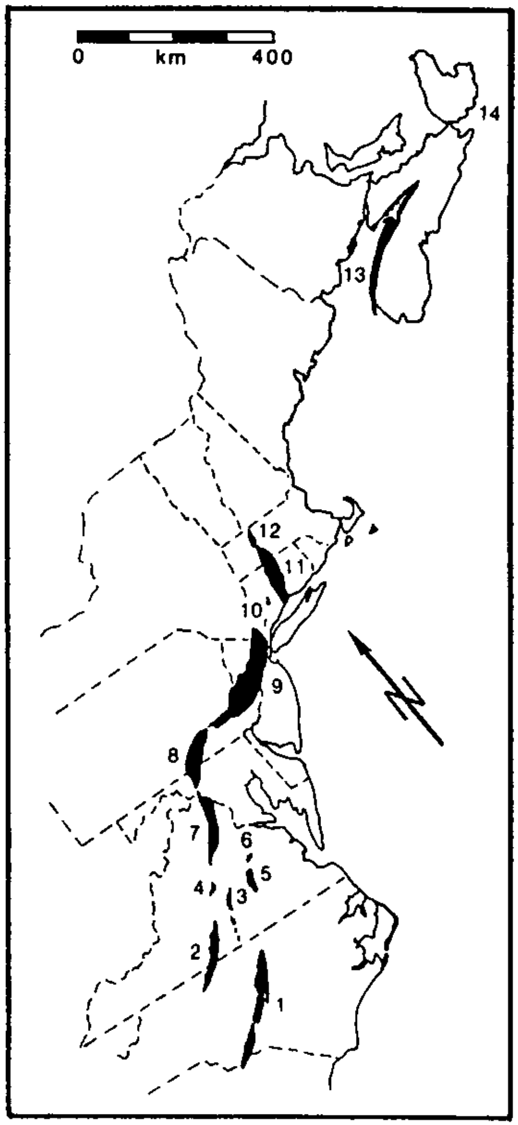


Figure 1. The Newark Supergroup of eastern North America (in black): 1) Deep River Basin; 2) Dan River Basin; 3) Farmville Basin Complex; 4) Scottsville Basin; 5) Richmond Basin; 6) Taylorsville Basin; 7) Culpeper Basin; 8) Gettysburg Basin; 9) Newark Basin; 10) Pomperaug Basin; 11) Hartford Basin; 12) Deerfield Basin; 13) Fundy Basin; 14) Chedabucto Basin (=Orpheus Graben).

to oceanic crust. The western zone consists of thin (2-10 km) allochthonous slices separated from relatively undeformed North American craton by a low angle detachment (corresponding to zone 2 of Allmendinger and others, 1987). This thin-skinned zone is separated from the thick-skinned zone by an east-dipping ramp which may extend to MOHO (zone 3 of Allmendinger and others, 1987). In proximity to the ramp (but regionally discordant to surface trends) is the Appalachian gravity gradient which is often considered to mark the edge of the Precambrian-Early Paleozoic continent. Above and immediately to the east of this ramp are rooted slices of cratonic rocks and to the east of this is the first of several apparent sutures with accreted terranes (within zone 4 of Allmendinger and others 1987). The eastern zone, transitional to oceanic crust, is as yet poorly known. It lies to the east of the basement hinge zone of the continental crust and probably consists of fragments of continental crust and Mesozoic igneous rocks (Klitgord and others, 1988).

Cook and others (1981) have plausibly argued that development of the Precambrian and earliest Paleozoic rifting and Iapetus passive margin established the configuration of the eastern edge of the North American craton. A succession of Paleozoic accretionary and collisional events reactivated the old Precambrian-Paleozoic continental margin as a ramp, with repeated emplacement of thin allochthonous slices toward the west in fold and thrust belts. The presumably thickened crust was then thinned during the formation of the present Atlantic margin in the Early

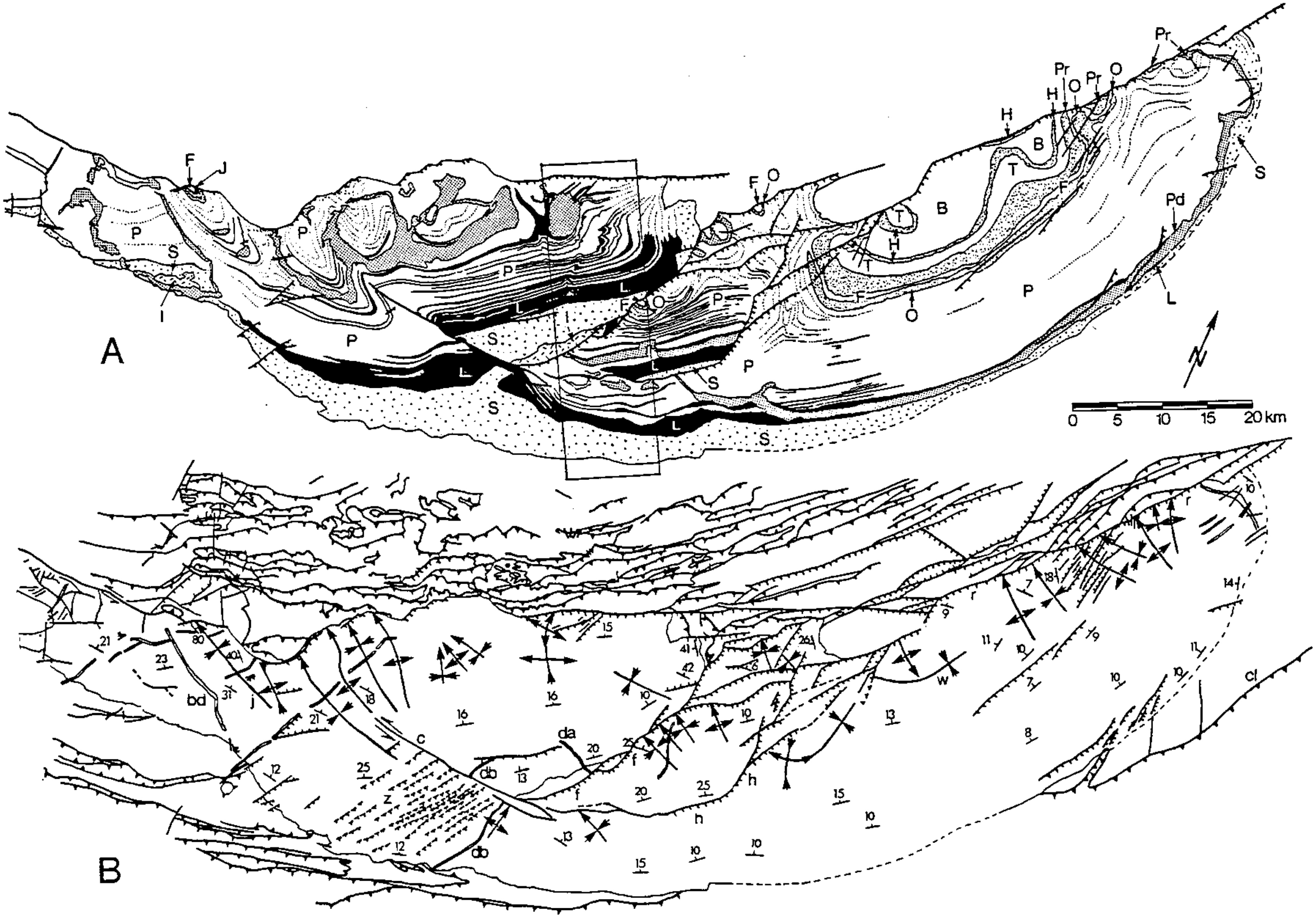


Figure 2. Geologic (**A**) and Structure (**B**) maps of the Newark Basin. Inset in (**A**) shows position of field stop key, Figure 7. Abbreviations as follows: **B**) Boonton Formation; **F**) Feltville Formation; **H**) Hook Mountain Basalt; **I**) Inlier of Paleozoic and Precambrian rocks; **J**) Jacksonwald Basalt (=Orange Mountain Basalt); **L**) Lockatong Formation; **O**) Orange Mountain Basalt; **P**) Passaic Formation (where conglomeritic called Hammer Creek Formation in Pennsylvania); **Pd**) Palisade Diabase; **Pr**) Preakness Basalt; **S**) Stockton Formation; **T**) Towaco Formation. Figure adapted from Schlische and Olsen (this volume).

Mesozoic, with the greatest thinning being manifest in the eastern zone of transitional crust, east of the basement hinge line.

The Newark basin is a block-faulted, deeply eroded, half-graben (Figures 2 and 3), very much like other Newark Supergroup basins. It seems likely it sits directly over the deep seated east-dipping ramp separating the thin skinned western zone from the thick skinned, allochthonous eastern zone. Most of the basin's northwest margin is bound by low angle (20°-65°) normal, right and left oblique faults which are reactivated thrust faults developed during the Paleozoic Taconic, Acadian, and Alleghenian Orogenies (Ratcliffe, 1980; Ratcliffe and Burton, 1985; Ratcliffe and others, 1986) (Figure 3).

The Newark Basin appears to have been a relatively simple half graben bound by these faults during most of it history (Olsen, 1985; Schlische and Olsen, this volume). However, the basin is also cut by two major intra-basin faults (Flemington-Furlong and Hopewell-Chalfont faults) which break the basin into three fault blocks, repeating most of the section three times (Figure 2). These two intra-basin faults probably post-date most basin sediments (Olsen, 1985; Schlische and Olsen, this volume; Faill, this volume). Most sediments in the basin dip 5° to 15° to the northwest toward the northwestern border faults. However, the hanging walls of most fault blocks are warped into a series of anticlines and synclines which die out toward the opposite sides of the blocks. Because of the block faulting and folds, most stratigraphic intervals can be examined at the surface in three

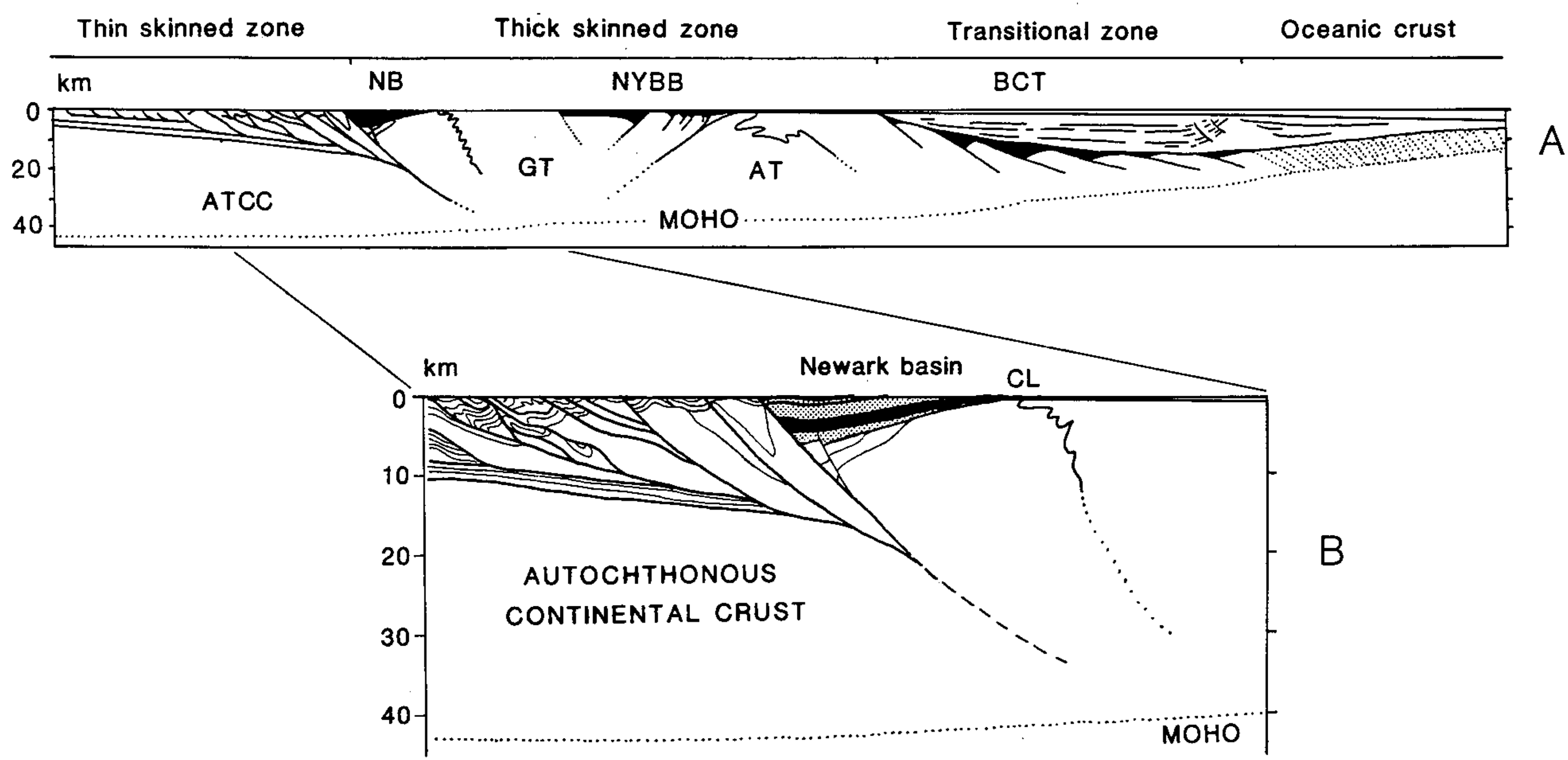


Figure 3. Generalized cross section through crust of eastern North America from New Jersey to the south east (see Figure 1) showing different crustal zones. In (**A**) rift basins are shown in black while in (**B**) The Lockatong Formation is the thick black layer in the Newark Basin while the thinner black layers higher in the section are the extrusive basalts. Abbreviations are: **AT**, Avalon Terrane; ATCC) Autochthonous continental crust **BCT**) Baltimore Canyon Trough; **CL**) Cameron's Line; **GT**) Gander Terrane; **NB**) Newark Basin; **NYBB**) New York Bight Basin. Based partly on sections in Klitgord and others (1988).

dimensions (Figure 2).

Sedimentation seems to have begun in the Middle Carnian or perhaps as early as the Ladinian of the Middle Triassic with the deposition of the predominantly fluvial Stockton Formation (Figure 4). The overlying Lockatong and Passaic formations were deposited as lacustrine ringed by fluvial sequences. Shortly after the Triassic-Jurassic boundary (preserved in the upper Passaic Formation) massive tholeiitic basalt flows covered the basin floor in three episodes, each consisting of two or more cooling units and thin sedimentary units (Orange Mountain, Preakness, and Hook Mountain basalts). These are separated by two sedimentary formations (Feltville and Towaco formations) and covered by the Boonton Formation. The syn- and post-extrusive formations were formed in similar environments to the Passaic and Lockatong formations except with a four- to six-fold increase in sedimentation rate. [The Passaic and overlying formations were included in the Brunswick Formation of Kümmel (1897) and the Watchung Basalt of Darton (1890); for a review of the stratigraphic nomenclature of the Newark Basin see Olsen (1980a and 1980b)]. Sedimentation seems to have ended by the Middle Jurassic during the period of block faulting, and this was followed by deep erosion, thermal subsidence, and coastal plain deposition along the entire Atlantic margin. On this field trip, only the Stockton, Lockatong and lower Passaic formations will be examined.

The Lockatong and Passaic formations together make up a natural

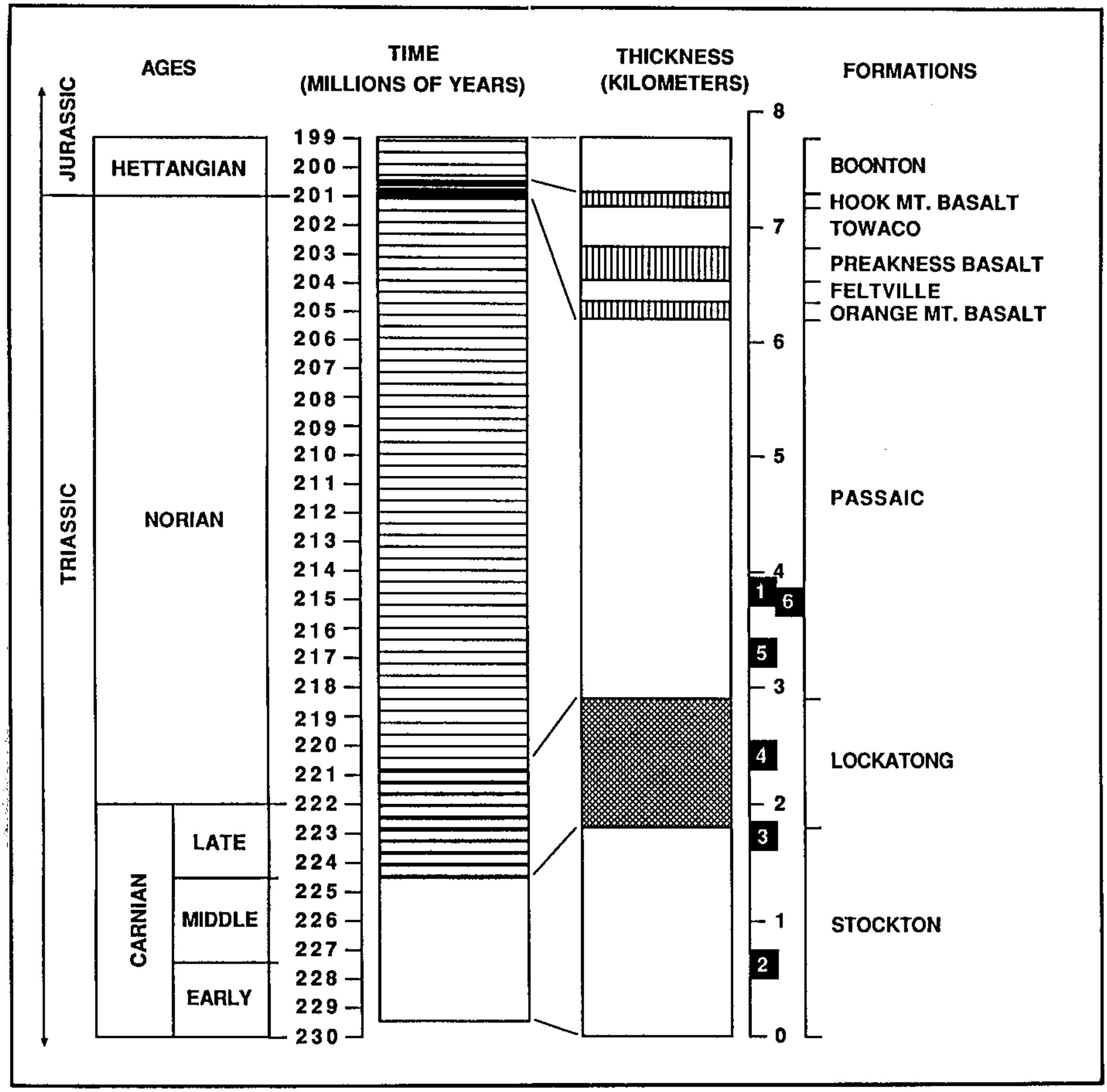


Figure 4. Stratigraphy of the Newark Basin in time based on calibration by Van Houten cycles (left) and thickness (right) showing stratigraphic position of the field stops.

facies package united by a common theme of repetitive and permeating transgressive-regressive lake level sequences called Van Houten cycles after their discoverer (Van Houten, 1964, 1969, 1980; Olsen, 1980a, 1980c, 1984, 1986). The fundamental Van Houten cycle consists of three divisions interpreted as lake transgression, high stand, and regression plus low stand facies (Figure 5).

1. Division 1 is a calcareous claystone to siltstone with pinch and swell lamination, thin bedding, occasional desiccation cracks, burrowed horizons, stromatolites, and oolites. This represents a lacustrine transgressive sequence with shoreline deposits.

2. Division 2 is a laminated to microlaminated calcareous claystone and siltstone, or limestone, with rare to absent desiccation cracks. This division has the highest organic content of the cycle and fish and other animal fossils are sometimes abundant; it represents the lake's high stand deposit.

3. Division 3 is a calcareous claystone and siltstone, with abundant desiccation cracks, burrowed and rooted horizons, vesicular and crumb fabrics, and reptile footprints. Division 3 represents the regressive and low stand deposit; the lake was at least occasionally dry with incipient soil development.

These Van Houten cycles were apparently produced by the rise and fall of very large lakes that responded to periodic climate changes controlled by variations in the earth's orbit (Van Houten, 1969; Olsen, 1984, 1986). Van Houten cycles also make up at least three orders of compound cycles (Figure 6). Fourier analysis of

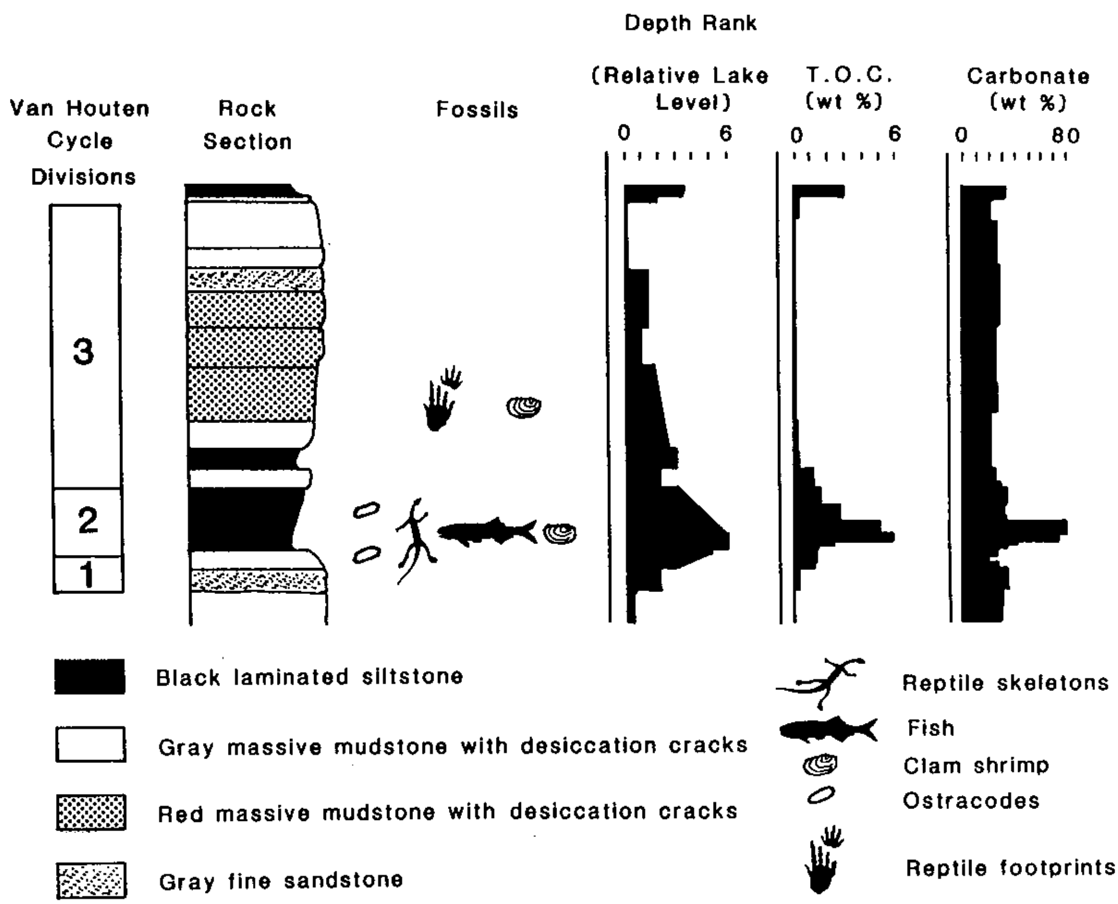


Figure 5. Generalized Van Houten cycle.

long sections of the Late Triassic Lockatong and Passaic formations show periodicities in time of roughly 21,000, 41,000, 95,000, 125,000 and 400,000 years, calibrated by radiometric time scales and varve calibrated sedimentation rates. These periodicities are in accord with the orbital theory of climate change.

The orbital theory of climate change states that celestial mechanical cycles of the earth's orbit, such as the precession of the equinoxes (21,000 years), the obliquity cycle (41,000 years), and the eccentricity cycle (95,000, 123,000, and 413,000 years), produce changes in the seasonal and geographical distribution of sunlight reaching the earth which, in turn, affects climate (Berger, 1984). The theory makes specific predictions about the periodicity of climate cycles which should be testable in the geological record. Recently, the predictions of the orbital theory have been convincingly confirmed through studies of climate-sensitive geochemical and biotic indices in deep sea sediments of Quaternary age, especially in middle and high latitudes (Hays, Imbrie, and Shakleton, 1976). The Lockatong periods are also in rather close agreement and are strong evidence for orbital forcing far into the past. Long sequences of sedimentary cycles similar to those in the Lockatong and Passaic formations occur through the Jurassic portions of the Newark basin and most of the rest of the Newark Supergroup, spanning a period of more than 40 million years. Ultimately these lacustrine cycles will allow for extremely tight controls on sedimentation rates, stratigraphy, and structural development (see Olsen, this volume).

The route of the field trip is shown in Figure 7. The trip covers the three central fault blocks of the Newark Basin, running up section though the Stockton, Lockatong, and Passaic formations. Apart from Stop 1, all the stops are in the northern (Hunterdon Plateau) fault block where the exposures are excellent.

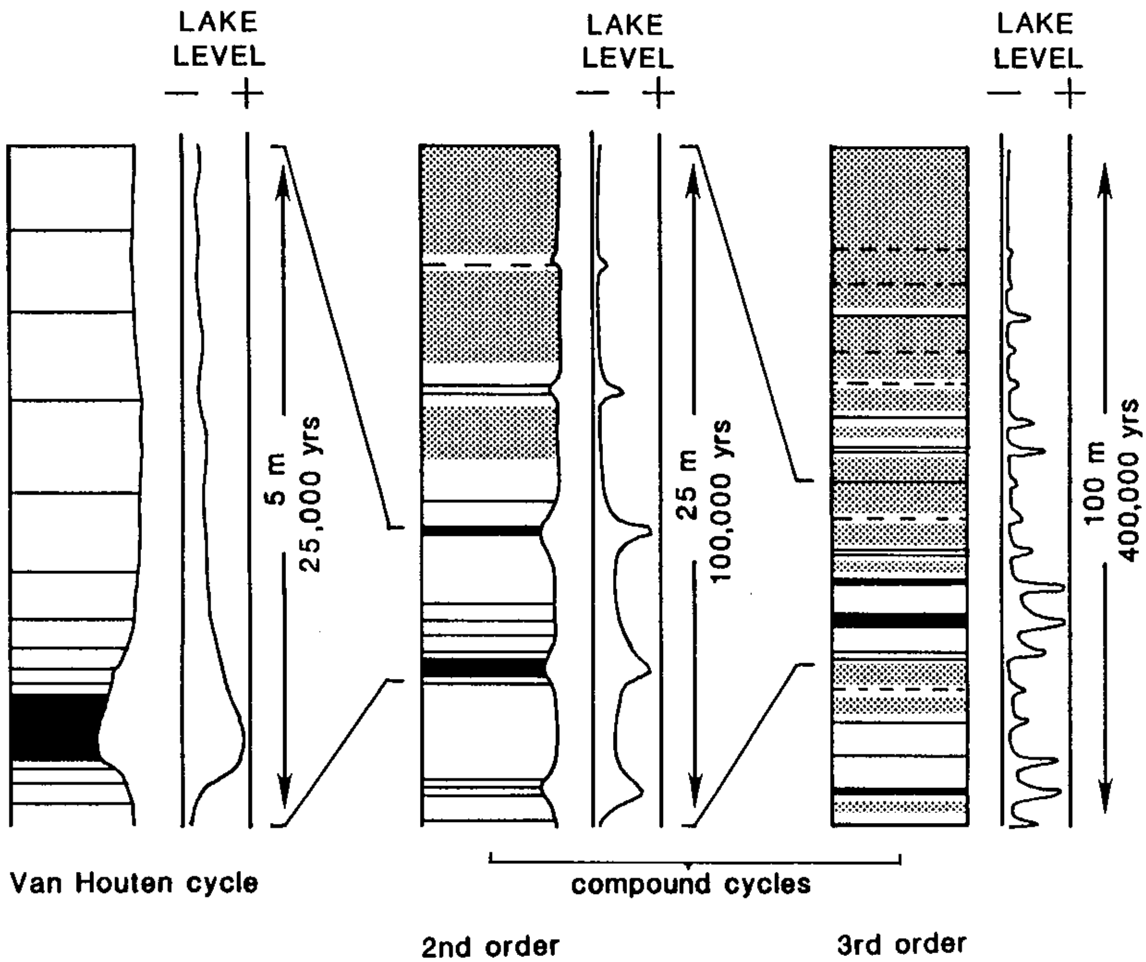


Figure 6. Compound cycles of the Newark Supergroup. Key as in Figure 5.

FIELD TRIP ROAD LOG

(Note: Mileage on left if cumulative mileage for road log; mileage on right is the increment between each mentioned mileage.)

MILEAGE FROM RIDER COLLEGE TO STARING POINT FOR FIELD TRIPS

0.0 0.0 Leave south gate of Rider College.

0.2 0.2 Leaving main entrance to Rider College.

0.7 0.5 Enter I-295 north following signs to I-95 south.

2.9 2.2 I-295 north ends, I-95 south begins.

8.1 5.2 Take exit number 1 for NJ-29 north. Mileage for field trip begins at this point.

MILEAGE FOR FIELD TRIP

0.0 0.0 Starting point - Junction I-95 - NJ-29: Scudder Falls Bridge.

West portal in Pennsylvania is in the uppermost Stockton Formation, with basal black platy mudstone of the Lockatong Formation exposed at N end of the northern access. East portal is about on the contact. Lockatong Formation in the southern fault block is about 500 m thick, about 600 m thinner than in the northern fault block (Hunterdon Plateau fault block). This thinning takes place by lateral replacement of massive gray mudstones by red mudstones in the upper Lockatong and by thinning of individual Van Houten cycles from 6.5 m in the Hunterdon Plateau fault block to 4.5 m in this area and the compound cycles they make up. This follows the typical pattern of updip thinning typical of half graben.

1.1 1.1 Analcime-rich reddish brown mudstone crosses road. Van Houten cycles well displayed in old quarry on west side of river.

1.3 0.2 Gray beds of the upper Lockatong Formation on right.

1.4 0.1 Crossing Jacob Creek. Uppermost Lockatong red mudstone beds outcrop in deep creek valley on right.

2.6 1.2 Washington Crossing State Park: thin green-gray siltstones making up bases of vague Van Houten cycles in lower Passaic Formation in low embankment on right.

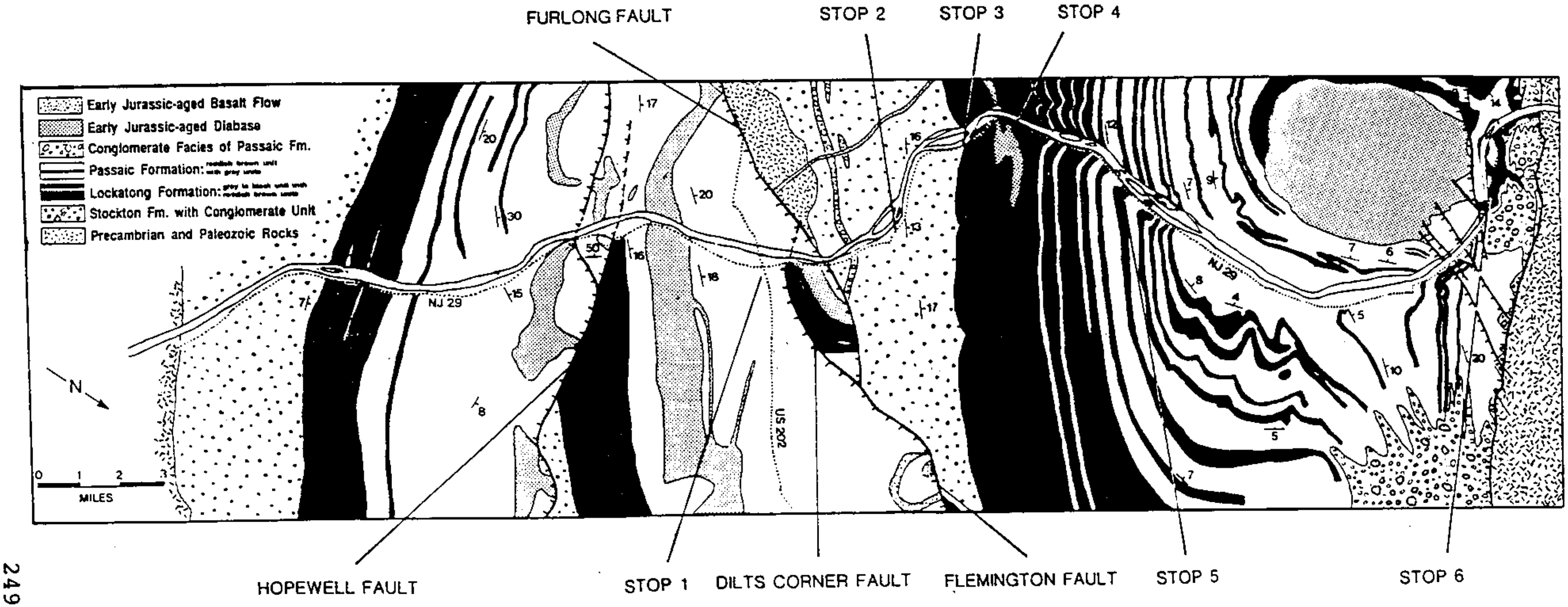


Figure 7: Geologic map of the Delaware River region of the Newark basin, showing the field trip stops (after Van Houten, 1969, and Ratcliffe *et al.*, 1986).

3.4 0.8 Van Houten cycles of lower Passaic Formation with a grey and black division 2 in back of lunch stand. These are unnamed and uncorrelated to sequences in other fault blocks.

5.6 2.2 Moore Station quarry in Baldpate Mountain diabase and metamorphosed Passaic Formation.

6.0 0.4 Moore Creek. Southern branch of Hopewell fault crosses river and continues west as the Buchmanville fault.

6.5 0.5 Mercer County Correctional Center Quarry in small Belle Mt. diabase body. Across the river this sheet forms Bowman Hill and has cross-cutting contacts (for additional details see Stop 1 of Husch and Hozik, this volume).

7.0 0.5 Crossing Hopewell Fault and entering Stockton Formation of middle (Sourland Mountain) fault block which is about 6.5-8 km wide along the river; section previously passed through is now repeated.

7.8 0.8 Lower Lockatong Formation poorly exposed in back of Flea Market.

8.9 1.2 Exposures of Lambertville Sill on right.

9.1 0.2 Crossing approximate top of Lambertville Sill (550 m thick)

9.4 0.3 NW-dipping hornfels of lower Passaic Formation on right.

9.7 1.0 Left onto Bridge Street in Lambertville and then right at first traffic light onto Main St. (NJ-29).

10.7 1.0 Right To US-202

10.9 0.2 Right onto US-202 N ramp. First of a series of four large outcrops of Passaic Formation stratigraphically close to the Perkasie Member of early Norian Age. These exposures show a number of Van Houten cycles which are extensively faulted at several scales, with considerable duplication of section. Division 2 of the most prominent Van Houten cycle at this outcrop is highly uraniferous as are several outcrops of what may be the same unit at mileage 11.3 and 12.1. M. J. Hozik has contributed the data presented in Figure 8 and this outcrop corresponds to Station 1 in that figure.

11.3 0.4 Section seemingly a duplication of that seen at previous outcrop. This is Station 2 of Figure 8.

12.1 0.8 Pull off on right next to large outcrops.

STOP 1
FAULTED PASSAIC FORMATION ADJACENT TO DILTS CORNER FAULT:
IMPLICATIONS OF STRUCTURES

Largest road cut of hornfels along US-202, again apparently duplicating the previous exposures, stratigraphically. This is Station 3 of Figure 8.

The following discussion of the structures visible at this outcrop was contributed by R.T. Faill:

Two statements can be made about the mesoscopic deformational structures seen in outcrops along the Delaware River: 1) the structures do not seem to reflect the megascopic basin tectonics; and 2) they appear only sporadically.

It is a geologic truism that large-scale megastructures that characterize a geologic terrane often appear in smaller form in the mesostructures that can be seen in the outcrops throughout the terrane. The Mesozoic basins of eastern North America apparently lack this characteristic. True, the rocks are not greatly metamorphosed or deformed, nor on the other hand are they undeformed and flat lying. This itself reveals a great deal about the basin history. Steeply dipping faults are locally common, but often they are in orientations quite different from the presumed orientation of the major faults; and those kinematic indicators, the slickenlines, occur in a multiplicity of orientations.

A good example of this apparent inconsistency occurs at the northwest margin at Monroe, on the west side of the Delaware River. (It is not one of the stops on this field trip.). The Mesozoic/Pre-Mesozoic contact is mapped as a fault (Drake and others, 1967). Recent drilling data (Ratcliffe and Burton, 1985; Ratcliffe and others, 1986) clearly demonstrate the presence of two Paleozoic thrusts at and just below the Mesozoic margin. These faults dip only moderately to the southeast, at 27 and 34 degrees respectively. Cataclastic structures associated with these faults suggest that they were apparently reactivated late in the basin history (Early Jurassic?), perhaps as listric faults. Cataclastic zones in the pre-Mesozoic rocks dip on the average 30 degrees to the southeast, and slickenlines trend northeast and northwest on southeast dipping faults and northwest dipping antithetic faults (Ratcliffe and Burton, 1985).

In the outcrop along PA-611, within 200 meters of

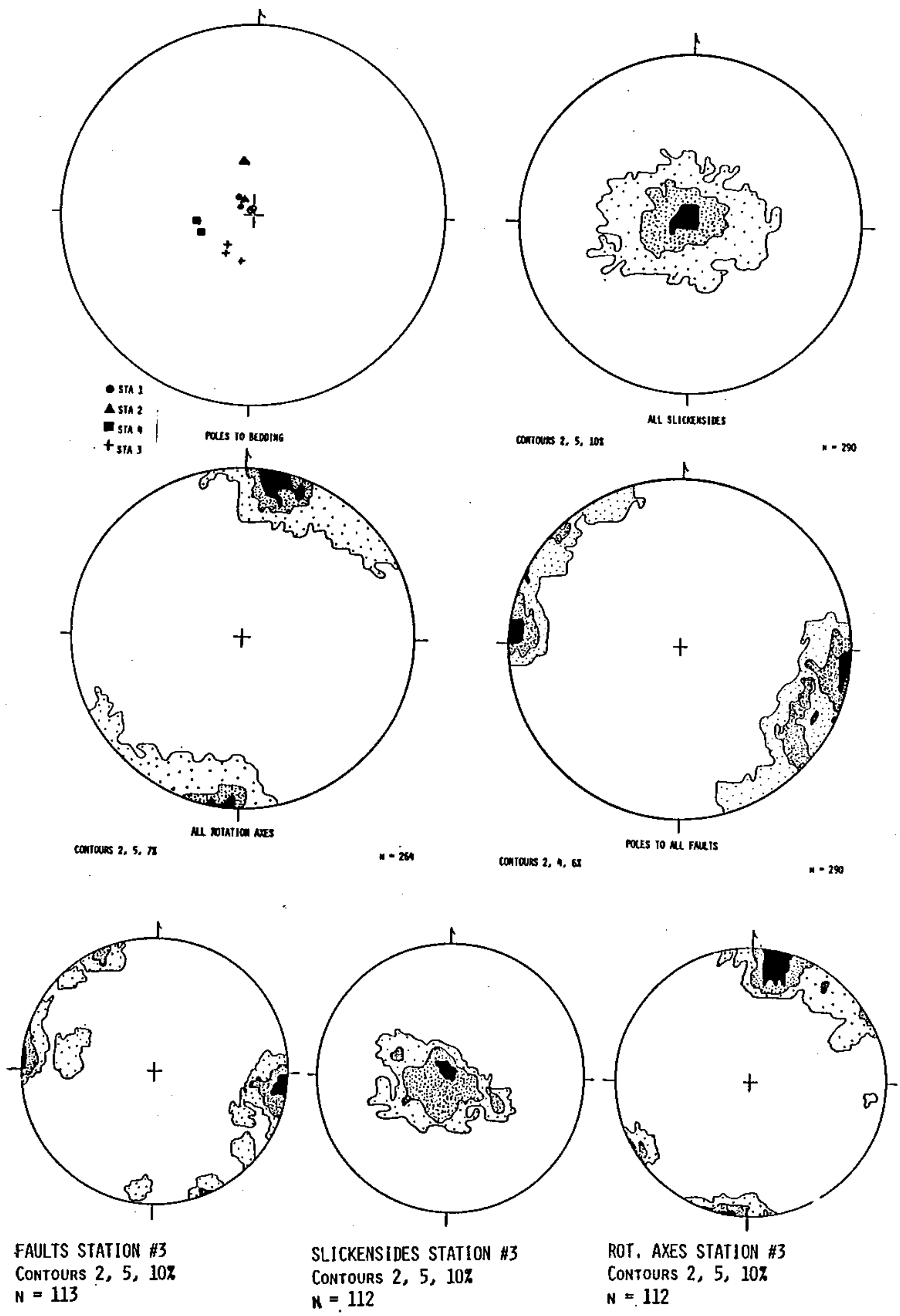


Figure 8. Equal area stereograms for bedding and fault data from exposures of lower Passaic along NJ-202. Upper four diagrams based on data from all exposures (stations 1-5). Lower 3 diagrams for station 3, Stop 1. Figure contributed by M.J. Hozik.

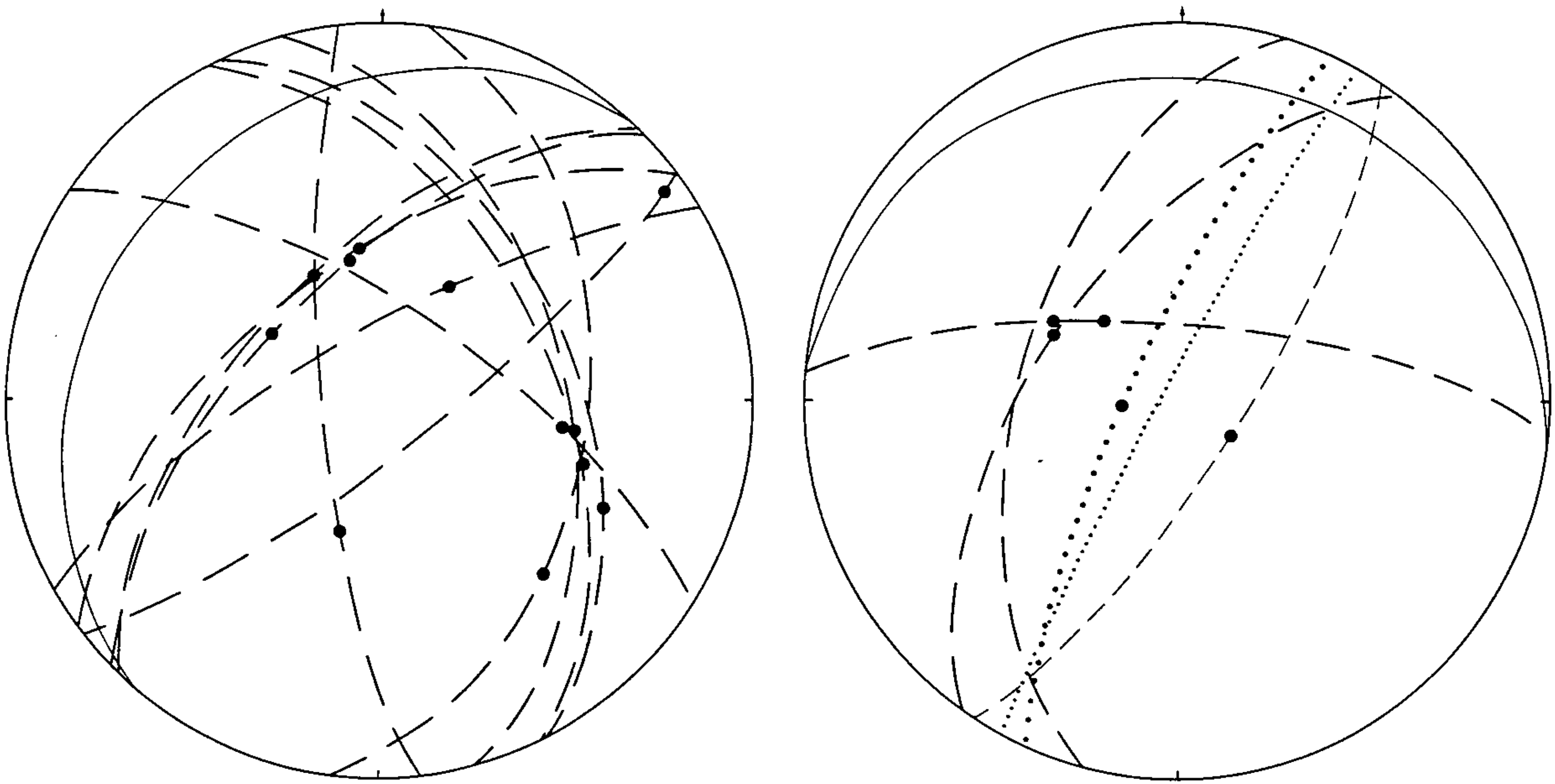


Figure 9. Data for Stop 1 contributed by R.T. Faill. Left, Equal-area stereogram (lower hemisphere), showing traces of bedding (solid line) and the slickensided fractures (dashed lines) at the basin margin, south of Riegelsville. Right, Equal-area stereogram (lower hemisphere), showing traces of bedding (solid line), faults, and slickensided fractures along U.S. Route 202, east of Delaware River. Light dot, fracture with plumose structure; heavy dot, sigmoidal fracture; light dash, east end fault; heavy dash; cleft faults; bullets, slickenline orientations.

the fault contact, interbedded carbonate-clast fanglomerates, siltstones, and argillaceous siltstones dip to the northwest at 22 degrees. (Relative to dip in the surrounding few kilometers, this dip is anomalously steep). One might expect that structures reflecting the basin margin tectonics would also be present here. Faults are present, but they do not dip moderately to the southeast as the margin faults do. Instead, they dip moderately to the east and northeast, and steeply to the northwest (Figure 9). In addition, the slickenlines plunge to the east and northwest. There is nothing in the outcrop that is consistent with the cataclastic extensional structures in the pre-Mesozoic rocks. In other words, the mesoscopic structures give little clue here to the overall basin tectonics.

This characteristic applies to the outcrops along US-202 at Stop 1 as well. Here, we are in the southern part of the Newark basin, separated from the northern part by the Chalfont-Furlong-Flemington fault complex and the Buckingham window. The Dilts Corner fault, a subordinate part of the complex, lies a few hundred meters north of this Stop. The structures seen in these road cuts may reflect the movements on the Dilts Corner fault rather than the basin as a whole, but this is not known.

In the third long road cut east from the river (on the south side of the road), bedding dips northward at 15 degrees. Subvertical fractures with a persistent north-northeast trend carry prominent plumose structures. And faults are common. Near the center of the cut, a pair of steep faults, one north dipping, the other west dipping, have formed a prominent cleft in the face of the road cut. Slickensides on both plunge steeply to the northwest, down their mutual intersection (Figure 9). To the east along the exposure is a set of steeply west-northwest dipping fractures that pass through a subhorizontal, meter-wide zone halfway up the face. Within this zone, the dip of the fractures decreases markedly, giving the fractures a sigmoidal form. Many of the fractures are covered with a dark green chlorite sheen, and slickensides indicate a dip slip movement. Even farther to the east, bedding is offset by tens of cm (down on west) by several steeply west dipping faults, again with subvertical slickensides.

Just what relationship these numerous structures have with the nearby Dilts Corner fault is unknown. And what role they and the steeply northwest plunging slickensides have with the overall basin tectonics is similarly not known.

Hozik (1985) offers a different interpretation of the same outcrops. According to Hozik, these outcrops provide ample evidence of normal faulting specifically associated with movement on the Flemington-Dilts Corner fault system. Fewer than 10 out of 150 of the small faults measured by Hozik (Figure 8) show a significant strike-slip component of movement (based on slickenlines. He notes that most of the faults strike northeasterly and dip steeply either to the northwest or to the southeast. Faults that strike N40°E and dip 70° to the northwest or southeast appear to define a conjugate set which would require σ_1 to be nearly vertical, σ_2 to be nearly horizontal and trending N40°E, and σ_3 to be horizontal and trending S50°E.

As pointed out by Hozik (1985) both Sanders (1962) and Manspeizer (1980) have argued that the intrabasin faulting was essentially strike slip in origin and specifically that Flemington fault should be a right lateral fault (Figure 8). Hozik's data suggest dominantly normal motion on the Flemington-Dilts corner system which is much more in line with the predictions of the Ratcliffe-Burton (1985) and Schlische and Olsen (this volume) models regional northeast-southwest extension (Figure 8). Of course, the Flemington fault is not exposed in the vicinity of these outcrops and the structures visible might, as Faill (above) argues, have nothing to do with the regional picture. However, the type of intense brittle fracturing seen at these outcrops is limited to the vicinity of the major faults and thus should be related to movement during some part of the history of the basin.

12.8 0.7 Section on both sides of roads expose faulted red beds of the Passaic Formation from a stratigraphic position somewhat above the previous outcrops. This outcrop corresponds to Station 4 of Figure 8.

13.0 0.2 Exit right for Mt. Airy - Dilts Corner

13.4 0.4 Left onto Queen Road and pass under US-202.

13.6 0.2 Left onto ramp for US-202 south and enter highway.

15.6 2.0 Exit right for NJ-29.

15.8 0.2 Left on Alexauken Creek Rd.

16.1 0.3 Right onto NJ-29 N.

16.8 0.7 Crossing Dilts Corner fault.

17.0 0.2 Quarry in Mount Gilboa Diabase, gabbro, and metamorphosed Lockatong Formation.

17.6 0.6 Lockatong nepheline-cancrinite hornfels above diabase intrusion. Nepheline and analcime syenites crop out along small creek to east (Van Houten, 1969). Barker and Long (1968, 1969) proposed that these syenites formed by reaction of Lockatong Formation mudstones with a granophyric differentiate of diabase (see also Husch and Hozik, this volume).

17.9 0.3 Crossing Flemington Fault. Entering northern or Hunterdon Plateau Fault Block. Across river in Pennsylvania is uplifted basement consisting of Precambrian metamorphic rocks, Cambro-Ordovician limestones and phyllite, all overlain by basal Stockton Formation.

18.0 0.1 Entering Stockton, New Jersey - type area of the Stockton Formation (Kümmel, 1897; McLaughlin, 1945). Exposures of conglomeratic Solebury Member of Stockton Formation (McLaughlin, 1960) on right.

18.4 0.4 School yard with exposures of Solebury Member

19.3 0.9 Crossing Wickecheoke Creek at junction with N.J. Rt. 515 in Prallsville, New Jersey, with extensive exposures of Stockton and Lockatong formations upstream. Lower Lockatong well exposed and typically fossiliferous. Old quarries along road and canal in vicinity are in Prallsville Member of Stockton Formation.

19.6 0.3 Turn right onto open dirt area and park.

STOP 2
QUARRY IN PRALLSVILLE MEMBER OF STOCKTON FORMATION, MIDDLE CARNIAN AGE.

Recently active quarry in upper part of the Prallsville Member of the Stockton Formation exposes about 60 m of section and begins about 760 m above the base of the formation (Van Houten, 1969, 1980).

Four main sediment types are obvious at these outcrops: 1) massive, well sorted medium gray to buff arkose with faint to prominent large scale cross bedding; 2) massive to crudely crossbedded (2 m) arkosic conglomerate units some of which are kaolinized and have small to large rip up clasts; 3) well bedded red coarse siltstone with small dune scale cross bedding, ripple cross lamination, and parallel lamination; 4) blocky, massive red mudstone intensely bioturbated by *Scoyenia* and roots. Scoyenia is the most common Newark burrow-type and it may have been made by crayfish-like crustaceans in environments with perennially high fresh water tables. The sediment types seen here make up thick

(+ 10 m), poorly defined upward fining cycles possibly deposited by large, perennial meandering rivers (J. Smoot, pers. comm.).

The large river systems implied by these outcrops are incompatible with a closed basin model for the Newark Basin and indicate an open basin for this part of the basin's history. This is in accord with the predictions of the basin filling model outlined by Olsen (this volume) and Olsen and Schlische (1988).

Other items:

1. Na-feldspar to K-feldspar ratio is commonly 2:1.
2. This interval is part of a large magnetically reversed interval (Witte and Kent, 1987).
3. Micro-placers of specular hematite (after magnetite) grains locally outline thin bedding. Probable implications for detrital remnant magnetism.
4. Small to large root mottles and goethite stained weathered vertical trains of nodules, probably following roots.

Stockton Vertebrates

The Stockton Formation remains almost entirely unprospected for fossils. Nonetheless, important reptile remains have been found. Sinclair (1917) described the lower jaw of a very large labyrinthodont amphibian from the basal Stockton near Holicong, Pennsylvania. This jaw, still not described in detail, may belong to a large capitosaur, rather than a metoposaur as usually thought (W. Seldon, pers. comm.). The age of the basal Stockton is unknown and it could be as old as Middle Triassic. Early Middle Triassic capitosaurs are known from fragments from the oldest parts of the Fundy basin sequence (Olsen and Sues, 1986).

All the rest of the Stockton vertebrates come from the upper beds of the formation of Late Carnian Age. These include a large headless phytosaur skeleton referred to *Rutiodon manhattanensis* (Von Huene, 1913) from Fort Lee, New Jersey, an impression of the fragment of a metoposaur interclavicle (Baird, 1986) from Princeton, New Jersey, phytosaur teeth from Assinetcong Creek in Flemington, New Jersey (Olsen and Flynn, in press), and a few small assemblages of footprints and very scrappy bones from Rockland County (Olsen and Flynn, in press). The following ichnotaxa occur: *Rhynchosauroides* cf. *hyperbates*, *Apatopus lineatus*, *Brachychirotherium eyermani*, and ?*Atreipus* sp. This assemblage does not differ appreciably from that of the Lockatong Formation and it is in fact possible that the fossiliferous portions of the Stockton of Rockland County may be a shoreward facies of the Lockatong.

The uppermost Stockton (Raven Rock Member) at Carversville, Pennsylvania has produced a rich assemblage of plants dominated by cycadeoides, especially *Żamites powelli* (Bock, 1969, Ash, 1980). Another upper Stockton assemblage, this time dominated by the conifer *Pagiophyllum simpsoni* has more recently been described by Axsmith (in press) from Phoenixville, Pennsylvania. Silicified wood is common in the middle and lower Stockton Formation and plant foliage and stem compressions have been found including *Zamites* cf. *powelli*, and *Neocalamites* sp. Again, there has been no systematic attempt to prospect for plants, although what has been found so far is obviously promising.

Return to vehicle and turn right onto NJ-29 north.

20.8 1.2 Crossing Lockatong Creek - extensive exposures of middle and upper Stockton Formation and type section of Lockatong Formation are exposed upstream (McLaughlin, 1945).

22.0 0.2 Old quarry in woods on right in Raven Rock Member of Stockton Formation. Lower parts of quarry expose gray uraniferous sandstones and siltstones which were the site of a uranium prospect during the 1950's (Turner-Peterson, 1980). This radon producing unit has been traced from central Bucks County, Pennsylvania to Wickecheoke Creek to the east (Turner-Peterson and others, 1985).

22.3 0.3 Cliffs of red mudstones and sandstones of upper Stockton Formation on right.

22.7 0.4 Crossing approximate Stockton-Lockatong transition. Turn right onto small rest area.

STOP 3
STOCKTON-LOCKATONG TRANSITION

Several good exposures of uppermost Stockton Formation transitional into Lockatong Formation occur on slopes above small rest area. Measured section of portion of the sequence is shown on Figure 10 Sequence resembles single Van Houten cycle. (Figure 10 and comments below contributed by K. W. Muessig and H.F. Houghton).

Notable at this outcrop are:

1. Well-bedded purplish-brown sandy mudstone.
2. Thin, interbedded units of indistinctly laminated, gray mudstones and sandstones with calcareous cements.

3. Uranium concentrated in grey mudstones and locally, possibly secondarily, in calcareous sandstones.
4. Elevated radon levels in homes built on this unit throughout the southern portion of the Newark basin.

23.1 0.4 Stumphs Tavern Road.

Stream along north side of road exposes gray and red beds of the ?Weehawken Member of the lower Lockatong Formation. The Weehawken and succeeding Gwynedd Member of the lower Lockatong are the most fossiliferous portions of the Lockatong. Precisely equivalent cycles are exposed in creeks which cut down cliffs on opposite side of river at Lumberville, Pa. Here a microlaminated and strongly metamorphosed limestone (bý overlying Byram Sill) of division 2 of a Van Houten cycle produces complete Turseodus and abundant conchostracans. Presumably equivalent beds produce copious fossils in the Princeton, Weehawken, and North Bergen areas of New Jersey to the north east. Unfortunately exposures of this part of the Lockatong are very poor in this part of the Newark Basin.

Note on Lockatong Formation Nomenclature.

McLaughlin (1943, 1945) mapped the distribution of the red portions of the 400,000 year cycles in the Hunterdon Plateau fault block and gave most of them informal member names. These names consist of a mixture of descriptive terms (First Thin Red Member), letter designations (member B), and place names (Smith Corner Red). The gray portions of the cycles in the Lockatong were not named with the exception of members A_1, A_2, and B. Because of the intense interest in many of the gray portions of these same cycles, it makes sense to apply some kind of informal, consistently applied name to each. Rather than add a new set of names for the gray portions of the cycles and retain McLaughlin's informal names for the red parts of the cycles, it would be more useful to revise the nomenclature of the Lockatong so that each member consisted of a lower gray portion and an upper red portion making up a single 400,000 year cycle, and receiving the same member name (Figure 11). Where McLaughlin used place names for a unit, the names are conserved. However, new names are needed both where he used letter or descriptive designations and for the units he never named (in accordance with the American Code of Stratigraphic Nomenclature 1961, 1983). Most of these names have already been used informally in previous works (Olsen, 1984, 1986). In the central Newark Basin, the Lockatong Formation is 1100 meters thick and is thus divided into 11, 400,000 year cycles, each about 100 m thick (Figure 11).

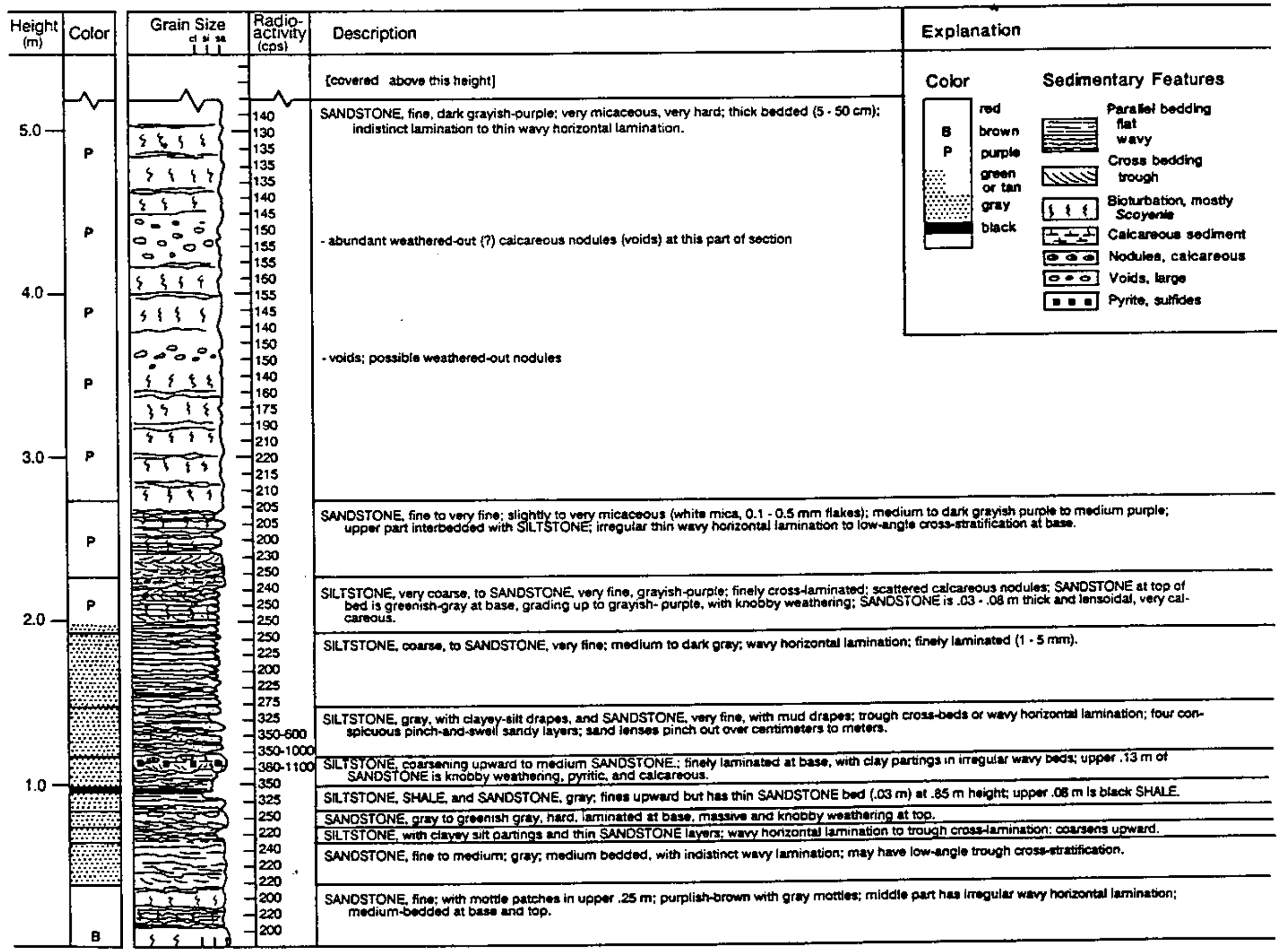


Figure 10. Measured section of upper Stockton (or basal Lockatong) north of Raven Rock, N.J., Stop 3 (Lumberville PA-NJ 7 1/2' quadrangle). Radioactivity measured by hand held scintillometer and given in units of counts per second. Figure contributed by K. W. Muessig. and H.F. Houghton.

23.5 0.4 Poor exposures on right of upper part of lower Lockatong Formation (upper ?Gwynedd Member).

23.8 0.3 Pull over on right next to culvert.

STOP 4
BYRAM SILL AND MIDDLE LOCKATONG FORMATION

This outcrop is one of the largest in the Newark Supergroup covering a stratigraphic thickness of over 500 m of section along a distance of 1.2 mi. Begin at the Byram sill and walk to the north.

Byram Sill on right (Figure 12) intruding zone between lower and middle Lockatong Formation.

Items:

1. Vague cyclic pattern of major ledges of albite-biotite hornfels; Nepheline occurs about 30 m below sill.
2. Lower contact of chilled border of diabase with sheared, weathered Lockatong pelitic hornfels.
3. Coarse-grained diabase, extensively fractured. Joint minerals have no apparent relation of compositional layering in the diabase.
4. Complex upper contact (Figure 12) on north, is a fault, probably post-intrusion as is common for cross-cutting intrusions. On the south is a welded curved contact. Nepheline, cancrinite, pyroxene and amphibole present in both blocks of Lockatong albite-biotite hornfels.
5. Relations of 21 m sequence from faulted diabase contact to outcrops north of creek obscure because of intervening faults.
6. The Byram Sill does not seem to extend down dip in this area as revealed by proprietary seismic reflection studies in Pennsylvania and a magnetic traverse along the railroad grade parallel to NJ-29 by R. Colombo (M.J. Hozik, pers. comm.

Walk to west.

23.9 0.1 Begin 410 m almost continuous section of middle Lockatong.

A "0" painted on outcrop marks beginning of undisturbed continuous sequence of middle Lockatong Formation (Figure 12). Measured section shown in Figure 13 is in meters but marks on outcrop are in feet with the correspondence shown in the figure. About 95 % (401

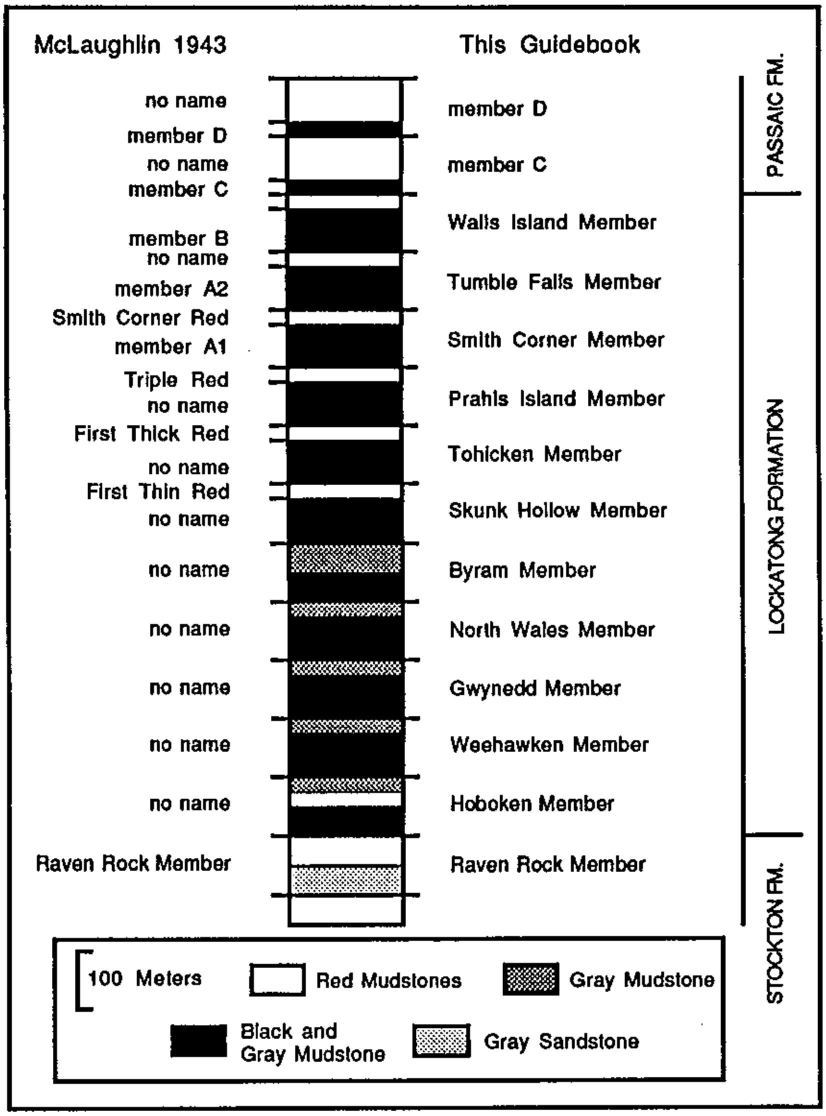


Figure 11. Subdivisions of the upper Stockton, Lockatong, and basal Passaic formations.

m) of this section consists of fine calcareous mudstone and claystone with the rest 5 % (5 m) consisting of very fine sandstone. In the terminology outlined above, these exposures include North Wales, Byram, Skunk Hollow, and Tohicken members. McLaughlin's (1944) "First Thin Red" makes up the upper part of the Skunk Hollow Member and the "First Thick Red" makes up the upper part of the Tohichen member.

This section shows the cyclic pattern typical of the Lockatong and Passaic formations perhaps better than any other section in the Newark Supergroup. Fourier analysis of sediment fabrics in this section (Figures 14, 15) and calibrated by varve sedimentation rates and radiometric time scales show the main thickness and time periodicities typical of both the Lockatong and Passaic formations in this part of the basin. At these outcrops there are prominent periods in thickness at 5.5 m, 6.1 m, 24.0 m, 32.0 m, and 96.0 m. The 5.5 and 6.1 m cycles are Van Houten' chemical and detrital cycles, respectively [*i.e.*, Olsen's (1986) Van Houten cycles], the 24.0 and 32.0 m cycles are the "25 m" long cycle, and the 96.0 m cycle corresponds to the "100 m" long cycle of Van Houten. In terms of time, these five thicknesses of cycles were evidently controlled by the precession cycle of 19,000 and 23,000 years, and the eccentricity cycles of roughly 95,000, 123,000 and 413,000 years.

The interval from the 115 to the 300 foot mark (35-91 m) is exposed in old quarry in upper part of North Wales and lowermost part of Byram members about 0.3 mi north of the 0 mark in this traverse (Figures 12, 13). Thomsonite is a common joint mineral along with analcime, calcite, and rarer ilmenite. Scapolite in the hornfels and laumontite on joints reported by Lewis (1909) have not been confirmed. This dark gray hornfels produces crushed stone megascopically almost indistinguishable from fine grained diabase. A quarry in this interval, across the river at Point Pleasant, has produced slab of large but sloppy *Apatopus* (D. Baird, pers. comm.).

The pattern of short and longer cycles can be best seen in the interval from the 550 ft mark to the northern end of the main outcrop (Figure 13).In order to facilitate comparisons with the succeeding stops and in order to illustrate the wide range of variation exhibited by the Van Houten cycles at this outcrop, we will compare two end-members of the range of sequences of sedimentary fabrics present in this section (Figure 16). These two end-members correspond, in part, to Van Houten's (1962, 1964, 1969) detrital and chemical short cycles.

A single Van Houten cycle is shown in Figure 16a (cycle SH2 - Figure 13) between the 620 and 815 marks

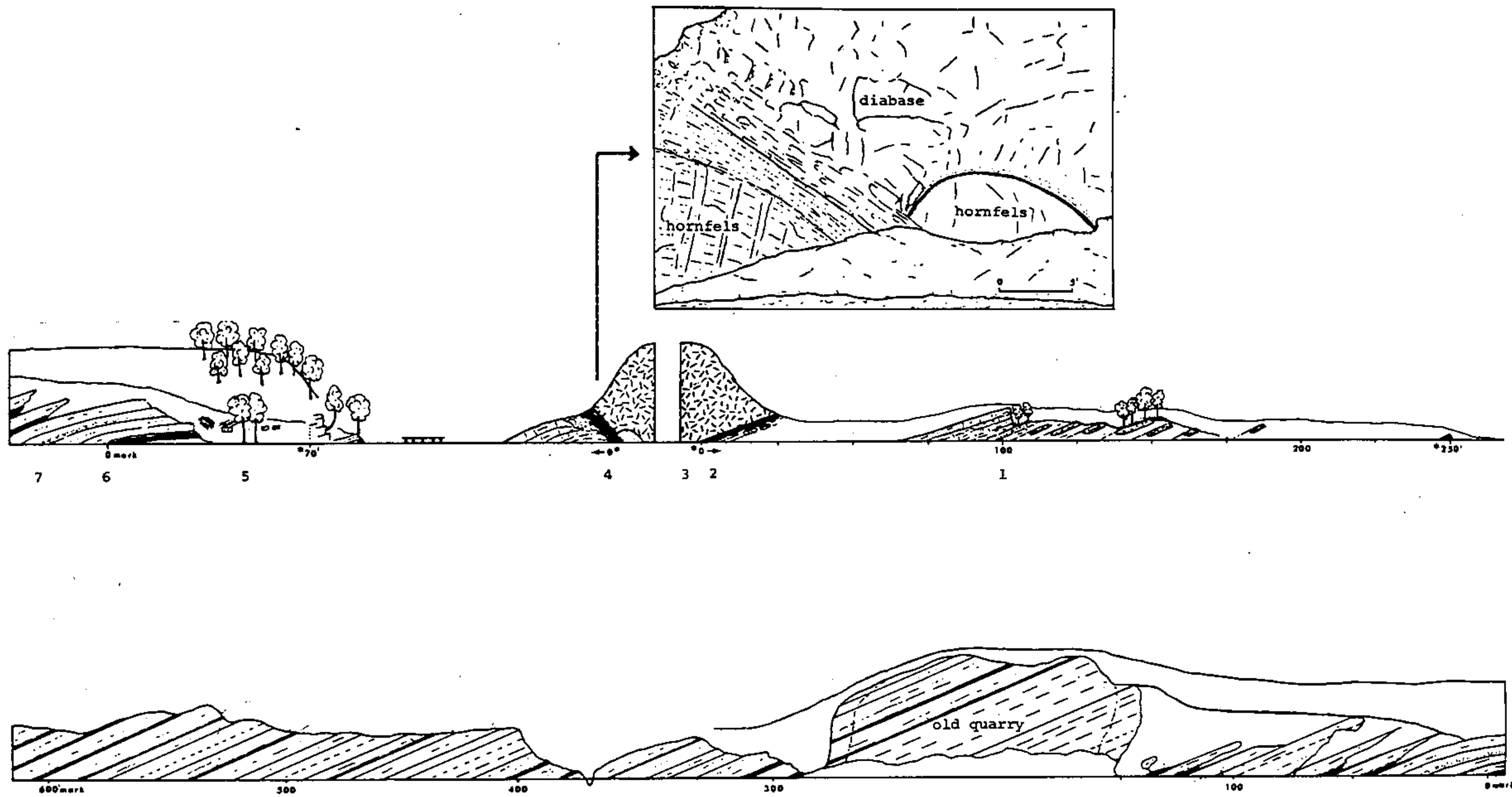


Figure 12. Sketch of road cut through middle part of the Lockatong Formation and Byram diabase sheet, NJ-29, Byram, N.J., Stop 4 (adapted from Van Houten, 1969).

Upper profile: Lockatong hornfels 76 m stratigraphically below and 21 m stratigraphically above diabase. Enlargement is of upper contact; 1-7, items mentioned in text; 0 mark, probablely about 21 m above diabase; base of continuous section measured in feet above base (0 mark).

Lower profile: Lockatong Formation 21 to 218 m above 0 mark (70 to 715 ft). Well displayed analcime rich Van Houten cycles (chemical cycles of Van Houten, 1969) between 137 and 168 m (450 - 550 ft marks).

(about 750 ft, 227 m) in the lower part of the Skunk Hollow Member. This is the most fossiliferous cycle exposed in this outcrop. Division 1 is a thin sand overlain by thin bedded mudstone with an upward increase in total organic carbon (T.O.C.). This represents the beginning of lake transgression over the mostly dry playa flat of the preceding cycle and this is followed by deposition under increasing water depth.

The transition into division 2 is abrupt and is marked by the development of microlaminated calcareous claystone consisting of carbonate rich and carbonate-poor couplets, an average of 0.2-3 mm thick. The basal few couplets of this interval contains articulated skeletons of the aquatic reptile *Tanytrachelos* (Figure 16a) and infrequent clam shrimp and articulated fossil fish (*Turseodus* and ?*Synorichthys*) and abundant conchostracans occur in the middle portions of the unit. In its upper part, the microlaminae are very discontinuous, perhaps due to microbioturbation. No longer microlaminated, the succeeding portions of division 2 become less organic-rich and less calcareous upward, and there is an upward increase in pinch and swell lamination. Deep, widely spaced, and sinuous desiccation cracks propagate down from the top of the unit. The average T.O.C. of division 2 (83 cm thick) of this cycle is 2.5 % which is a rather common value for black siltstones and limestones of the Lockatong.

The absence of syndepositional desiccation cracks, and the presence of extremely fine microlamination and articulated aquatic vertebrates, suggest that division 2 is the high stand deposit of lake which formed this Van Houten Cycle. Studies of similar microlaminated units in other parts of the Lockatong suggest deposition in water in excess of 100 m deep, below wave base, and in perennially anoxic water (Manspeizer and Olsen, 1981; Olsen, 1984).

As noted by Turner-Peterson and others (1985) microlaminated and adjacent units such as this one are often uraniferous and radioactive. According to K. W. Muessig and H.F. Houghton (pers. comm.) homes in the southern portion of the Newark Basin built on similar units are notably elevated in radon.

The presence of the articulated reptiles at the base of the microlaminated unit may represent the transgression of the chemocline of the chemically stratified lake as the lake expanded. It is common for articulated reptiles to occur right at the base of the microlaminated portions of division 2 in many other cycles of the Lockatong (Olsen, 1980c) and in other Triassic formations as well (Olsen *et al.*, 1978). This pattern is of considerable use in prospecting for these little reptiles. The decrease in microlamination toward the top of division 2 suggests the action of

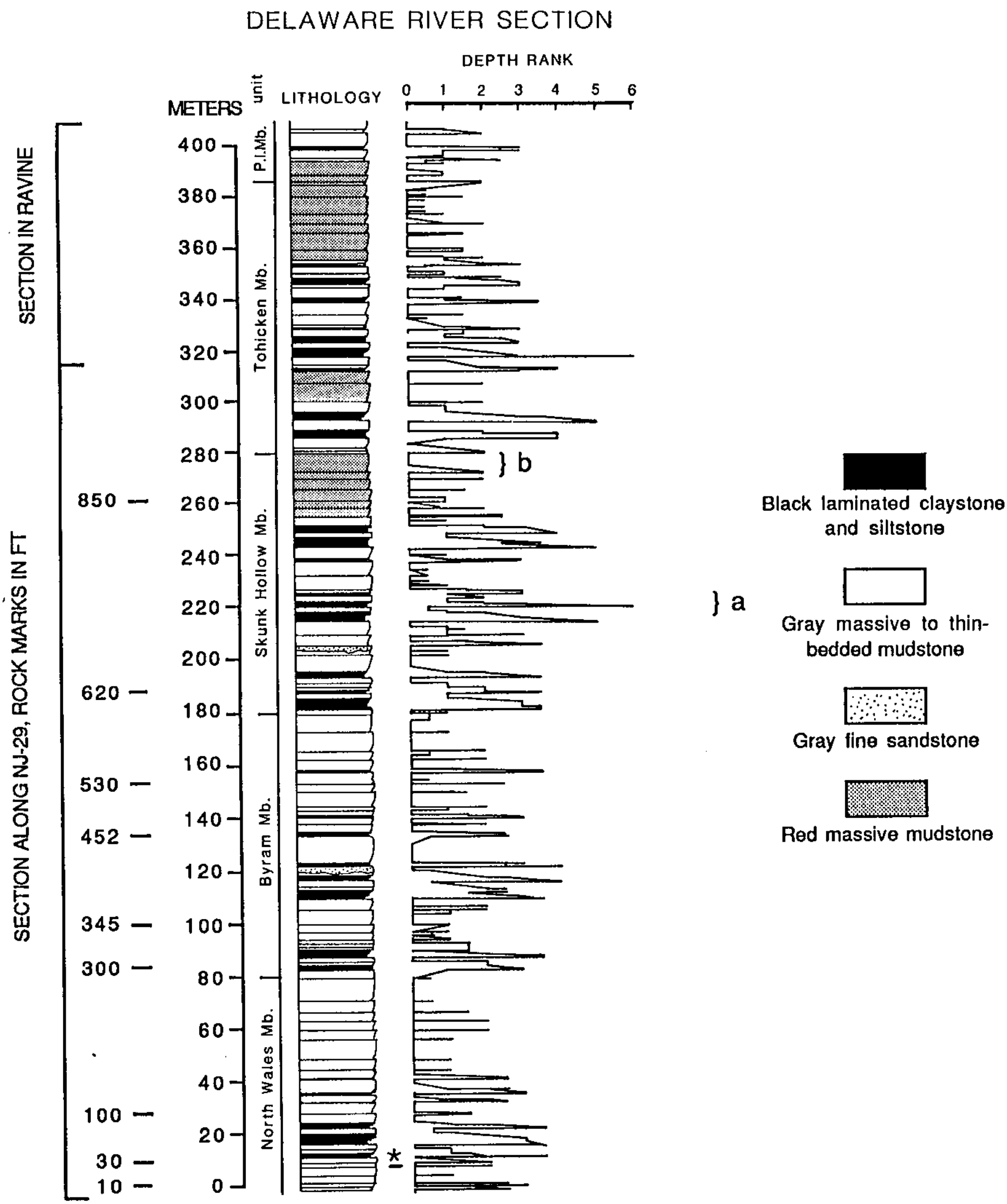


Figure 13. Measured section through the middle Lockatong Formation along NJ-29 at Byram (Stop 4) showing depth rank and positions of detailed measured sections shown in Figure 16. (**a**) Corresponds to Van Houten cycle in Figure 16a, and (**b**) corresponds to cycle in Figure 16b.

Table 1.
Key to units in Figure 16.

Figure 16A

N	+30 cm black laminated claystone (2.9% TOC).
M	10 cm gray mudstone with widely spaced desiccation cracks.
L	96 cm massive seemingly structureless monosulfide-bearing gray mudstone (0.6% TOC).
K	30 cm gray fine oscillatory rippled sandstone.
J	240 cm gray decimeter-scale siltstone and fine sandstone beds with abundant contorted desiccation cracks, soft sediment deformation, reptile footprints (*Apatopus*), and rare clam shrimp (*Cyzicus*)(,0.5% TOC).
I	30 cm dark dray laminated mudstone with rare fish bones (0.1% TOC).
H	30 cm dark gray centimeter-scale mudstone beds with widely spaced, deep, contorted desiccation cracks (1.3 % TOC).
G	83 cm black, laminated organic-rich (5.0% TOC) black limestone (77% carbonate) grading up into less calcareous (34% carbonate) and less organic rich (1.6% TOC) siltstone with deep, widely spaced, and sinuous desiccation cracks originating in the overlying beds. Abundant pinch and swell laminae present. Clam shrimp (*Cyzicus*) and ostracodes (*Darwinula*) abundant at the base but absent at top.
F	3 cm black, laminated, organic-rich (6.6% TOC) blebby limestone (76% carbonate) with occasional microlaminae. Abundant ostracodes (*Darwinula*) and clam shrimp (*Cyzicus*) present.
E	8 cm black, microlaminated limestone with even microlamination and some pronounced stylolites passing up into limestones with very discontinuous blebby microlaminae. Clam shrimp (*Cyzicus*), ostracodes (*Darwinula*), fish (*Turseodus* and *Synorichthyes*) , and coprolites present.
D	3 cm black, microlaminated blebby calcareous siltstone with extremely abundant clam shrimp (*Cyzicus*), articulated fish (*Turseodus)*, and coprolites.
C	4 cm black, microlaminated calcareous claystone (1.4% TOC, 33% carbonate). Basal few couplets with articulated reptiles (*Tanytrachelos*) and rare clam shrimp (*Cyzicus*).
B	12 cm black well bedded mudstone showing an upward increase in thin-bedding and total organic content. Rare fish bones (*?Diplurus*) and clam shrimp (*Palaeolimnadia* and *Cyzicus*).
A	39 cm gray, climbing ripple-bedded fine sandstone (0.3% TOC), Desiccation cracks are absent in contrast to underlying division 3 of the preceding cycle.

Figure 16B

F	+20 cm gray massive mudstone with crumb fabric.
E	491 cm consisting of 7 massive mudstone sequences, each 150 to 20 cm thick, showing upwards transition from thin-bedded claystone to breccia fabric to crumb fabric.
D	31 cm red and green laminated to thin-bedded claystone with widely spaced desiccation cracks projecting down from overlying unit,
C	8 cm gray claystone becoming laminated upwards with widely spaced desiccation cracks at top.
B	82 cm gray massive mudstone with well developed crumb fabric with analcime filled vesicles.

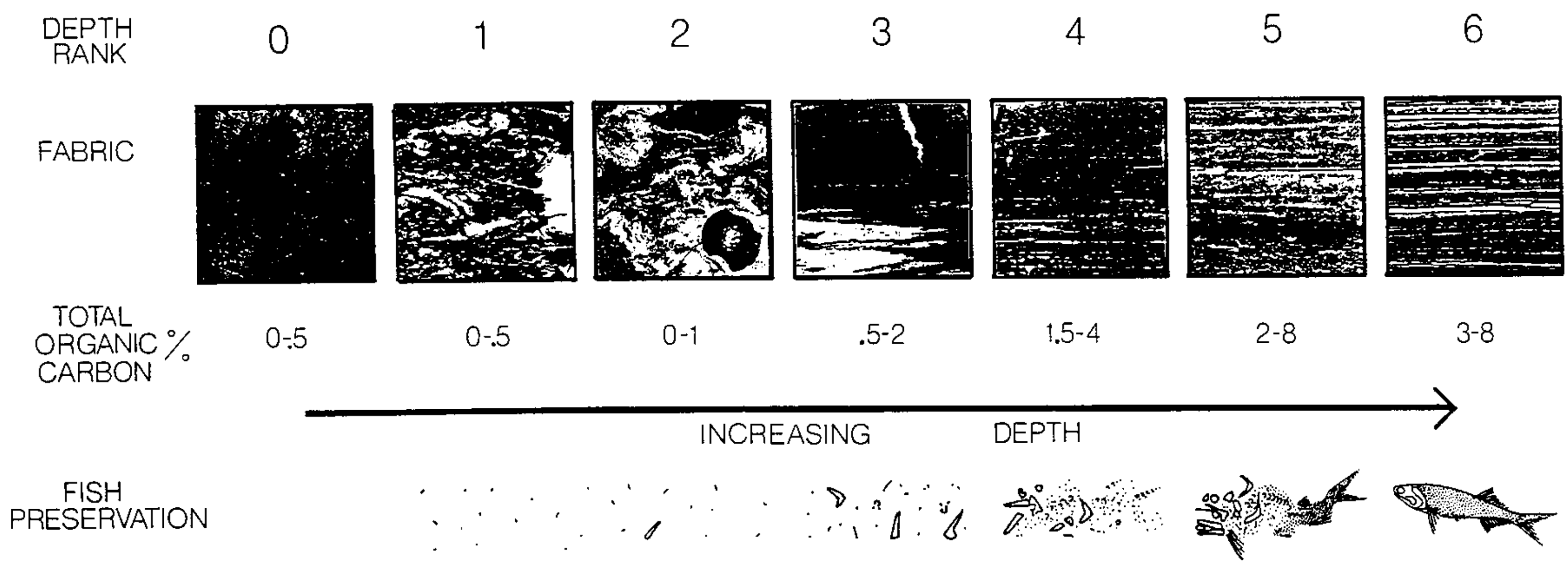


Figure 14, Sedimentary fabrics characteristic of "depth ranks" showing a range from massive, intensely desiccated breccia and crumb fabrics (depth rank 0) to completely undisturbed, microlaminated calcareous siltstone and claystone (depth rank 0). Complete descriptions of these depth ranks are in Olsen (1986).

microbioturbation and increased water column ventilation which we attribute to slowly decreasing water depth (Olsen, 1982, 1985a).

Division 3 is characterized an overall upward increase in the frequency and density of desiccation cracked beds and an decrease in thin bedding. Reptile footprints (?*Apatopus*) occur near the middle of the sequence (Figure 16). Organic carbon content drops through division 3 is low (1.3 % - 0.5 %). This interval shows the greatest evidence of exposure in the cycle and represents the low stand of the lake. However, as evidenced by the relatively low frequency of desiccation cracked intervals and the generally wide spacing of the cracks themselves, submergence was much more frequent than emergence during the deposition.

The cycle below SH2 (SH1, Figure 13) also contains a division 2 with a microlaminated base, fish and clam shrimp. The total thickness of division 2 of this cycle is much thicker than the succeeding cycle.

Division 2 in other cycles at this outcrop are not as organic rich or as well laminated as SH2 and SH1, and the other cycles also show greater degrees of desiccation of division 3. The only fossils present tend to be rare clam shrimp and indeterminate fish bones and plant scraps.

Figure 16b shows a Van Houten cycle in McLaughlin's "First Thin Red" about 40 m above the cycle shown in Figure 16a (top of cycle about 1.5 m below the 900 ft mark, 274 m). This cycle contrasts dramatically with the Upper Skunk Hollow Fish Bed in consisting of 75 % red massive mudstone and completely lacking a black organic-rich division 2. Division 3 of the previous cycle (also mostly red) consists of red massive almost homogeneous appearing mudstone with a faint but highly vesicular crumb fabric which according to Smoot (1985) formed by breaking up, by desiccation, of aggrading mud in a playa lake with the vesicles representing cement-filled voids (now dolomite and analcime) formed during aggradation and cemented shortly after deposition. The vesicles do not seem to be related to an evaporitic texture as suggested by Van Houten (1969).

The overlying massive mudstone of division 1 is gray and contains a better developed crumb fabric. This unit was probably deposited under much the same conditions as the unit underlying it; its iron was reduced, however, by the interstitial waters from the lake which deposited the next unit. The transition from an aggrading playa to a short lived perennial lake is marked by an upward increase in lamination, a disappearance of the crumb fabric and a decrease in the density of desiccation cracks.

Division 2 of this cycle consists of red and green laminated to thin-bedded claystone lacking

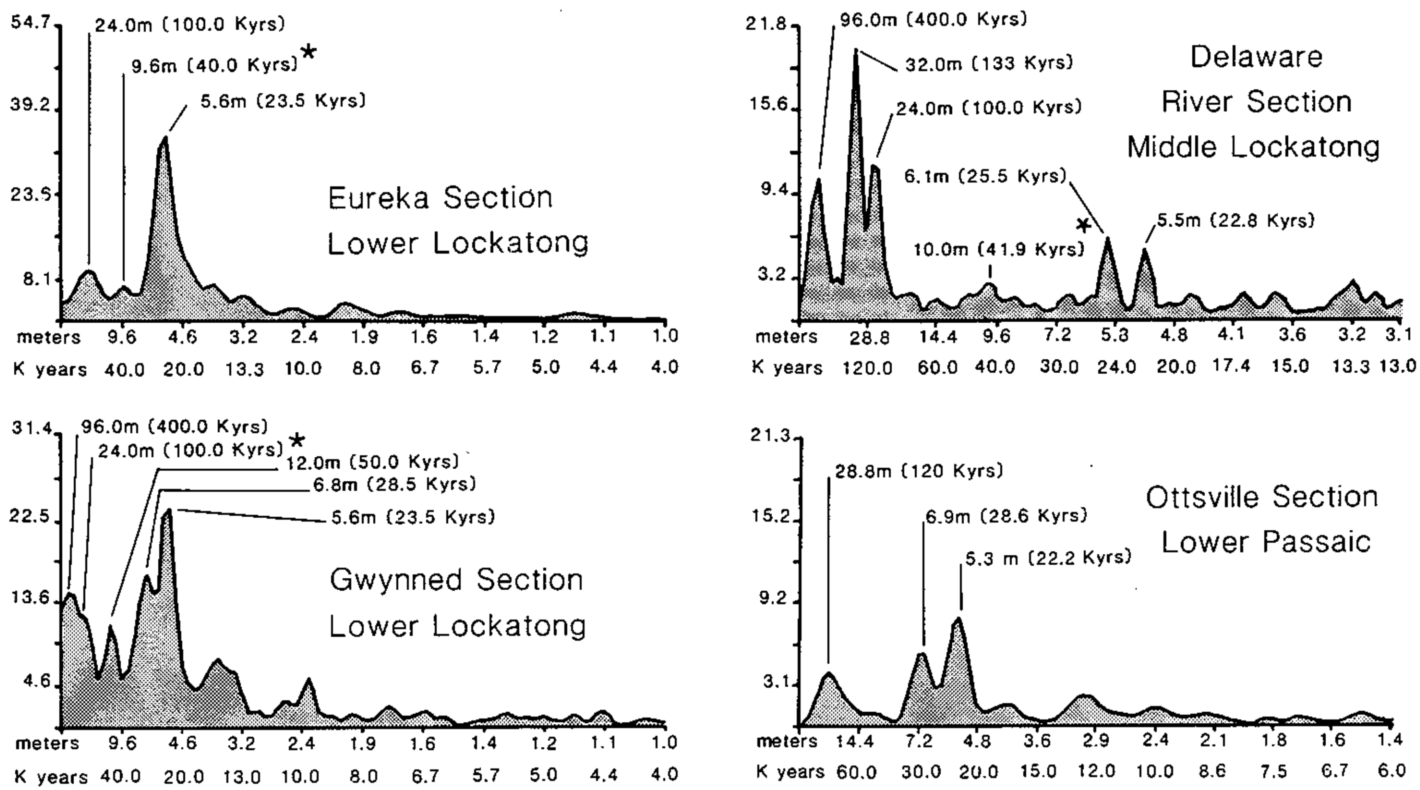


Figure 15. Power spectra of sections in the Lockatong and Passaic formations (from Olsen, 1986). The power spectrum for Stop 4 is the "Delaware River Section" on the upper right. Section is calibrated to time using varve counts yielding a sedimentation rate of 0.24 mm/yr.

syndepositional desiccation cracks. This is the high stand deposit of this cycle and is most comparable to the fabric seen in the lower part of division 3 of the cycle shown in Figure 16a.

Division 3 consists of massive mudstone (argillite) sequences consisting of a basal green-gray mudstone passing into a very well developed breccia fabric (Smoot, 1985) and then upwards into a well developed crumb fabric. Breccia fabrics consist of densely spaced anastomosing cracks separating non-rotated lumps of mudstone showing some remnant internal lamination. They are very common in the Lockatong Formation and are directly comparable to fabrics produced by short periods of aggradation separated by longer periods of desiccation and non-deposition (Smoot and Katz, 1982). The breccia fabric passes upward into a well developed vesicular crumb fabric. The transition into the overlying similar sequence is abrupt. These thin sequences, several of which here make up division 3 of a Van Houten Cycle, are very common in the fine grained facies of the Newark Supergroup (Smoot in Olsen and Gore, in press).

The upper half of these thin sequences are cut by dish-shaped, upward concave clay-lined surfaces with radial slikensides. These "dish structures" were described by Van Houten (1964, 1969) and interpreted as a possible gilgai-type soil feature (Van Houten, 1980). A mechanical analysis of these structures suggests they are coulomb fractures produced in relatively rigid, but not lithified, mud by more or less isotropic expansion in a three dimensional analogue of Davis, and others' (1983) "bulldozer" model for the development of thrust faults in accretionary wedges. The volume increase was probably due to eolian deposition of mud in desiccation cracks followed by rewetting. Because the bed was confined laterally, it fractured along listric, dish-shaped faults and compensated for increased volume by increasing in thickness. As a mechanical consequence of volume increase, the dish-structures are probably not specific to any particular environment in larger sense, but could be indicative of a limited range of environments in a particular basin setting such as the Lockatong (J. Smoot, pers. comm).

Clearly, this particular Van Houten cycle represents the transgression and regression of a lake which never became as deep as the cycle shown in Figure 16a and when lake level dropped the lake desiccated for longer intervals. Unfossiliferous here, cycles of this type contain clam shrimp in division 2 and sometimes abundant reptile bones in division 3 in coarser facies (see Stop 4). Most of the cycles at this stop fall somewhere between these two extremes.

At the north end of the outcrop is a creek with excellent exposures of addition 100 m of section of the upper Tohicken and lower Prahls Island Member (Figure 13).

Return to vehicle and head north on NJ-29.

25.0 1.1 Upper part of Tohicken Member (McLaughlin's "First Thick Red" exposed on right.

25.2 0.2 Old Busik Quarry on right is in the upper part of Prahls Island Member (McLaughlin's "Triple Red"). Section in quarry has been described by Van Houten (1969, Stop 3).

25.5 0.3 Ravine on left exposes lower parts of Smith Corner Member with well developed black and gray Van Houten cycles. The lowest exposed produces *Cyzicus* and indeterminate fish scraps.

25.6 0.1 Red and Gray Van Houten cycles comprising the upper parts of the Smith Corner Member

25.9 0.3 Tumble Falls Road. Stream on right exposes most of Tumble Falls Member of upper Lockatong Formation, the upper the red parts of which correspond to McLaughlin's member B. Minor transgressive interval in middle of division 3 of a gray Van Houten cycle exposed in stream at road level contains abundant and well preserved *Rhynchosauroides* sp.

26.2 0.3 Upper red beds of Walls Island Member (member B plus High Rocks member of McLaughlin). These red beds have usually been considered the basal beds of the Passaic Formation (Van Houten, 1969) even though they are mineralogically more similar to the Lockatong (Van Houten, 1969). They are regarded here as the uppermost beds of the Lockatong and the base of the Passaic Formation is defined at the base of the gray beds of member C.

26.3 0.1 Exposures of gray and red Van Houten Cycles of McLaughlin's member C. Section described by Van Houten (1969, 1980).

McLaughlin (1944, 1946) split the Passaic Formation into members differently than the Lockatong and named the gray portions of the 400,000 year cycles, not the red. Thus, this gray interval is named C of the Passaic Formation (Brunswick of Kümmel, 1897), while the underlying red interval of the Lockatong was named B. Pending detailed revision, we use McLaughlin's member names for the Passaic, but include the overlying red half of each 400,000 year cycle in the named unit.

26.5 0.2 Large exposures of upper parts of Member C of Passaic

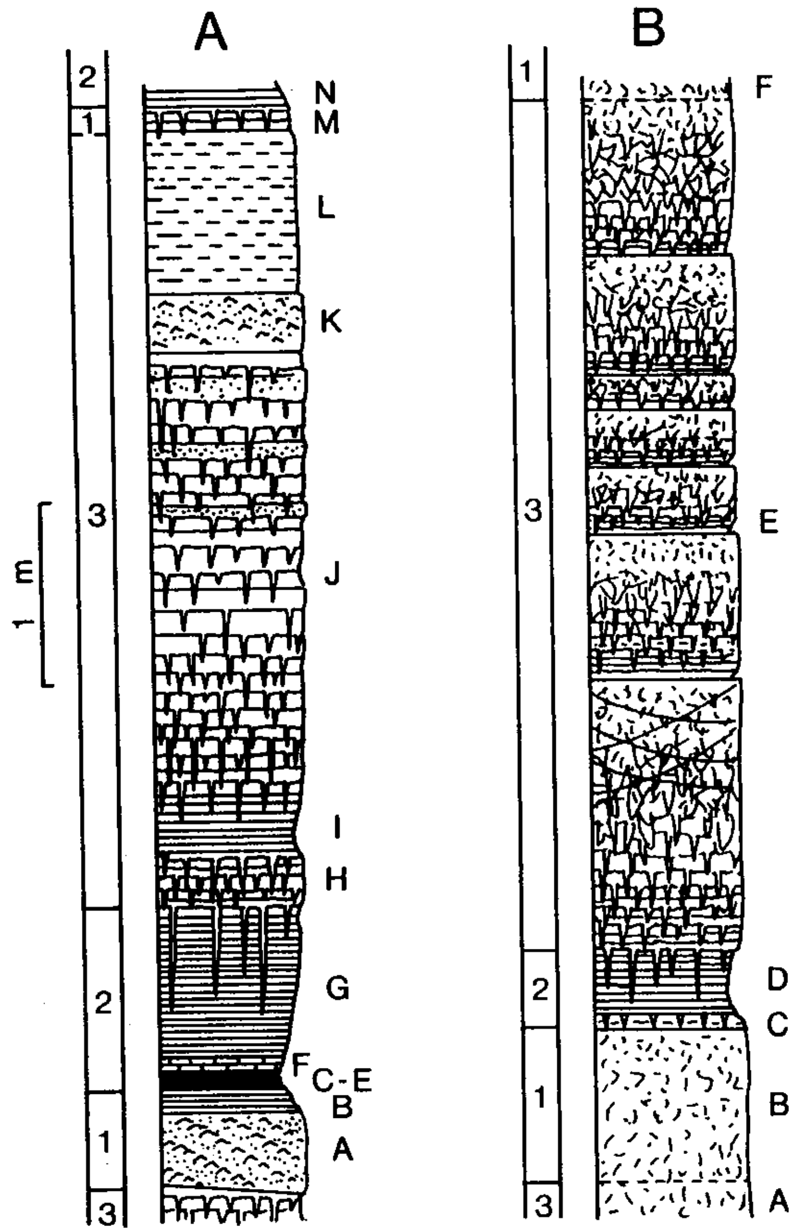


Figure 16. Two end members of the cycle types seen at Stop 4 along NJ-29. (A) corresponds to Van Houten's detrital cycle type while (B) corresponds to Van Houten's chemical cycle type (Van Houten, 1964). Key to units in Table 1.

Formation showing continuation of pattern of Van Houten cycles expressed as variations in red mudstone fabric resembling the red mudstone cycle described in detail at Stop 2. *Rhynchosauroides* sp. occurs in some of the red laminated mudstones of division 2 of these cycles.

27.2 0.7 Pull off on right just in front of culvert for Warford Creek.

STOP 5
MEMBER D OF PASSAIC FORMATION - WARFORD CREEK

Extensive exposures of a single Van Houten cycle in the lower part of Member — of Passaic Formation extend for over 0.5 mi. upstream (Figure 17). The lower part of division 2 of this cycle is microlaminated and contains articulated and disarticulated *Semionotus* cf. *braunii* and *Paleolomadia*-type clam shrimp. Division 3 contain *Scoyenia*, abundant tool marks, and possibly poor *Rhynchosauroides*. This same cycle outcrops 6 mi. to northeast along east branch of Nishisackawick Creek where it still contains fish, although they are disarticulated (Olsen and Gore, in press). Although mapping is as yet incomplete, member D evidently is traceable more than 30 mi to the south west to Linfield, Pennsylvania, near the Limerick nuclear power plant. At a small exposure along Longview Road the lower part of division 2 of the same cycle exposed along Warford Creek consists of black siltstone with well developed oscillatory ripples. Articulated and disarticulated *Semionotus* cf. *braunii* are common in calcareous nodules in the black claystone laminae. Oscillatory ripples adjacent to the fish are also carbonate cemented and much less compacted than other ripples showing the nodules to be of early diagenetic origin.

The uranium concentration is unusually high in sandstone beds of division 1 of this Van Houten cycle at Warford Brook, and moderately high in part of the thinly laminated black siltstone. No radon problem has been identified at this time in association with this unit, although, this may be due to the small number of houses located on it.(Paragraph contributed by K. W. Muessig and H.F. Houghton).

27.8 0.6 Excellent exposures of members E and F in creek to right. In this case, McLaughlin (1944) named two the gray parts of two successive 100,000 year cycles in the lower part of a 400,000 year cycle members E and F.

30.7 2.9 Members G and H of McLaughlin are poorly exposed in hills on right. G and H are two 100,000 years cycles

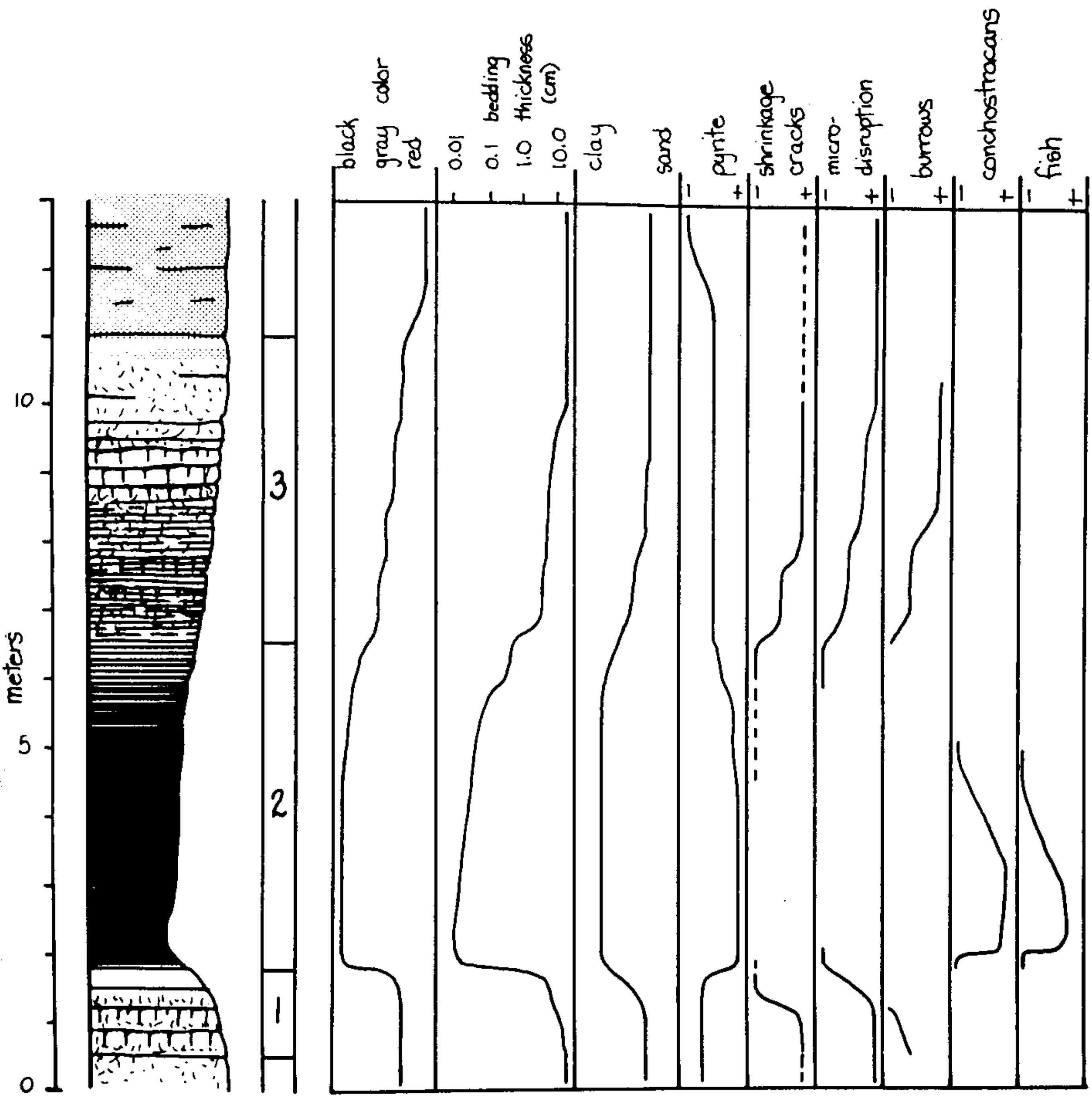


Figure 17. Van Houten cycle in member D of the lower Passaic Formation, Stop 5 (from Manspeizer and Olsen, 1981).

named in the same way as E and F; however, McLaughlin (1933) provided an alternate name, the Graters Member. McLaughlin (1943, 1944, 1946) mapped the Graters and its constituents over much of the basin.

30.9 0.2 Crossing Nishisakawick Creek.

31.0 0.1 Turn left (west) onto Main St. and then right (N) onto Hunterdon County Road 615 (Harrison Street) and head north.

32.6 1.6 Small quarry on right in Passaic Formation. Much burrowing by *Scoyenia* and some very poor footprints are present (*Brachychirotherium*). Outcrop described by Van Houten (1980, Stop 3).

34.6 2.0 Turn left at intersection with County Road 515 at traffic light, Milford center.

About 0.8 mi north of here are several good exposures of the Perkasie Member of the Passaic Formation and adjacent beds (Olsen and Flynn, in press). A road cut on the east side of Water Street (see Olsen and Flynn, in press) exposes gray and red claystones of division 2 of a Van Houten cycle about 5 m above the base of member O containing *Semionotus*, the clam shrimp *Cyzicus* sp. and cf. *Ellipsograpta* sp. and non-darwinulid as well as darwinulid ostracodes. These and a succeeding Van Houten cycle make up Cornet's (1977) pollen and spore localities M-3 and M-4. About 100 m above the Perkasie further north up the road are outcrops of another gray member.

McIntosh *et al.* (1984) have identified the boundary between a lower thick normally magnetized zone and an upper thick reversed zone in this same outcrop. Here the base of the Perkasie lies about 20 m above boundary between magnetozones. This boundary has been identified at several outcrops to the southwest in the Hunterdon Fault Block and in the Watchung Syncline 60 km to east (Olsen and Gore, in press).

The Smith Clark Quarry is located on the east side of the large hill to the east of Water Street (Olsen and Flynn, in press). It is the locality for much of the type material of reptile footprints characteristic of the Triassic parts of the Newark Supergroup. The types of *Atreipus milfordensis*, *A. sulcatus* (Baird, 1957; Olsen and Baird, 1986), *Brachychirotherium parvum* (C.H. Hitchcock, 1889), *B. eyermani* (Baird, 1957), *Apatopus lineatus* (Bock, 1952), and *Rhynchosauroides hyperbates* (Baird, 1957) were collected from this site. Examples of Grallator (*Anchisauripus*) *parallelus* (Baird, 1957), *Rhynchosauroides brunswickii* (Baird, 1957). *Coelurosaurichnus* sp. (Olsen and Baird, 1986), and an uncertain tridactyl form (Baird, 1957) have also been found along with abundant plants. The stratigraphy and detailed paleontology of this site have been described in Olsen and Flynn (in press).

McLaughlin's members L and M are exposed about 100 m below the Perkasie Member on west side of York Street, at the base of the hill below and to the south of the the Smith Clark Quarry (Olsen and Flynn, in press). Outcrops of these gray members contain clam shrimp and constitute Cornet's localities M-1 plus M-2 and M-6, respectively. Pollen and spores form these beds indicate an early Norian age (Cornet, 1977; Cornet and Olsen, 1985) for the Perkasie Member and members L and M.

34.7 0.1 Right onto Church Street and then left on continuation of Church Street.

34.8 0.1 Right onto Spring Glen Rd. (Hunterdon Co. 627).

THIS ROAD IS VERY NARROW AND HAZARDOUS; BEWARE OF ONCOMING VEHICLES.

35.1 0.3 Culvert on right and adjacent exposure of member M, described by Turner-Peterson (1980). Gray shale at base of division 2 of Van Houten cycle at this outcrop has produced disarticulated *Semionotus* and *Cyzicus*.

35.7 0.6 First of a series small faults which are down to the east.

35.9 0.2 Estimated position of second fault.

36.0 0.1 Cyclical mudstone beds of Passaic Formation below members L and M.

Cycles are obvious at several scales. Picard and High (1963) report that this exposure show an alternation of massive ledge forming units of poorly sorted siltstone with less resistant and finer-grained units, with the couplets ranging from 1.1 to 4.8 m and averaging 2.6 m thick. Most units show intense bioturbation and common desiccation cracks merging into breccia fabrics. The thinner of these alternations seem to us to be grouped into larger cycles averaging 5 to 7 m thick. We hypothesize that these larger cycles are probably homologous to the Van Houten cycles in the more arid portions of the 400,000 year cycles as seen at Stop 4, while the shorter alternations correspond to the "aggradational cycles" also seen in the Lockatong at Stop 4. This hypothesis needs testing by careful measurement of this section combined with Fourier analysis. This section is also described in Van Houten (1969, Stop 5).

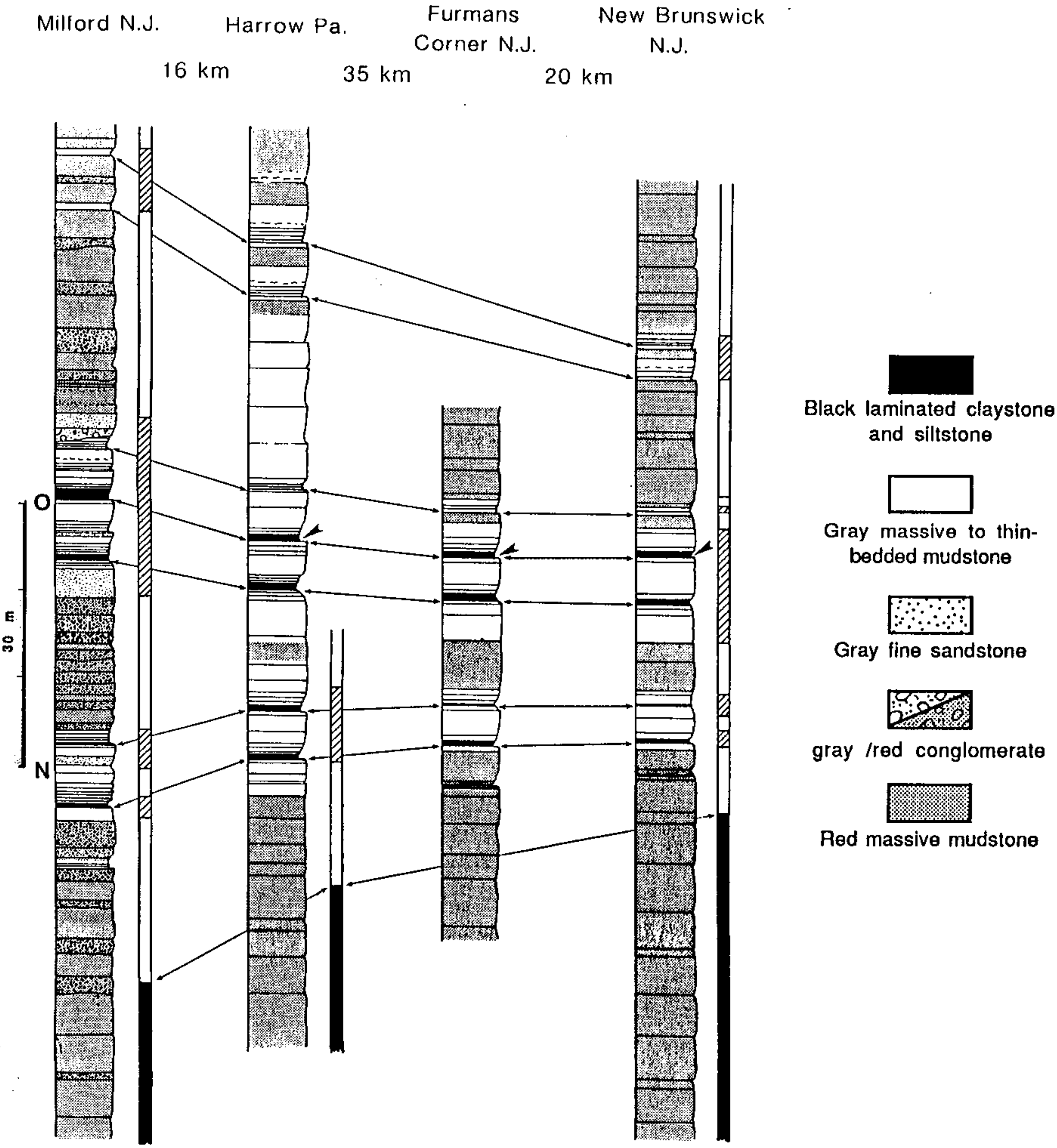


Figure 18. Stratigraphy and lateral correlation of the Perkasie Member from New Brunswick (on right) to Milford on left (from Olsen, in press). Vertical black, white, and diagonal ruled bars represent normal, reversed, and uncertain polarity zones, respectively. Large chevrons denote intervals enriched in uranium.

36.3 0.3 Park in dirt area off road to right at railroad milepost 37 (adjacent to tracks and road).

STOP 6
PEBBLE BLUFF EXPOSURES OF CONGLOMERATE AND LACUSTRINE BEDS AT EDGE OF NEWARK BASIN

These outcrops [also described by Van Houten (1969, Stop 6; 1980, Stop 1) and Arguden and Rodolpho (1987)] consist of extensive exposures of thick sequences (+20 m) of red conglomerate and sandstones alternating with cyclical black, gray, and red mudstone, sandstones. Dip averages 10-15° NW and there several faults between the largest outcrops which drop down to the east. The largest gray beds (see Figures 18, 19) appear to be the Perkasie Member.

The Perkasie Member consists of Van Houten cycles and compound cycles showing the same pattern as the Lockatong (Figure 18) (Olsen, in press; Olsen and Baird, 1986). It consists of two sequential 100,000 year cycles, each containing two well developed Van Houten cycles and one weakly developed red and purple Van Houten cycle, succeeded upwards by red clastics (Figure 20). McLaughlin named the lower set of gray beds member N and the upper member O.

Once thought to be the youngest strata in the Basin, it is now clear that the Perkasie Member and surrounding strata actually lie in the lower Passaic Formation roughly 2,800 m below the top of the formation. All of the overlying strata have been eroded in this area. Correlation of the Milford area with other fault blocks of the Newark basin where these younger beds are still preserved is afforded by magnetostratigraphy, lithological matching, and a palynological correlation web.

Most striking at these outcrops of the Perkasie and adjacent units are the large amounts of conglomerate present (Figure 19). According to Van Houten (1969, 1980), most of the conglomerates are made up of fining-upward sequences averaging about 7-9 m thick. Each sequence consists of poorly sorted conglomerate passing upward into poorly sorted sandstone with carbonate nodules (arranged in vertical columns or streaks) which become more dense upward. Most clasts are less than 8 cm in diameter, but some are as large as 23 cm (Figure 20). According to Van Houten (1969) these sequences resemble those of alluvial fans and the calcareous horizons are probably caliches. Some of the mudstones and sandstones show cross bedding at the more eastern outcrops, but elsewhere, they appear massive (Van Houten, 1969). We

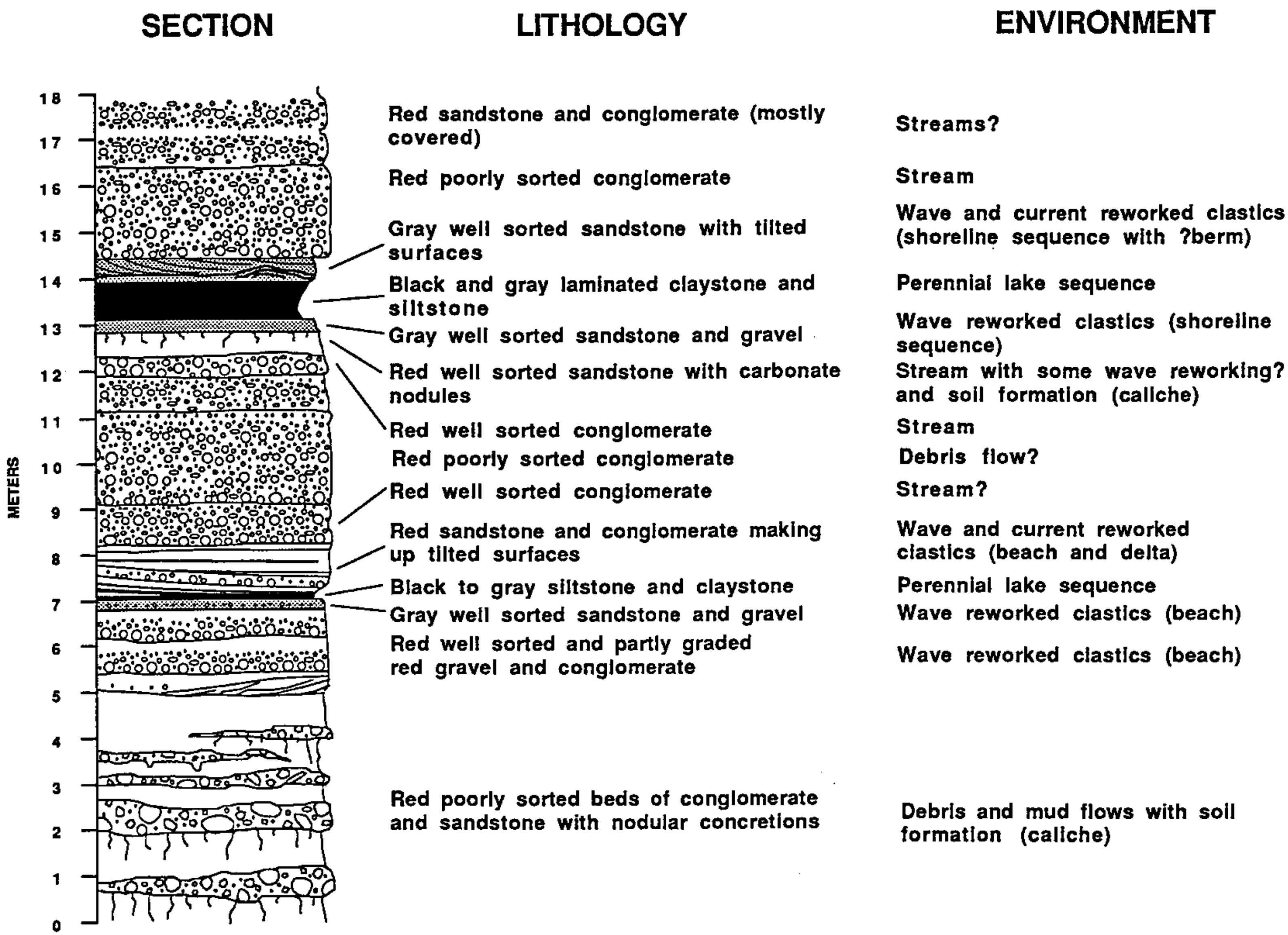
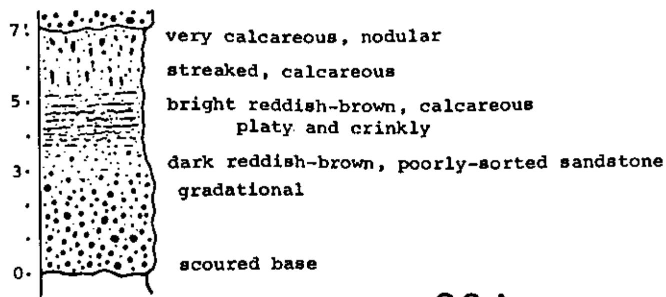
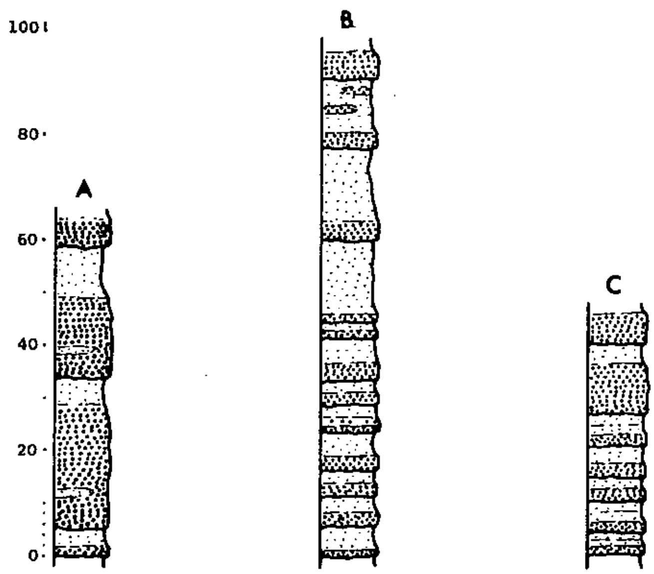


Figure 19. Section through small part of section at Stop 6 showing interfingering between perennial lake sequences, shoreline deposits, and alluvial fan sequences.



20A



20B

Figure 20. Conglomerate and caliche from Pebble Bluff area, Stop 6: 20A, Model of fining-upwards conglomerate-sandstone unit, showing sequence of elements commonly present (from Van Houten, 1969); 20B, Vertical sections of lenticular fining-upwards conglomerate-sandstone units [A is 0.5 mi northwest of C, maximum clast size is about 33 cm for (A) and about 23 cm for (B)](from Van Houten, 1969).

wonder if these sequences might not be related in some way to the smaller scale cycles seen at mileage 36.0.

Arguden and Rudolpho (1987) recognize three types of conglomerates at these outcrops: 1) matrix supported, poorly sorted conglomerate often showing reverse grading, which they interpret as debris flow deposits; 2) poorly sorted clast supported conglomerate, which they interpret as a hyperconcentrated stream facies; and 3) better sorted clast supported conglomerates which they interpret as short-lived braided stream deposits. They interpret all three conglomerate types as the products of alluvial fans. At least some beds which fit the criteria of their last type of conglomerate, however, are very closely packed, tabular beds with crude normal grading, and are intimately associated with black and gray Van Houten cycles of the Perkasie; these closely resemble wave reworked fluvially supplied gravels as described by Bourgeois and Leithold (1984) and LeTourneau (1984). Within these conglomerates J. Smoot (pers. comm.) recognizes the presence of patches of oscillatory ripples, preferential sorting of larger particles adjacent to pebbles, interbedded sets of internally well sorted thin beds of different clast size.

Laterally continuous black and gray siltstones and claystones within division 2 of Van Houten cycles (Figure 18) contain pinch and swell laminae abundant burrows and rare conchostracans and are definitely lacustrine deposits, almost certainly marginal to the finer grained facies more centrally located in the basin. These lacustrine sequences mark transgressions of perennial lakes over the toes of alluvial fans, much as LeTourneau (1984) has described in marginal facies of the Portland Formation in Connecticut. Division 2 of cycles comprising the Perkasie member were produced by lakes which were almost certainly shallower than those which produced the microlaminated, whole fish- and reptile-bearing units in other parts of the Newark basin section. They evidently were deep enough, however, to transgress over at least the relief caused by the toes of alluvial fans. To what degree have more subtle transgressions of shallower lakes modified clastic fabrics elsewhere in this section? A number of researchers (Van Houten, pers. comm.; Smoot, pers. comm., Manspeizer, pers. comm.) have commented how many of the conglomerate clasts in many of the debris flow units are relatively well rounded, and mixed with angular clasts. We suggest that it is possible that these debris flows originated by degradation of shoreline deposits formed high above the basin floor during high stands of the great Perkasie Member lakes.

In this reagion individual Van Houten cycles of unit O are an average of 7 m thick, as compared with a mean of 6.5 for the same two cycles to the southeast and 4.5 for the cycles at New Brunswick (Figure 20). The large, comound cycles show the same trend as well. This thickening toward the border fault is typical of most of the Newark Supergroup. The trend to greater fossil richness is evident at this outcrop as is typical for more marginal facies of the Passaic Formation, and the conchostracans present in the upper parts of division 2 are absent in the more basinward exposures.

The border fault is less than 2.4 km (1.5 mi.) to the northwest of this series of outcrops. As mapped by Ratcliffe and others (1986) on the basis of surface geology, drill cores, and Vibroseis profile the border fault in this area is a reactivated imbricate thrust fault zone dipping 32° SE. Core and field data reveal Paleozoic mylonitic fabrics of ductile thrust faults in Precambrian gneiss and early Paleozoic dolostone overprinted by brittle cataclastic zones of normal Mesozoic faults which form the border fault system of this part of the Newark basin. According to Ratcliffe and others (1986, p. 770), "Field evidence indicates reactivation of the imbricate thrust faults in an extensional mode, a dominant extension direction being S40° to S50°E. These data suggest strongly (1) that reactivation of Paleozoic thrusts controlled the formations of the border fault in eastern Pennsylvania and (2) that fault patterns may have formed by passive response to extension rather than by active wrench fault tectonics as proposed by Manspeizer (1980). This differs from the interpretation of Drake and others (1967) who proposed that the border fault of the Newark basin is a relatively high-angle reactivated, overturned thrust fault of the Musconetcong nappe. Ratcliffe and others' model is in agreement with proprietary seismic lines and drill hole records in this area and to the southwest.

A series of very gentle but rather complex folds are present in this area close to the border fault (Figure 7). They are the most northeast folds in the hanging wall along this part of the border fault system. On the New Jersey side of the Delaware River their axes are parallel to the fault but the folds have axes which switch to high angles with the border fault on the Pennsylvania side of the river. What might this transition have to do, if anything, with the formation of folds in the basin?

Additional Comments on Structure (contributed by R.T. Faill)

The northward dip of bedding, and the various fractures dominate the structure here. One set of fractures is noteworthy. These fractures dip southward, and are quite common and persistent. The distinctive aspect is that some of these bend sharply into the bedding planes. No offsets are apparent, but at least one of them has subhorizontal slickenlines, parallel to the strike (northeast). This suggests that these fractures are strike slip faults, and therefore the absence of bed offsets is not surprising.

Spring Garden Road is 600 m south of the nearby railroad milepost 37. The Passaic Formation crops out for 450 m south of Spring Garden Road, with beds dipping northward at approximately 10 degrees. As another illustration of the sporadic distribution of structures, cleavage and numerous faults occur near the south end of that outcrop, a distinct contrast to here, at Stop 5. There, the most common features are steeply northeast to east dipping fractures, some of which offset bedding by several centimeters each, down on the east. Slickensides are steeply to moderately northwest plunging, an orientation also seen at Stop 1.

In addition, a few of the argillaceous siltstone beds have developed a spaced cleavage in them that dips steeply to the east, subparallel to the faults. Spacing of this fracture cleavage is 1 to 5 cm, and a lineation on the cleavage plunges steeply to the north. This lineation is more a crenulation than slickenlines. The cleavage is not present in the more argillaceous beds, which themselves possess a fair fissility.

Many of the outcrops along the Delaware River do not exhibit much structure. But they do support the two contentions: 1) the mesostructures that are present show little direct relevance to the overall extensional tectonics of the basin; and 2) the occurrence of specific structures is sporadic.

END OF FIELD TRIP

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MESOZOIC DIABASES ALONG THE DELAWARE RIVER:

A ROAD LOG FOR A HALF-DAY FIELD TRIP

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INTRODUCTION

This field trip is intended to familiarize you with a variety of intrusive Mesozoic diabasic rocks exposed in the central Newark Basin of west-central New Jersey and eastern Pennsylvania along the Delaware River (Fig. 1). The rocks occur in a number of separate sheet-like bodies, including the Point Pleasant or Byram diabase, the Stockton diabase, the Lambertville sill, the Belle Mountain diabase, the Baldpate Mountain diabase, the Pennington Mountain diabase, and the Rocky Hill diabase, and a small dike known locally as the Quarry dike. Although it will be impossible to examine the entire range of diabasic rock types found in the central Newark Basin, the four stops discussed in the road log will provide an understaning of regional rock diversity. In addition, the proposed petrogenetic processes summarized in following sections and discussed in more detail elsewhere in this volume (see Husch and others, 1988) should lead to lively and fruitful discussions in the field.

The hypabyssal, diabasic nature of the Mesozoic igneous rocks of the central Newark Basin has been recognized since before the beginning of this century (see Phillips, 1899; Bascom and others, 1909). Whole-rock compositions range from mafic-rich pyroxene cumulates to highly fractionated granophyric differentiates. Texturally, the rocks vary from nearly glassy, fine-grained basalts to coarsely crystalline, granular gabbros. Diabasic textures are common in almost all intrusions. Although no diabase intrusion from the central Newark Basin region has

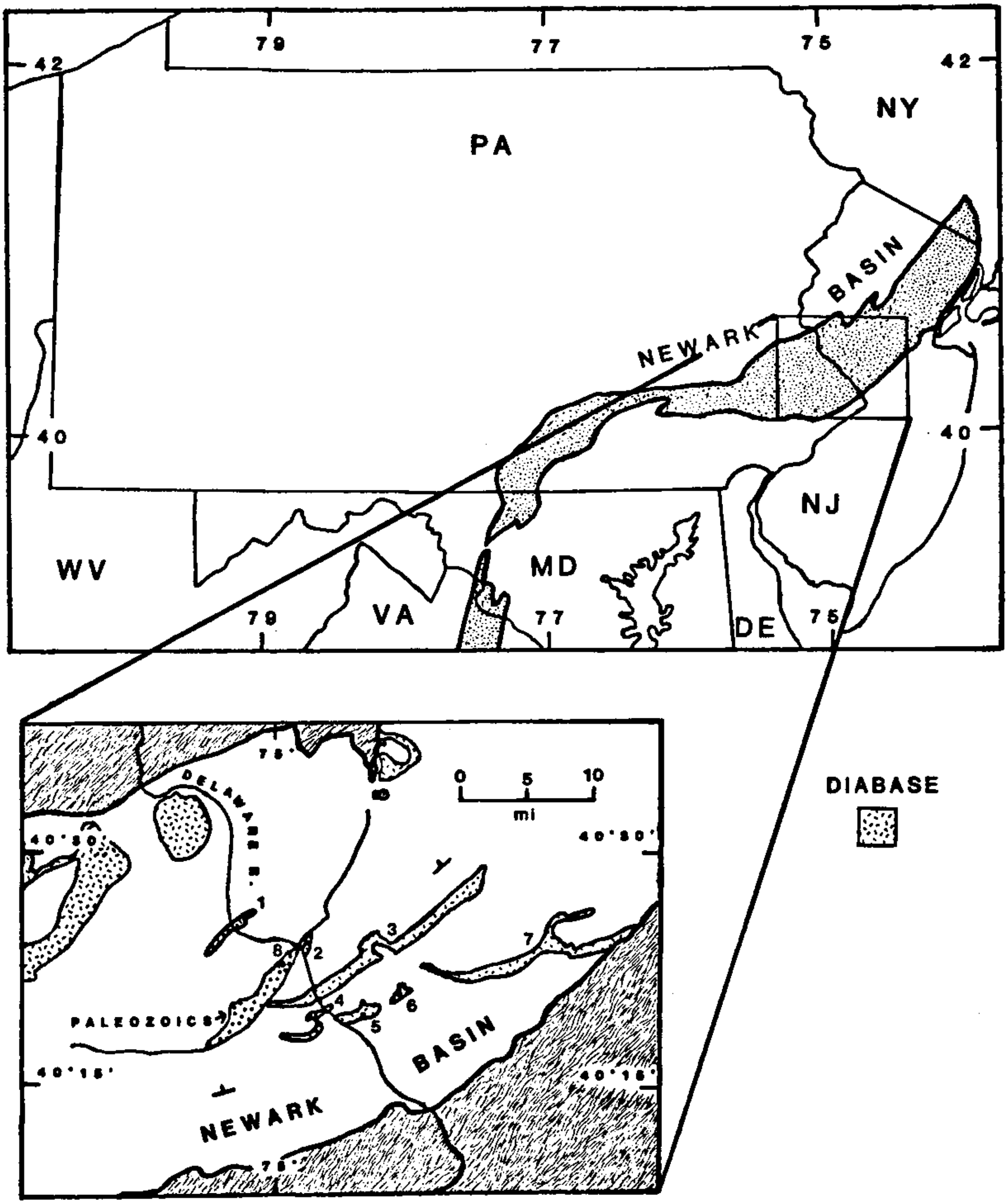


Figure 1. Location of Newark Basin and geologic sketch map of Mesozoic diabase in the Delaware River region. Individually numbered intrusions shown are: 1) Byram (Point Pleasant) diabase; 2) Stockton diabase; 3) Lambertville sill; 4) Belle Mountain diabase; 5) Baldpate Mountain diabase; 6) Pennington Mountain diabase; 7) Rocky Hill diabase; 8) Quarry dike locality.

been directly dated by radiometric means, their similarity to Mesozoic diabases whose ages are known (Dallmeyer, 1975; Seidemann and others, 1984; Sutter, 1988) makes it most likely that they were intruded at essentially the same time (~200 my) during the earliest Jurassic. Paleomagnetic (Hozik and Columbo, 1984) and sedimentological (Olsen and Fedosh, 1988) data support the interpretation that Newark Basin igneous activity was confined to a very short span of time at the beginning of the Jurassic period. Indeed, many of the intrusions from the region may be lateral extensions of the Palisades sill re-exposed by subsequent normal faulting (Lewis, 1907; Van Houten, 1969).

PETROGRAPHY

Most diabase samples from the central Newark Basin are composed predominantly of plagioclase (An 50-70), clinopyroxene (augite and pigeonite), orthopyroxene (En 70-80), and Fe-Ti oxides. Accessory minerals include biotite, apatite, quartz, alkali feldspar, pyrite, and sphene. In highly fractionated granophyres, clinopyroxene often is replaced by deuteric hornblende and biotite, plagioclase compositions become more sodic (An 20-50), particularly near grain boundaries, and quartz and alkali feldspar usually occur in graphic micro-pegmatitic intergrowths. Olivine is very rare and there are no known olivine accumulation zones. Where olivine is found, it is present in small amounts and commonly is embayed and anhedral in form, suggesting an extensive peritectic reaction with the host basaltic liquid.

Accumulations of large (up to 5 mm), euhedral to subhedral clinopyroxene and/or orthopyroxene are found in a number of samples, particularly those from just above the base of the Lambertville sill and in the Byram (Point Pleasant) diabase. In addition, glomeroporphyritic masses of clinopyroxene are common in most chill margins.

GEOCHEMISTRY

Major- and trace-element whole-rock concentrations have been determined for nearly 100 samples (Husch and others, 1984; Husch and Schwimmer, 1985; Husch, 1988; Husch and Roth, 1988). In general, diabase compositions are quite variable. They include: 1) high magnesium (>9% MgO) diabase; 2) less mafic (<9% MgO) diabase extremely close in bulk chemistry to the quartz normative, high-titanium tholeiite (HTQ) of Weigand and Ragland (1970) and York Haven tholeiite of Smith and others (1975); 3) high-iron type diabase derived from a HTQ parent; 4) even more highly fractionated, residual granophyres. Samples from fine-grained chill margins exhibit remarkably uniform compositions and are similar in composition to numerous other HTQ magmas from other localities in the Mesozoic Eastern North America (ENA) Province.

Chill margin samples also contain the lowest initial Sr isotopic values ($\approx$.706). These findings support the assumption that the chill margin compositions are the best available approximation of actual parental (although not necessarily primary) magma compositions and that only one major basalt type

(HTQ) is recognized with any certainty in the central Newark Basin (Husch 1987). Although recent whole-rock geochemical data from the Lambertville sill suggests the presence of a second magma type similar in composition to the quartz normative, low-titanium tholeiite (LTQ) of Weigand and Ragland (1970) and Rossville tholeiite of Smith and others (1975), no LTQ chills have been identified.

Harker-type and ternary variation diagrams exhibit regional differentiation trends consistent with crystal fractionation. Overall, observed variation trends for diabasic rocks of the central Newark Basin region are nearly the same as those seen for whole-rock samples collected from vertical sections through the Palisades sill (Walker, 1969). For central Newark Basin diabases, fractionation initially was controlled by the removal of pyroxene from parental melts, followed by the removal of plagioclase (≃An 65), Fe-Ti oxide, and finally apatite. Although the early pyroxene fractionation was dominated by the removal of augite, pigeonite and orthopyroxene (≃En 80) removal also occurred. High magnesium (>9% MgO) samples represent the accumulation of these early crystallizing mafic phases, presumably by gravity settling since they are found toward the base of an intrusion beneath complementary, more felsic residual compositions.

In the Lambertville sill, magma injection and subsequent fractionation appears to have occurred in at least two distinct pulses, one of which may have been of a LTQ-like magma (Husch and Roth, 1988). This is indicated by the presence of a second mafic-rich zone approximately 700 feet above the base of the intrusion.

In contrast to the overwhelming majority of samples whose compositions are compatible with crystal fractionation, there are at least six samples from three sheet-like bodies which contain anomalously high alkali concentrations for their corresponding MgO content. These same samples also contain anomalously low CaO concentrations. These relationships were interpreted by Benimoff and Sclar (1984; sample D1) and Husch and Schwimmer (1985) to result from the contamination of variably differentiated diabase by local country rock whose compositions are best represented by hornfels from the Lockatong Formation. Contamination apparently was controlled by the selective diffusion of alkalis from the contaminant into the magma and of Ca from the magma into an anatectic melt derived from the contaminating material.

The Quarry dike contains no textural or geochemical evidence for fractional crystallization. However, inversely correlated variations for CaO and the alkalis, supported by Sr isotopic evidence, suggests that the compositional variability found across the dike was controlled solely by contamination (Husch and Schwimmer, 1985; Husch and others, 1988).

Finally, calculated average compositions for many of the differentiated sheet-like intrusions do not match their chill margin compositions (Husch, 1987). Intrusiòns with abundant mafic-rich compositions often lack highly evolved granophyres and visa-versa. Generally, intrusions rich in granophyre are found higher in the stratigraphy than those with large mafic-rich zones. This suggests that residual liquids were displaced laterally up-section from early cumulates. Their displacement may have been driven by the multiple injection of relatively fresh, undifferentiated magma from below.

PALEOMAGNETISM

The diabasic and basaltic rocks of the Newark Basin have been the objects of numerous paleomagnetic investigations (Opdyke, 1961; Beck, 1972; Hozik and Colombo, 1984; McIntosh and others, 1985, Stuck and others, 1988; Hozik and Opitz, 1988). A comparison of the results of the these studies supports several conclusions which place constraints on the timing of both the igneous activity and the deformation of sediments within the basin.

By way of introduction, a virtual geomagnetic pole is the geographic position of the earth's magnetic pole required to produce the declination and inclination of the remanent magnetism measured for a given igneous body. Ideally, the virtual geomagnetic poles for all rocks magnetized at the same time would coincide. This is in marked contrast to the magnetic vectors for individual intrusions of the same age which vary as a function of the geographic position of each rock unit. To compensate for this variation, virtual geomagnetic poles were calcualted for all of the Newark Basin igneous bodies studied and were used in all comparisons. In addition to the declination and inclination of the magnetic vector for each sample, Table 1A lists the derived latitude and longitúde of the virtual geomagnetic pole for each site, rotated and unrotated for subsequent tilting, and the mean pole for each igneous body. Table 1B shows the latitude and

Table 1A: Paleomagnetic data collected by Hozik and students.

		UNCORRECTED FOR STRUCTURE				VGP		CORRECTED FOR STRUCTURE				VGP	
Sample	N	K	A	D	I	LAT	LONG	K	A	D	I	LAT	LONG
By 1	6	121	6.1	12.6	40.0	69.4	70.1	110	6.4	3.0	33.0	67.5	97.5
By 2	4	586	3.8	13.8	42.2	70.2	65.0	392	4.6	3.5	35.0	68.8	95.8
By 4	7	257	3.7	-1.4	37.9	70.8	108.9	245	3.8	-8.7	28.0	63.5	124.1
By 5	5	1188	2.2	6.5	37.8	70.0	87.0	432	3.7	-1.0	29.0	65.2	107.3
By 8	3	181	9.0	10.0	41.0	70.9	75.1	235	9.0	2.8	28.0	65.7	99.1
By 1-8	25	136	2.5	7.2	39.6	71.1	83.5	121	2.6	-1.0	30.8	66.3	107.1
By 23	3	390	6.0	5.1	33.7	67.6	92.1	446	6.0	0.7	20.4	60.3	103.6
By 24	1	-	-	4.0	35.0	68.8	115.4	-	-	-8.0	23.0	60.9	121.2
By 25	3	428	6.0	21.7	42.1	65.7	49.8	621	5.0	13.3	32.3	64.5	74.2
By-Pa	7	67	7.4	10.5	37.8	69.0	76.8	64	7.6	4.5	26.0	63.2	95.2
All By	**32**	**113**	**2.4**	**8.0**	**39.2**	**70.7**	**81.8**	**94**	**2.6**	**0.2**	**29.8**	**65.7**	**104.1**
RH 14	3	68	15.0	-2.6	43.9	75.1	114.4	78	14.0	4.4	23.4	61.6	96.1
RH 16	5	26	15.2	-7.4	61.4	84.1	-140.6	27	15.0	6.0	41.0	72.4	86.7
All RH	**8**	**26**	**11.1**	**-5.1**	**54.8**	**83.6**	**145.3**	**27**	**10.9**	**5.3**	**34.4**	**68.0**	**91.6**
S 7	**3**	**121**	**11.0**	**17.5**	**37.8**	**65.7**	**62.0**	**108**	**11.9**	**11.5**	**29.5**	**63.4**	**79.6**
SB 17	3	104	12.0	-6.1	45.7	75.8	127.8	142	10.4	-3.7	47.9	78.1	122.3
SB 18	3	156	10.0	6.9	50.8	79.4	71.2	220	17.0	-7.2	53.1	81.1	62.4
SB 19	3	25	25.0	1.8	40.4	72.5	99.6	25	25.0	3.9	42.0	73.5	92.4
ALL SB	**9**	**139**	**11.0**	**0.7**	**46.0**	**76.7**	**102.4**	**54**	**7.0**	**4.0**	**47.0**	**60.4**	**103.1**
JI 1	8	53	14.2	-14.4	53.1	76.7	168.7		14.2	-0.6	10.0	54.7	105.2
JI 5	8	188	4.0	- 0.6	13.2	56.4	105.3		4.0	-4.1	21	60.3	112.3
JI 6	6	36	11.3	- 9.9	20.3	58.9	123.3		11.3	-18	27	59.6	140.5
ALL J	**22**	**14**	**34.8**	**- 7.4**	**28.8**	**64.2**	**120.8**	**45**	**18.7**	**-7.3**	**19.5**	**59**	**118.3**
JL 2	6	42	10.5	10.1	63.7	81.1	-23.3		10.5	12	32.2	64.9	76.3
JL 3	6	16	17.4	-23.6	39	62.9	158.8		17.4	-37.6	38	52.9	174.8
ALL JL	**12**	**13**	**36.2**	**-21.1**	**45.1**	**67.8**	**162.6**		**51.6**	**4**	**31.9**	**66.7**	**94.5**
P 1	8	42	8.7	11.7	15	55.2	85.2	41	8.8	7.8	11.1	54	92.5
P 2	12	5	21.5	14.6	21.2	57.4	78.4	9	15.1	20.6	18.4	53.6	69.9
P 3	15	35	6.5	7.8	45.1	74.4	78.9	36	6.4	-7.9	45.3	74.5	133.1
ALL P	**35**							**10**	**8.1**	**6.7**	**29**	**64**	**90.9**

By = Byram Diabase--Samples 1, 2, 4, 5, & 8 collected in New Jersey; Samples 23, 24, & 25 collected in Pennsylvania; Their mean is designated as By-Pa.
RH = Rocky Hill Diabase
S = Sourland Mountain Diabase (Lambertville Diabase)
SB = Sand Brook Lava Flow
JI = Jacksonwald Syncline Sill
JL = Jacksonwald Syncline Lava Flow
P = Preakness (I, II, III, first, second, and third flow units)

N = Number of specimens
K = Fisherian precision parameter
A = Radius of cone of 95% confidence
D = Declination
I = Inclination
VGP = Virtual geomagnetic pole

Table 1B: Virtual Magnetic Pole Positions From Triassic Basins

UNIT	N	K	A	D	I	LAT	LONG	SOURCE
Orange Mt.	12	85		3	34	67.9	97.8	Opdyke (1961)*
	8	50		3	23	61.2	99.4	Opdyke (1961)*
	31	251	4.8	359.8	29.9	65.2	106.1	M, H, W **
Preakness	10	157		18	25	58.3	70.5	Opdyke (1961)*
	8	350		19	14	52.5	73.4	Opdyke (1961)*
	55	21	11.5	13.9	27.9	61.4	76.4	M, H, W **
	35	10	8.1	6.7	29	64	90.9	Hozik & Opitz (1988)
Hook Mt.	12	52		10	31	64.3	83.1	Opdyke (1961)*
	34	55	12.5	4.0	27.1	63.3	97.0	M, H, W **
Haycock	13	21		360	11	55.1	104.5	Opdyke (1961)*
Baldpate	12	85		4	36	69.4	94.4	Opdyke (1961)*
Rocky Hill	12	27		354	25	66.3	109.8	Opdyke (1961)*
	8	27	10.9	5.3	34.4	68.0	91.6	Hozik & Colombo (1984)
Belle Mt.	12	31		339	31	60.2	148.8	Opdyke (1961)*
Sourland Mt.	8	88		358	31	66.3	109.8	Opdyke (1961)*
	3	109	11.9	11.5	29.5	63.4	79.6	Hozik & Colombo (1984)
Palisades	12	73		4	30	65.0	96.9	Opdyke (1961)*
Byram	32	94	2.6	0.2	29.8	65.7	104.1	Hozik & Colombo (1984)
Sand Brook	9	54	7.0	4.0	47.0	60.4	103.1	Hozik & Colombo (1984)
Jacksonwald Int	22	45	18.7	-7.3	19.5	59	118.3	Stuck & others (1988)
Jacksonwald flo	12		51.6	4	31.9	66.7	94.5	Stuck & Others (1988)
Pa Intrusives	78	118	3.2			62.0	104.5	Beck (1972)
Newark Group						63	108	Opdyke (1961)
Gettysburg	7	54	8.3			65.0	103.0	Beck (1972)
Yorkhaven	37	237	1.5			61.0	105.5	Beck (1972)
Birdsboro	26	85	3.2			62.0	105.5	Beck (1972)
Quakertown	9	143	4.3			66.0	105.0	Beck (1972)
Mass Lavas						55	88	Irving & Banks (1961)

* Lat and long of pole calculated for this study from data presented in Opdyke (1961).
** Lat and long of pole calculated for this study from data presented in McIntosh, Hargraves, and West (1985)

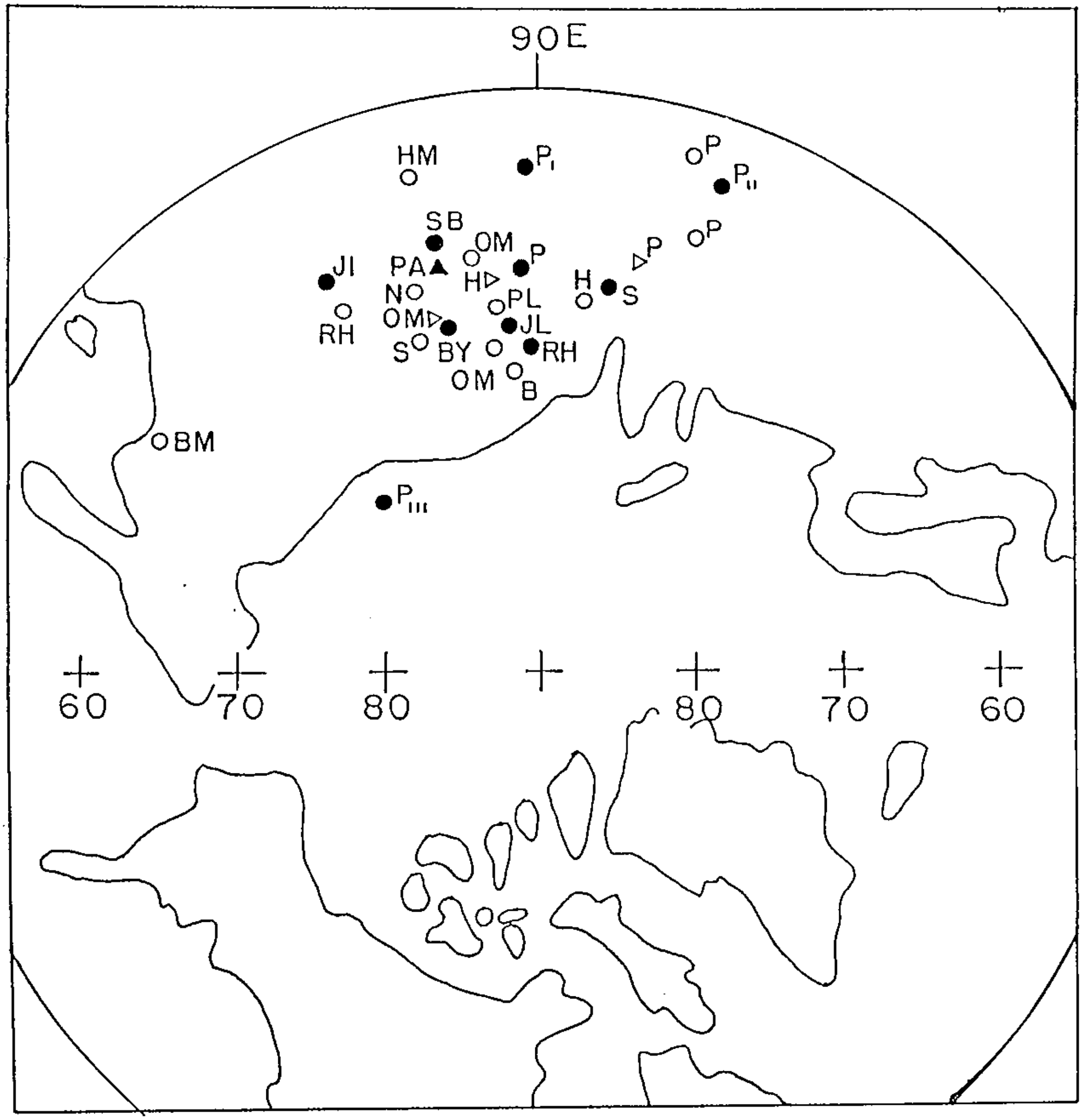


Figure 2A. Virtual geomagnetic poles for rocks of the Newark Basin. Solid circles indicate data from Hozik and Columbo (1984), Hozik and Opitz (1988), or Stuck and others (1988). Open circles are from Opdyke (1961). Open triangles are from McIntosh and others (1985). Solid triangle is from Beck (1972). Key to letters: SB-Sand Brook basalt; RH-Rocky Hill diabase; By-Byram diabase; S-Lambertville (Sourland Mountain) sill; HM-Haycock Mountain diabase; BM-Belle Mountain diabase; PA-Pennsylvania intrusives; PL-Palisades sill; OM-Orange Mountain Basalt; P-Preakness Basalt; PI, PII, PIII-first, second, and third flow units of the Preakness Basalt, respectively; JI-Jacksonwald intrusion; JL-Jacksonwald lava flows; H-Hook Mountain Basalt; N-Newark Basin sediments.

longitude of all the virtual geomagnetic poles from our work, important poles calculated from the data of other workers, as well as the sources of the information.

Figure 2A is a plot of virtual magnetic poles calculated for each of the igneous bodies after correcting for the tilting of the adjacent sedimentary units. The important point of this figure is that, with very few exceptions, all of the virtual geomagnetic poles plot as a diffuse cluster. Such a cluster is usually interpreted as indicating that the igneous bodies cooled below the Curie point at approximately the same time, suggesting that the Newark Basin igneous activity was a short-lived event. This is consistent with the interpretation of Olsen and Fedosh (1988) who believe that the three Watchung Basalts were extruded over a period of less than 600,000 years.

Figure 2B shows a more restricted data set that has not been corrected for the dip of the surrounding sedimentary units. The large amount of scatter for these data, as compared to Figure 2A, is obvious. Figure 2C shows these same data after the application of structural corrections. Clearly, the amount of scatter has been lessened significantly. The conventional interpretation of these observations is that tilting of the sediments occurred after the igneous bodies had cooled below the Curie point. This implies that the bulk of deformation within the Newark Basin occurred late in its history, in accord with the conclusions of Schlische and Olsen (1988) and Faill (1988).

With the exception of the igneous rocks in the Jacksonwald syncline, where the structural correction was applied by

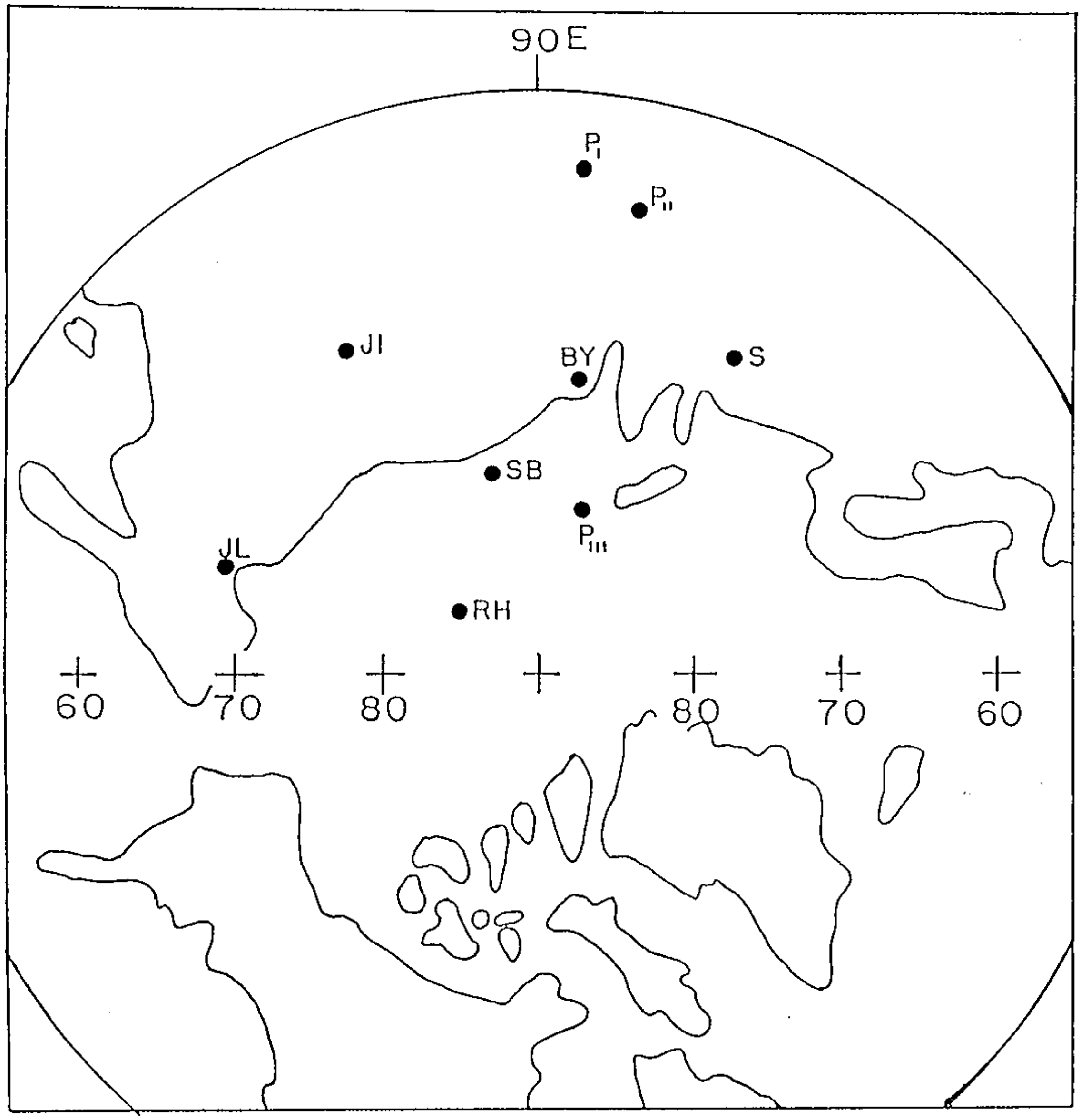


Figure 2B. Virtual geomagnetic poles determined by Hozik and Columbo (1984), Hozik and Opitz (1988), or Stuck and others (1988) without any structural corrections. See Figure 2A for key to letters.

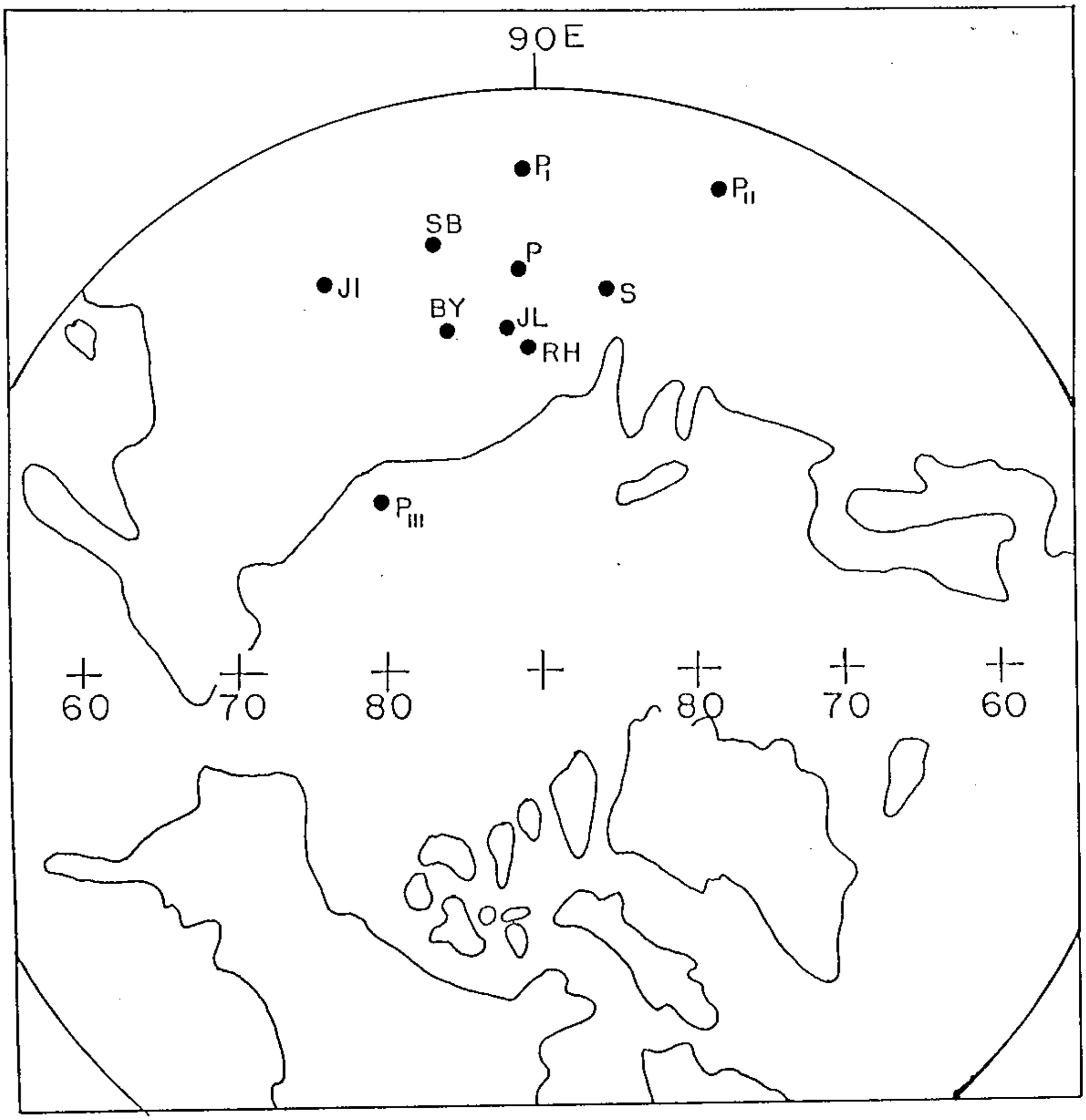


Figure 2C. Structurally corrected virtual geomagnetic poles for data plotted in Figure 2B. See Figure 2A for key to letters.

unfolding about the fold axis, all structural corrections for Newark Basin igneous bodies were determined simply by rotating the sedimentary beds to horizontal around an axis parallel to their strike. That the corrected data agree as well as they do is a strong argument against large amounts of rotation of fault blocks around subvertical axes, as would be required by models in which the curved Flemington and Hopewell Fault systems were active as strike-slip faults.

In summary then, paleomagnetic data indicate that:

1) Igneous activity in the Newark Basin was confined to a relatively short time period.
2) Deformation of rocks within the Newark Basin occurred late in basin development, after the conclusion of the bulk of igneous activity.
3) Most fault motion was associated with dip-slip rather than strike-slip faulting.

ROAD LOG

Stops 1 and 2 are in the Lambertville 15' quadrangle, stop 3 is in the Stockton 15' quadrangle, and stop 4 is in the Lumberville 15' quadrangle. The field trip starts at the junction of I-95 and NJ-29 (Exit 1 of I-95; Fig. 3). The following directions are to get you there from the parking lot in front of the Rider College Student Center.

1) Leave the parking lot and exit the Rider College campus via the south gate.

2) Turn left onto US 206 North. Mileage begins at gate.

   0.0 Mileage at South Gate of Rider College.

   0.2 North Gate (Main Entrance) to Rider College.

   0.7 Enter I-295 North following signs to I-95 South.

   2.9 End I-295 North and begin I-95 South.

   8.1 Junction NJ-29 and I-95. Exit here and follow signs for NJ-29 North. Prepare to rezero mileage.

Mileage Total	Interval	
00.0	00.0	Junction of I-95 and NJ-29. Mileage begins at expansion joint in bridge over Delaware-Raritan Canal. Follow signs for NJ-29 North toward Lambertville.
2.6	2.6	Traffic light at junction with Mercer County Route 546; Washington Crossing State Park.
3.8	1.2	Traffic light at Titusville.
4.7	0.9	Pass milepost 14. Hill to right is the Baldpate Mountain diabase, an intrusion with abundant granophyric material.
5.6	0.9	Entrance to Moore's Station Quarry.

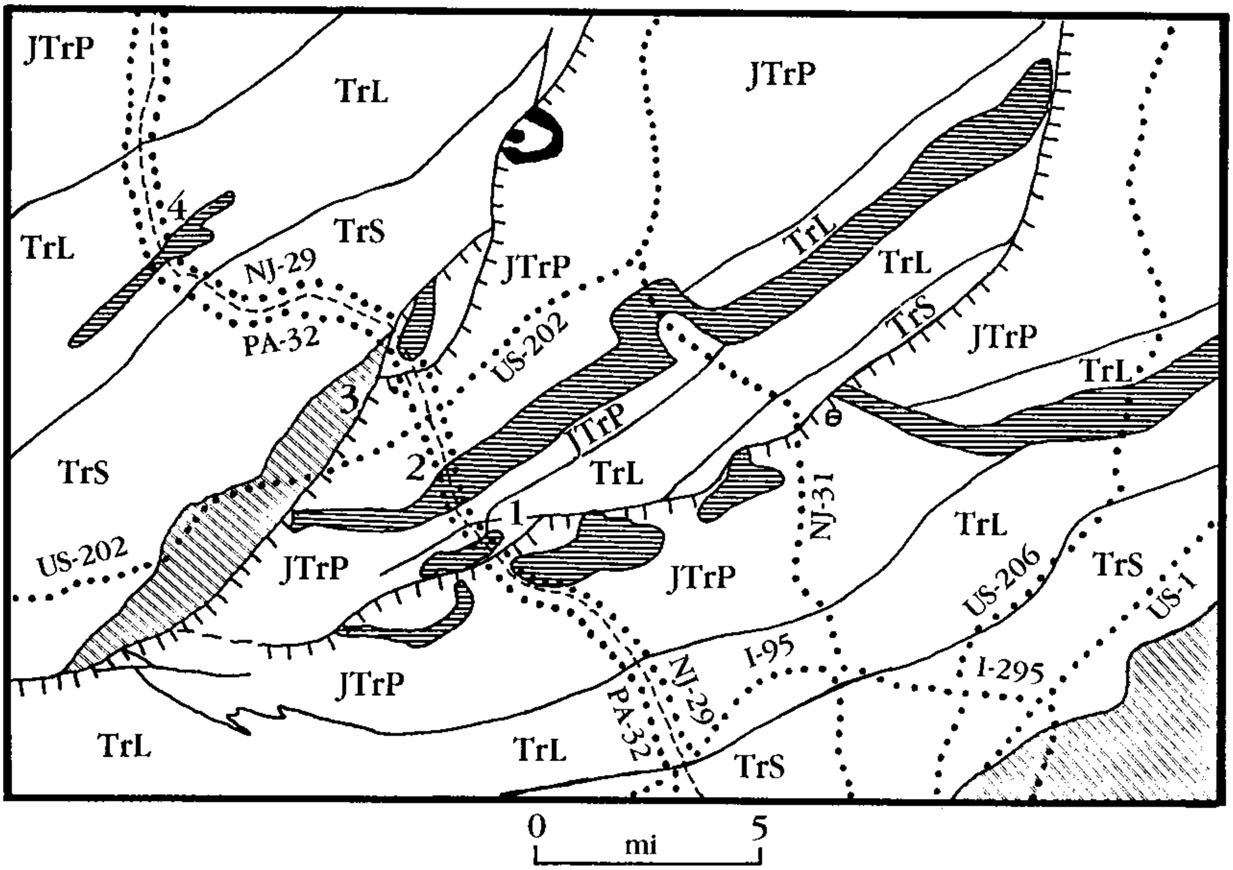


Figure 3. Geologic sketch map (after Lyttle and Epstein, 1987) of field trip region. Stop locations (numbered) and primary roads (dots) are shown. Horizontally ruled areas represent intrusive diabase sheets, diagonally ruled areas represent pre-Triassic or post-Jurassic rock units, and black-filled areas represent Lower Jurassic basalt flows. Hopewell and Furlong-Flemington fault systems are shown with hachures on the hanging wall or down side. Abbreviations used are: TrS, Stockton Fm.; TrL, Lockatong Fm.; JTrP, Passaic Fm.

5.8	0.2	Best view of quarry in Baldpate Mountain diabase.
6.6	0.8	Primrose Restaurant and entrance to Department of Public Safety, Mercer County Correctional Facility Quarry. HARD HATS, SIGNED RELEASES, AND PROOF OF INSURANCE ARE REQUIRED IN ORDER TO GAIN PERMISSION TO ENTER THIS QUARRY. For more information, contact Mr. Walter Krwatschenko, Mercer County Department of Insurance and Property (609-989-6655).

STOP 1: MERCER COUNTY CORRECTIONAL FACILITY QUARRY (WORKHOUSE QUARRY)

This is the Belle Mountain diabase. This intrusion appears to be a large sliver of diabase caught in a major fault (Hopewell Fault) zone. The rock exposed in the quarry is highly fractured, and many of the faults are thoroughly serpentinized with polished and slickensided surfaces. A prominent fault orientation on the eastern wall is N20E, dipping 65NW with slickenlines trending N10W and plunging 40 degrees. These fault surfaces frequently show a second set of slickenlines which is nearly horizontal and trends N20E.

Along the north wall the rock generally is more felsic in composition. Near the northwest corner of the quarry, the rock contains clumps of amphibole crystals with a faint radiating texture. Potassium feldspar is locally abundant and in places gives the rock, now approaching a granophyric composition, a distinctive pinkish hue. Faults in the north wall of the quarry strike N35W and dip 70NE. Slickenlines trend due east and plunge 65 degrees. A faint second set of slickenlines is nearly horizontal and trends N35W.

TABLE 2

WORKHOUSE QUARRY WHOLE-ROCK COMPOSITIONS

	1	2	3
SiO_2	52.99	53.44	52.90
TiO_2	1.37	1.32	1.30
Al_2O_3	16.40	16.85	16.32
Fe_2O_3	2.13	1.82	1.98
FeO	9.08	7.73	8.43
MnO	0.17	0.16	0.19
MgO	4.07	5.19	5.63
CaO	8.37	9.92	7.33
Na_2O	3.57	2.86	4.46
K_2O	1.74	0.60	1.33
P_2O_5	0.11	0.11	0.13
Ba	281	174	303
Co	43	NA	NA
Cr	27	62	94
Cu	168	117	119
Ni	45	64	66
Rb	53	18	36
Sc	31	35	34
Sr	225	221	204
V	271	257	266
Zr	109	115	108

Analyses 1-3 are samples WH1, WH2, and WH3, respectively. Major elements normalized to 100 percent anhydrous. Fe2O3/FeO set at 0.235. NA-not analysed.

Three samples from this quarry have been collected and analysed for major- and trace-element compositions (Table 2, analyses 1-3). One, WH3, has been analysed for initial Sr isotopic ratios. Whole-rock compositions show these samples to be mildly fractionated; estimates are that these rocks formed after approximately 25% crystallization. In general, they show a progressive change from more felsic (WH1) to more mafic compositions (WH3) as one traverses the quarry, north to south. Also noted are the unusually high alkali concentrations and low Ca contents for WH3 (and to a lesser extent for WH1) when compared to WH2. WH3 also has an initial 87Sr/86Sr value of $\simeq$0.710. These data are consistent with there being significant contamination of WH3 (and perhaps WH1) by local country rock. Although no obvious xenoliths now are observed in the quarry, Bascom and others (1909) reported the presence of one. Further, the highly fractured nature of the rocks makes interaction with enclosing sediments likely. Consistent with this are the paleomagnetic data of Opdyke (1961). His determined virtual geomagnetic pole for this quarry is very different from results for most Mesozoic ENA diabases (Hozik and Colombo, 1984), probably reflective of extensive diabase/wall rock interaction.

		Return to NJ-29 North where mileage resumes.
7.2	0.6	Junction with Valley Road.
8.3	1.1	Passing through the center of the Lambertville (Sourland Mountain) sill. Good exposures of this large diabase are seen on the right.

9.5	1.2	Junction with South Main Street, Lambertville. Bear left onto South Main Street, follow the signs for NJ-29 North (very easy to miss!).
9.7	0.2	Traffic light. Turn left onto Bridge Street.
9.9	0.2	Cross Delaware River.
10.1	0.2	Enter New Hope, Pennsylvania.
10.2	0.1	Traffic light at junction of Bridge Street and South Main Street, New Hope. Note mileage. Because the next outcrop has become heavily overgrown (including poison ivy!) THE DIRECTIONS AND MILEAGE THAT FOLLOW ARE FOR AN OPTIONAL STOP.
	0.0	Turn south onto South Main Street (PA-32).
	0.1	Pass Buck's County Playhouse on left.
	0.5	Junction with PA-232; continue south on PA-32.
	0.6	Odette's Restaurant on left. Park in lot. Outcrop is on the opposite side of the road.

STOP 2 (OPTIONAL): UPPER CONTACT ZONE OF THE LAMBERTVILLE SILL

The fairly continuous series of outcrops seen here exposes the upper contact zone of the Lambertville sill and overlying thermally metamorphosed hornfels. The sill is approximately 1300 feet thick and can be seen along both sides of the Delaware River. Immediately to the west of this location the sill narrows considerably and cuts up-section before terminating near or against the Furlong Fault. Also seen here is a thin (~5 ft thick) sill separated from the main body of the Lambertville sill by a seemingly in-place layer of hornfels (alternatively, this could be a xenolith of the roof rock). A recent discovery in this small sill is the presence of relatively large (~3 mm) phenocrysts of orthopyroxene (?) in a fine-grained groundmass. The phenocysts appear to be concentrated towards the center of the sill, perhaps due to flow differentiation mechanisms.

The diabase immediately at the upper contact is very fine grained. Grain size increases rapidly towards the interior of the sill (to the south) so that at the southern end of the exposure the diabase (now located approximately 150 ft below the upper contact) is fairly course grained. Major- and trace-element concentrations of the upper chill margin (Table 3, analysis 1) and the adjacent small sill are typical for the ENA HTQ magma type. Whole-rock samples from the southern end of the exposure (Table 3, analysis 2) are decidedly more fractionated and probably represent residual liquids complementary to the mafic-rich rocks (Table 3, analysis 3) found near the bottom of the sill (not exposed at this location). Highly fractionated granophyric compositions, however, are unknown from anywhere in the Lambertville sill.

Analyses from both vertical sections through the sill, one from along the Delaware River (this stop locality) and the other from near the sill's northeast terminus at Sourland Mountain, show two distinct mafic zones overlain by more felsic, residual zones. Geochemical variation trends for the two cross sections are consistent and indicative of at least two separate (though probably closely spaced in time) magma pulses (Fig. 4). The Sourland Mountain section, however, contains whole-rock compositions that suggest the magma of the second pulse had LTQ, rather than HTQ, affinities. Finally, although no xenoliths are seen in the immediate vicinity, a contaminated sample (Table 3,

TABLE 3

LAMBERTVILLE SILL WHOLE-ROCK COMPOSITIONS

	1	2	3	4
SiO_2	52.60	53.39	52.19	52.52
TiO_2	1.16	1.41	0.90	1.14
Al_2O_3	14.11	15.27	11.50	15.20
Fe_2O_3	1.97	2.05	1.93	1.92
FeO	8.36	8.71	8.23	8.14
MnO	0.17	0.17	0.17	0.17
MgO	7.81	5.14	11.97	6.68
CaO	10.65	9.90	10.94	9.03
Na_2O	2.13	2.86	1.67	2.69
K_2O	0.86	0.89	0.47	2.33
P_2O_5	0.18	0.21	0.03	0.18
Ba	148	178	125	272
Co	56	NA	52	NA
Cr	278	49	470	173
Cu	115	131	79	111
Ni	96	56	110	72
Rb	33	34	13	93
Sc	37	33	40	36
Sr	235	227	121	262
V	257	290	241	281
Zr	107	113	71	95

Analyses 1-4 are samples ELS5, ELS23, LS2, and ELS6, respectively. Major elements normalized to 100 percent anhydrous. Fe2O3/FeO set at 0.235. NA-not analysed.

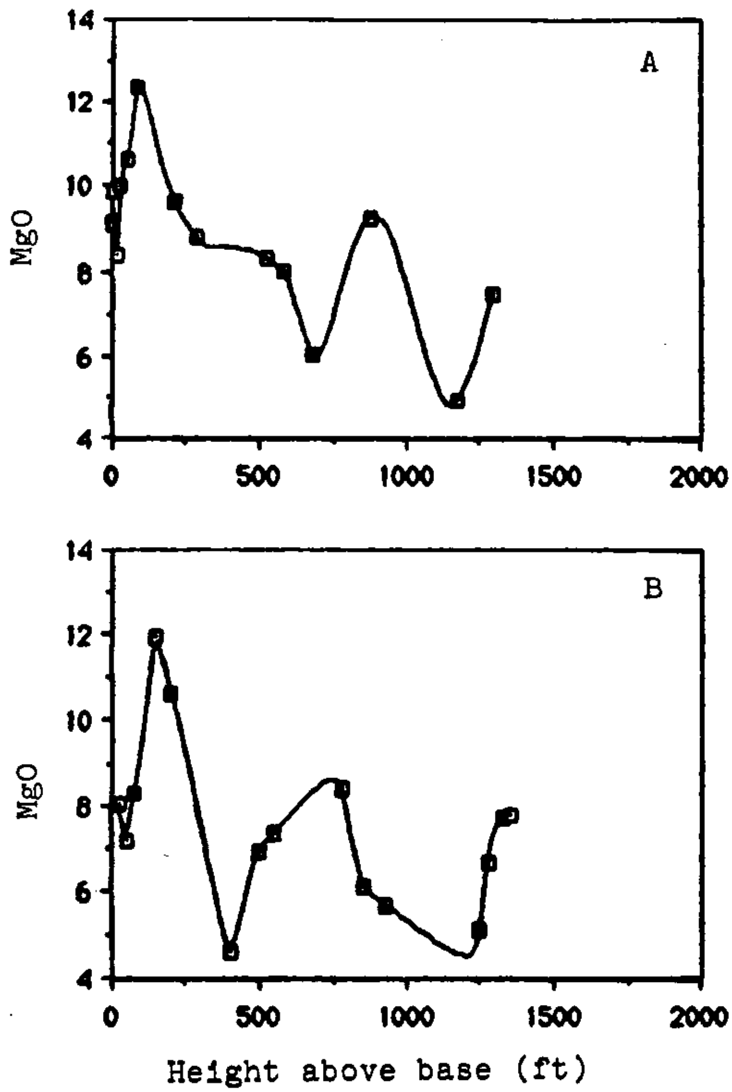


Figure 4. MgO (weight percent) versus height (feet) above the base of the Lambertville sill. A) Sourland Mountain (Northeast) section near Zion, NJ; B) Delaware River (Southwest) section. Upper contact region samples from the Delaware River section were collected at Stop 2. Note the reversals in MgO depletion for both sections, although at slightly different stratigraphic levels. Similar patterns are found for other compatible elements such as Ni and Cr. Mirror-image patterns are found for many of the incompatible elements, particularly alkalis.

analysis 4) was collected about 75 feet below the upper contact at this stop locality.

		Return north to traffic light at intersection of Bridge Street and Main Street in New Hope.
10.2	0.1	Intersection of Bridge Street and South Main. Street. Proceed on North Main Street (PA-32).
11.0	0.8	Narrow bridge on blind curve. Be very careful!
11.4	0.4	US-202 overpass.
11.6	0.2	Power line.
11.8	0.2	Proceed to second left after power line. This is the junction with Phillips Mill Road. Turn left onto Phillips Mill Road. Road goes through a series of right angle bends for approximately the next half mile.
12.6	0.8	Entrance to New Hope Crushed Stone Quarry. Turn left into quarry, and proceed to scales area and field office immediately adjacent. After checking in, proceed to the south pit. HARD HATS, SIGNED RELEASES, AND PROOF OF INSURANCE ARE REQUIRED IN ORDER TO GAIN PERMISSION TO ENTER THIS QUARRY. For further information contact Mr. Edwin Spencer, New Hope Crushed Stone Co. (215-862-5288).

STOP 3: QUARRY DIKE AND PALEOZOIC CARBONATES

The Quarry dike is exposed spectacularly in both the north and south pits of the quarry. However, because the dike was sampled by Husch and Schwimmer (1985) in the south pit only and the exposure there is most illustrative of the structure of the dike, we will confine our visit to that location.

The Quarry dike was first discussed by Lessentine (1952) along with a series of other small basaltic dikes which strike north-northeast across a lower Paleozoic section of mostly carbonate rock. The Quarry dike intrudes Conococheague (Limeport) dolomitic limestone of Cambrian age. This unit is part of the local pre-Triassic basement complex which Mesozoic sedimentary rocks unconformably overlie. These Cambro-Ordovician units are exposed in this area as the result of erosion on a crustal block uplifted by as much as 10,000 feet along the normal Furlong Fault (part of the Chalfont-Flemington Fault system; McLaughlin, 1959). Although the Quarry dike has not been age-dated by radiometric means, it is assumed from its structural setting and petrographic and geochemical similarities (Table 4) to be close in age to numerous other early Jurassic ENA HTQ basalts such as the Palisades sill, the Orange Mountain (First Watchung) basalt, and Talcott basalt.

In the south pit the Quarry dike intrudes the gently folded and well-layered Conococheague limestone in an en echelon fashion, both along its strike (N15E) and dip (nearly vertical). Along the south wall the first section of the dike extends down approximately 30 feet from the surface. At this point the first section terminates and the dike shifts abruptly about 25 feet to the east where it continues down the south wall and across the floor of the pit. Prior to reaching the north wall, the second section ends (or at least narrows to a very thin dike) and a third section is seen in the upper portions of the quarry some 200 feet to the east.

In outcrop, the Quarry dike is a dark green to black rock which erodes to sharp, often fist sized, angular fragments. Fractures in the dike are often filled with carbonate material,

Table 4

ENA-HTQ BASALT AND QUARRY DIKE WHOLE-ROCK COMPOSITIONS

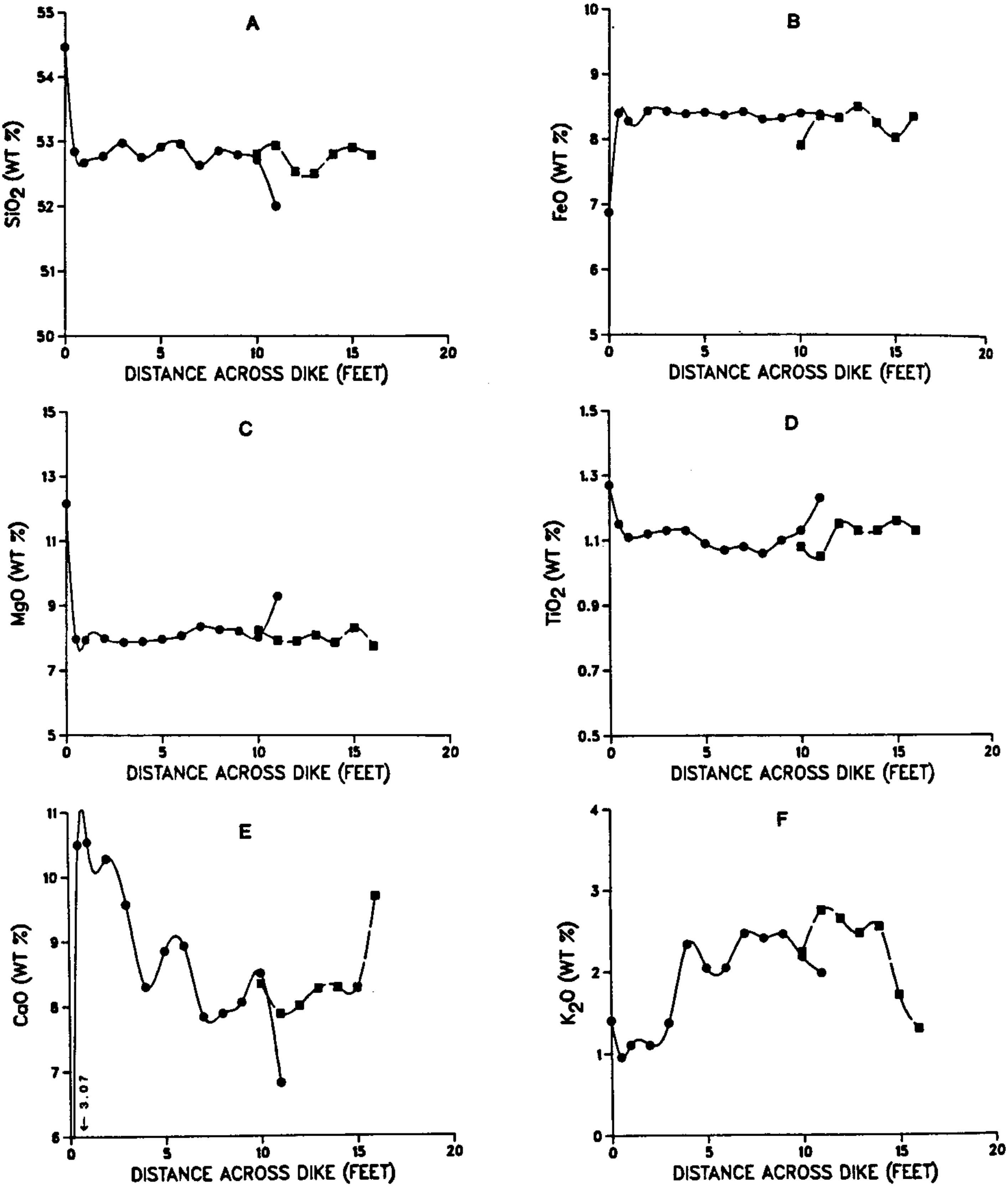
	1	2	3	4	5	6
SiO_2	52.55	51.61	51.86	52.84	52.91	52.85
TiO_2	1.22	1.09	1.07	1.15	1.09	1.06
Al_2O_3	14.64	14.34	14.27	14.01	14.34	14.24
Fe_2O_3	---	---	---	1.97	1.97	1.94
FeO	---	---	---	8.40	8.40	8.31
FeO*	10.40	10.20	10.86	---	---	---
MnO	0.16	0.15	0.16	0.17	0.17	0.17
MgO	7.67	8.32	7.98	7.97	7.97	8.27
CaO	10.44	11.41	11.24	10.50	8.85	7.89
Na_2O	2.06	2.15	2.06	1.90	2.10	2.73
K_2O	0.84	0.59	0.50	0.96	2.05	2.41
P_2O_5	0.14	0.13	0.12	0.13	0.14	0.13
Ba	195	182	174	145	187	227
Co	53	45	47	NA	NA	NA
Cr	315	260	322	282	304	374
Cu	110	127	123	90	74	82
Ni	95	61	72	86	89	105
Rb	NA	37	22	28	55	76
Sc	37	NA	NA	35	36	40
Sr	175	183	186	190	200	229
V	235	272	270	241	257	281
Zr	120	116	87	101	98	106

Major elements normalized to 100 percent anhydrous.
Fe2O3/FeO ratio set at .235 for analyses 4, 5, and 6.
1) Palisades sill chill (Walker, 1969).
2) Average Orange Mountain Basalt (Puffer and Lechler, 1980).
3) Average Talcott Basalt (Puffer and others, 1981).
4) Quarry dike chill (QD2).
5) QD7 (intermediate contamination).
6) QD10 (extensive contamination).
NA-not analysed.

probably dolomitic. Serpentine is noted along many of the fracture surfaces. Pyrite also is common, particularly near the contacts with surrounding dolomite. In thin section, the dike is seen to be somewhat altered, with the most pervasive alteration occurring near the contacts and along carbonate filled fractures. Glomeroporphyritic masses of clinopyroxene (predominantly augite) are common throughout the dike with no significant changes in modal distribution across the dike.

A series of 20 whole-rock samples, collected in the south pit across the second dike section, have been analysed for major- and trace-element concentrations (Fig. 5). In addition, 3 of these samples have had their strontium isotopic ratios determined (Fig. 6). These data, along with derived variation trends (Fig. 7), support the conclusions of Husch and Schwimmer (1985) that the Quarry dike formed from the injection of typical HTQ magma as it was being subjected to progressive contamination from a Mesozoic sedimentary source, presumably Lockatong argillite. (Figs. 6 and 7). The least contaminated parts of the dike are found near the chill margins (Table 4, analysis 4) and the most contaminated are found in the central zone (Table 4, analysis 6); intermediate levels of contamination are found in the intervening areas (Table 4, analysis 5). Presumably, additional samples from other central Newark Basin diabase intrusions exhibiting similar geochemical characterisitics were affected by similar processes.

Return to quarry entrance. Turn right onto Phillips Mill Road.



A
SiO_2 (WT %)
DISTANCE ACROSS DIKE (FEET)
B
FeO (WT %)
DISTANCE ACROSS DIKE (FEET)
C
MgO (WT %)
DISTANCE ACROSS DIKE (FEET)
D
TiO_2 (WT %)
DISTANCE ACROSS DIKE (FEET)
E
CaO (WT %)
← 3.07
DISTANCE ACROSS DIKE (FEET)
F
K_2O (WT %)
DISTANCE ACROSS DIKE (FEET)

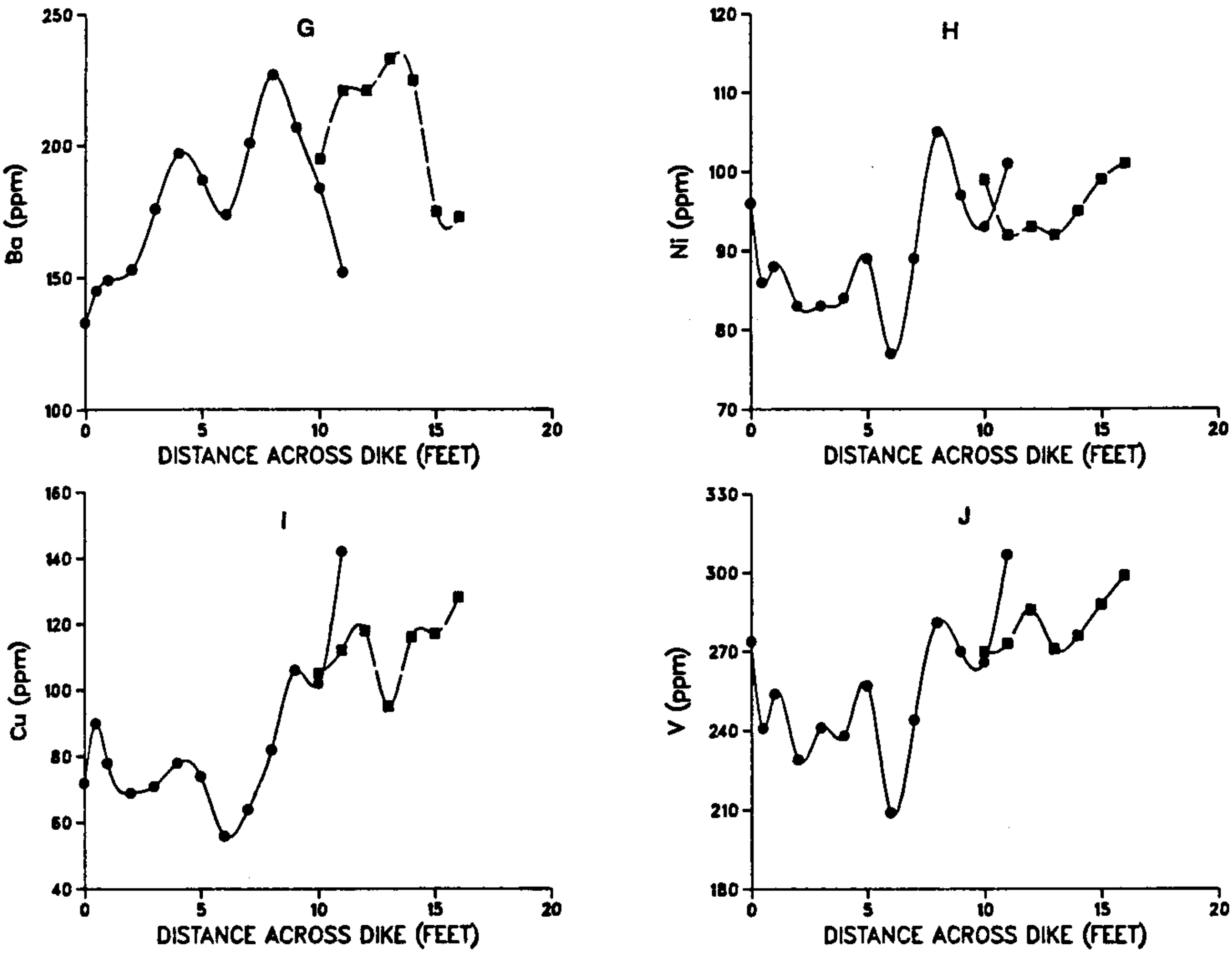


Figure 5. Compositional variations of selected major- and trace-elements across the Quarry dike. Western contact of the dike is to the left for each diagram. Traverse shown by filled circles is for samples 1-13; traverse shown by filled squares is for samples 14-20 (right to left).

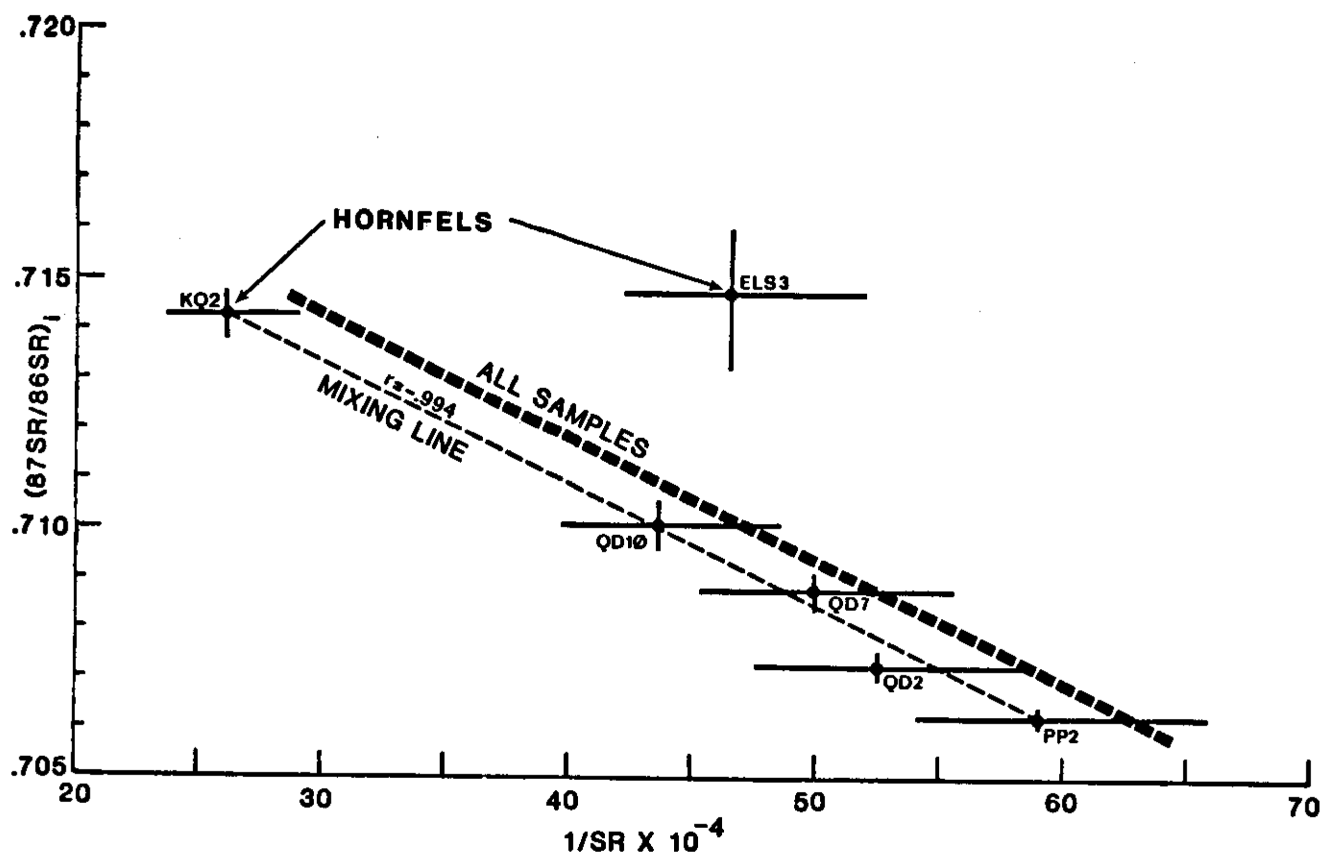


Figure 6. Calculated initial 87Sr/86Sr versus 1/Sr variation diagram. Calculated strontium isotopic ratios are determined for 200 m.y. (the best estimate for the actual age of the diabase intrusions). Sample PP2 represents typical HTQ type basalt. Samples QD2, QD7, and QD10 represent Quarry dike chill, intermediately contaminated, and highly contaminated samples, respectively. The mixing line connecting KQ2 with PP2 is indistinguishable from the best least-squares fit (correlation coefficient shown) for samples PP2, QD2, QD7, QD10, and KQ2. Error bars are given for each sample.

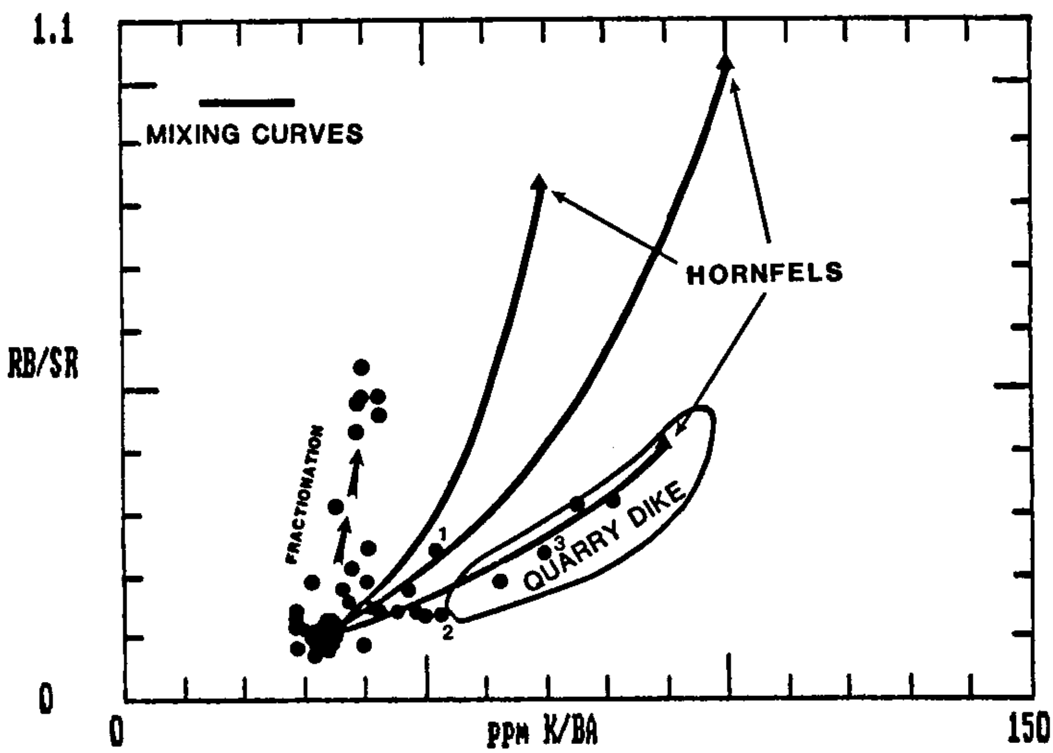


Figure 7. Rb/Sr versus K/Ba variation diagram. Three mixing curves extend from the average chill composition to each of three analysed hornfels. Mixing curves are generated using general equations of Langmuir and others (1978). Quarry dike field from Husch and Schwimmer (1985). Numbered samples showing effects of contamination and referred to in text are: 1) WH1; 2) sample D1 of Benimoff and Sclar (1984); 3) ELS6. Note that most contaminated sample compositions (excluding WH1) and the Quarry dike field plot along the mixing curve produced by having a single hornfels composition (sample KQ2 from the Lockatong Formation) as the contaminating source.

13.4	0.8	Junction of Phillips Mill Road and PA-32 (North Main Street). Turn right (south).
15.0	1.6	Traffic light at junction of Main Street and Bridge Street (PA-179) in New Hope. Turn left onto Bridge Street, cross the Delaware River, and proceed through Lambertville to second light.
15.5	0.5	Junction of Bridge Street and Main Street. Turn left (north) onto Main Street and follow NJ-29 North.
16.7	1.2	US-202 overpass.
17.4	0.7	Mt. Gilboa Quarry in Stockton diabase.
18.9	1.5	Stockton Schoolyard.
19.2	0.3	Junction with Hunterdon County 523. Bear left and follow NJ-29.
19.8	0.6	Junction of NJ-29 and Hunterdon County 519, just after crossing Wickecheoke Creek. Continue on NJ-29.
20.1	0.3	Quarries in Stockton Formation.
22.0	1.9	Enter small village of Raven Rock.
22.8	0.8	Raven Rock outcrop.
24.3	1.5	Outcrop of Byram diabase. Proceed to the north end of the outcrop and park off the road on the right shoulder.

STOP 4: BYRAM (POINT PLEASANT) DIABASE

The contact of the Byram diabase with surrounding Lockatong Formation is well exposed at the northern end of the outcrop along the east side of NJ-29. The diabase here is very fine grained with phenocrysts of clinopyroxene. Also found here are rare anhedral phenocrysts of olivine which are highly embayed and apparently were undergoing reaction with the surrounding liquid. The composition of the chill (Table 5, analysis 1) is nearly identical to the other fine-grained diabases seen already and is typical of the ENA HTQ basalt found in the central Newark Basin.

Although it appears that there is normal faulting along the contact and the adjacent Lockatong strata are deformed, there are no indications of cataclasis in the chilled diabase itself, even within an inch or two of the contact. However, in outcrop near the contact the diabase is fractured pervasively. Structural studies, both from near the road outcrop and from farther east along the adjacent stream, show the Byram diabase to be offset by at least two generattions of faults; one is a northeast striking normal fault (the one seen at the road) and the other is a north-south striking set of oblique to strike-slip faults.

Moving away from the contact zone, both south along the road and upwards along the face of the outcrop, the diabase quickly becomes noticeably coarser grained. Most of the exposed diabase is a medium-to coarse-grained diabase with a good subophitic texture. Grain size decreases again at the southern end of the outcrop as the basal contact is approached. Although this contact is not exposed, the outcrop pattern seen does suggest that the Byram diabase is discordant with the surrounding Lockatong Formation.

Rock compositions (Table 5, analysis 2) at road level remain fairly constant, regardless of grain size, with little deviation from chill margin values. An exception to this unchanging nature of the diabase is found in a horizon approximately 50 feet above road level which exhibits faint horizontal layering, particularly when viewed from across the road. This layering is marked by

TABLE 5

BYRAM (POINT PLEASANT) DIABASE WHOLE-ROCK COMPOSITIONS

	1	2	3
SiO_2	52.42	52.58	52.51
TiO_2	1.11	1.10	0.76
Al_2O_3	14.19	14.26	10.29
Fe_2O_3	1.93	1.95	1.87
FeO	8.22	8.30	7.95
MnO	0.16	0.16	0.18
MgO	8.14	8.03	15.39
CaO	11.00	10.95	9.03
Na_2O	2.08	1.98	1.56
K_2O	0.61	0.55	0.35
P_2O_5	0.14	0.14	0.11
Ba	147	140	90
Co	48	NA	59
Cr	281	278	482
Cu	105	100	101
Ni	89	89	122
Rb	16	18	10
Sc	36	36	37
Sr	169	161	109
V	264	258	229
Zr	94	96	65

Analyses 1-3 are samples PP2, PP6, and PP5, respectively. Major elements normalized to 100 percent anhydrous. Fe2O3/FeO ratio set at .235. NA-not analysed.

centimeter thick layers of alternately pyroxene-rich and pyroxene-poor diabase. Mafic-rich samples collected here contain very high MgO values (Table 5, analysis 3) and large (up to 4 mm) phenocrysts of euhedral orthopyroxene (En 80) and augite. Many of the orthopyroxenes are rimmed with augite, a texture resembling those reported by Philpotts and Reichenbach (1985) in the Talcott Basalt of the Hartford Basin. They believe these textures and associated pyroxene compositions are indicative of high pressure crystallization at upper mantle/lower crust depths and occurred early in the crystallization sequence of the basalts prior to their intrusion to present stratigraphic levels. Finally, it is worth noting the lack of any evolved, granophyric compositions within the Bryam diabase which would be complementary to the mafic compositions seen at this location. This argues either for the lack of _in situ_ differentiation or the removal of locally generated fractionated liquids to granophyre-rich intrusions, such as the Baldpate Mountain diabase, located higher up in the Triassic section.

END OF TRIP. Return to Rider College by taking NJ-29 south, I-95 north, I-295 south, and US-206 south.

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