Guide to the Mesozoic Redbeds of Central Connecticut

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STATE GEOLOGICAL AND NATURAL HISTORY SURVEY
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DEPARTMENT OF ENVIRONMENTAL PROTECTION

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GUIDEBOOK NO. 4

On the cover:

In the early morning along the shore of an East Berlin Lake, the 7-m phytosaur *Rutiodon* snatches a *Semionotus* from the shallows. The tall horsetail *Equisetum* and cycad *Otozamites* thrive in the wet mud of the lake strand. Stands of the conifer *Araucarioxylon* tower 60 m high along the distant horizon on sandy soils of the well drained uplands. The dinosaur *Eubrontes* passed this way the previous evening.

Sketch by Amy S. Hubert

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GUIDE TO THE MESOZOIC REDBEDS OF CENTRAL CONNECTICUT

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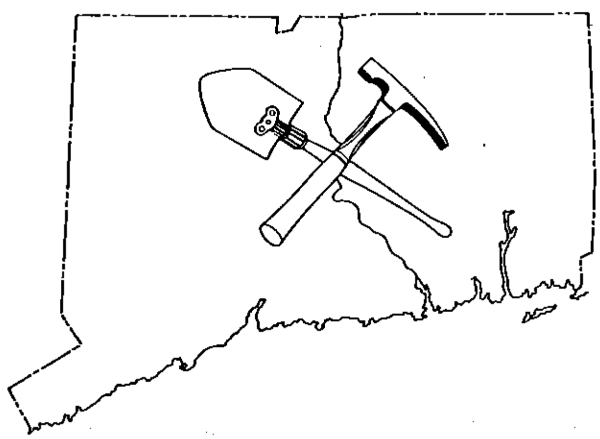

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GUIDEBOOK NO. 4

STATE GEOLOGICAL AND NATURAL HISTORY SURVEY
OF CONNECTICUT

THE NATURAL RESOURCES CENTER
DEPARTMENT OF ENVIRONMENTAL PROTECTION

This guidebook is a reprint of "Guide to the Redbeds of Central Connecticut: 1978 Field Trip, Eastern Section of the Society of Economic Mineralogists and Paleontologists." It was originally published as Contribution No. 32, Department of Geology and Geography, University of Massachusetts, Amherst, Massachusetts.

For information on ordering this guidebook and other publications of the Connecticut Geological and Natural History Survey, consult the List of Publications available from the Survey, Dept. of Environmental Protection, State Office Building, Hartford, Connecticut 06115.

DEDICATION

This guidebook is dedicated to the memory of Paul D. Krynine (1902-1964), superb teacher and researcher at The Pennsylvania State University and the foremost sedimentary petrographer of his day. His pioneering 1950 monograph "Petrology, stratigraphy, and origin of the Triassic sedimentary rocks of Connecticut" provides inspiration and a solid foundation for the subsequent advances in our understanding of the sedimentology of the redbed sequence of the Connecticut Valley.

Special Note

The fact that a locality is described in this guidebook does not imply that the public has access to the locality. Stopping on a limited access highway is forbidden by a regulation of the State Traffic Commission, which prohibits all vehicles from stopping or parking on any part of the highway. These regulations also prohibit pedestrians on any limited access highway. Field trip features on these highways can be viewed from other ground. In other instances, stops on private property require permission of the owner. Anyone planning to go on this field trip should check carefully the suggested stops.

Hugo F. Thomas
State Geologist

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ACKNOWLEDGMENTS

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OBJECTIVES OF THE TRIP

During this trip to central Connecticut, we will try to present an overview of the history of the sedimentary and volcanic fill of the rift valley in Late Triassic and Early Jurassic time. There is at least one stop at each of the stratigraphic units (Fig. 1).

Portland Arkose (stop 1)
Hampden Basalt (stop 7)
East Berlin Formation (stop 7)
Holyoke Basalt (stops 4, 6)
Shuttle Meadow Formation (stop 6)
Talcott Basalt (stop 3)
New Haven Arkose (stops 2, 5)

The focus of the tirp is interpretation of depositional environments using primary sedimentary structures and stratigraphic sequences. The paleoenvironments emphasized are: alluvial-fan conglomerate (stop 1); paleosol caliche profiles (stop 2); braided-river sandstone and flood-plain mudstone (stops 2, 5); symmetrical cycles of gray mudstone-black shale-gray mudstone that accumulated in carbonate-producing alkaline lakes (stop 7); and redbeds of lacustrine origin (stop 6). The field observations are combined with laboratory data and then applied to several of the classic problems of the Connecticut Valley: the broad terrane hypothesis that the Hartford and Newark Basins are remnants of one rift valley, the paleoclimate, and the origin of the color of the redbeds.

REGIONAL SETTING

The fault-bounded basins of the Newark Supergroup (Olsen and Galton, 1977, p. 983; Olsen, 1977) comprise a linear zone about 450 km wide and 2000 km long from Florida to the Grand Banks off Newfoundland (Fig. 2). The basins are unified by their fill of redbeds and basalts of Late Triassic to Early Jurassic age. Especially impressive are the areal extent and number of the buried basins, many only recently discovered (Ballard

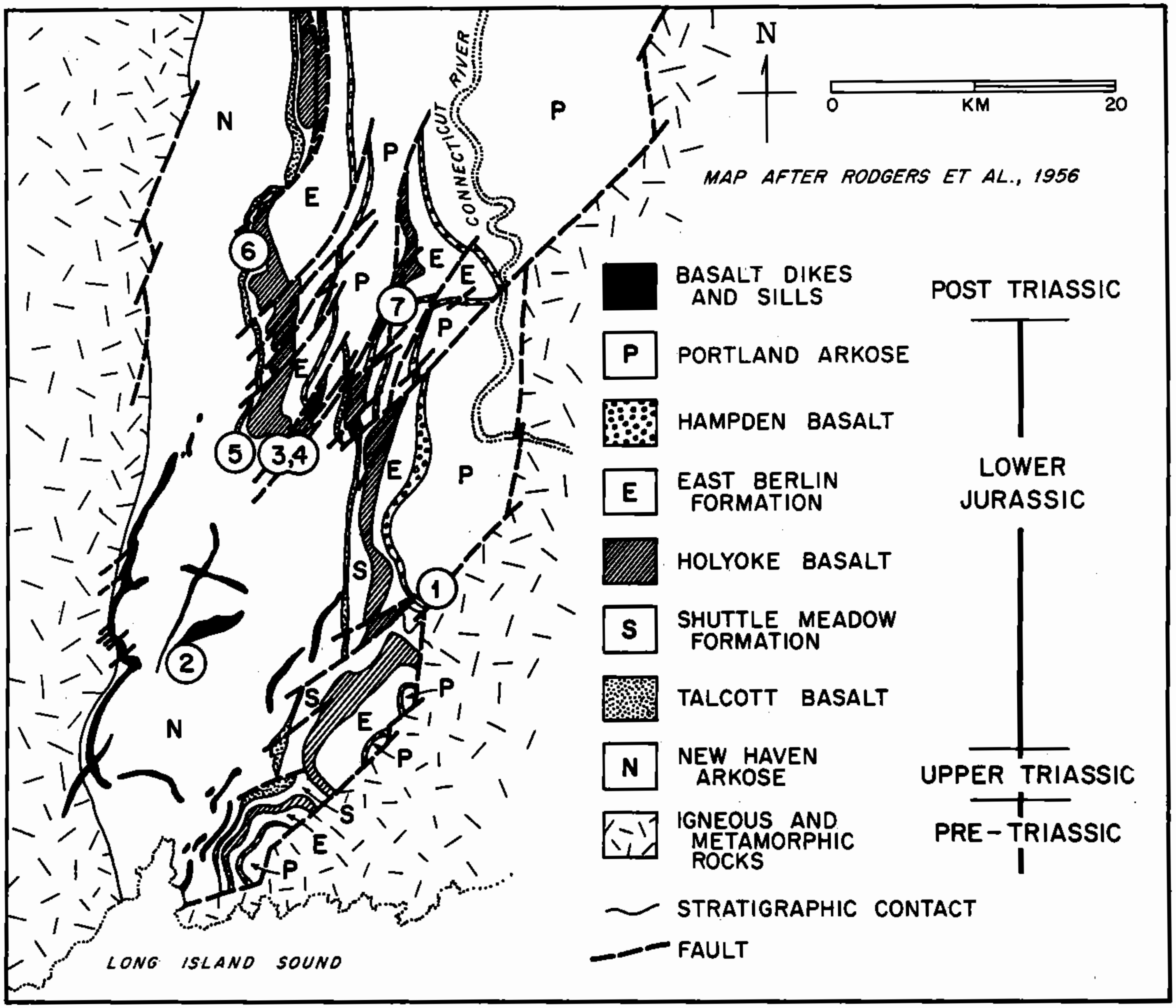

Fig. 1. Geological map of central Connecticut showing the location of the seven steps of the field trip.

and Uchupi, 1972, 1975; Marine and Siple, 1974; Uchupi et al., 1977). The Hartford Basin is on the western margin of a grand display of basins of Late Triassic to Early Jurassic age that extended from North America to Africa on continental crust before opening of the North Atlantic. The rift basins formed in response to tensional forces associated with initial rifting of North America from Europe and Africa (May, 1971, p. 1289).

The strata in the exposed basins in eastern North America have long been thought to be of Late Triassic age based on radiometric ages (Armstrong and Besancon, 1970). Some of the basins, however, have been discovered to contain Early Jurassic strata as evidenced by spores and pollen in lacustrine gray mudstone (Cornet et al., 1973; Cornet and Traverse, 1975; Cornet, 1977). The basins with only Upper Triassic rocks are the Durham-Wadesboro (1 on Fig. 2), Davie County (2), Farmville (3), four small areas south of the Farmville Basin (4), Dan River and Danville (5), Richmond (6), and Taylorsville (8). The basins where sedimentation extended from Late Triassic into Early Jurassic time are the Culpepper (7), Scottsville (9), Gettysburg (10), Newark (11), Pomperaug (12), Hartford with Cherry Valley outlier (Platt, 1957) (13), Deerfield (14), Fundy (15), and Chedabucto (16). The pattern is that the southern basins are the oldest, with only Upper Triassic rocks (Middle and Upper Carnian). Deformation proceeded northward until the strata in the Hartford Basin span Late Triassic (Norian) through Early Jurassic and perhaps into Middle Jurassic (Bajocian) time.

The Hartford Basin of Connecticut and southern Massachusetts is a half graben 140 km long, filled during the Late Triassic and Early Jurassic by about 4 km of sedimentary rocks and basaltic lavas and intrusives (Figs. 1, 3). The strata dip eastward toward west-dipping normal border

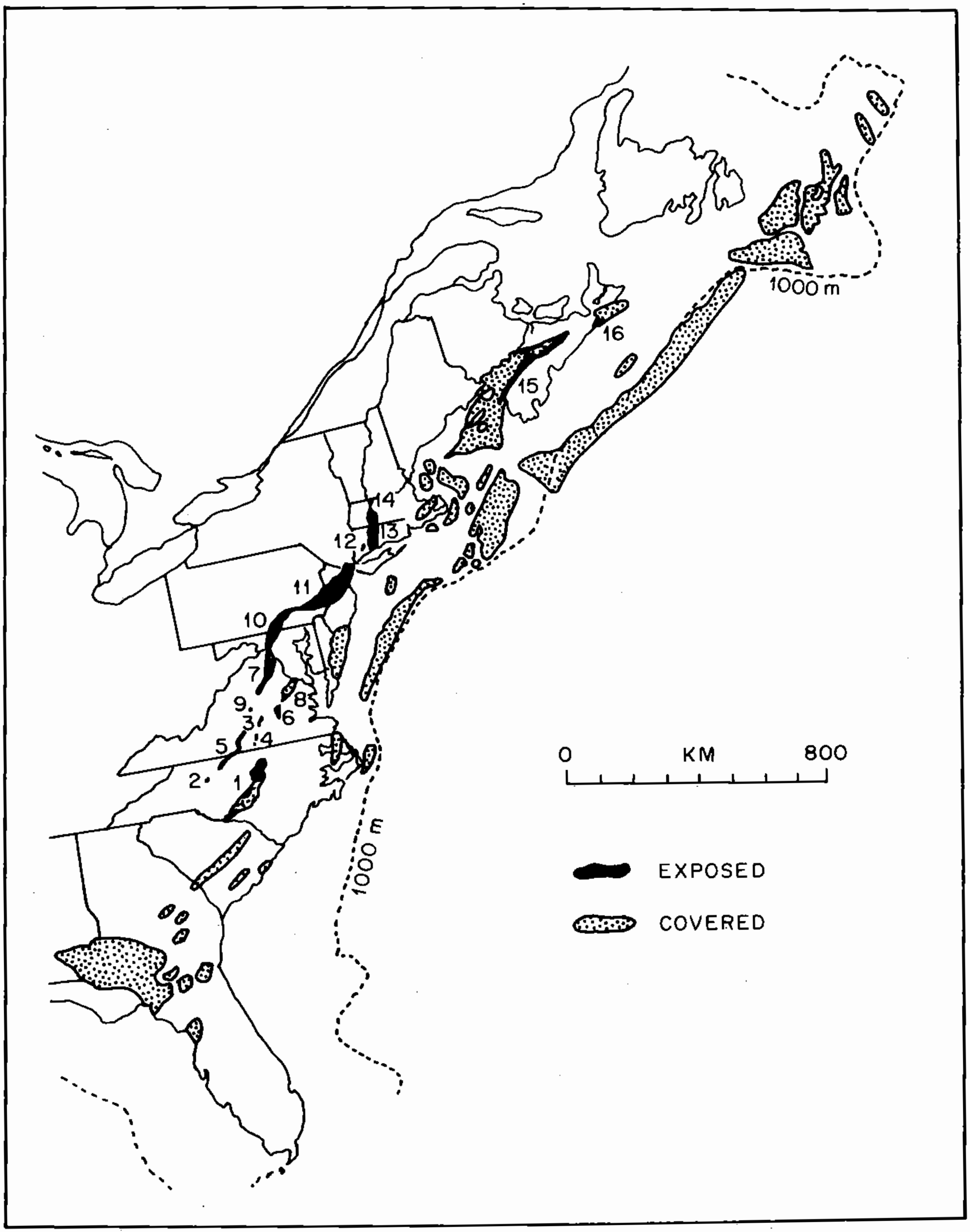

Fig. 2. Basins of the Newark Supergroup along eastern North America (Van Houten, 1977, p. 80). The exposed basins are numbered as explained in the text.

faults. The detritus was eroded from highlands of Lower Paleozoic metamorphic rocks east of the rift valley.

Several excellent reviews of the Hartford Basin are in the literature. Krynine (1950, p. 11-15) summarized the contributions made by Silliman, Percival, Hitchcock, Dana, Russell, Davis, and Hobbs in the 19th century and Barrell, Lull, Longwell, and Thorpe in the first half of the 20th century. Lull (1953) described the plant and animal life in a monograph. In two guidebooks to the Hartford Basin in Connecticut, Rodgers (1968, p. 1-2) and Sanders (1970, p. 1-3) integrated the stratigraphy, sedimentology, and structure of the Mesozoic rocks. Sanders (1974, p. 1-3, 15-22) in a guidebook to the northern end of the Newark Basin in Rockland County, New York, compared the stratigraphic sequences and structure of the Hartford and Newark Basins. Sanders concluded that they are the margins of one rift valley with the central portion removed by erosion (the broad terrane hypothesis of Russell, 1878, p. 230; 1880, p. 704). Van Houten, 1977, p. 89-93) placed the Newark Basins of eastern North America in a depositional framework with correlative rift valleys in Morocco and summarized the history of the opening of the North Atlantic.

ABSTRACT OF THE PALEOGEOGRAPHIC HISTORY

The paleogeographic history of the part of the Hartford Basin visited by us is briefly as follows. Supporting data and literature citations are presented with the descriptions of the field trip stops. The thickness of the stratigraphic units are generalized.

In pre-continental drift position, the rift valley was located in the tropics at about 15 degrees north paleolatitude. The valley was floored by multiply-deformed, high-grade metamorphic rocks of Lower Paleozoic age. The initial sedimentary fill was the 2000-m New Haven Arkose

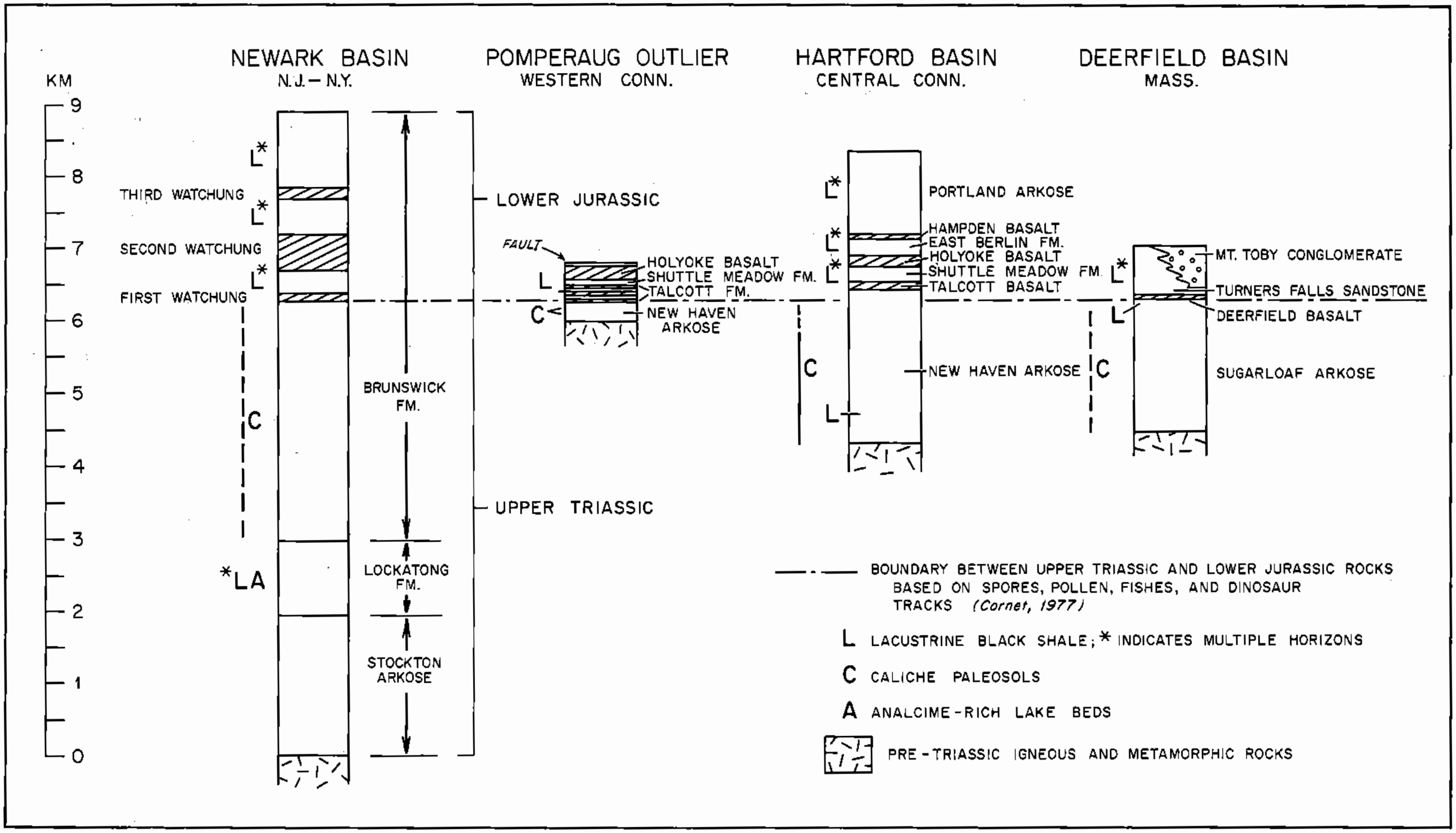

Fig. 3. Stratigraphy of Upper Triassic and Lower Jurassic rocks in the Newark, Hartford, and Deerfield Basins and in the Pomperaug Outlier. The thickness for each formation is the maximum value so that at any one locality the total section is not as thick as shown on the diagram.

of Late Triassic to probably Early Jurassic age. Rivers flowed from the eastern highlands, depositing conglomerate and sandstone in alluvial fans along the base of a fault-bounded escarpment. The rivers coursed southwest, constructing an alluvial-plain sequence of braided-river sandstone and pebbly sandstone and floodplain red mudstone. Caliche paleosol profiles are abundant, reflecting a paleoclimate dominated by tropical semiaridity with seasonal precipitation of about 100 to 500 mm.

In Early Jurassic time, tholeiitic basaltic magma rose along deep crustal fractures to form the fissure flows and interbedded volcanic agglomerate of the 65-m Talcott Basalt. The flows substantially lowered the gradient of the valley floor. The overlying 100-m Shuttle Meadow Formation is dominated by playa and perennial lakes that existed during intervals of relatively increased precipitation. The lacustrine rocks include laminated dolomite-gray mudstone, gray sandstone, limestone, and thin, evenly bedded redbeds. Thin fluvial sequences of redbeds separate the lacustrine rocks. The famous dinosaur tracks of the Connecticut Valley make their first appearance in the mudstones of the Shuttle Meadow Formation. They also are found in the East Berlin and Portland Formations.

Volcanic activity then resumed with huge outpourings along fissures of highly fluid basalt that form the 100-m Holyoke Basalt. The lava flows were succeeded by the lacustrine and fluvial strata of the 170-m East Berlin Formation. A third of the formation consists of lacustrine cycles of gray mudstone and sandstone-black shale-gray mudstone and sandstone. The lakes were perennial with alkaline hard water. At times the lakes lapped onto the alluvial fans along the eastern escarpment and extended westward beyond the present faulted and eroded margin of the Hartford Basin. Paleowinds blew from the west and northwest across the lakes recorded in the Shuttle Meadow and East Berlin Formations.

Next, thin lava flows spread from fissures and vents located southwest of central Connecticut to form the 60-m Hampden Basalt. The overlying Portland Arkose is a 1200-m sequence consisting mostly of braided-river sandstone and floodplain red mudstone. Alluvial fans continued to coalesce along the front of the eastern highlands. Thin lacustrine beds of gray mudstone and sandstone are present in the lower half of the Portland Arkose.

The strata in the rift valley have been intruded by basalt dikes and sills, tilted to dip to the southeast, locally folded and faulted, and subjected to erosion.

STOP 1. PORTLAND ARKOSE, DURHAM

*... do you not see that stones even
are conquered by time, that tall turrets
do fall and rocks do crumble ...?*

Lucretius

Location

This outcrop of the Portland Arkose is at the Y-intersection of routes 17 and 77 in Durham, Connecticut. The 13-m cliff is a few meters into the thick woods on the east side of, and directly opposite, the road intersection.

Introduction

The thickness of the Portland Arkose can be only approximated due to faulting and erosion of the top of the formation. In central Connecticut, it is somewhat thicker than 1200 m (Krynine, 1950, p. 69). In the Middletown quadrangle north of Durham, the formation is estimated at 900 to 1050 m (Lehmann, 1959, p. 26).

Along the eastern border fault, the Portland Arkose is interbedded sandy conglomerate and coarse sandstone long recognized as the record of alluvial fans at the base of the eastern escarpment of the rift valley (Longwell, 1937, p. 437; Krynine, 1950, p. 69). A few kilometers to the west, the alluvial-fan deposits pass into an alluvial plain sequence of braid-bar channel sandstone and overbank red mudstone. Several intervals of lacustrine gray mudstone and sandstone occur in the lower part of the formation, including the well known locality for fossil fish at Middlefield, northwest of Durham (McDonald, 1975, p. 77).

Objective of Stop 1

At stop 1 we shall see primary sedimentary structures in conglomerate and sandstone that match those of modern alluvial fans (Bull, 1972; Schumm and Ethridge, 1976). The eastern border fault lies only 0.6 km to the southeast of stop 1.

Description of the Alluvial-fan Sequence

At stop 1, the sequence is mostly sandy conglomerate and pebbly sandstone (Fig. 4). Rapid lateral transitions in texture occur within individual layers along the 40-m outcrop. The boulders exceed 60 cm, reflecting the short distance of 0.6 km to the eastern escarpment of the rift valley. Most of the pebbles are dispersed in sand, implying joint deposition rather than infiltration of sand into the interstices of previously deposited gravel. The beds are horizontally laminated with slight undulations, some of which evidently were the floors of shallow, wide gulleys.

Lenses of sandstone up to 60 cm in thickness are interbedded with the sandy conglomerate. The conglomerates cross-cut, and in turn are cross-cut by the sandstone lenses. Horizontal lamination predominates in the sandstone with some planar cross-bed sets 4 to 8 cm in thickness. The number and thickness of the sandstone lenses decrease upward in the section, evidently reflecting progradation of the fan in response to movement along the west-dipping border fault.

Discoidal clasts of metamorphic rocks are abundant in the conglomerate. The clasts include schist, gneiss, and quartzite. Pegmatitic quartz is also common.

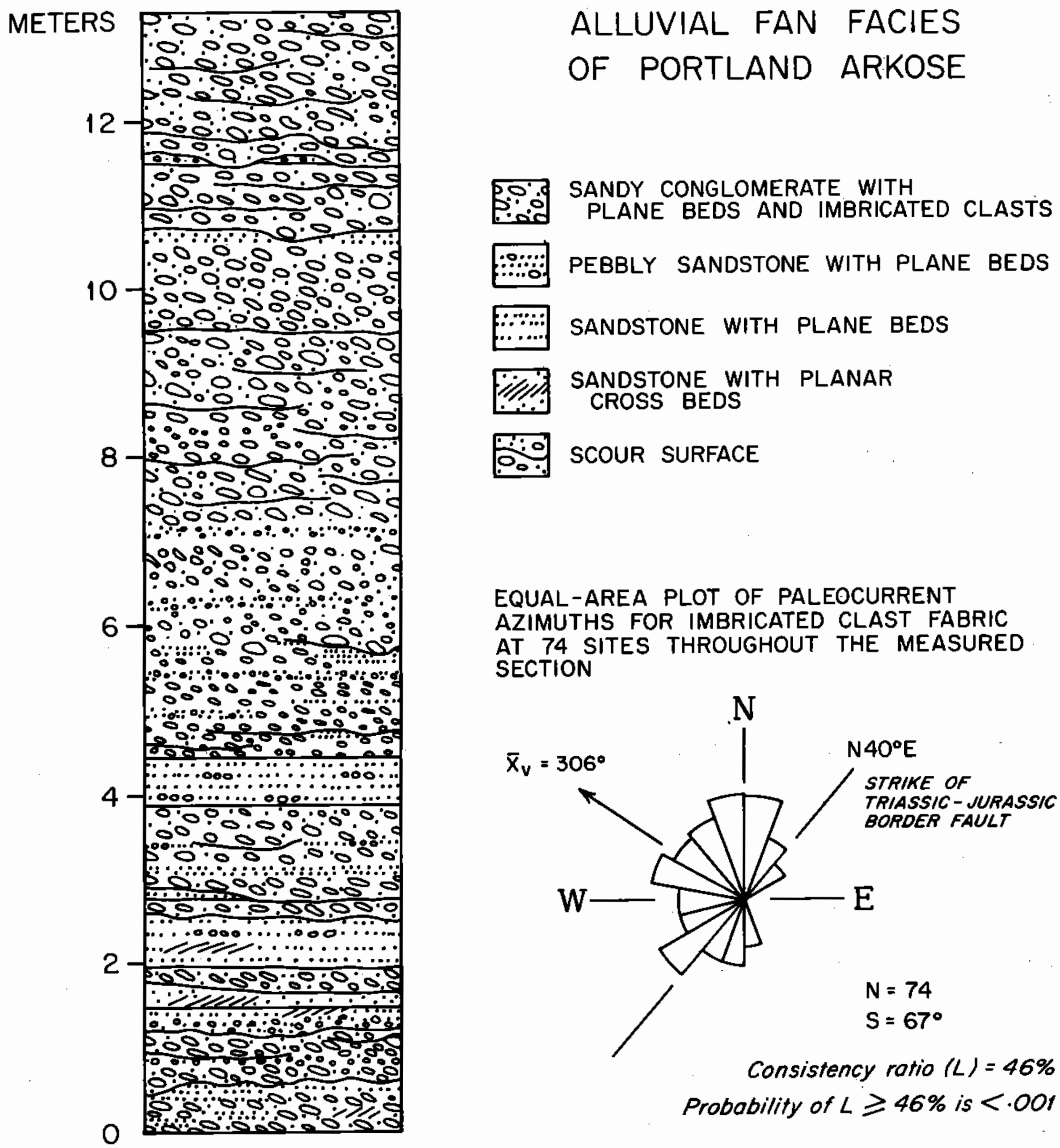

Fig. 4. Measured section of the alluvial-fan facies of the Portland Arkose at stop 1, Durham.

The upcurrent-dipping discoidal clasts produce a prominent imbrication. An equal-area geometric plot of the paleocurrent azimuths of the average imbrications at 74 sites throughout the section at stop 1 is shown in Figure 4. The paleocurrent azimuths compose a semicircular arc with a vector mean to the northwest. The azimuths of the planar cross-bed sets also trend in this direction. The alluvial fan built northwestward, perpendicular to the N40°E strike of the border fault.

Round cobbles and boulders of basalt compose less than one percent of the clasts in the conglomerate. A source of basalt must have existed east of the border fault. The basalt boulders in alluvial-fan conglomerate in the lower part of the Portland Arkose immediately above the Hampden Basalt and adjacent to the border fault at North Branford suggested to Krynine (1950, p. 70) that the Hampden flows extended some distance east of the border fault. The Hampden Basalt abuts the fault with no sign of thinning, supporting this view. The basalt dike that traverses the Haddam quadrangle east of Durham might possibly have been a further source of basalt clasts if the dike is older than the Portland Formation. Basalt cobbles decrease in abundance in the Portland conglomerates south of Durham in the Guilford quadrangle (Mikami and Digman, 1957, p. 62).

Also intriguing are basalt cobbles in the fluvial conglomerate of the Shuttle Meadow Formation along the southwest end of Beacon Hill, east of the trolley line of the Branford Trolley Museum in East Haven (stop 7 of the guidebook of Sanders, 1970, p. 10). This occurrence of basalt clasts is 0.2 km from the eastern border fault.

Sedimentation on the alluvial fan at stop 1 was episodic, characterized by flash floods after heavy rains. Runoff surged down the constricted highland valleys, debouching with great force across the fan apex. We interpret the setting as a tectonically active scarp-fan system with only rare apex entrenchment (Class I system of Bull and McFadden, 1977, p. 124). Sand and gravel were deposited on the apex by sheet floods or by intermittent streams whose width to depth ratio may have exceeded 300. This would account for the lack of deep cut and fill surfaces typical of mid-fan and distal-fan environments.

The shallow depth and high velocity of the water commonly produced large Froude numbers with the turbulent, rapid (shooting) motion of supercritical flow. Consequently, more than 95 percent of the sedimentary rocks at this stop have horizontal lamination developed in the plane bed phase of the upper flow regime. Decreasing flow deposited sandstone layers over the initial coarser flood deposits. The sandstones contain planar cross-bed sets which formed at lower Froude numbers in the dune phase of the lower flow regime.

The absence of a mud matrix in the conglomerates suggests that debris flows and mudflows were not a factor in formation of the 13-m section.

STOP 2. NEW HAVEN ARKOSE, NORTH HAVEN

I slip, I slide, I gloom, I glance,
Among my skimming swallows;
I make the netted sunbeam dance
Against my sandy shallows

Song of the Brook *Tennyson*

Location

Stop 2 is an unusually large exposure of the New Haven Arkose along route 40 in North Haven, Connecticut. The road cut was opened during the spring of 1977 when route 40 was built as a limited access highway linking route 10 on the west with the Wilbur Cross Parkway and I-91 on the east. Exposures are excellent along both sides of route 40. The 72-m section was measured along the north side (Fig. 5).

Objectives of Stop 2

After allowing for some possible duplication of strata due to faulting, the thickness of the New Haven Arkose is about 1950 to 2250 m in the area between Mount Carmel and New Haven, which includes stop 2 (Krynine, 1950, p. 43). Stratigraphically, this outcrop is just below the middle of the formation.

At stop 2, we shall examine an alluvial-plain sequence of channel pale red sandstone and conglomerate interbedded with floodplain red sandy mudstone (Figs. 6, 7). The average direction of flow of the braided rivers was to the southwest. Cross-bed sets of pebbly sandstone commonly exceed 0.5 m in thickness, suggesting avalanche deposition on prograding slipfaces of braid bars.

Also interesting are numerous caliche profiles produced by paleosol calcification of channel sand and floodplain mud. A combination

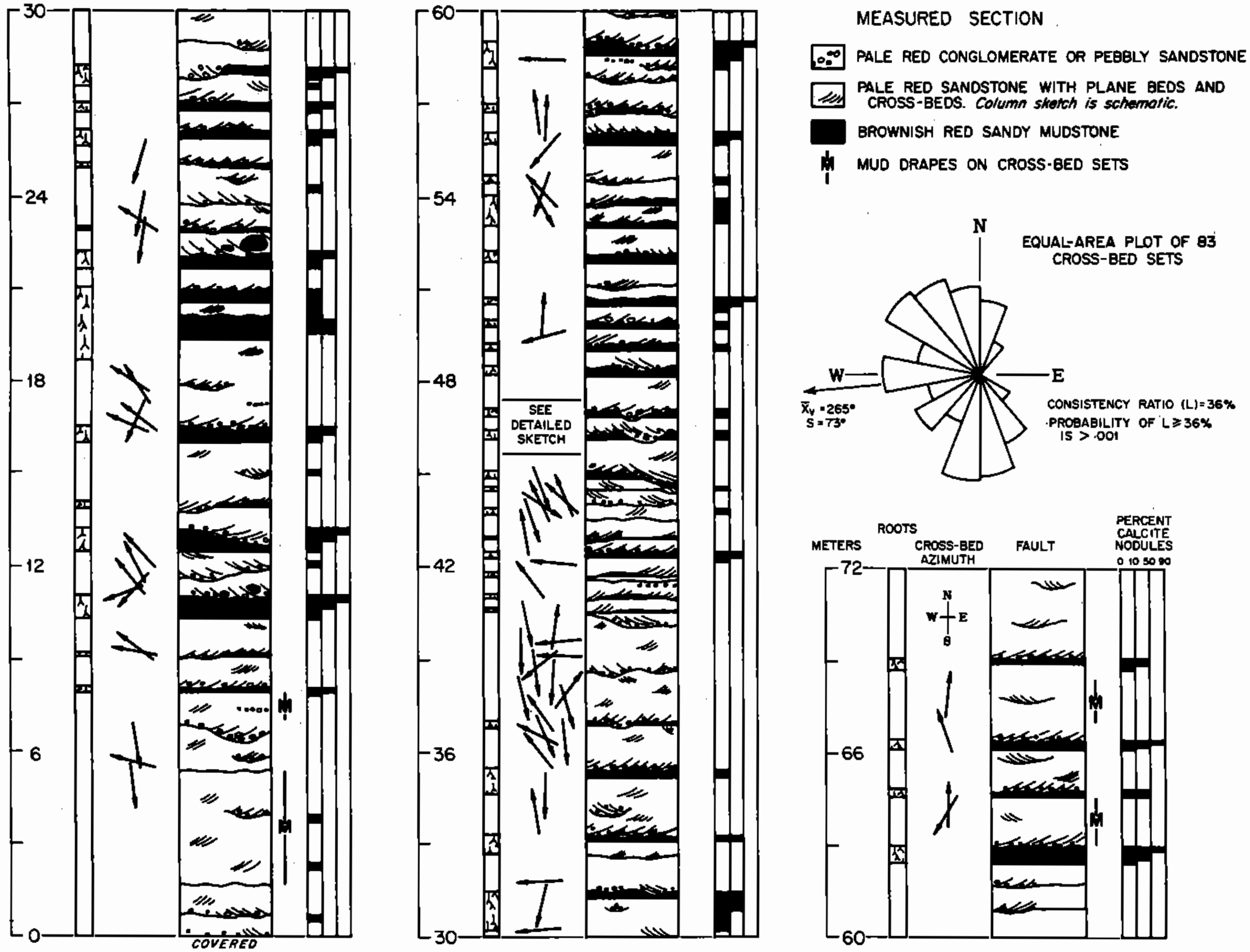

Fig. 5. Measured section of the New Haven Arkose at stop 2 along route 40, North Haven. The detailed sketch is Fig. 8.

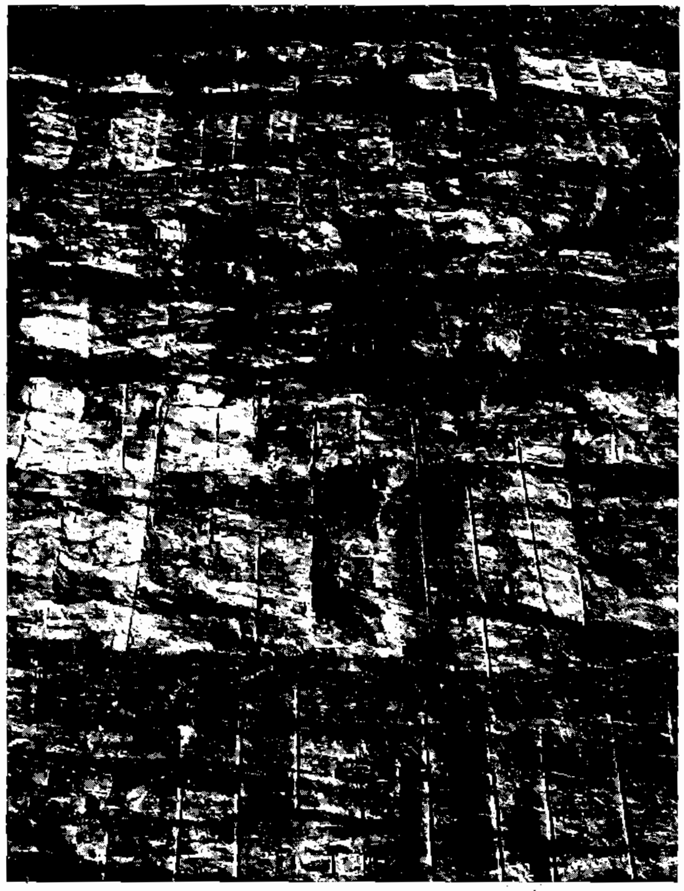

Fig. 6. Braided-river sandstone and floodplain red mudstone in the New Haven Arkose at stop 2.

Fig. 7. Close-up of the lower portion of Fig. 6. Common to Figs. 6, 7, and 8 is the red mudstone boulder with white caliche (next to geology pick, bottom center).

of a slow rate of sedimentation with a paleoclimate dominated by semi-aridity generated 35 caliche horizons in the 72-m section.

Sedimentation on the Braided-river Alluvial Plain

From the evidence presented in the following paragraphs, we infer that the alluvial-plain sequence at stop 2 was deposited by braided rivers. Our interpretation of the hydrologic and topographic characteristics may be summarized as follows. The rivers were ephemeral, with large fluctuations in water discharge, shallow relative to their width, floored by bars and channels, of high gradient and low sinuosity, and with a coarse bedload of pebbly sand. Reviews of the braided-river environment have been prepared by Boothroyd (1976), Collinson (1970), Miall (1977), and Smith (1971, 1972).

The average direction of river flow was to the southwest. The vector mean of 265° is statistically significant at better than the 99 percent level when tested by the Rayleigh statistic. The depositional processes, however, introduced a wide scatter in cross-bed azimuths, as shown by the standard deviation of 73° and consistency ratio of 36 percent (Fig. 5). A high variability of cross-bed azimuths is also characteristic of individual beds of sandstone. For example, in the sketch of Figure 8, the average paleoflow is westerly and the 14 azimuths fan in a 180° arc from south to north.

The bedload of the rivers was sand and pebbly sand. The deposits are a complex pattern of plane beds, festoon cross-beds, and planar tangential cross-beds. The planar tangential sets commonly exceed 30 cm in thickness, reaching a maximum of 135 cm at stop 2 (Fig. 9). The deposits are cut by numerous scour surfaces, 1 to 2 m deep, which are overlain by pebbly sandstone or conglomerate. The scours and thickness of the planar cross-bed

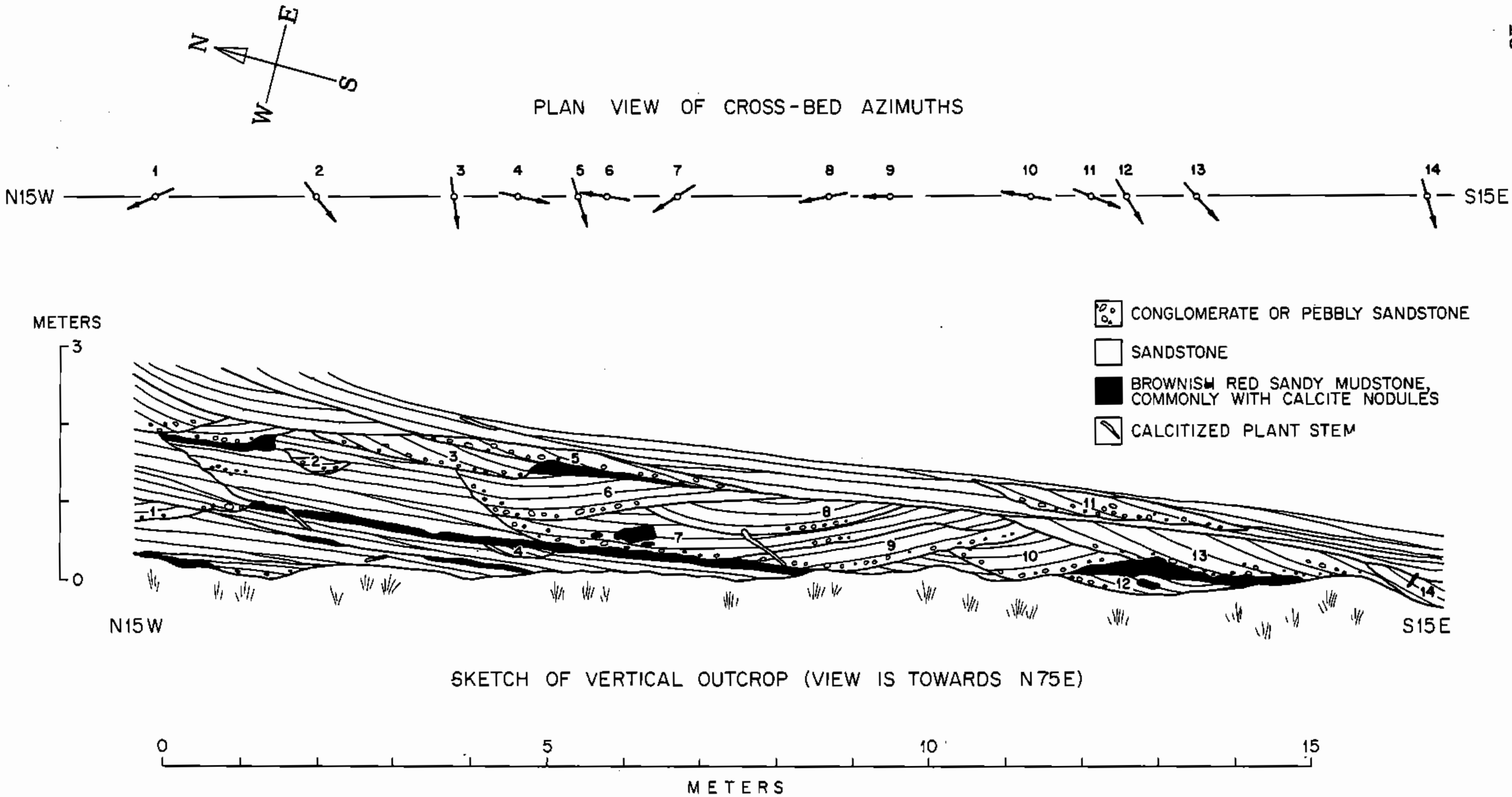

Fig. 8. Braid-bar complex in the New Haven Arkose at stop 2, North Haven. The base of the sketch is at 46 m in the measured section (Fig. 5). The cross-bed sets are numbered to correspond to the azimuths plotted in plan view across the top of the diagram.

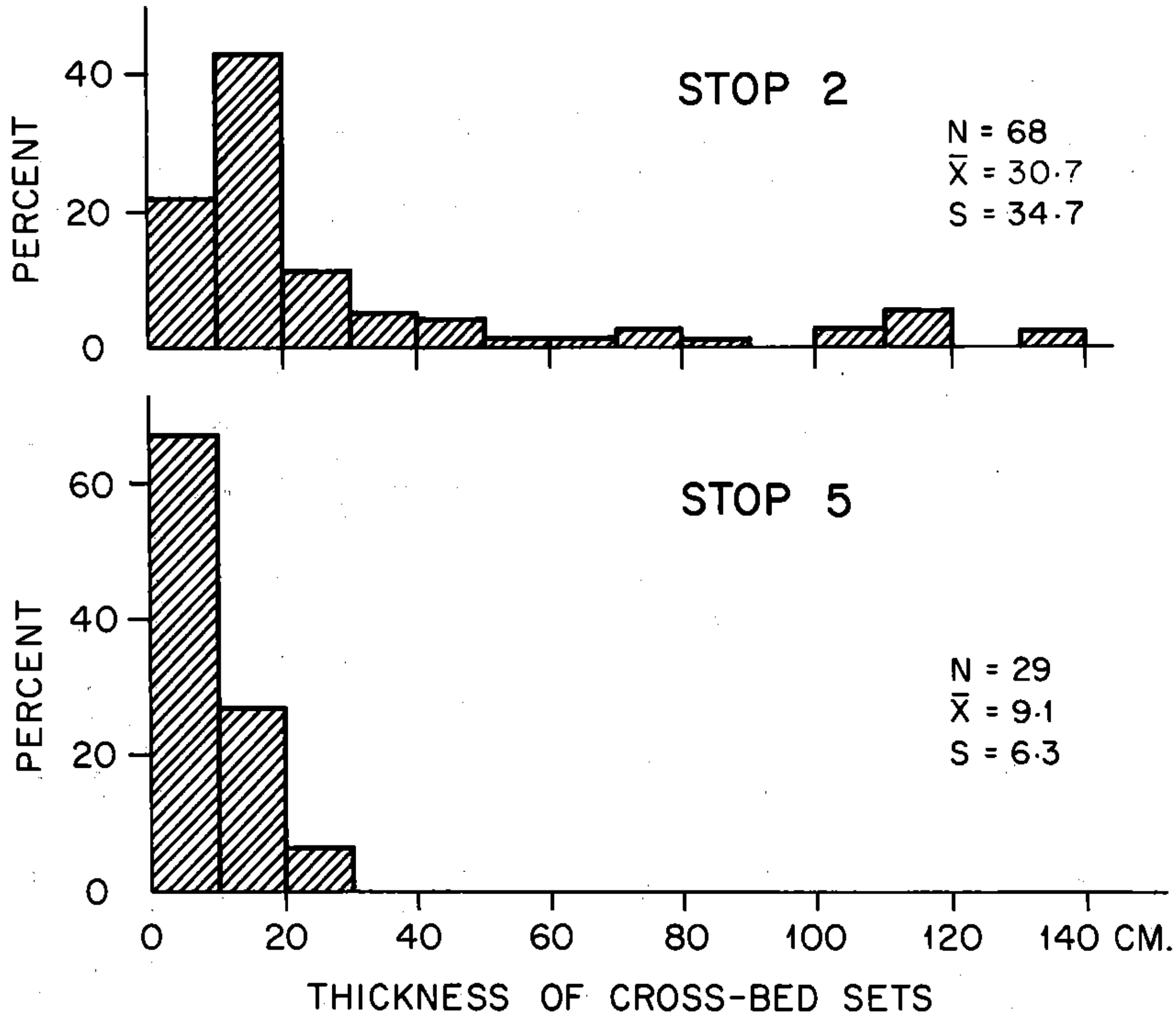

Fig. 9. Histograms of the thickness of cross-bed sets in the New Haven Arkose at stops 2, 5.

sets suggest substantial relief of about 2 m from the tops of bars to the bottom of channels between the bars.

At high-water stages, rushing water occasionally swept away large areas of sand, leaving steep erosional faces about a meter high. A typical scour face is shown in Figure 8 directly above the 4-m mark on the horizontal scale. Low angle cross-beds abut against this scour, with a slight upturn of the ends of the laminae against the face. These structures may record the advance of a bar margin up to the scour face and subsequent filling in of the relief on the channel floor.

At flood stage, the river topped the bars. As the water level fell, pebbly sand and sand were deposited as plane beds in the upper flow regime on the surface of the bars. Some of the sweeping, subhorizontal laminae accumulated on bar margins, which in braided rivers commonly have low depositional dips (Miall, 1977, p. 31). Sand and pebbles avalanched down the occasional steep slipfaces of the bars to form thick planar tangential cross-bed sets with considerable divergent orientation. The shape of the bars is unknown from direct evidence. The presence of gravel in the bedload of the river suggests longitudinal bars of diamond shape, elongated in the direction of the river flow (Fig. 10; Boothroyd, 1976, p. 18; Miall, 1977, p. 31). The bars in braided rivers that transport sand are commonly linguoid or modified linguoid (transverse).

As water fell in the channels between the bars, festoon cross-beds formed at the advancing front of sinuous crested sand waves; sand filled pits scoured by water "boils" created by flow separation over the bed form. Planar cross-bed sets were generated by migration of straight-crested sand waves without scour pits. There are also reactivation surfaces where a cross-bed set is deposited, partially eroded, and then another cross-bed set is deposited on the scoured surface. A reactiva-

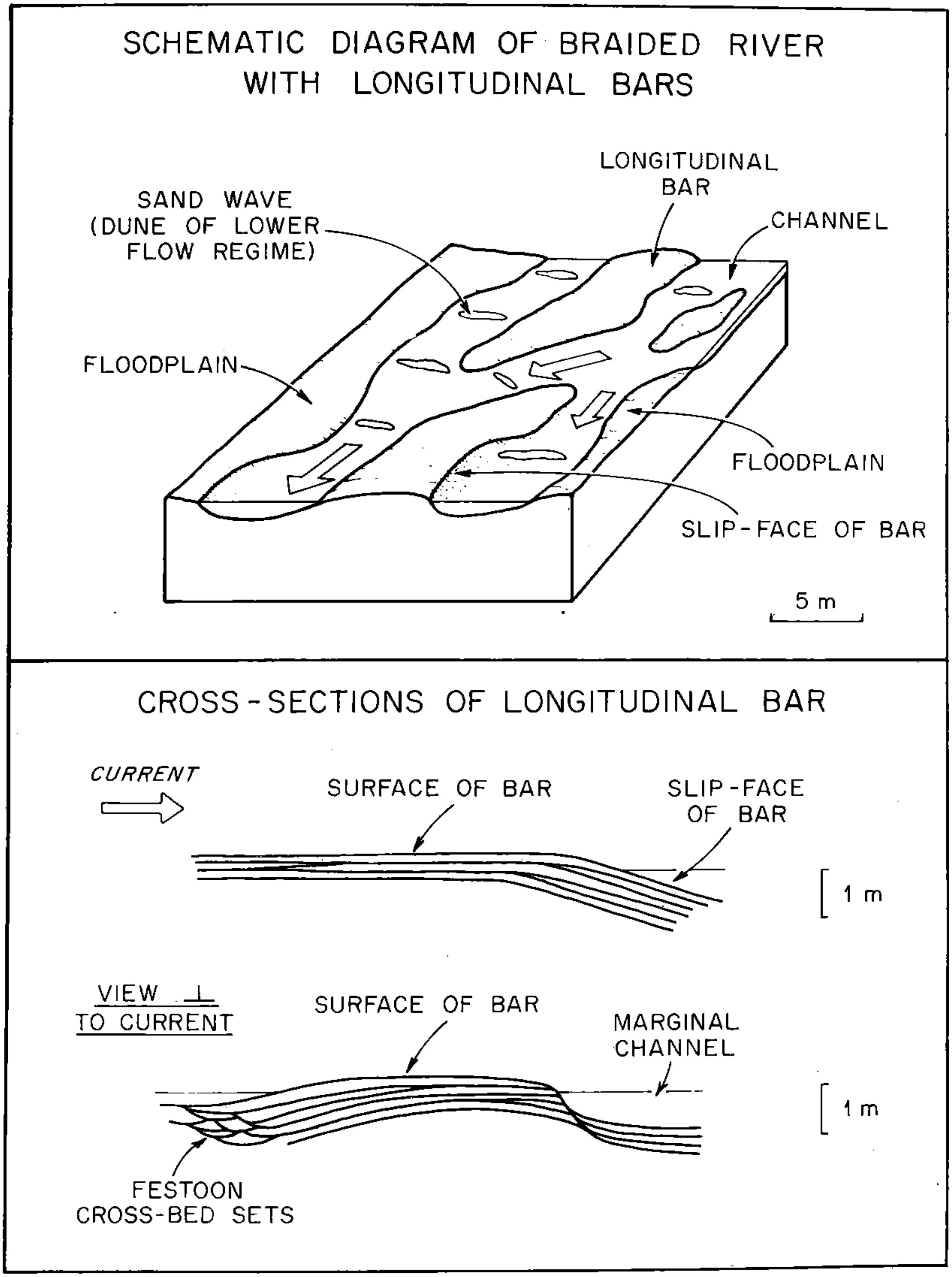

Fig. 10. Sketch of braided river with longitudinal bars (modified from Fig. 10 of Brown et al., 1973, p. 14).

tion surface is present between cross-bed sets 3 and 5 on Figure 8.

During the falling-water stage, currents dissected the bars, producing scour surfaces cut through the sand. Ripples and mud drapes formed on the lower parts of the bars and the channel floors. The juxtaposition of mud drapes on plane-bedded pebbly sandstone of the upper flow regime testifies to the rapid fluctuation in discharge characteristic of braided streams. Mud drapes are well developed in the lower part of the section at 2-8 m and again high in the section at 63-68 m (Fig. 5).

Numerous layers of red sandy mudstone with gray calcareous horizons in their upper portions can be seen extending across the outcrop at stop 2. The layers are cut through in places by channel sandstone and some appear as discontinuous thin stringers in the channel sandstone because they are erosionally truncated at both ends. These sandy mudstones were deposited by flood water that spread over floodplains. Mudstone comprises 14 percent of the section, a low proportion compatible with the interpretation that only a relatively small volume of mud was provided by semiarid weathering in the highlands along the rift valley.

Next to be discussed are paleosol caliche profiles in the red mudstones that indicate a semiarid paleoclimate with 100-500 mm of seasonal precipitation and a long dry season. Flash floods were severe in the rainy season, as shown by the scour surfaces and conglomerate lenses in the fluvial sandstone and by the chunks of red mud larger than 1 by 2 m that were ripped by rushing floodwaters from the collapsing banks of the rivers. Extensive calcification of the mud helped maintain cohesiveness of the blocks as they moved at least short distances down river.

Caliche Paleosol Profiles

Description of the Caliche

The section at stop 2 contains numerous calcareous horizons with densely packed root casts in floodplain red mudstone and channel pale red sandstone (Fig. 5). The calcareous horizons and their petrographic microstructures match caliche profiles of Quaternary age (Gile *et al.*, 1966, p. 348; Reeves, 1976, p. 120) and caliche occurrences in ancient terrestrial sequences (Allen, 1974a, p. 114-120; Steel, 1974, p. 353-354) and thus are interpreted as ancient caliche profiles (Hubert, 1977, p. 302, and 1978). The problems associated with recognition of paleosols are reviewed by Yaalon (1971), Buurman (1975), and Valentine and Dalrymple (1976).

Four sequential stages are present in the calcareous horizons in the New Haven Arkose (Fig. 11). The initial stage of calcification is the most common, consisting of irregularly shaped calcite nodules a few centimeters in diameter that compose up to 10 percent of the mudstone or sandstone. They are made of 1- to 100-μ calcite crystals in a crystic plasmic (microspar) fabric with residual patches of red mud characteristic of paleosols (Brewer, 1964, p. 317). Many sand grains in the mudstones are encrusted by acicular calcite; in sandstones, pseudo-ooids result from multiple crusts around sand nuclei. With progressive development, the calcite nodules became larger, constituting 10-50 percent of the rock, commonly as vertical cylindroids (Fig. 12). A third stage of development increased the calcite to 50-95 percent of the rock as nodules, veins, and diffuse patches of calcite (Fig. 13). In the most advanced stage, with more than 90 percent calcite, the limestone has a complex fabric of nodules, veins, laminae, clasts, and cement. These layers of limestone are thin (0 to 10-cm) sheets with

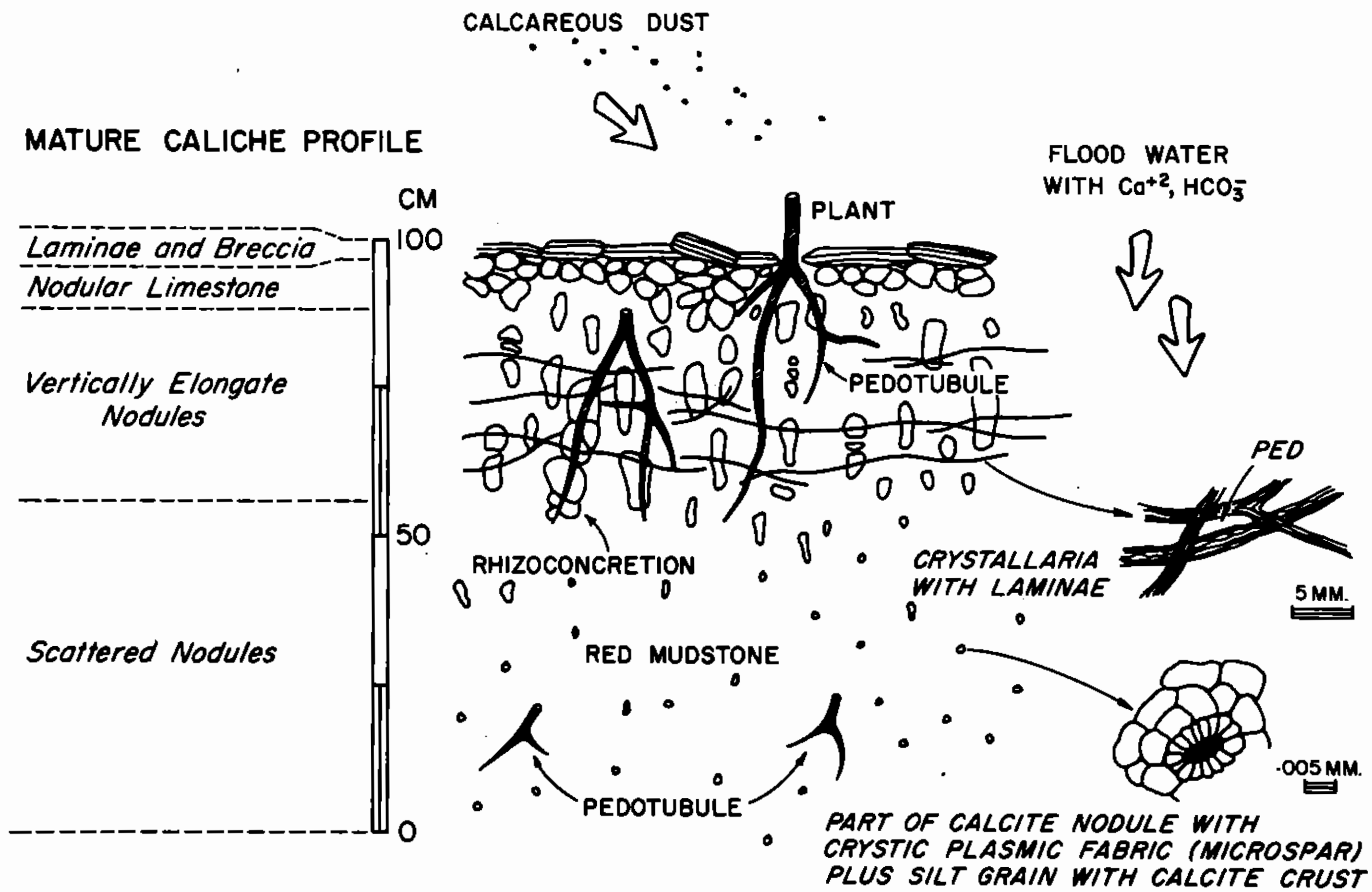

Fig. 11. Field and microscopic features of caliche paleosol profiles in the New Haven Arkose. From Hubert, 1977, Fig. 4.

thin places and holes. Locally a limestone is capped by discontinuous calcite laminae that total a few mm in thickness. Mature caliche profiles with nearly pure beds of limestone are absent at stop 2, but may be seen along I-84 near exit 29 in Southington, Connecticut (Fig. 14). The calcified red mudstone and sandstone contain conspicuous green patches and layers produced by local removal of limonite grain coatings as ferrous-organic complexes in reducing soil water, probably due to decaying plant material.

The limestones contain multiple generations of calcite veins that bound areas of calcite microspar and red mud in a paleosol fabric of crystallaria and peds (Brewer, 1964, p. 150, 284). The wider (5 to 10 cm) crystallaria have internal calcite laminae with crinkled, convoluted, and pseudo-pisolitic forms. The centers of some are void-filling calcite mosaics; others contain clusters of gypsum crustals, as at stop 2. Wetting and drying of the calcareous muddy soil produced planar openings, mostly parallel to the ground surface, in which calcite was precipitated from downward percolating soil water enriched in calcium, bicarbonate, and sulphate.

Fig. 12. Caliche profile in red mudstone of New Haven Arkose at 28 m in the section at stop 2, North Haven. The top of the geology pick (left center) is at the base of a laterally persistent layer of densely packed root casts. The roots bifurcate downwards and are replaced by calcite that weathers white. A nodular limestone forms the upper part of the horizon of root casts. The paleosol profile is erosionally overlain by cross-bedded channel sandstone. From Hubert, 1977, Fig. 2.

Fig. 13. Caliche profile in floodplain red mudstone of the New Haven Arkose at 63 m in the section at stop 2, North Haven. Nodular limestone of uneven thickness passes transitionally down to vertically stacked calcite nodules that outline former plant roots. The red mudstone low in the profile contains smaller nodules and is densely packed with pedotubules. The cross-bedded channel sandstone (coin) cuts into the limestone and contains numerous clasts of reworked caliche. From Hubert, 1977, Fig. 3.

Fig. 12.

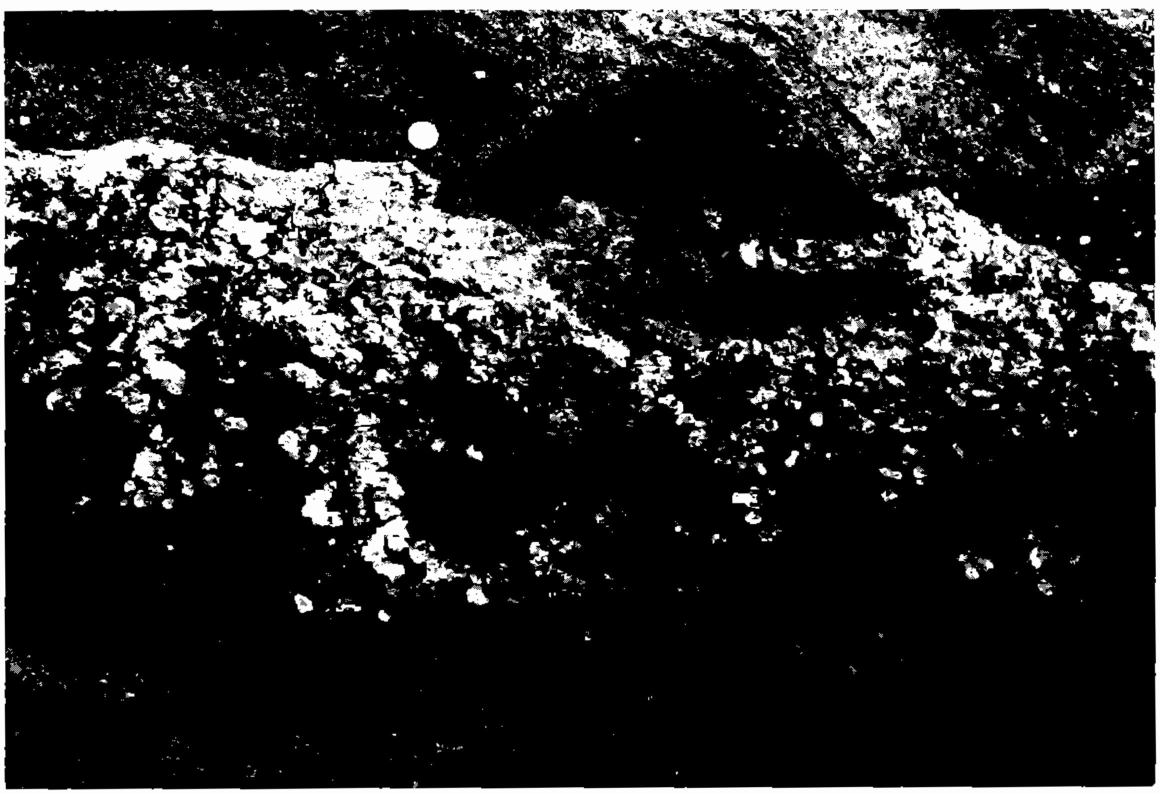

Fig. 13.

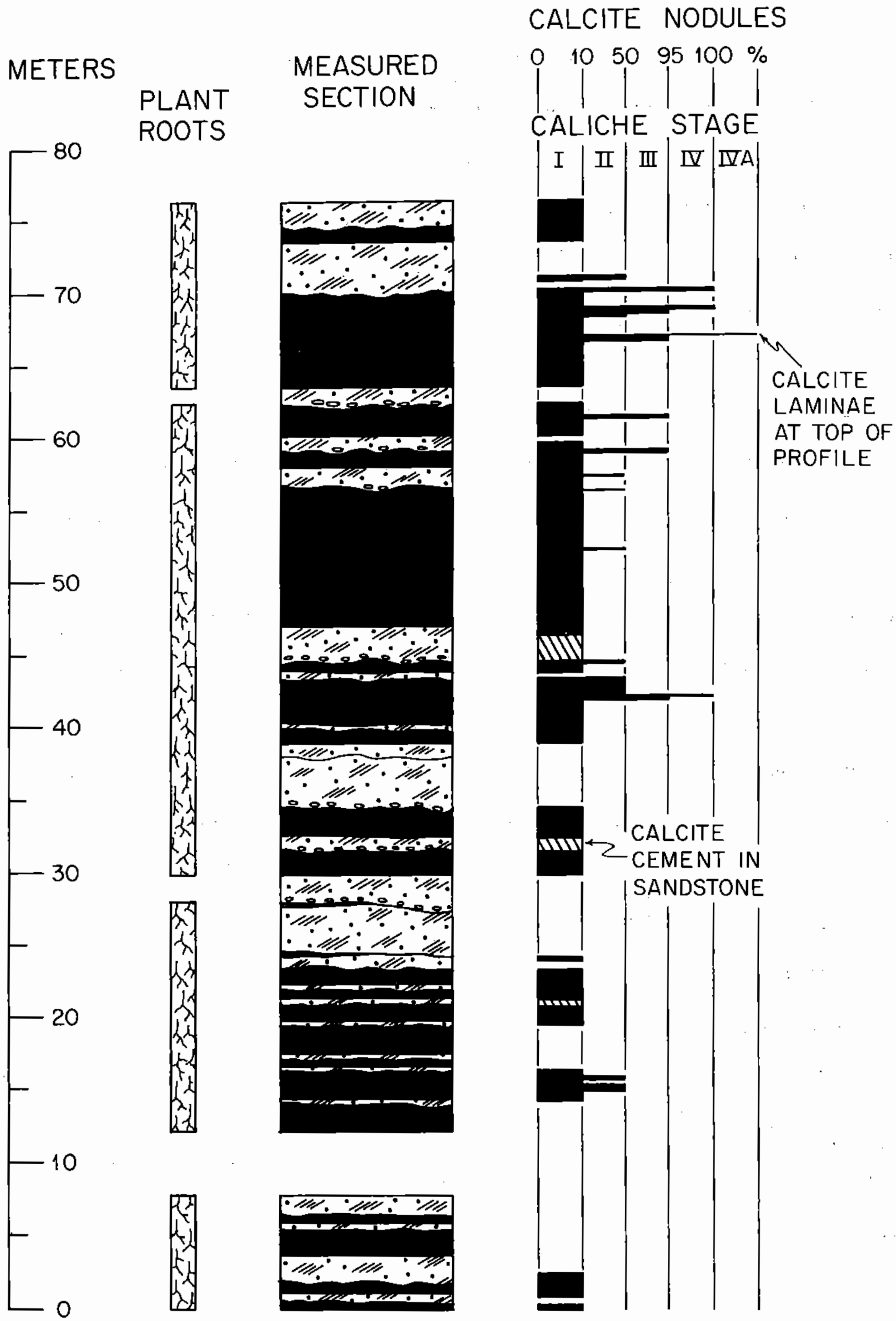

Fig. 14. Measured section of New Haven Arkose along I-84 near exit 29, Southington, Connecticut. Symbols as in Fig. 5 (from Hubert, 1977, Fig. 5).

Soil Processes and Paleoclimate

Carbonic acid plays an important role in the calcification of soil. It is produced when rain or river water infiltrates soil and combines with CO_2 generated by decomposition of organic matter and respiration of roots and micro-organisms. The soil concentrations of CO_2 are 10 to 100 times the atmospheric concentration. The carbonic acid dissolves carbonate particles in the upper part of the soil, enriching the downward percolating water in calcium and bicarbonate.

At some depth determined by permeability and other factors, carbonate, usually calcite, is precipitated due to rising pH, drop in temperature, decrease in CO_2 partial pressure, and higher ion concentrations due to evapotranspiration. In the New Haven Arkose the calcite commonly is non-ferroan, as shown by petrographic thin sections stained with alizarin Red-S and potassium ferricyanide.

The older concept that capillary forces draw calcium-rich water up from the water table is now discounted because the distance of capillary rise as measured experimentally is only about 1 m. Suitable perched or very shallow water tables are rare in semiarid regions.

Calcite was extensively precipitated in both sand and mud at stop 2 and the section at Southington (Fig. 14). Such extensive pedogenic caliche is characteristic of semiarid alluvial plains with 100-500 mm of seasonal rain (Reeves, 1976, p. 85). The very low precipitation of arid regions is less favorable. Before later Mesozoic opening of the North Atlantic, the Connecticut rift valley was in the hot tropics at about 15 degrees north paleolatitude (Van Houten, 1977, p. 93).

The thicker caliche profiles required several thousand years of soil formation and low rates of detrital sedimentation (Leeder, 1975, p. 212;

Allen, 1974b, p. 664). Wide lateral shifts or pronounced incisions of the braided-river belt evidently produced areas removed from sediment accumulation. Also favorable would be slowing down of the rate of vertical movements along the fault-bounded eastern escarpment.

The caliche profiles are particularly impressive because the alluvium lacks detrital limestone and dolostone grains that could be sources for calcium. Calcium instead was provided by southwest-flowing rivers that drained the metamorphic-igneous highlands along the rift valley. Plant roots concentrated the calcium, releasing it on decay to soil water. Another source may have been calcareous dust blown from the carbonate sequence of Lower and Middle Paleozoic age that crops out west of the rift valley. Ripple marks in lacustrine beds of the Shuttle Meadow and East Berlin Formations show that the Early Jurassic paleowinds predominantly blew from the west and northwest (stops 6, 7; Hubert et al., 1976, p. 1200). Calcareous dust will dissolve in carbonic acid generated by soil water combining with CO_2 released by decomposition of plants.

During the millions of years spanned by the New Haven Arkose, there must have been numerous fluctuations in the paleoclimate, at times perhaps very pronounced, affecting annual precipitation, length of the dry season, and mean annual temperature. Nevertheless, the wide areal and stratigraphic distribution of caliche horizons in the mudstone and sandstone imply that much of the time conditions were favorable for formation of calcareous soils. The paleoclimate was hot and commonly semiarid, perhaps with 100 to 500 mm of seasonal rain and a long dry season. The paleoclimate was much drier than the tropical and wet paleoclimate postulated by Krynine (1950, p. 182), who favored a regime with a minimum of 1250 mm of seasonal rain.

Figure 15 is a map of paleoclimatic zones for Late Triassic time. The zonal patterns are inferred from the distribution of climate-sensitive rocks of Late Triassic age, namely eolian sandstone, evaporites, and coal, together with analysis by Robinson (1973) of the theoretical positions of the paleoclimatic zones that would be predicted by the shape and paleolatitude of the continent and Tethys Sea. Figure 15 is modified from Robinson (1973, p. 466) to include an equatorial humid region on the west coast of Laurasia-Gondwana. This belt of increased annual precipitation reflects the coal beds in the redbed sequence of the Durham-Wadesboro rift valley, North Carolina. Precipitation evidently was relatively high during formation of the coal beds, but semiaridity returned during accumulation of the redbeds which contain playa-lake limestone and chert (Wheeler and Textoris, 1977, pers. comm.). Lateral shifts in the paleoclimatic zones could account for the long-term variations in precipitation. The semiaridity and seasonal precipitation implied by the caliche in the New Haven Arkose are compatible with the location of the rift valley in the transition zone between an area of aridity to the east recorded in the salt basins of Morocco (Van Houten, 1977, p. 85) and higher precipitation required for coal beds in the Durham-Wadesboro area (Bain and Harvey, 1977, p. 10) (Fig. 15).

Caliche of Late Triassic age also occurs in the fluvial Hammer Creek Conglomerate, New Jersey (Van Houten, 1969, p. 337), a 240-m core drilled near Thomasville, Pennsylvania (Cloos and Pettijohn, 1973, p. 529), and the fluvial redbeds of the Wolfville Formation, Nova Scotia (Jensen, 1975, p. 80).

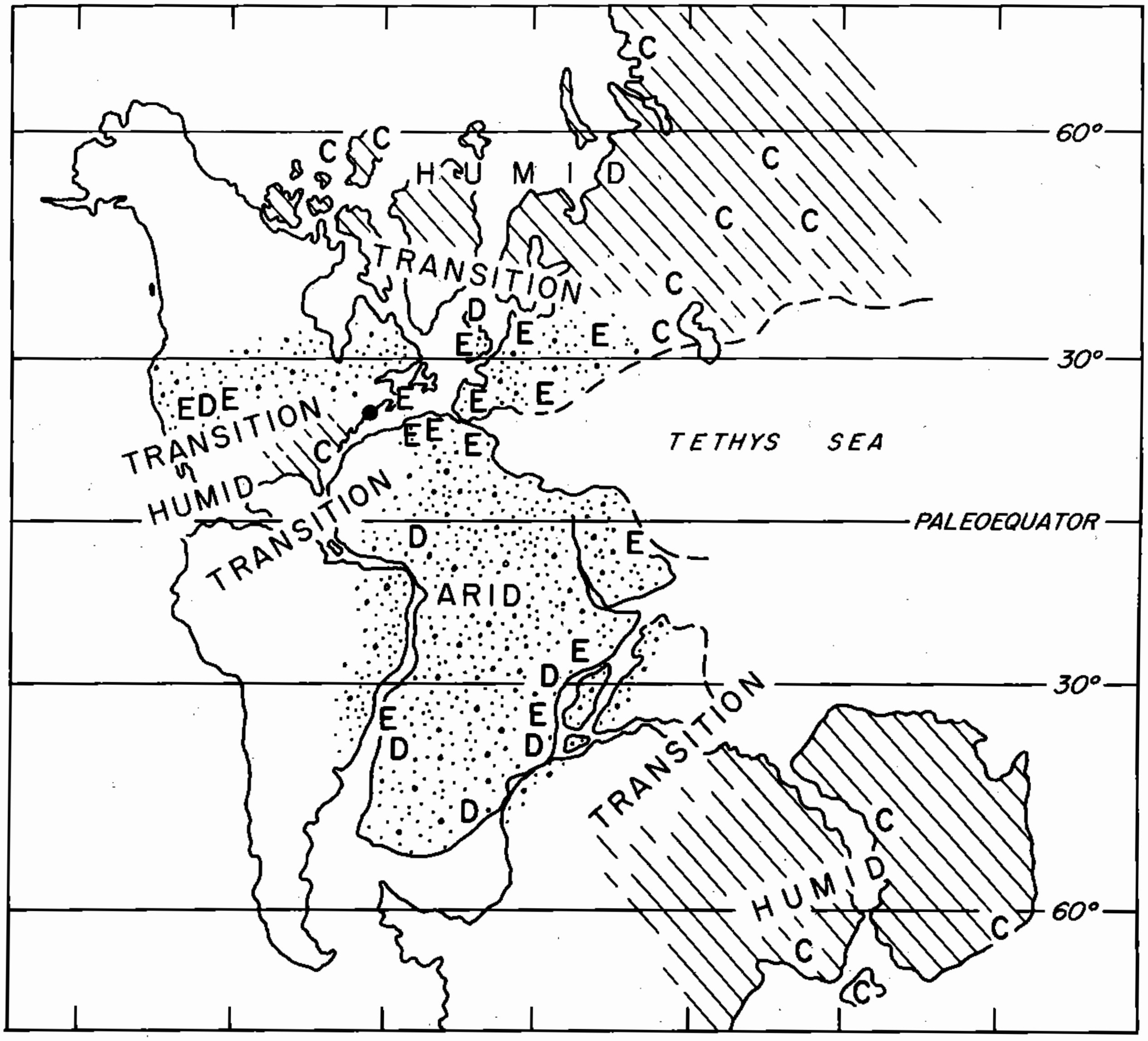

Fig. 15. Late Triassic paleoclimatic zones in Gondwanaland and Laurasia (modified from Robinson, 1973). The climate-sensitive rocks are coal (C), dune sandstone (D), and evaporite sequences (E). The Connecticut rift valley is located by the small circle.

Broad Terrane Hypothesis

Introduction

The combination of the east dip of the redbeds in the Hartford Basin with the west dip of the redbeds in the Newark Basin has long teased geologists with the possibility that the basins are remnants of a formerly continuous rift valley (Fig. 16). The idea is reinforced by the presence of normal faults and alluvial-fan conglomerate on both the west side of the Newark Basin and east side of the Hartford Basin. After mentioning that others before him had liked the concept, Russell (1878, p. 230) wrote the first paper proposing a broad terrane hypothesis. He elaborated the proposal in a second paper in 1880. The concept then was accepted by a number of geologists, for example Longwell (1922, 1928). Additional supporting evidence, largely structural, was advanced by Wheeler (1937, 1938) and Sanders (1960, 1963, 1974).

The reconstructed rift valley is about 60 km wide (Fig. 17). Near the center of the inferred rift is the Pomperaug Outlier, with its Upper Triassic to Lower Jurassic sequence of fluvial and perennial lake strata and lava flows (Figs. 3 and 17; Scott, 1974). A gap of 60 km separates the outcrops in the northern Newark Basin from the Pomperaug Outlier.

Modified Concept

As first proposed, the broad terrane hypothesis envisioned a graben bordered on each side by normal faults. The faults remained active throughout the time that terrestrial sediments and lava flows accumulated on the subsiding floor of the rift. This model should be modified to allow for widening of the rift valley as it evolved from Late Triassic to Early Jurassic time. Under this suggestion, the graben widened as new border faults successively developed at sites further from the axial region of the rift. Perhaps at times there were horsts within the graben.

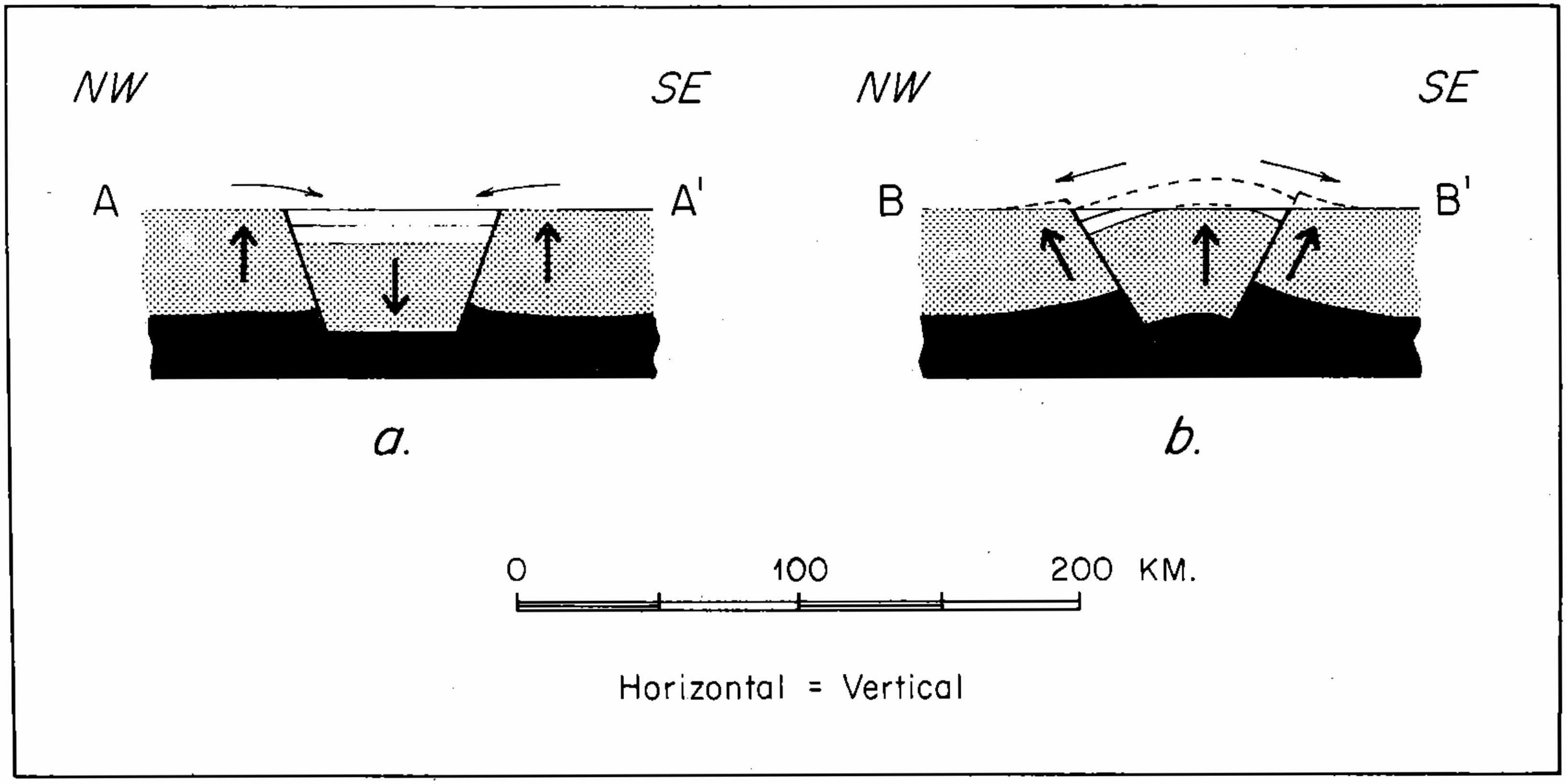

Fig. 16. Schematic diagram showing the Newark and Hartford Basins as remnants of a former rift valley (broad terrane hypothesis of Russell, 1878, p. 230). (a) Late Triassic to Early Jurassic deposition in the rift valley. (b) Post-depositional uplift and erosion.

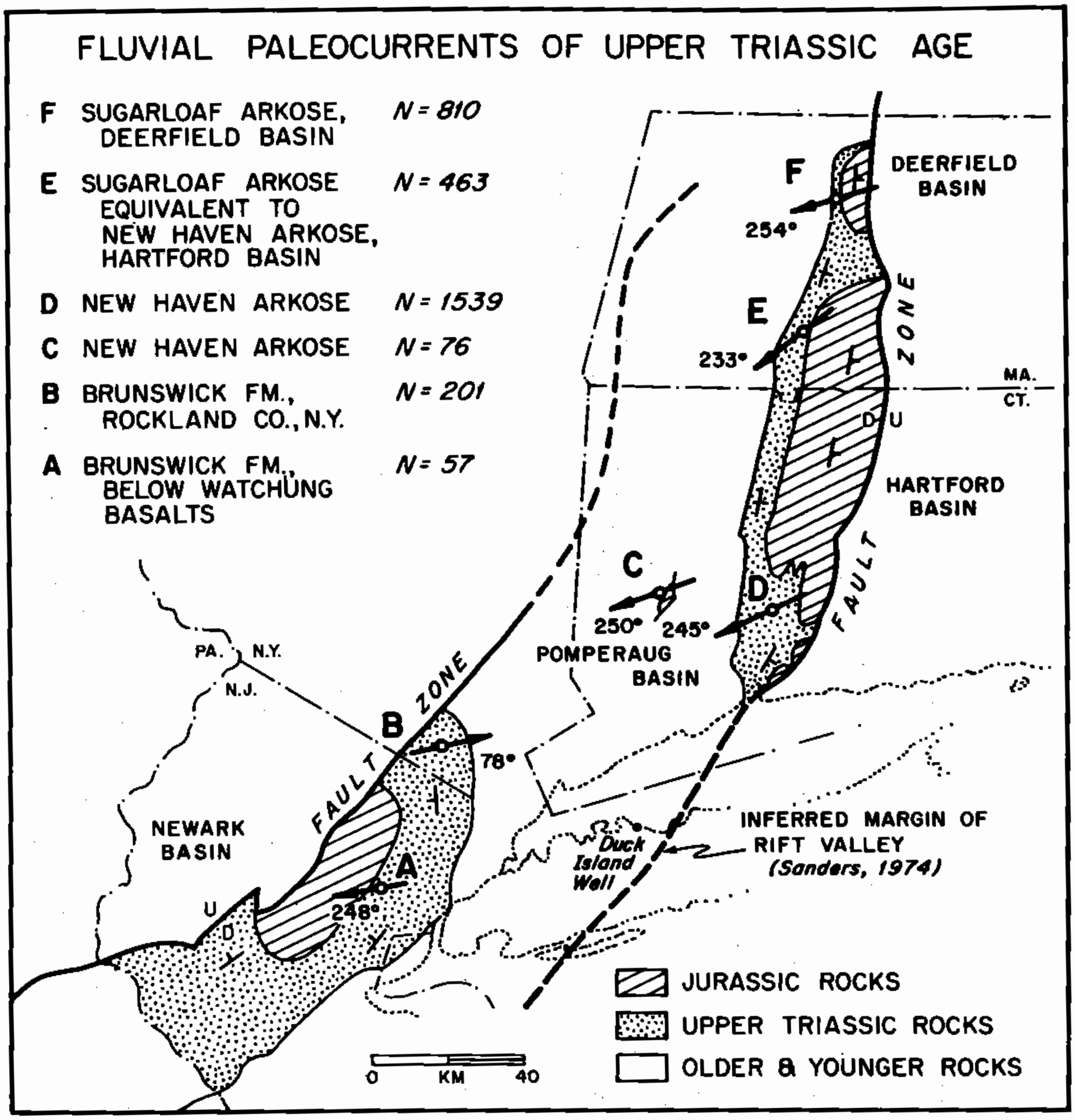

Fig. 17. Reconstruction of a Newark-Hartford rift valley under the broad terrane hypothesis. Vector means for Late Triassic paleocurrents are: (A) Manspeizer, 1977, pers. comm.; (B) Savage, 1968; (E) Franz, 1977, pers. comm.; and (F) Stevens, 1977. The unknown margins of the rift valley are generalized as dashed lines.

Support for this model where normal faults "step-down" basinward comes from Chang (1968, p. 94-96), who explains his gravity map for southern Connecticut by modeling a buried fault scarp near the middle of the Hartford Basin. The scarp extends in a north-south direction at least from New Haven to Meriden, where map control ends. Chang estimated displacement on the scarp to be 1800 to 2400 m. West-dipping, closely spaced normal faults are inferred along the scarp, controlling fluvial deposition of the lower part of the New Haven Arkose. Beginning with the upper part of the New Haven Arkose, the detritus came from east of the present border faults and buried the earlier scarp. The deepest part of the basin is not adjacent to the exposed border faults, an idea also supported by the gravity data of Eaton and Rosenfeld (1960, p. 176) for central Connecticut in the vicinity of Middletown. To evaluate the significance of these concepts, additional gravity stations should be occupied and the work expanded to cover the Hartford and Deerfield Basins.

Similar step faulting with buried scarps seems to be common in the basins of the Newark Supergroup. Buried scarps are inferred in the Deerfield Basin from geological map evidence (Willard, 1951). Gravity surveys suggest buried scarps on the northern margin of the Newark Basin in New Jersey (Dunleavy, 1975), the Newark-Gettysburg Basin of southeastern Pennsylvania (Sumner, 1977, p. 941), and the Durham-Wadesboro Basin of North Carolina (Bain and Harvey, 1977, p. 22).

The relative timing of sedimentary and volcanic events in the Hartford and Newark Basins can be estimated by the eight palynofloral zones present in the lacustrine gray mudstone (Fig. 33 of Cornet, 1977). The sedimentary record in the Hartford Basin began later and finished later than in the Newark Basin. The strata in the Hartford Basin span about 24 million years, whereas those of the Newark Basin span about 36 million years.

In the Hartford Basin, the New Haven Arkose began to accumulate during the Early Norian (Upper Triassic) at 201 million years B.P. (Fig. 33 of Cornet, 1977). The Portland Formation at the top of the sequence extended into the Early Jurassic and perhaps into the Bojocian (Earliest Middle Jurassic) at about 165 million years B.P. Summing the maximum thickness of each formation, the column in the Hartford Basin is about 4 km thick (Fig. 3). At any one locality the section is substantially less.

In the Newark Basin, sedimentation began with the Stockton Arkose during the Late Middle Carnian (Upper Triassic) at about 208 million years B.P. The sequence ends in the upper part of the Brunswick Formation during the Late Sinemurian (Early Jurassic) at about 184 million years B.P. The maximum column is about 9 km thick.

In both basins, the interval of outpouring of basalt lavas was during the Early Jurassic (Fig. 3; Cornet, 1977). The flows were mostly fissure eruptions of great fluidity (Faust; 1975, p. 33). In the Hartford Basin, the Talcott, Holyoke, and Hampden lava-flow units have a mean age of 184 million years B.P., with a standard deviation of 8 million years, as evidenced by 14 K-Ar whole rock age determinations (Reesman, et al., 1973, p. 211). The Palisades sill in New Jersey was intruded about 190 million years B.P. (Dallmeyer, 1975, p. 244). The boundary between the Triassic and Jurassic is about 192 million years B.P. (Van Hinte, 1976, p. 490).

The Hartford and Newark Basins differ in time of initial and terminal deposition and in total thickness of the sedimentary-volcanic column. If the basins are remnants of a single rift valley, its evolution was characterized by substantial irregularity in spacing and timing of normal faulting and associated areal patterns of sediment accumulation.

The concept of a broad terrane remains a working hypothesis useful to suggest future lines of research. Our discussion of it must end not with a period, but a question mark.

Isolated Basin Hypothesis

The broad terrane hypothesis is criticized by geologists who favor an isolated basin hypothesis. In this view, the strata in the Newark Basin were never continuous with the sequence of the Hartford Basin-Pomperaug Outlier.

An effective argument for isolated basins, if substantiated, would be that some of the detritus in the Newark or Hartford Basins can be traced to specific source rocks between the basins (Klein, 1968, p. 1827). This area is the floor of the inferred rift valley. In the Newark Basin, a large proportion of the detritus in the Upper Triassic rocks of northern New Jersey and Rockland County, New York, has K-Ar ages of about 340 million years B.P. (Abdel-Monem and Kulp, 1968, p. 1237). Rocks of this age are absent on the northwest side of the Newark Basin, which led Abdel-Monem and Kulp to favor isolated basins. Metamorphic rocks imprinted with Acadian ages, however, crop out in southern Connecticut east of the Hartford Basin and they may continue southward across Long Island in the subsurface. The limited paleocurrent data in the northern Newark Basin show that rivers in Late Triassic time flowed from west of the basin and also from northeast to southwest (Fig. 17). These data are compatible with an interpretation that some detritus was transported across the valley from eastern highlands.

Further southwest in central New Jersey, in the vicinity of the Delaware River, much of the detritus in Late Triassic time came from southeast of the basin. The evidence consists of the high proportion of sodic plagioclase among the grains (Van Houten, 1965, p. 836 and 1969, p. 315) and the map patterns of conglomerate and feldspar percentages (Glaeser, 1966, p. 17).

The broad terrane hypothesis would be strengthened if Triassic-Jurassic redbeds are present beneath the Upper Cretaceous cover on western Long Island. Especially interesting is the report of Wheeler (1938, p. 141) that the "Duck Island well" on the north shore of western Long Island encountered bedrock at 121 m and drilled through 314 m of "typical brown sandstone of the Triassic", bottoming in this unit. This is well S-34, drilled in 1914 (Fig. 17). Unfortunately, the driller's original log is lost. The published description of well S-34 quotes the description of the "bedrock" in the driller's log as consisting of the one word "sandstone" (deLaguna and Brashears, 1948, p. 32). deLaguna and Brashears say (1948, p. 8), "No Triassic rocks have been identified in the well records of Long Island."

The regional contours of the depth to the bedrock surface beneath the Cretaceous strata show that at well S-34 the depth to bedrock is about 174 m (Jensen and Soren, 1974). Jensen, the present U.S.G.S. groundwater geologist for the area, suggests (1977, pers. comm.) that well S-34 penetrated a buried valley in the metamorphic rocks. Narrow subsurface buried valleys with relief that exceeds 180 m are common beneath the Cretaceous rocks along the north shore of Long Island, including the area of well S-34 (Grim *et al.*, 1970, map on p. 662). The total depth of well S-34 is 495 m, which would imply an unusually deep buried valley of 321 m relief. Perhaps the driller's total depth of the well is in error. Jensen (1977, pers. comm.) says that none of the many wells drilled to the pre-Upper Cretaceous basement in central and western Long Island has encountered Triassic-Jurassic redbeds. In brief, the log of well S-34 is at best ambiguous. The existence of Triassic-Jurassic redbeds on Long Island has yet to be established.

Pomperaug Outlier

In the Pomperaug Outlier, 138 m of redbeds crop out along the Pomperaug River in South Britain, Connecticut. The section correlates with part of the New Haven Arkose of the Hartford Basin (Fig. 3; Scott, 1974, p. 34; Cornet, 1977, p. 124).

At this section, eight of the channel sandstones show pronounced fining-up sequences. In each cycle, the basal erosional surface is commonly overlain by pebbles, followed by an interval of festoon cross-beds, and then plane beds (Fig. 18). These sandstone bodies are point-bar deposits. The other channel sandstones lack clear fining-up sequences but contain abundant festoon cross-beds and some planar cross-beds.

Floodplain brownish-red mudstone is interbedded with the channel sandstone (Fig. 18). The mudstone is sandy and most beds contain casts of plant roots. Caliche paleosols occur at two horizons.

These Late Triassic strata in the Pomperaug Outlier were deposited by meandering rivers that flowed to the southwest (Figs. 18, 19).

Regional Maps of Hartford Basin-Pomperaug Outlier

The regional maps of the Hartford Basin and Pomperaug Outlier for Late Triassic time show river morphology and dispersal patterns (Fig. 20), maximum thickness of cross-bed sets (Fig. 21), and mean size of the five largest igneous and metamorphic clasts (Fig. 22). The information provided by the maps is necessarily generalized because they are based on

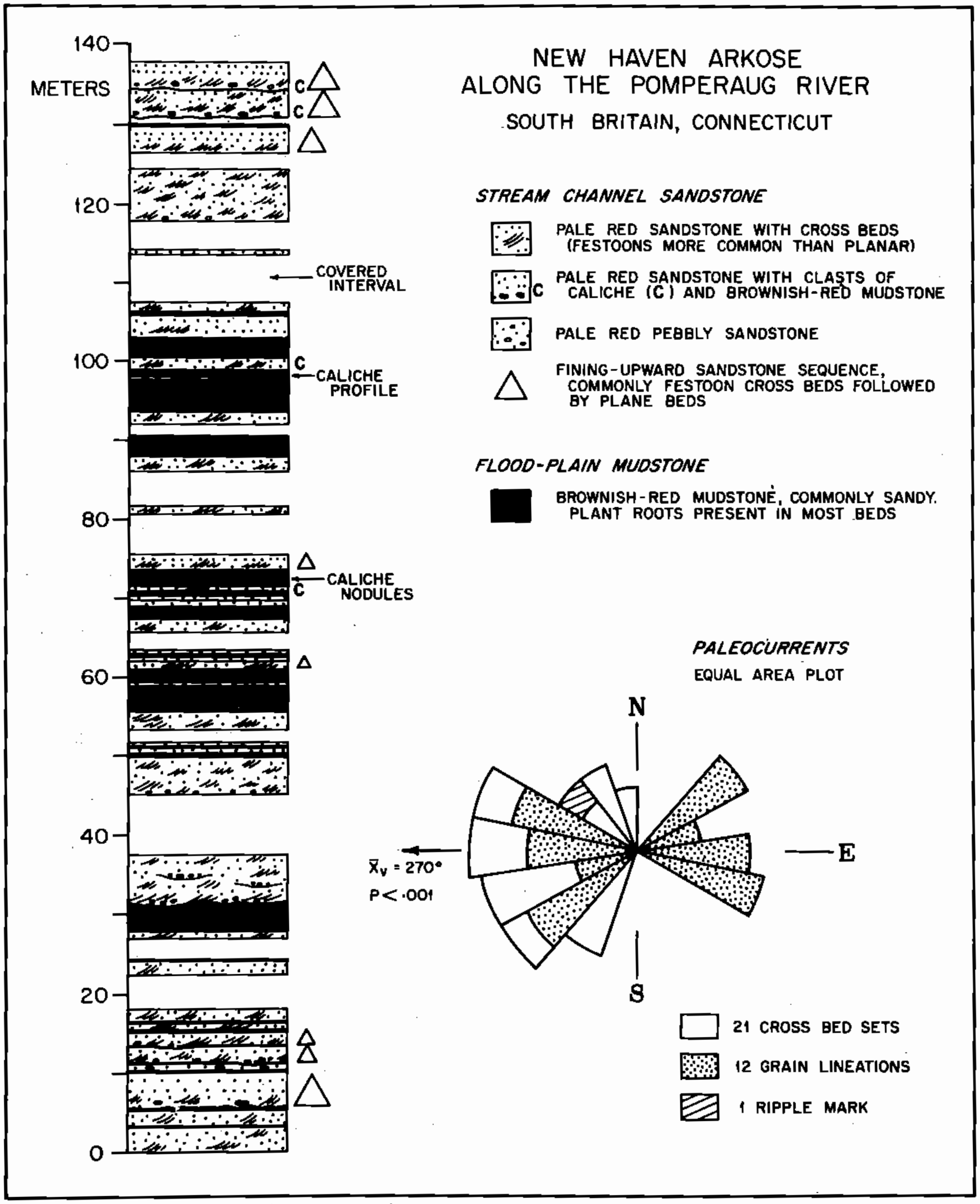

Fig. 18. Measured section and paleocurrents of the New Haven Arkose along the Pomperaug River, South Britain.

the 1500 to 2500 m of Upper Triassic rocks of the New Haven Arkose and the age-equivalent part of the Sugarloaf Arkose. The time span covered by each map is about 12 million years (Cornet, 1977, and 1977, pers. comm.). Each map is constructed using all available outcrop control points.

During Late Triassic time, the rivers flowed down the Eastern Highlands, over alluvial fans banked along the eastern escarpment, and across the valley floor (Figs. 12, 20). The mean paleoflow direction shown by the vector means of the 75 outcrops was southwest towards an azimuth of 241 degrees. The rivers continued beyond the present faulted and eroded western margin of the basin.

North of Hartford, the vector means of the 19 outcrops show that the rivers consistently flowed to the southwest (Fig. 19). The only exception is near Amherst where flow to the northeast may reflect a basement high of metamorphic rocks between the Hartford and Deerfield Basins.

South of Hartford, the pattern of river flow is more complex. The vector means for 47 outcrops show river flow to the southwest, west, and northwest, but nine outcrops demonstrate paleoflow to the east and southeast. The variability in the vector means of the outcrops evidently records the back and forth shifting of the braided-river belts in response to alluviation on the gradually subsiding valley floor. Especially interesting are four outcrops near New Haven where the rivers flowed southeast, directly toward the alluvial fans inferred to have been present along the border escarpment only 5 km distant.

An eastern source of the detritus is evident in the westward decrease in mean size of the five largest igneous and metamorphic clasts at each outcrop (Fig. 22). The values progressively decline from 40 cm to less than 10 cm proceeding away from the border fault. The four outcrops in

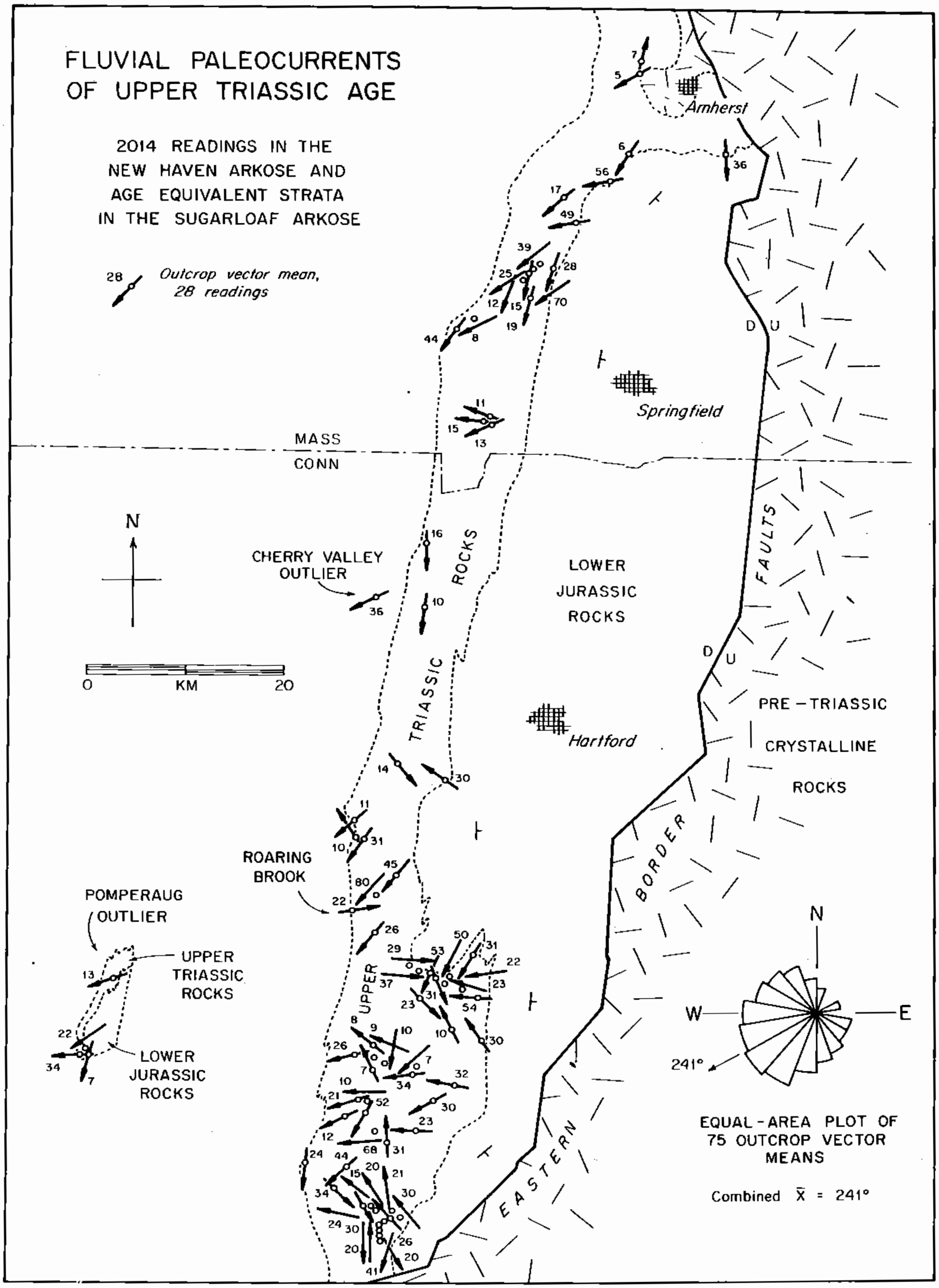

Fig. 19. Fluvial paleocurrents of Late Triassic age in the Hartford Basin and Pomperaug Outlier. Massachusetts data in part from Franz (1977, pers. comm.).

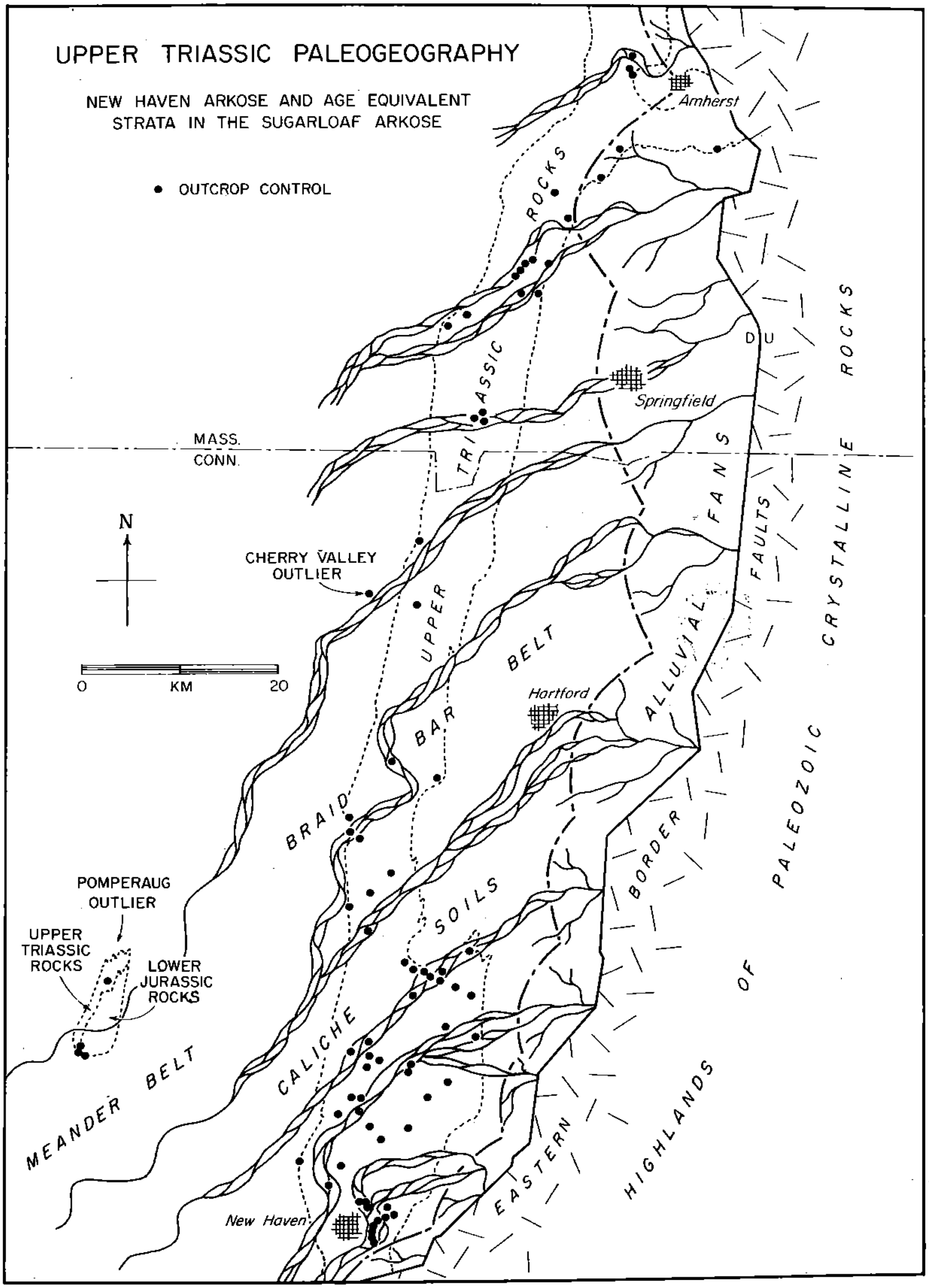

Fig. 20. River morphology and dispersal patterns generalized for Late Triassic time in the Hartford Basin and Pomperaug Outlier.

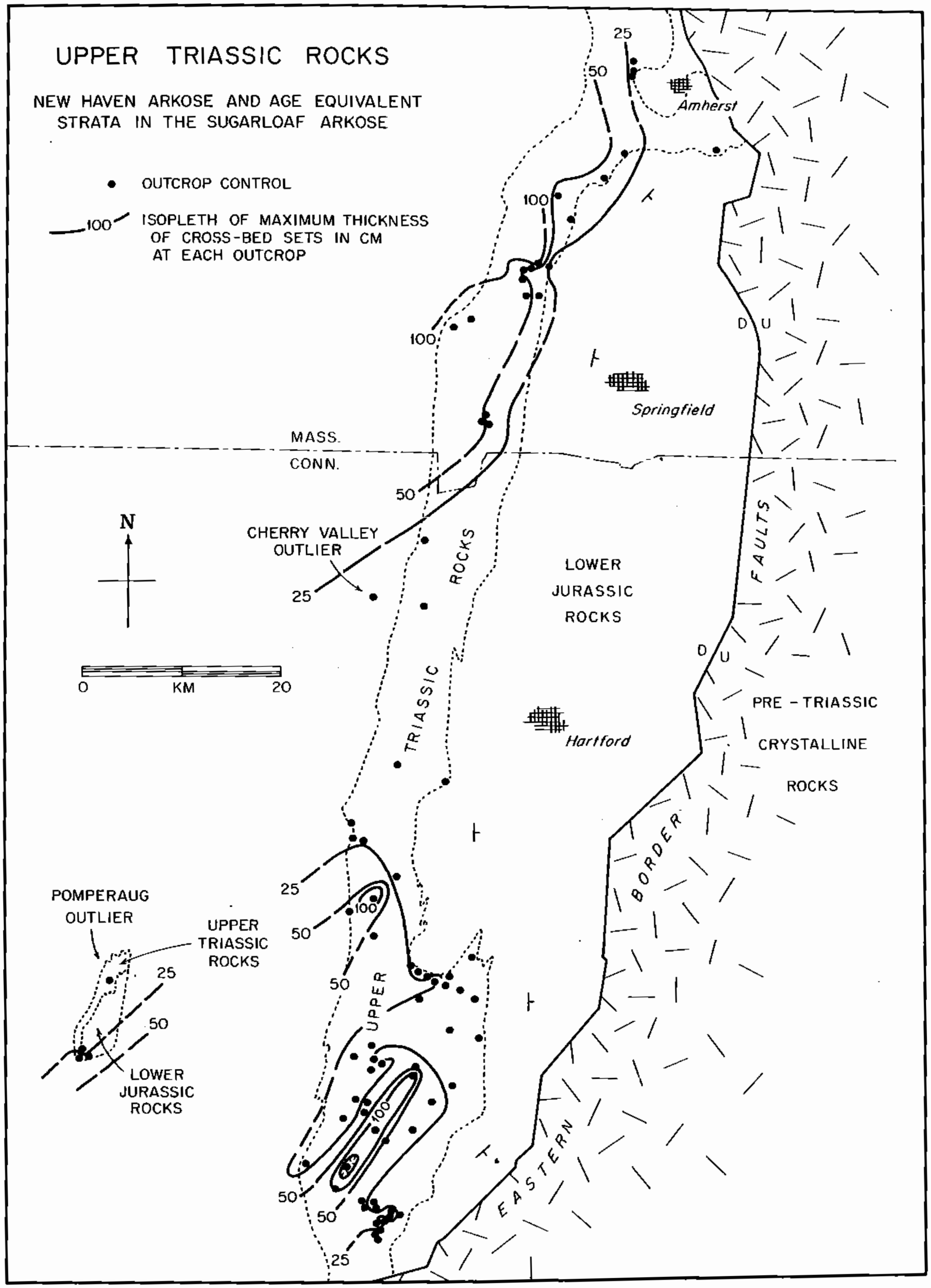

Fig. 21. Maximum thickness of cross-bed sets contoured in cm for outcrops of Upper Triassic fluvial rocks.

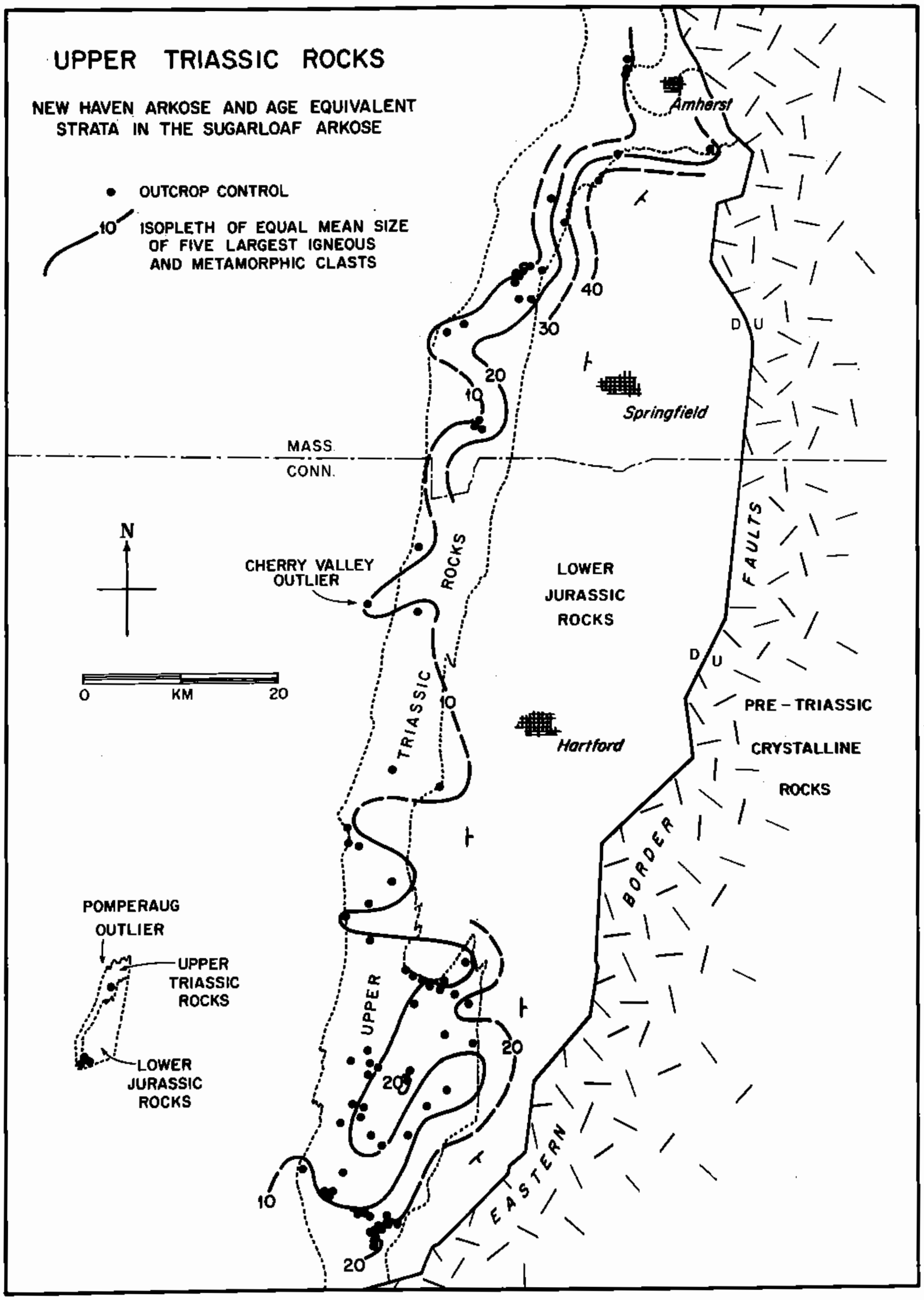

Fig. 22. Mean size of five largest igneous and metamorphic clasts contoured in cm for outcrops of Upper Triassic fluvial rocks.

the Pomperaug Outlier have values less than 10 cm, reinforcing the interpretation that the strata in the outlier were once continuous with those of the Hartford Basin.

The paleogeographic map (Fig. 20) shows the Eastern Highlands of Paleozoic crystalline rocks bordered on the west by normal faults and alluvial fans. The fans are sketched to extend about 15 km into the valley in southern Massachusetts, with fan width decreasing southward to New Haven. This interpretation is based on alluvial-fan sequences exposed along I-91 in the vicinity of Mt. Tom, Massachusetts. Furthermore, the largest crystalline clasts occur at these outcrops (Fig. 22). The fans grade westward into a braided-river belt and then into a meander belt preserved in the Pomperaug Outlier. Caliche paleosols are best developed in southern and central Connecticut, decreasing in number, thickness, and stage of maturity into northern Connecticut and Massachusetts.

STOP 3. TALCOTT BASALT, MERIDEN

The interest in a science such as geology
must consist in the ability
of making dead deposits represent living scenes.

Hugh Miller

Location

This excellent exposure of the Talcott Basalt is located along the west side of the parking lot directly behind the G. Fox department store at the Meriden Square. The store is easily seen from route 66. The Meriden Square is reached by leaving route 66 at exit 6. The limited access portion of route 66 and the Meriden Square are both too new to be shown on the 1967 Meriden quadrangle.

Objectives of Stop 3

At stop 3 in Meriden, the Talcott Basalt is about 67 m thick (Hanshaw, 1968, p. 2), comprising two flows with interbedded volcanic agglomerate (Figs. 23, 24). We shall examine a continuous exposure of 49 m of Talcott Basalt, including almost all of the lower lava sheet, an overlying 20 cm of volcanic agglomerate, and the lower 12 m of the upper flow (Fig. 23). Especially interesting are the pillow structures (Fig. 24) and pipe vesicles in the lower flow.

The thickness and complexity of the Talcott Basalt increase from north to south in Connecticut. In south-central Connecticut in the Gaillard area, the Talcott consists of four flows and interbedded sedimentary rocks with an aggregate thickness of 170-330 m (de Boer, 1968a, p. 1; Sanders, 1970, p. 2). In north-central Connecticut in the vicinity of the Newgate Prison Mine in East Granby, the Talcott is a single flow of 15 to 30 m thickness (Perrin, 1976, p. 10). The Talcott Basalt ends

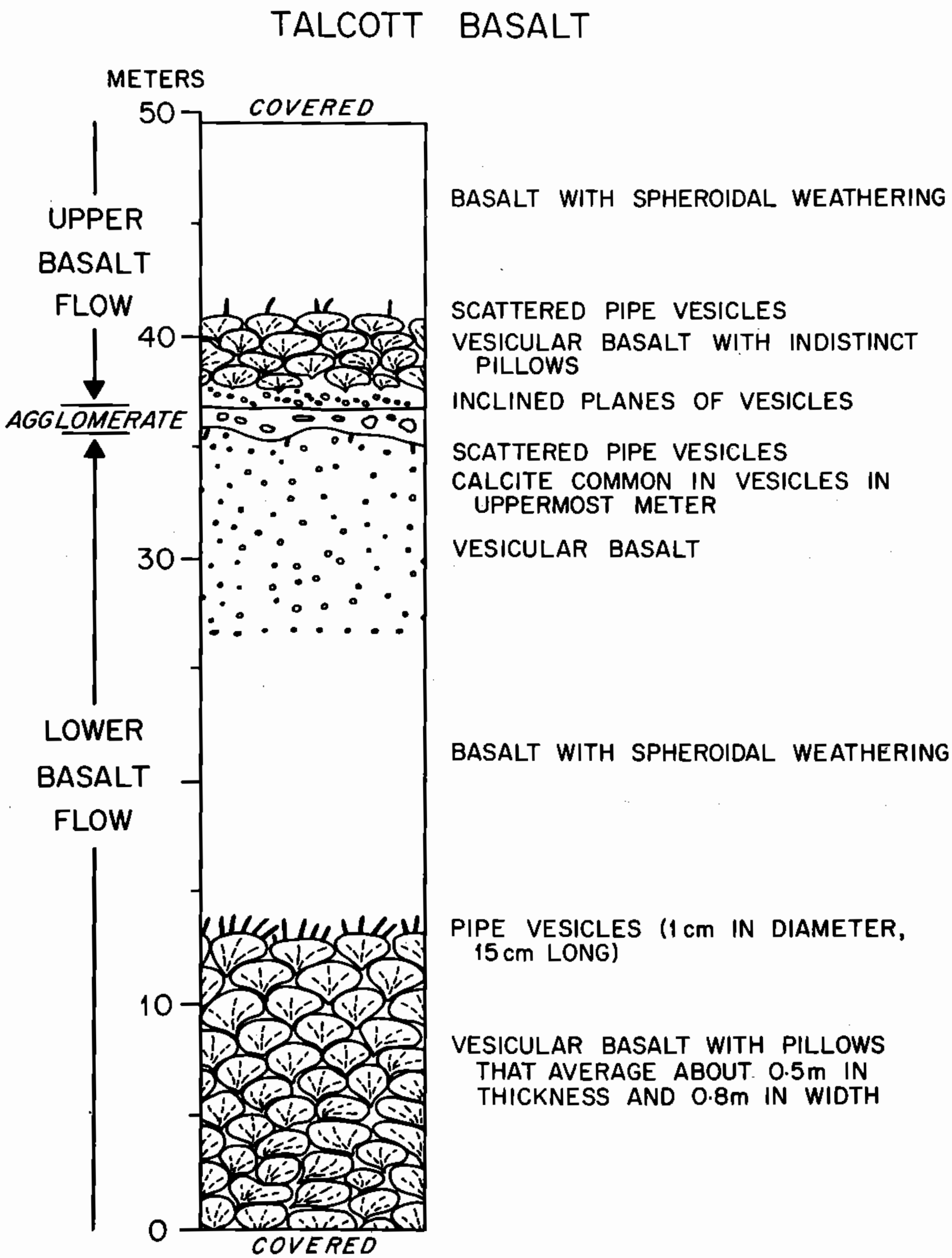

Fig. 23. Measured section of the Talcott Basalt at stop 3, Meriden.

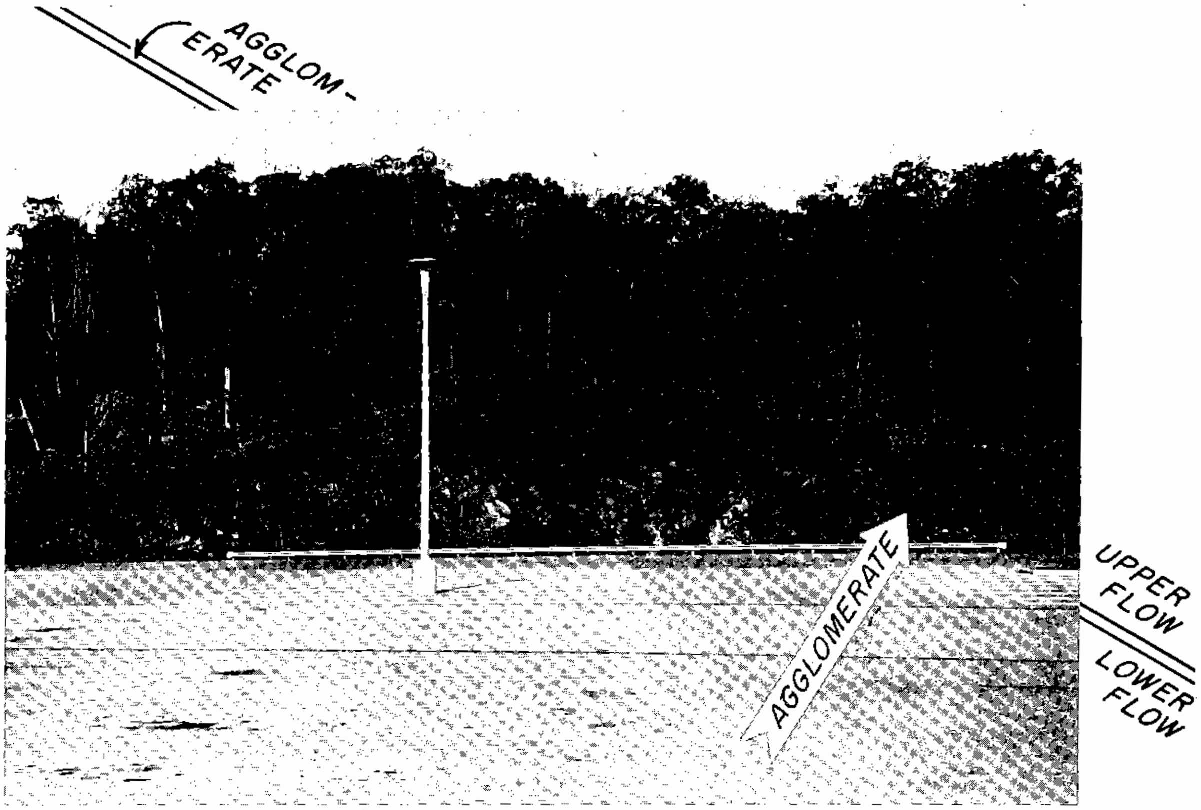

Fig. 24. The Talcott Basalt at stop 3, Meriden.

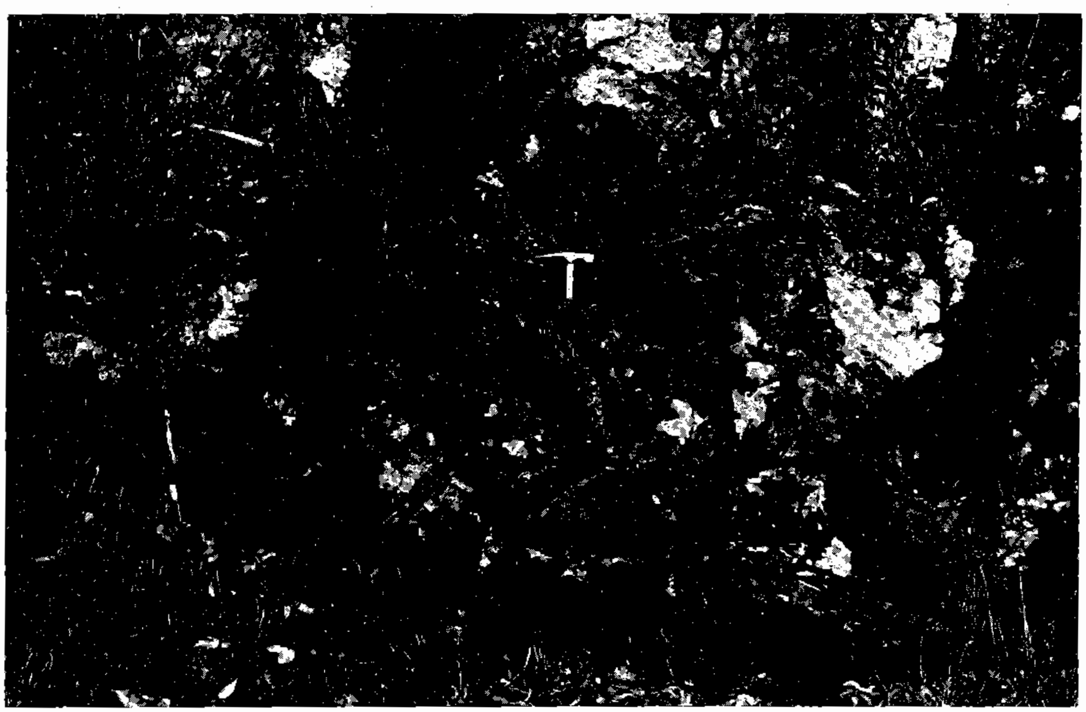

Fig. 25. Cross sections of pillows in the lower lava flow of the Talcott Basalt, stop 3, Meriden.

by erosional truncation or nondeposition 2.2 km north of the Newgate Prison Mine. The internal stratigraphy of the Talcott Basalt is complex. Tongues and sheets of variable thickness and areal extent crystallized from highly fluid lava (Foye, 1924, p. 332; Gray, Norman, 1977, pers. comm.).

Description of the Talcott Basalt at Stop 3

The Talcott flows at Meriden are tholeiitic basalt with labradorite, clinopyroxene, altered glass, opaques, and traces of leucite (Weed, 1976, p. 53). Weed concluded that the basalt is an early differentiate, extruded and cooled rapidly in thin flows so that the leucite was not reabsorbed in the magma.

Little is known of the trace elements in the Talcott Basalt, but 11 random samples from Connecticut average 74 ppm nickel, 86 ppm boron, and 293 ppm chromium (Hanshaw and Barnett, 1960, p. 171).

Pillow structure in the lower flow may be examined on the south end of the outcrop (Fig. 25). The pillows average about 0.5 m in thickness and 0.8 m in width. The outer 2 to 8 cm of the pillows are commonly a former glass rind now devitrified and heavily altered. The pipe vesicles in most pillows radiate outward toward the rind, dying out just beneath the shell.

To form pillows, it appears necessary to have a juxtaposition of hot fluid magma with a cold fluid, generally water or mud (Snyder and Fraser, 1963, p. 2). The pillow basalts at stop 3 evidently formed when the two lavas flowed along the floor of a lake. The process of pillow formation was "budding", where the chilled glassy rind of an advancing tongue is punctured by pressure of the expanding magma within the tongue. Extrusion of magma containing steam bubbles through the rupture leads to the formation of a new pillow. A layer of pillow basalt accumulates by continuation of the process. The pillow horizons in the Talcott Basalt can be seen to pass laterally into basalt without pillows when a number of outcrops are examined.

The Talcott flows commonly contain pillow structure but only poorly developed columnar joints. In contrast, the flows of the Holyoke and Hampden Basalts lack pillows but display prominent columnar joints. The contrast evidently is due to the extrusion of the Talcott flows into a lake, whereas the Holyoke and Hampden flows moved over more or less dry ground.

The vesicular zone at the top of the lower flow is due to degassing of the lava. The vesicles are mostly spherical to subspherical with smooth walls. They are commonly empty, but some contain calcite, drusy quartz, or quartz crystals. The shape of the vesicles and lack of any obvious clinker along the upper surface of the flow hint that the flow was of the pahoehoe type. Aa flows tend to have irregularly shaped vesicles with angular projections (Macdonald, 1972, p. 89).

Paleogeography

The paleoflow direction of some lava flows can be determined because the long axes of pillow-like tongues are parallel to the flow direction and pipe vesicles are tilted so they plunge toward the upflow direction (Waters, 1960, p. 359). In the lower flow at stop 3, the average orientation of the long axes of four pillows is N10°E-S10°W when the tectonic tilt of the sheet is removed. A slight inclination toward the southwest of seven pipe vesicles suggests that the flow travelled northeast. Also interesting is the dip to the northeast of planes of vesicles at the base of the upper flow.

Manspeizer (1969, p. 142) reported that the average direction of flow of the Talcott, Holyoke, and Hampden lava flow units in Connecticut was towards N28°E, based on a combined total of 40 readings of pipe vesicles. Manspeizer suggested that the source of the lava flows lay to the southwest.

The Talcott Basalt was extruded onto an alluvial plain recorded by the braid-bar sandstone and overbank red mudstone of the uppermost beds of the New Haven Arkose. The average direction of river flow was southerly in the Meriden area, as will be noted in the discussion at stop 5. The Talcott Basalt rests on these fluvial strata without evidence of lacustrine deposits. Perhaps the initial volcanism was some distance southwest of Meriden, damming the drainage network. Water would back up on the low-gradient alluvial plain to create a lake of appreciable size, but so new that it lacks a stratigraphic record. As subsequent lava flows reached Meriden, they would flow into the lake with extensive development of pillow structure.

A lake is recorded by the lacustrine gray mudstone and sandstone that rest directly on the surface of the upper flow along the east side of the northbound lane of I-91 at exit 19 in the eastern part of Meriden. The basal 1.5 m of the lacustrine sequence is exposed, with the younger portion covered. The lake evidently formed on the nearly level upper surface of the lava.

So far it has proved impossible to correlate the Mesozoic extrusive basalts with intrusive basalts by petrographic, geochemical or radiometric methods. Using the azimuth and angle of inclination of paleomagnetic pole vectors, de Boer (1968b, p. 618) grouped the rocks into four volcanic events. From oldest to youngest they are (1) the

Talcott Event; (2) the Holyoke Event, comprising the Holyoke Basalt and the Mount Carmel, West Rock, and Barndoor sills; (3) the Hampden Event, composed of the Hampden Basalt and Cheshire dikes; and (4) the Higganum Event, incorporating the Higganum, Bridgeport and Foxon-Fair Haven dike systems.

Based on his gravity map of southern Connecticut, Chang (1968, p. 81) postulated the existence of a buried "Cheshire mafic body." His gravity models are compatible with an interpretation that the top of the inferred mafic body is at a depth of 3.2 km. The width of the mafic body is about 300 m and the length more than 10 km. Chang (1968, p. 92) associates emplacement of the "Cheshire mafic body" with extrusion of the Hampden lava flows.

STOP 4. EAST PEAK OF THE HANGING HILLS OF MERIDEN

... and some rin up hill and down dale,
knapping the chucky stanes to pieces wi' hammers,
like sae money road-makers run daft--
they say it is to see how the world was made!

St. Ronan's Well — *Sir Walter Scott*

Location

We are at the stone tower overlook on East Peak of the Hanging Hills of Meriden. To reach the stone tower, turn north from West Main Street into Hubbard Park on the west side of Meriden. This is a particularly blind corner coming from the east. Proceed straight north 1.3 miles (2.1 km) on Reservoir Road, through the park and up the hill past the Talcott Basalt to Lake Merimere. Travel along the east side of the lake to the north end, turn left, and follow the narrow, poorly paved Percival Park Road up the dip slope of Holyoke Basalt. The irregularity of the upper surface is probably the result of numerous small faults. After 1.5 miles (2.4 km) take the left fork to the parking area below the stone tower on East Peak. The right fork leads to the steel fire tower and the television installation on East Peak.

Objective of Stop 4

The view from the stone tower provides a fine panorama of the Connecticut Valley (Fig. 26).

Geology Seen from the Stone Tower

This description of the topography and geology visible from the stone tower is adapted in part from the excellent guidebook description of Rodgers (in de Boer, 1968a, p. 7-9). Additional extensive remarks on the view are contained in the guidebook of Rice and Foye (1927, p. 113-120).

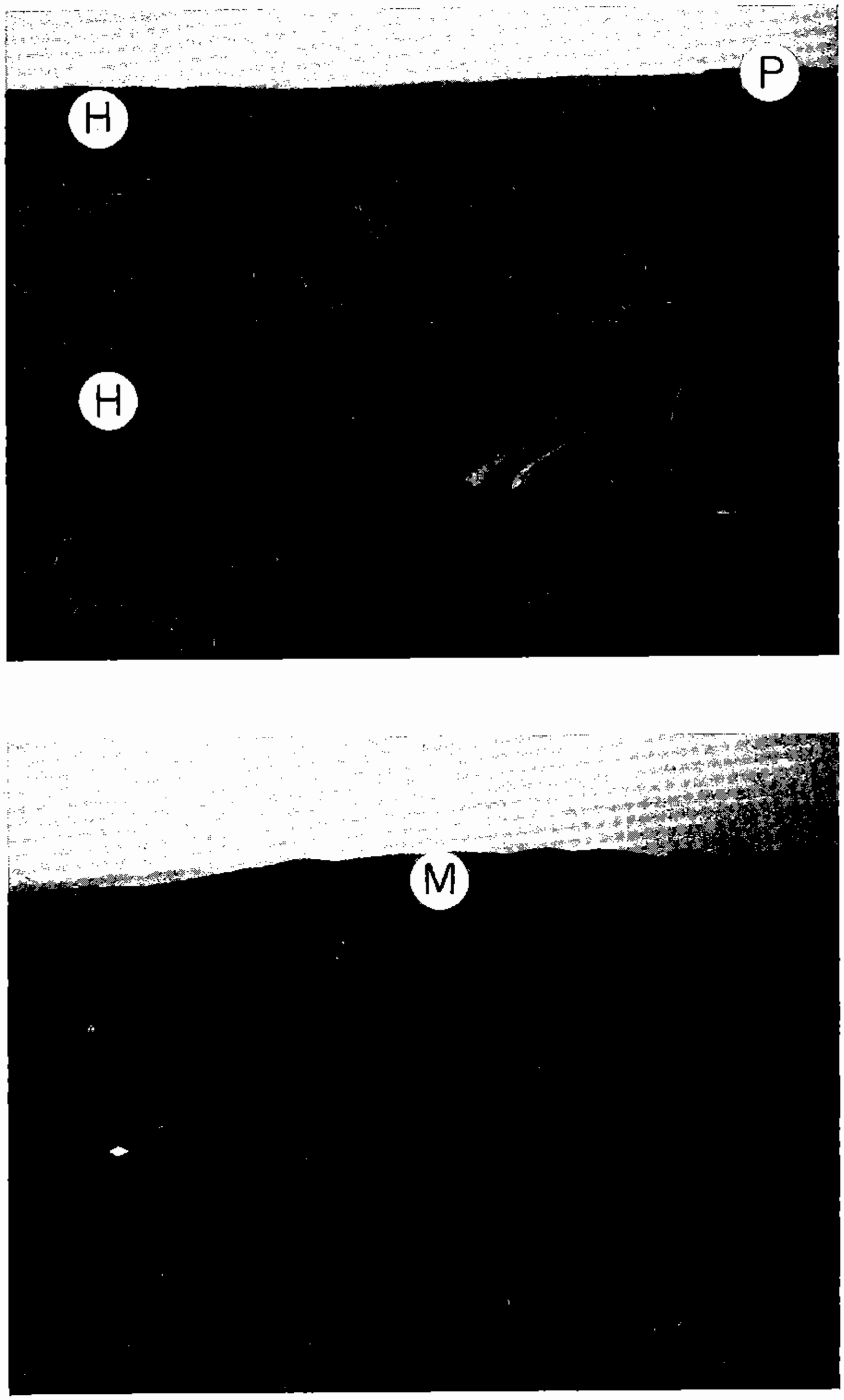

Fig. 26. Views from the stone tower on East Peak of the Hanging Hills of Meriden. The upper photograph looks east along route 66 toward the Holyoke Basalt (H) and Paleozoic crystalline rocks (P). The lower photograph looks south toward the Mount Carmel intrusive basalt (M), "The Sleeping Giant".

The stone tower is built on the upper surface of the Holyoke Basalt at an elevation of about 950 feet (285 m). The Holyoke Basalt has two sets of joints. The older set is columnar with narrow silicified selvages that weather in low relief above the basalt. The younger joints cut the columnar joints. Rodgers (1968, p. 7) noted that large boulders of Holyoke basalt with this distinctive joint pattern occur over the countryside from here to Long Island Sound. A group of these boulders forms the "Judges Cave" at West Rock in New Haven.

Looking east, divided Route 66 passes through Meriden (Fig. 26). The Talcott Basalt forms the bench below the Holyoke Basalt and north of Route 66. The area south of Route 66 is underlain by the New Haven Arkose.

Here in the Hanging Hills, the Holyoke Basalt and the underlying strata strike nearly east-west and dip gently north in contrast to the normal north-south strike and moderate east dip in most of the Connecticut Valley. The change in strike is associated with a zone of many northeast-southwest trending faults. Especially prominent from the stone tower is the 12.8 km of apparent offset in the Holyoke Basalt ridge caused by the largest of the faults. The offset is clearly visible looking east.

Strike-slip motions along these faults are suggested by the presence of nearly horizontal slickensides on 95 percent of the exposed minor fault surfaces (Wise et al., 1975, p. 45). Nevertheless, the dominant motions probably were dip-slip in that no significant strike-slip displacements are evident in the structures bordering the basin on the northeast or southwest projections of these fault zones (Wise, D.U., 1977, pers. comm.). Wise suggests that the rocks ground together along the fault planes during a late period of subordinate strike-slip movement to produce the bulk of the slickensides. In contrast, the dominant dip-slip displacements may have occurred earlier in response to extensional forces which left little record of slickensiding because of the reduced normal stress across the fault planes.

Lake Merimere is directly below us to the east, lying along the trace of a fault between East Peak and South Mountain. On the far horizon to the east are the Eastern Highlands, metamorphic and igneous rocks separated from the rift valley by the eastern border fault. This area was the source of the Triassic-Jurassic detritus. Also due east of us, one can make out the break in the highlands where the Connecticut River leaves the valley to flow through the highlands to Long Island Sound.

The hill out in the valley just south-southwest of us is Mount Carmel, known as "The Sleeping Giant." He sleeps with his head to the west and feet to the east. Mount Carmel is a basalt sill, or perhaps a stock, higher in the New Haven Arkose than the West Rock sill. The Mount Carmel intrusion may lie nearly above a feeder dike in the basement of the rift valley (de Boer, 1968a, p. 4; Chang, 1968, p. 77).

Looking more to the southwest, the north-south ridge that rises in elevation to the south is the West Rock sill of basalt intruded into the New Haven Arkose. Behind the sill are the Western Highlands composed of metamorphic rocks of Lower Paleozoic age. The western margin of the Connecticut Valley from the latitude of Meriden south to New Haven is mostly an unconformity, but locally is faulted (Wheeler, 1937, p. 14-24). Northward, the border is mostly faulted.

The Fall Zone in southern Connecticut is a belt of country where the elevations of the hills decline steadily southward, beginning at the Hanging Hills here at stop 4 and proceeding past the hills held up by the Holyoke Basalt and the Mount Carmel and West Rock sills. The former surface indicated by the hill summits reaches sea level around New Haven Harbor. The southerly slope of the former surface is about 8 m per km. In contrast, the hill summits north of the Hanging Hills do not reach 300 m until Mt. Tom and the Holyoke Range in Massachusetts. The slope from West Peak in the Hanging Hills to Mt. Tom is less than 1 m per km. Even the highest hills north of the hanging hills are eroded to heights well below those of the Fall Zone. The Fall Zone is continuous with the erosional surface, cut mostly on Paleozoic metamorphic rocks, that underlies the Cretaceous strata on Long Island.

Cretaceous strata evidently reached as far north as Meriden. Flint (1963, p. 695) believed that the Connecticut River leaves the Connecticut Valley at the latitude of Meriden because it occupies a channel in crystalline rock that existed prior to the inferred deposition of marine beds over the Fall Zone beginning in Late Cretaceous time. Flint noted that the Fall Zone surface under Long Island has valleys with more than 100 m of relief. The alternative possibility, less favored by Flint, is that the southeast flow of the lower Connecticut River through the crystalline Eastern Highlands is due to superposition from above the coastal plain sequence.

STOP 5. NEW HAVEN ARKOSE, MERIDEN

There is nothing constant in the Universe,
All ebb and flow, and every shape that's born
Bears in its womb the seeds of change.

Ovid

Location

This outcrop of the New Haven Arkose is at the west edge of Meriden along limited access route 66 at the Meriden-Southington corporate boundary. The outcrop is immediately east of the change from limited access to two-way traffic on route 66. The 25-m section was measured along the roadcut on the north side of route 66. The roadcut and limited route 66 are not on the Meriden topographic quadrangle or the USGS bedrock map GQ-738.

Objectives of Stop 5

At stop 5 we shall see a sequence of river-channel sandstone and overbank red mudstone typical of the upper part of the New Haven Arkose in central Connecticut (Figs. 27, 28, 29). The roadcut is oriented almost parallel to the southeast-flowing paleocurrents, providing a longitudinal cross section of the alluvial plain. The primary sedimentary structures and stratigraphic sequence suggest that the rivers were braided with low-relief linguoid bars. Incipient caliche paleosols are developed in the upper portions of some of the floodplain mudstones. The paleosols imply that the rate of sedimentation was slow.

Description of Stop 5

The ratio of channel sandstone to red mudstone is high, about 3 to 1. The channel sandstones range up to about 3 m in thickness and are separated by red mudstone (Fig. 27). Each channel sandstone is a compound body made of sandstone layers mostly 10 to 30 cm thick, separated by scour

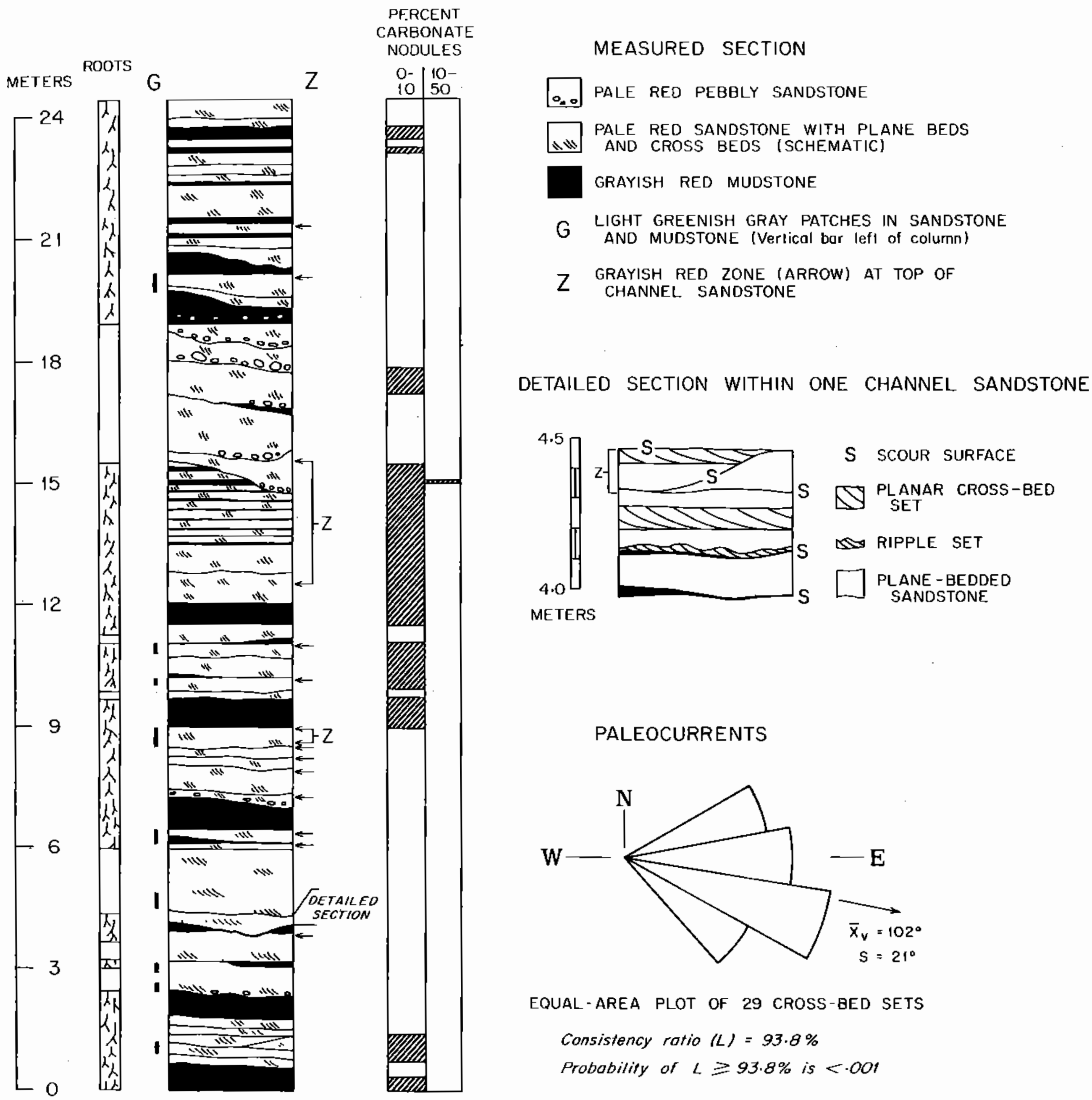

Fig. 27. Measured section and paleocurrents for the New Haven Arkose at stop 5, Meriden.

Fig. 28. River-channel sandstone and floodplain red mudstone in the New Haven Arkose, stop 5. The base of the prominent light-colored channel sandstone is 15 m in the measured section; the sandstone body thins rapidly to the left in the photograph.

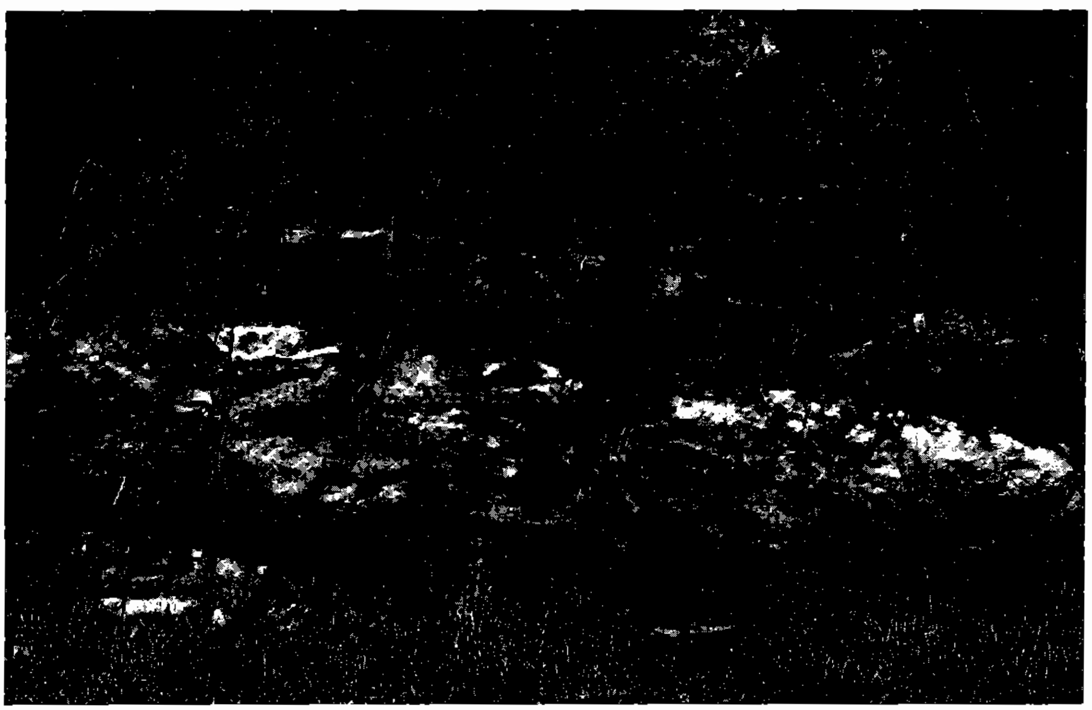

Fig. 29. Fluvial sequence from 3-11 m above the base of the section of New Haven Arkose, stop 5. Top of hammer (circle) is at 4 m.

surfaces or thin, eroded layers of red mudstone. The sandstone layers in turn are constructed of plane beds, planar and festoon cross-beds, and ripple cross-lamination in no obvious preferred vertical sequence. The details of a typical channel sandstone are sketched on Figure 27.

The rivers consistently flowed to the southeast. The paleocurrent azimuths of 29 cross-bed sets all lie in the northeast and southeast quadrants (Fig. 27). The standard deviation of the azimuths is only 21° and the consistency ratio is 94 percent. The cross-bed sets vary from 7 to 28 cm in thickness (Fig. 9). Planar cross-beds are more abundant than festoon.

Root casts occur in 73 percent of the section, commonly densely packed in red mudstone, but also present in channel sandstone (Fig. 27). Roots have nearly destroyed the lamination of some of the sandstones. At 20 m in the section, root casts vertically penetrate sandstone for 70 cm, suggesting that at times the root systems were extensive.

Also common in the red mudstone are burrows of the trace fossil _Scoyenia_. These tubes, about 1 cm in diameter, comprise tunnel networks that penetrated up to 50 cm below the sediment surface. The burrows evidently were constructed by fresh water crayfish, based on analogy with _Scoyenia_ burrows associated with fossils of the crayfish _Clytiopsis_ sp. in Upper Triassic red mudstone of the Durham Basin (Olsen, _in_ Bain and Harvey, 1977, p. 60). The life cycle of modern crayfish includes eggs that require the burrows to be continuously filled with aerobic water, either standing water surface water or ground water.

Irregularly shaped patches of maroon and green are present in the lower and upper portions of the pale-red channel sandstones. The maroon color seems to be due to minute organic particles. The green evidently resulted from removal of iron in ferrous iron-organic complexes in reducing water. The maroon and green colors may have their origin in plant debris that accumulated along the margins of the streams.

Small carbonate nodules and rhizoconcretions occur in 33 percent of the column in both mudstone and sandstone, and are commonly accompanied by root casts (Fig. 27). We interpret the carbonate as paleosol caliche that formed on floodplains and channels of the alluvial plain when the local drainage network was inactive or delivering only small amounts of sediment. Although widespread, the carbonate is less than 10 percent of the volume of the mudstone and sandstone except for a more mature caliche paleosol at 15 m in the measured section. The original calcite of the nodules is extensively altered to dolomite. This contrasts to the calcite nodules at stop 2 and to the calcite caliche generally present in the New Haven Arkose. The dolomitization may be hydrothermal as suggested by green patches with copper, uranium and dolomite mineralization in sandstones of the New Haven Arkose to the east along route 66.

Clasts of red mudstone, some containing carbonate nodules, are present in the lower portions of the channel sandstones above the basal scour surfaces. As river channels shifted laterally, the semilithified mud banks with paleosol caliche were undercut and collapsed. Clasts of caliche limestone near the base of a prominent channel sandstone at 15 m in the section do not match any rocks in the measured column. The limestone fragments were transported from nearby areas with thicker, more mature caliche paleosols.

Interpretation of the River-channel Sandstones

Stop 5 is the record of river-channel sand and overbank mud. Bed relief in the channels was relatively low, as shown by the absence of cross-bed sets greater than 0.3 m that could be interpreted as slip-off faces of large bars. We suggest that the sandstone layers and scour surfaces that comprise each channel sandstone are generated by interaction of linguoid bars with fluctuating stages of the river. Seasonal, semiarid precipitation is implied by the caliche paleosols. During the infrequent flood stages, more or less flattopped linguoid bars of lobate shape could build downstream, generating planar cross-beds on the avalanche face at the front of each bar and mostly plane beds on the upstream part (Smith, 1971, p. 3410 and 1972, p. 625; Collinson, 1970, p. 43). At high river stage, planar and festoon cross-beds form by downstream migration of dunes in the lower flow regime. The dunes occur between bars along the channel floor, and to a lesser extent over the deeper upstream ends of the bars. As the river stage falls, the bars are dissected, producing a transient braid pattern in the channel sand at low water. These scour surfaces are present in a few beds at stop 5, but are more common at nearby outcrops along route 66.

Stop 2 in the lower part of the New Haven Arkose is the record of a braided river with a gravel-sand bed load and high-relief bars, many probably longitudinal in plan view. In contrast, at stop 5 in the upper part of the New Haven Arkose, the braided rivers had a sand bed load with probable linguoid bars. The two stops are about the same distance from the eastern border faults so that distance of transport seems unlikely as the cause of the contrast.

The change in type of braided river perhaps reflects progressive lowering of the valley gradient during alluviation, combined with less relief in the eastern highlands. An additional contributing factor may be the trend to higher annual precipitation from Late Triassic to Early Jurassic time. Higher precipitation in the Early Jurassic is implied by the presence in the Shuttle Meadow Formation of perennial lake sequences together with the absence of paleosol caliche profiles. More precipitation is also suggested by the fact that the caliche profiles are much more mature at stop 2 in the lower part of the New Haven Arkose than at stop 5.

STOP 6. SHUTTLE MEADOW FORMATION, PLAINVILLE

Out of the earth to rest or range
Perpetual in perpetual change,
The unknown passing through the strange.

The Passing Strange   *John Masefield*

Location

Stop 6 is the upper 62 m of the Shuttle Meadow Formation and the contact with the Holyoke Basalt. The outcrop is reached by leaving I-84 at the route 72 interchange in Plainville, Connecticut, 1 mile (1.6 km) west of the New Britain corporate boundary. This is exit 34 going west on I-84, but exit 33 proceeding east of I-84. On leaving I-84, turn north on the connector road to route 72. Turn right (east) on route 72. Immediately on the left (north) side of the road is a quarry in redbeds of the Shuttle Meadow Formation. All these roads are shown on the New Britain topographic quadrangle and GQ-494 (Simpson, 1966). The quarry is used to obtain fill for construction sites. The base of the measured section is on the hillside behind the cafe immediately west of the gas station that abuts the quarry.

Important. Before visiting the quarry, obtain permission from the owners, Paul and James Manafort of Manafort Construction Company, Plainville, Connecticut. Do not climb the rock faces.

Objectives of Stop 6

We shall examine primary sedimentary structures that allow assigning a playa-lake model for the origin of most of the red mudstone, siltstone, and sandstone. There are also a perennial lake sequence of gray to tan mudstone and sandstone at 43 m in the section (Fig. 30) and beds of limestone micrite at 17 m. The hornfels zone in the redbeds below the 75-m Holyoke Basalt is exposed in the upper part of the quarry. In the New Britain quadrangle, the Holyoke Basalt locally consists of as many as nine closely sequential lava flows (Simpson, 1966).

Evidence for Lacustrine Redbeds

Seven converging lines of evidence suggest that most of the red mudstone, siltstone, and sandstone at stop 6 accumulated in shallow, oxidizing lakes, flooded playas or on the surrounding mudflats.

1. The thin beds of red siltstone and very fine sandstone are separated by smooth, laterally extensive bedding planes. The surfaces can be traced for more than 100 m along the access road of the trap rock quarry in the Holyoke Basalt opposite stop 6 on the south side of I-84. Looking south from stop 6 over the distance of 0.6 km, the evenness of the bedding gives a striking impression of the low relief of the floor of the rift valley. Krynine (1950, p. 33) and Sanders (1968, p. 289) cited the fine-grained, well bedded character of some of the redbeds in the Shuttle Meadow Formation as evidence of shallow lakes.

2. The red sandstone and siltstone were not deposited by streams flowing downslope. The paleocurrents, as discussed below, flowed to the northeast, east, and southeast, driven by paleowinds up the slight inclination of the playa or lake floors.

3. Flaser bedding occurs in some beds, as at 35 and 47 m in the section (Figs. 30, 31). The mud flasers fill troughs, lie on ripple foresets and extend as drapes over ripple crests. Flaser bedding is characteristic of intertidal and subtidal zones where currents alternate with slack water, but is also found after flooding on playas and in lakes (Reineck and Wunderlich, 1968, p. 104).

4. From 30 to 35 m in the section, the thin, evenly-bedded red sandstones contain lenses of intraformational clasts of red mudstone with scattered fish bones and <u>Semionotus</u> scales. The interbedded red mudstones contain trackways of the small crocodilian <u>Batrachopus</u>, impressions of the horse-

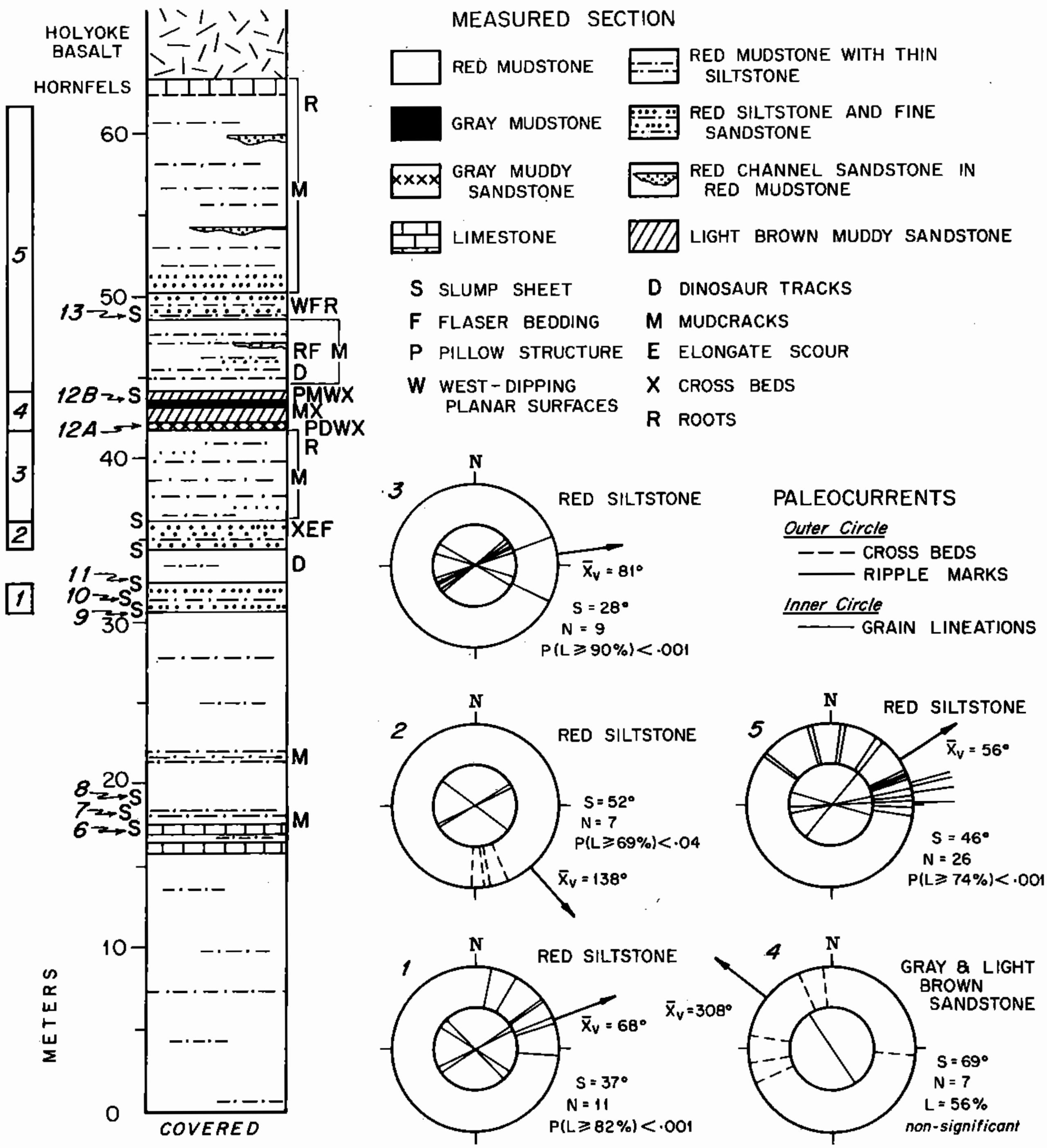

Fig. 30. Measured section of the Shuttle Meadow Formation, stop 6, Plainville. The circular plots show paleocurrent azimuths for the lacustrine redbeds and lacustrine gray to tan sandstone.

tail Equisetum, and Scoyenia burrows constructed by fresh water crayfish.

5. The redbeds are relatively fine grained and occur in association with lacustrine limestone micrite and lacustrine gray mudstone and sandstone. Dewatering of the red very fine sandstone produced synsedimentary sand volcanoes and sandstone dikes. Channel sandstones interbedded with red mudstone in the upper 20 m of the section are less than 20 cm thick. They seem to be watercourses that meandered across the lacustrine mudflats.

6. The red color implies destruction of the organic matter in an oxidizing environment. The former presence of organic material is indicated by root casts at many levels in the redbeds (Fig. 30). Although some of the inferred lakes probably persisted for many seasons, they were of oxidizing shallow depth.

In almost every climate, the surfaces of detrital grains are stained yellow-brown by limonite produced by weathering. These stains are readily removed as ferrous iron-organic complexes in the reducing bottom water of oligomictic perennial lakes. The resulting rocks are gray or black, like the lacustrine sequence at 44 m in the section (Fig. 30). In contrast, the oxidizing environment of the redbeds preserved the limonite surface stains; the limonite dehydrated to hematite during burial diagenesis over millions of years (Van Houten, 1973, p. 51). The rarity of green patches and streaks in the redbeds also indicates the persistence of oxidizing conditions.

Fig. 31. Flaser bedding in lacustrine red sandstone in the Shuttle Meadow Formation, stop 6. The base of the bed contains intraformational clasts of red mudstone.

Fig. 32. Scour surface (head of hammer) cut into lacustrine red sandstone at 35 m in the Shuttle Meadow sequence, stop 6 (Fig. 30).

Fig. 33. "Inclined surfaces" in bed of lacustrine sandstone with ripples and climbing ripples at 49 m in the Shuttle Meadow Formation, stop 6 (Fig. 30). The surfaces dip to the left.

Fig. 31

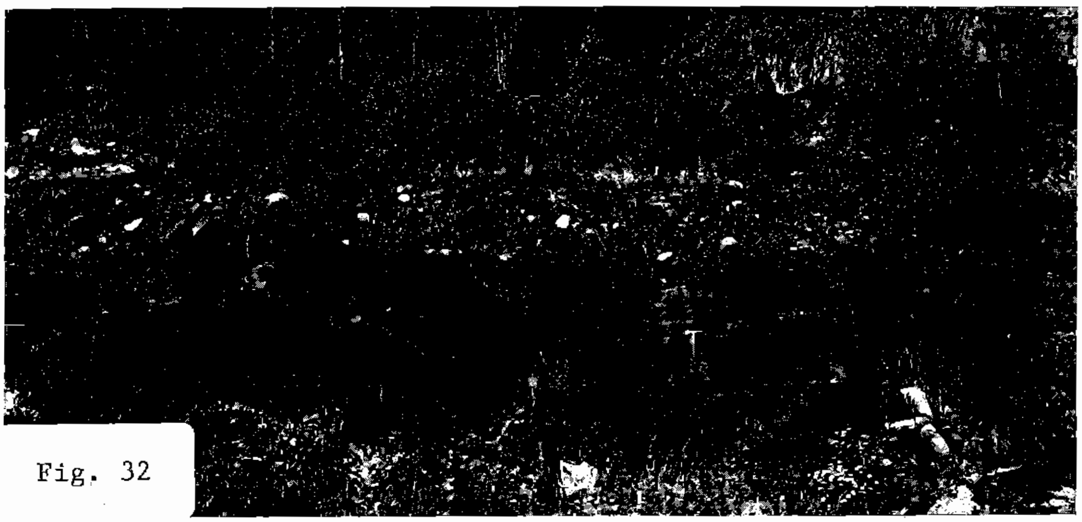

Fig. 32

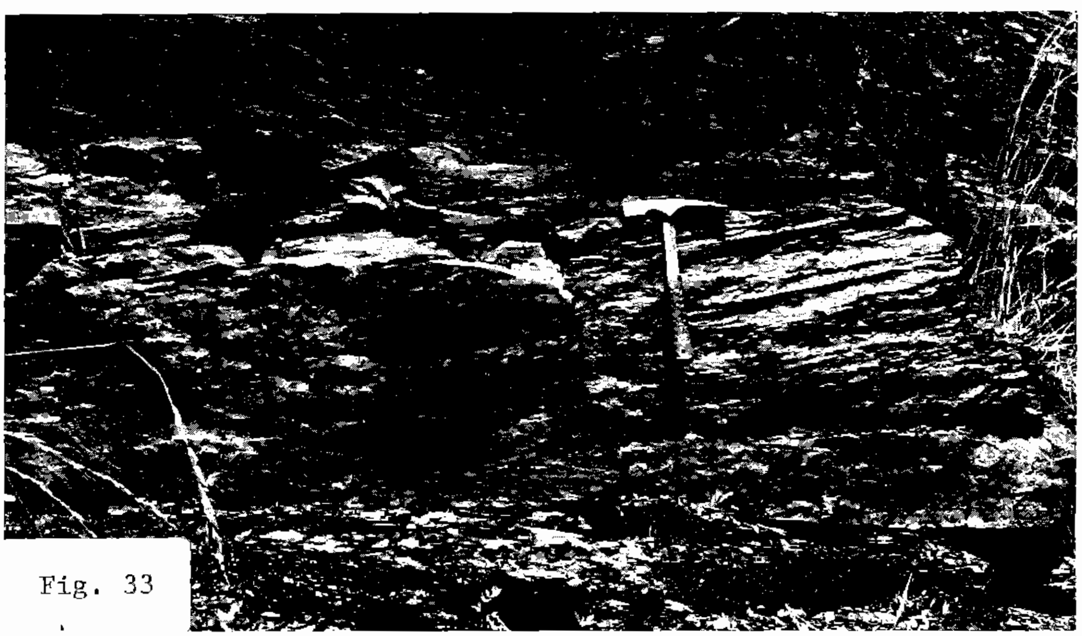

Fig. 33

7. Some of the redbeds are calcareous. Restriction of calcite to thin (5-15 cm) beds of sandstone, siltstone, and mudstone suggests precipitation of calcium carbonate in shallow lakes. In tropical lakes, the combination of heat, intense evaporation, and plant photosynthesis result in elevated surface temperature and increased pH and ion concentrations that favor precipitation of calcite or aragonite. Minor dolomitization during burial diagenesis suggests that the Mg/Ca ratios of the lake waters rose at times to more than 7, but commonly were relatively low (Müller _et al_., 1972, p. 163; Folk and Land, 1975, p. 63).

It is not known whether the lakes were closed or through-flowing with spillways. Probably a playa-lake model is applicable where a lake will persist for many years when precipitation exceeds evaporation during a prolonged wet interval.

Paleoslopes

The paleoslope directions of the floors of some of the lakes can be determined using thin slump sheets in the lacustrine redbeds, gray sandstone, and limestone micrite. The deformation is not tectonic because the slump folds are bevelled by erosion beneath undeformed strata deposited over them.

Hansen (1966) developed the general solution for determining the slip line of a slump sheet. He noted that an outcrop located anywhere in a slump sheet can be used to determine the direction of movement of the slide. The paleoslope direction is determined by plotting the orientation and sense of rotation of the fold axes on a Schmidt net after the tectonic tilt is removed by rotating the beds to a horizontal position. By convention the clockwise, or counter-clockwise, rotation of each fold axis is plotted looking down the plunge of the axis. The sense of rota-

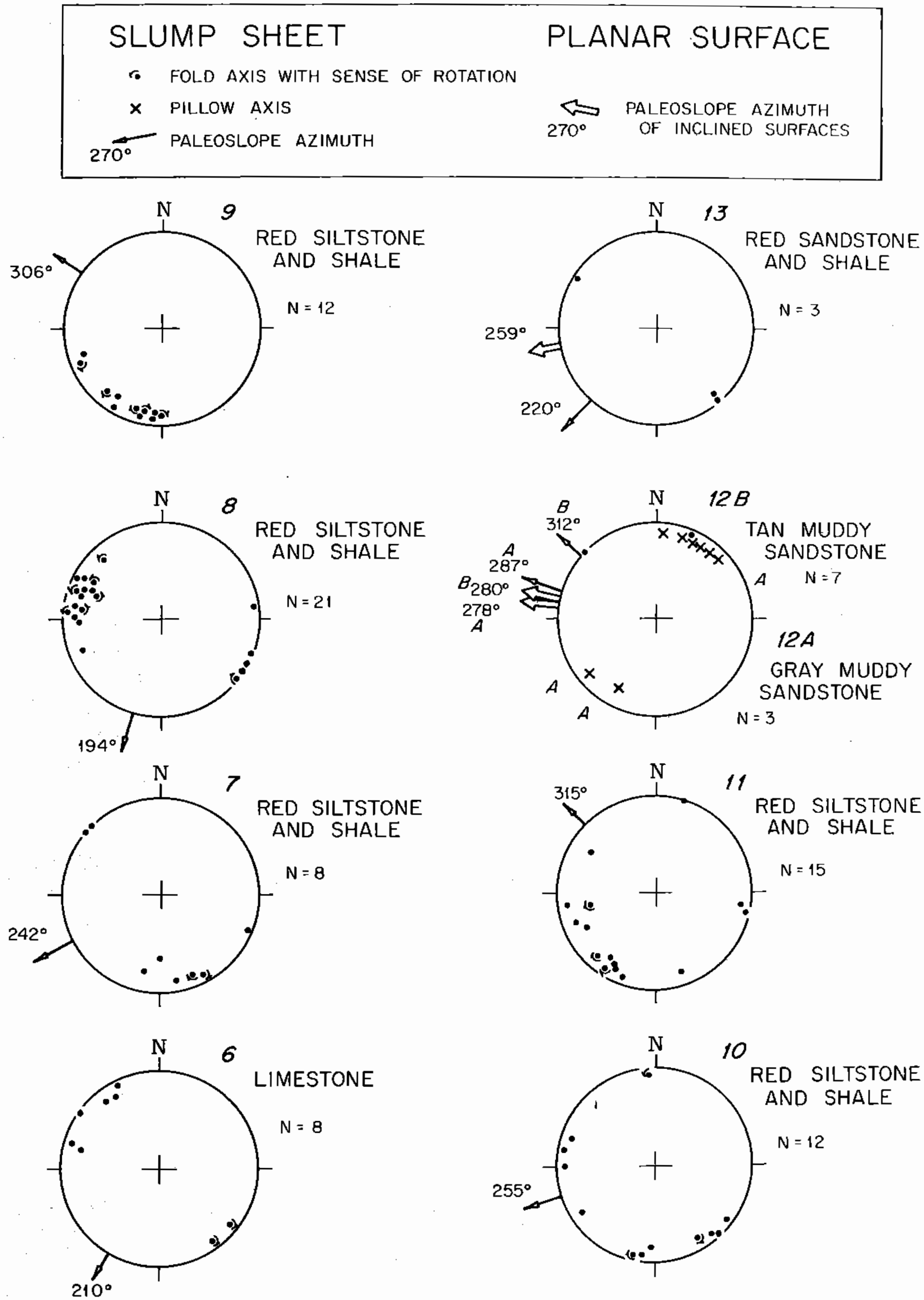

Fig. 34. Lacustrine paleoslope azimuths for redbeds, limestone micrite, and gray sandstone (Fig. 30) in Shuttle Meadow Formation, Plainville. The data at each stratigraphic level are plotted on a Schmidt net with the tectonic tilt removed.

tion of some folds is indeterminant and only the orientation of the axis is plotted. Although there is substantial scatter in the orientation of the axes, the paleoslope direction is determined if the rotation sense of some of the folds can be recorded. The axes form two groups characterized by clockwise and counter-clockwise rotation of the folds. The axes tend to cluster in two groups that lie along the paleocontour direction. The slip line of the slump sheet (paleoslope vector) bifurcates the two groups of axes.

The nine paleoslope azimuths of the floors of the lakes recorded by redbeds, gray to tan sandstone, and limestone micrite form a 105° arc that comprises paleoslopes to the southwest, west, and northwest (Fig. 34). The westerly component to the paleoslope persisted over many thousands of years required for accumulation of the 62-m section and during numerous advances and retreats of the lake strands.

Paleocurrents

The paleocurrent azimuths shown by the ripple marks in the lacustrine redbeds are to the northeast, east, and southeast at the four available levels in the section (Diagrams 1, 2, 3, and 5 on Fig. 30). Each of the four vector means is statistically significant at the 95 percent level when tested by the Rayleigh statistic, the magnitude of the vector mean expressed as a percentage. The distributions are unimodal with variances from 784 to 4624.

We interpret the paleocurrents as due to paleowinds that blew over the shallow lakes. The floors of the lakes sloped to the southwest and northwest (Diagrams 9, 10, 11, and 13 on Fig. 34). The opposing directions of paleocurrents and paleoslopes suggest that as waves approached the shore, wave refraction oriented ripple crests parallel to the paleocontours, a process commonly observed in modern and ancient lakes (Picard and High, 1972, p. 129).

Although wave refraction makes determination of the paleowind directions somewhat approximate, the paleowinds evidently were dominantly from the west in pre-continental drift orientation. These west paleowinds closely parallel paleowinds from the northwest recorded by ripple marks in perennial lake sequences of gray mudstone and sandstone and in lacustrine redbeds of the East Berlin Formation. The problems of regional interpretation of Early Jurassic paleowinds are discussed at stop 7.

Enigmatic Inclined Surfaces and Elongate Scour

Description of Inclined Surfaces

Lacustrine gray to tan sandstone and mudstone (41.8 - 44.1 m). Two laterally extensive beds of laminated sandstone with inclined surfaces occur in the gray to tan sandstone near the top and bottom of the perennial lake sequence (Fig. 30). A 40-cm layer of gray mudstone separates the two beds of sandstone. In the sandstone below the gray mudstone, the inclined surfaces are 30 cm high and the laminae sweep down the surfaces from top to bottom over a horizontal distance of 1.3 m. The strata deposited on the inclined surfaces are alternating laminae of very fine sandstone and muddy sandstone. In places, laminae are thrown into small slumps about 1 cm high with fold axes subparallel to the strike of the inclined surfaces. The horizontal red strata that lie over the ends of the inclined beds are, at different places, plane beds of sandstone, flaser-bedded ripples of fine sandstone, and mudstone. The surfaces slope northwest at an azimuth of 278° with a maximum dip of about 15° when the tectonic tilt is removed (Diagram 12A on Fig. 34).

In the upper sandstone bed, the inclined surfaces are 80 cm high and the inclined laminae extend for a horizontal distance of 3.5 m. The surfaces slope northwest at an azimuth of 280° with a maximum dip of 15°.

In both layers of sandstone, the inclined surfaces slope northwest, subparallel to the slope of the lake floor which was toward a 387° azimuth at 43 m, and 312° azimuth at 44 m, based on the orientation of the axes of pillows and slump folds in sandstone that does not have the inclined surfaces (Diagrams 12A and B on Fig. 34).

Five cross-bed sets of 6 to 16 cm thickness show paleocurrents that flowed southwest and northwest, whereas an 8-cm cross-bed set was deposited by flow to the southeast (Diagram 4 on Fig. 30). These directions are approximately down and up the paleoslope.

Lacustrine redbeds (48.7 - 50.2 m). This interval of redbeds contains a 60-cm sequence of ripples and climbing ripples that accumulated on prominent surfaces that slope down over a 4-m horizontal distance (Fig. 33). Thin films of mudstone with mudcracks occur in the inclined sequence. The strata with inclined surfaces overlie red mudstone and are covered by 90 cm of ripples, climbing ripples and flaser-bedded ripples. The inclined surfaces slope southwest at 259° azimuth with a maximum 10° dip in pre-tilt orientation (Diagram 13 on Fig. 34). Small slump folds in the redbeds show that the floor of the lake also sloped southwest at 220° azimuth (Diagram 13 on Fig. 34).

Paleocurrents for the lacustrine redbeds from 44 to 62 m flowed northeast with a vector mean of 56° (Diagram 5 on Fig. 30).

Description of Elongate Scour

Red sandstone and mudstone (36.2 - 34.5 m). A prominent scour channel, 1 m deep and 9 m wide, is cut into flaser-bedded ripples at 35 m in the section (Figs. 30, 32). The axis of the scour trends about N20°E-S20°W. Lying directly on the base of the scour surface is a 15-cm cross-bed set deposited by paleocurrents that flowed southeast towards a 98° azimuth. The cross-beds are made of alternating laminae of sandstone and muddy sandstone, suggesting repeated episodes of slack water as the cross-beds built southeast. Extending over the top of the channel-fill is a layer of mostly plane-bedded red sandstone.

Interpretation

The three planar surfaces slope northwest and southwest approximately perpendicular to the paleocontours. The elongate scour trends northeast-southwest, subparallel to the paleocontours.

The origin of these features is debatable. Perhaps the planar surfaces are inclined bedding planes cut during storms that removed substantial volumes of mud and sand in shallow water just off the lake strand. Could the elongate scour be a pathway for longshore transfer of water masses pushed against the shallows of the lake? Or, is the elongate scour a river channel cut at a time of low lake level? Might the planar surfaces be bedding planes in an equilibrium profile developed close to the strand line in response to a rapid drop in lake level? Another possibility is that the planar surfaces and elongate scour are the partial record of a ridge and runnel system similar to those observed in some modern lakes and common along mesotidal sandy beaches (Davis et al., 1972). A further suggestion is that the planar surfaces developed on the flanks of near-shore bars.

Let's look at the rocks and discuss your interpretations!

Limestone

Two beds of limestone separated by red mudstone interrupt the red-bed sequence at 17 m (Fig. 30). The 19-cm lower limestone is a sandy, clayey, slightly dolomitic micrite. The calcite crystals average about 4 microns with patchy areas of 4 to 30-micron microspar. Some laminae consist almost entirely of round fragments of micrite, many containing gypsum crystals, cemented by sparry calcite. Abraded lumps of algal tufa are abundant and there are a few ostracod shells and fish bones. There is some authigenic pyrite.

The upper bed, 23 cm thick, is a sandy, clayey micrite, heavily dolomitized. This bed also contains abraded micrite fragments with gypsum crystals. Slump folds at the top of the bed moved down the lake floor to the southwest (Diagram 6 on Fig. 34).

These limestones accumulated in shallow lakes of alkaline, hard water with substantial dissolved Mg^{+2}, Ca^{+2}, HCO_3^-, and SO_4^{-2}. The fragments of gypsum-bearing micrite evidently were transported from mudflats around the lakes. During burial diagenesis, the micrite was partially dolomitized.

STOP 7. EAST BERLIN FORMATION, CROMWELL

The seasons change, the winds they shift and veer ...

Sir William Watson

Location

Sixty-two m of the upper part of the East Berlin Formation, plus the contact with the overlying 50-m Hampden Basalt, are exposed along the excavated but unpaved access roads of the interchange between I-91 and route 9 in Cromwell, Connecticut (Fig. 35). The access roads connect from the west to the southbound lane of I-91. The interchange is shown on the Middletown quadrangle at the center of the north edge of the map and on the Hartford South quadrangle at the center of the south edge. Lehmann (1959) published the bedrock geologic map of the Middletown quadrangle.

Introduction

The East Berlin Formation is about 170 m thick in the Middletown quadrangle (Lehmann, 1959, p. 15). At stop 7, the rock types are 24 percent gray mudstone, black shale, and gray sandstone (perennial lakes), 61 percent red mudstone (floodplains), 8 percent thin, evenly bedded red sandstone and siltstone with abundant ripple marks (shallow oxidized lakes), and 7 percent pale-red channel sandstone (river channels). These

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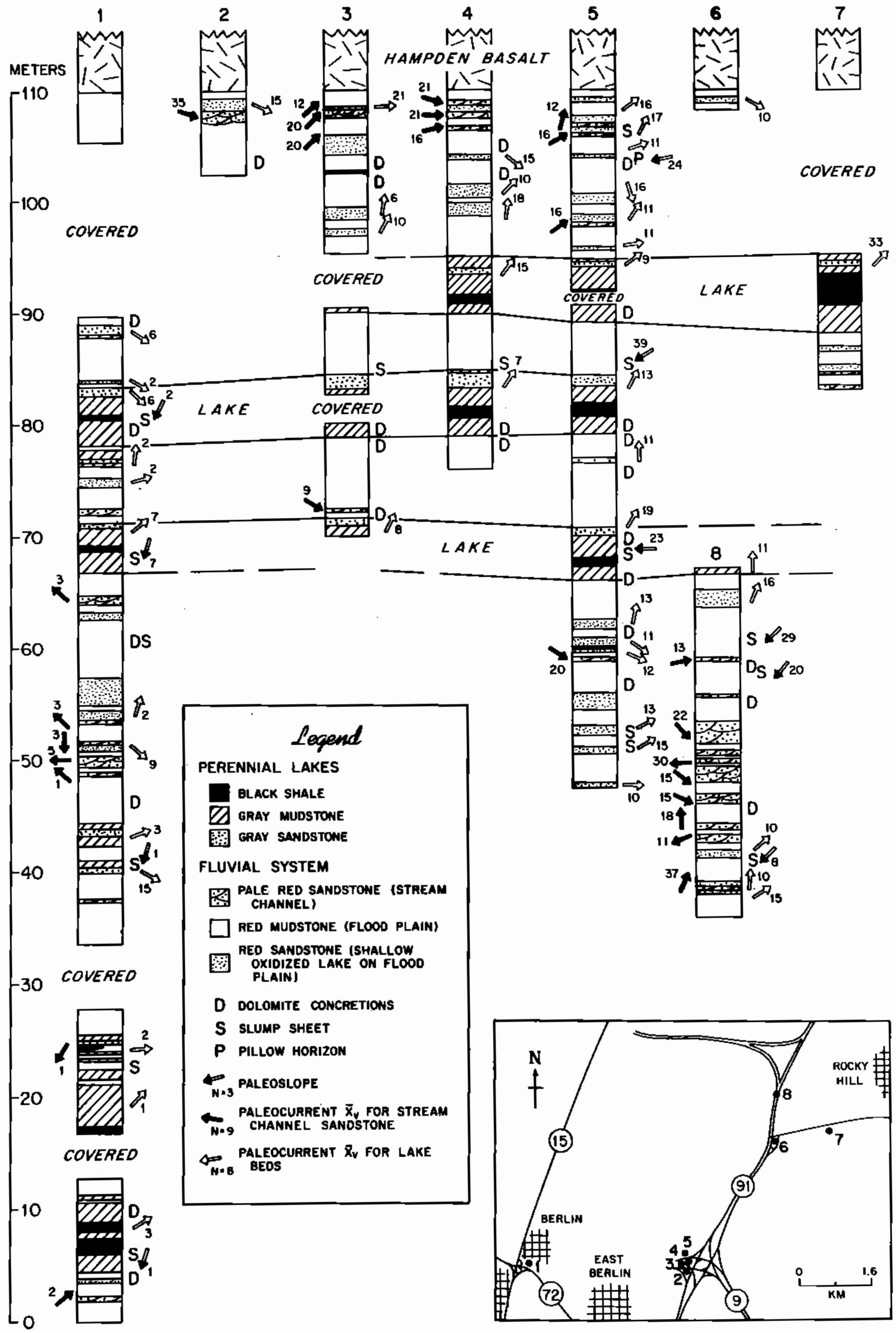
Fig. 35. Stratigraphy, depositional environments, paleoslopes, and paleocurrents for the East Berlin Formation, central Connecticut. Stop 7 comprises sections 2, 3, 4, and 5. Section 1 is the type section, 16.5 km south of the center of Hartford. Section 6 is on I-91 in Rocky Hill. Section 8 is at Dinosaur State Park in Rocky Hill. Paleoslope directions for lacustrine gray-black beds show the slopes of the floors of perennial lakes. The paleoslopes for red mudstone with thin sandstone layers are the slopes of the floodplains on the valley. From Hubert et al., 1976, Fig. 2.



HAMPDEN BASALT
METERS
COVERED
LAKE
Legend
PERENNIAL LAKES
BLACK SHALE
GRAY MUDSTONE
GRAY SANDSTONE
FLUVIAL SYSTEM
PALE RED SANDSTONE (STREAM CHANNEL)
RED MUDSTONE (FLOOD PLAIN)
RED SANDSTONE (SHALLOW OXIDIZED LAKE ON FLOOD PLAIN)
D DOLOMITE CONCRETIONS
S SLUMP SHEET
P PILLOW HORIZON
PALEOSLOPE
PALEOCURRENT $\bar{X}_v$ FOR STREAM CHANNEL SANDSTONE
PALEOCURRENT $\bar{X}_v$ FOR LAKE BEDS
N
ROCKY HILL
BERLIN
EAST BERLIN
KM

Fig. 35

values are the proportions deposited on the valley floor, away from the conglomerate and sandstone of the alluvial fans that existed along the fault-bounded escarpment on the eastern side of the valley.

Objectives of Stop 7

The upper third of the East Berlin Formation is exposed at stop 7. We shall observe three symmetrical cycles of gray mudstone and sandstone-black shale-gray mudstone and sandstone that accumulated in perennial lakes of alkaline, hard water (Figs. 36, 37, 38, 39). Also interesting are lacustrine redbeds and stream-channel sandstone and floodplain red mudstone that separate the lacustrine gray sequences.

Lacustrine Black Shale, Gray Mudstone and Gray Sandstone

Symmetrical Lacustrine Cycles

The black shale and gray mudstone record perennial lakes that existed from time to time in the rift valley (Krynine, 1950, p. 35, 60, 160; Sanders, 1968, p. 295; Klein, 1968, p. C14; Byrnes, 1972, p. 183; Hubert, *et al.*, 1976, p. 1193). At some localities in central Connecticut these beds contain fossil fish, comprising representatives of three groups: advanced chondrostean redfieldiids (to 20 cm long); holostean semionotids (to more than 30 cm); and the coelacanth *Diplurus* (to 69 cm) (McDonald, 1975, p. 100).

The black shale and gray mudstone form symmetrical cycles, mostly 2 to 7 m in thickness (Figs. 35, 36, 37). The center of each cycle is pyritic black shale that accumulated in the deeper, more central parts of a lake. Above and below is gray mudstone with structures indicative of shallower water, including dolomite concretions, ripple marks, mudcracks, and dinosaur footprints. Ferroan dolomite laminae are present in some of the black shale and gray mudstone (Fig. 38). The terrigenous grains in these drab-colored rocks were originally coated with limonite stains which were removed in solution, evidently as organic-ferrous iron complexes.

Thin beds of gray, fine- to very fine-grained sandstone occur in some of the gray mudstones, forming intervals of thin-bedded sandstone and gray mudstone. Most of the sandstone is horizontally laminated, but there are some festoon and planar cross-beds and ripple marks. The sandstones commonly are near the top and bottom of the cycles, implying accumulation in shallow water near the lake shores.

The symmetrical cycles of gray mudstone-black shale-gray mudstone require expansion and contraction of perennial lakes. Precipitation on the average exceeded evaporation as each lake initially formed and continued to expand. The larger lakes covered most of the rift valley, nearly filling the tectonic depression (Fig. 40). Each lake then contracted to complete the cycle.

The Early Jurassic palynoflorule in the East Berlin, Shuttle Meadow, and Portland Formations comprises more than 90 percent Corollina pollen from conifers that lived most abundantly on sandy areas of the alluvial fans and highlands (Cornet and Traverse, 1975, p. 30). The rarity of xeromorphic cuticular adaptions in the flora, which includes large-leaf forms of Clathropteris, plus the many kinds of cryptogams based on spore diversity, suggest to these authors a humid climate with a short dry season. The palynoflorule is found in lacustrine gray mudstone and black shale, recording periods of increased rainfall that coincided with the existence of large perennial lakes.

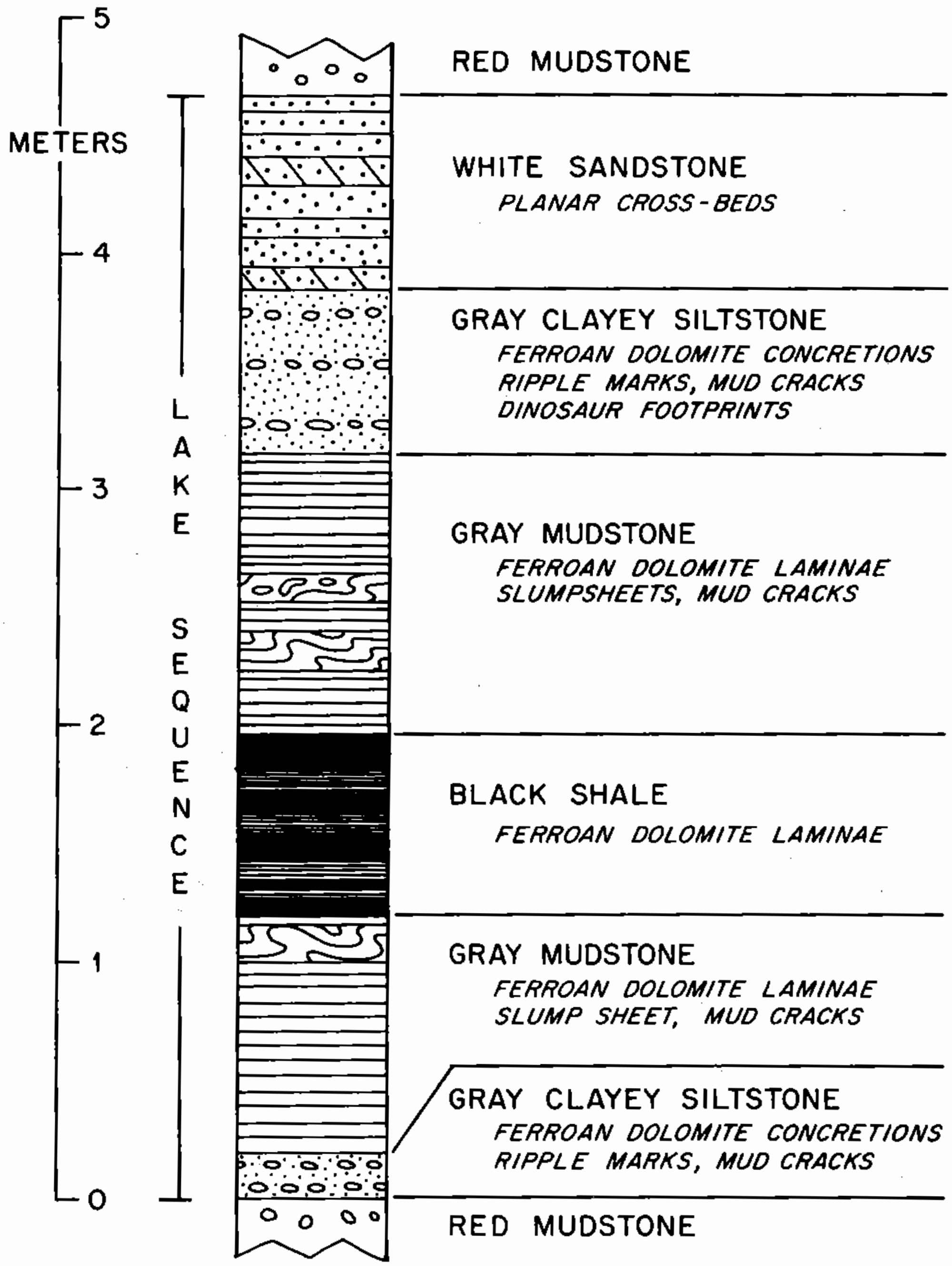


Fig. 36. Perennial lake cycle in the East Berlin Formation, stop 7, Cromwell. Fish scales, spores, and pollen occur in the black shale and gray mudstone. This cycle is 66.3-71.0 m at section 5 (Fig. 35). From Hubert et al., 1976, Fig. 7.

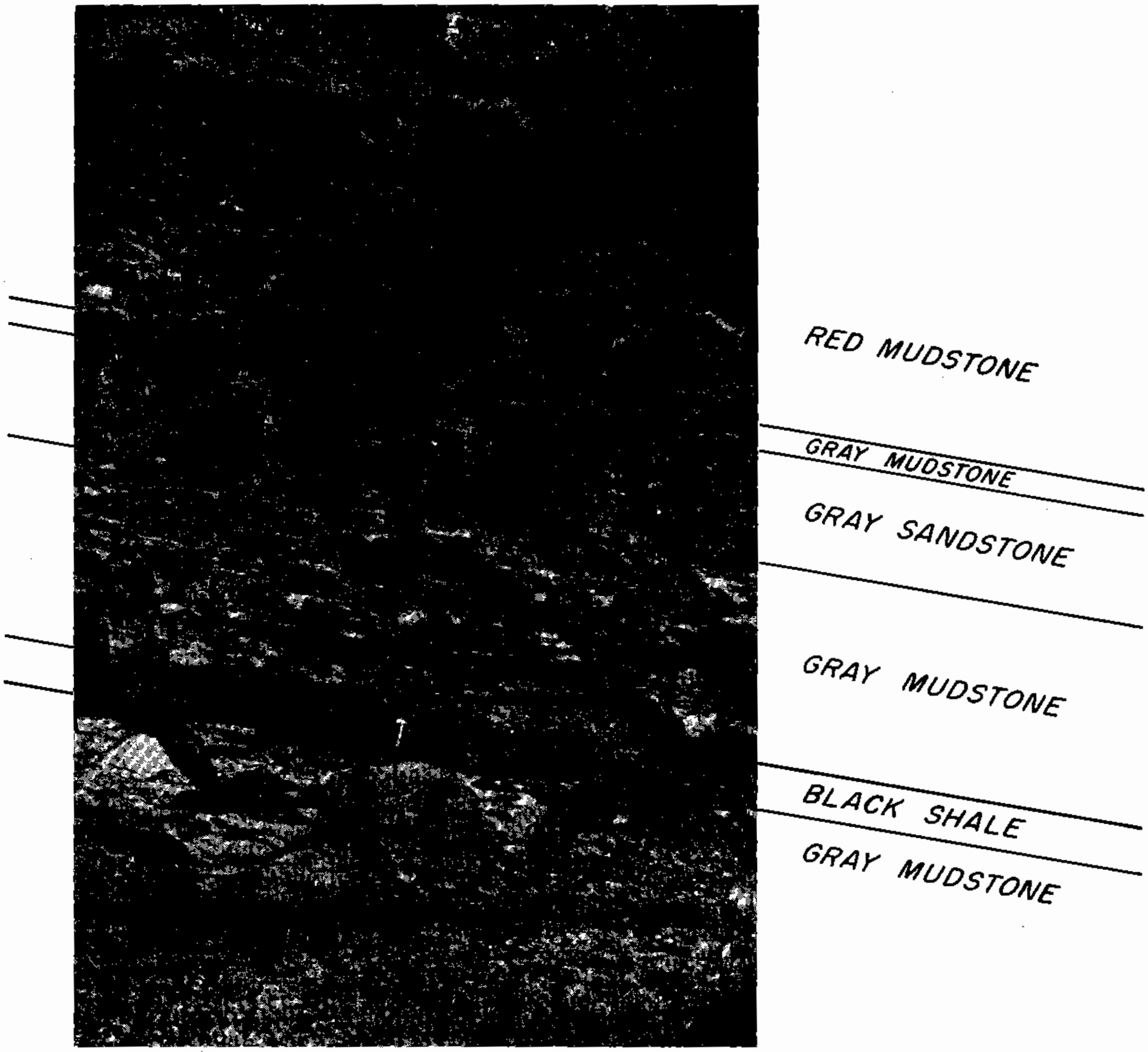


Fig. 37. Perennial lake sequence and overlying red mudstone in the East Berlin Formation, stop 7. The black shale (hammer) is at 80 m in section 4 (Fig. 35).

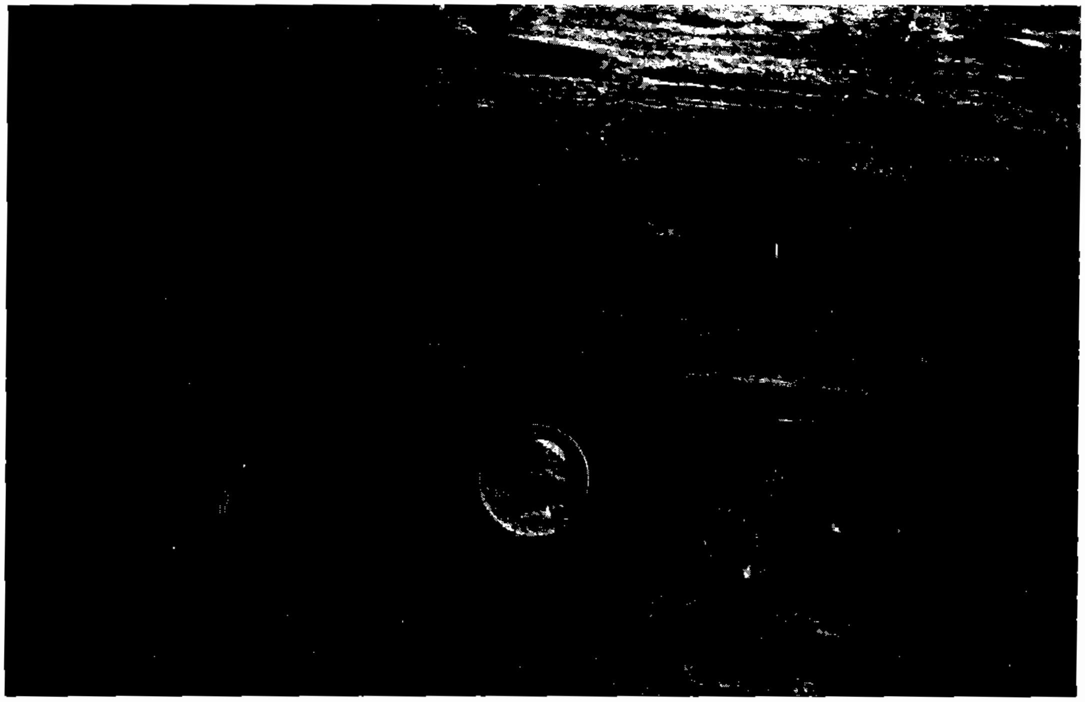


Fig. 38. Gray mudstone with dolomite laminae in perennial lake sequence of East Berlin Formation, stop 7. The bed is at 82 m in section 5 (Fig. 35).

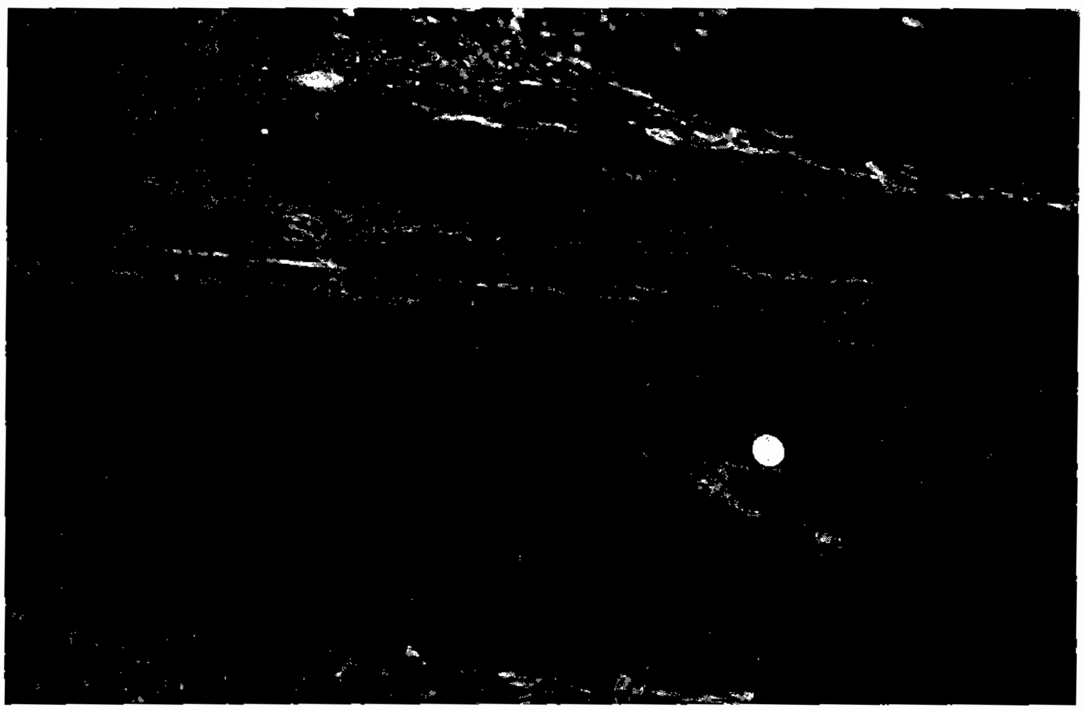


Fig. 39. Slump sheets in gray mudstone with dolomite laminae in a perennial lake sequence in East Berlin Formation, stop 7. These beds are 1 m from the bottom of the lacustrine cycle shown in Fig. 36.

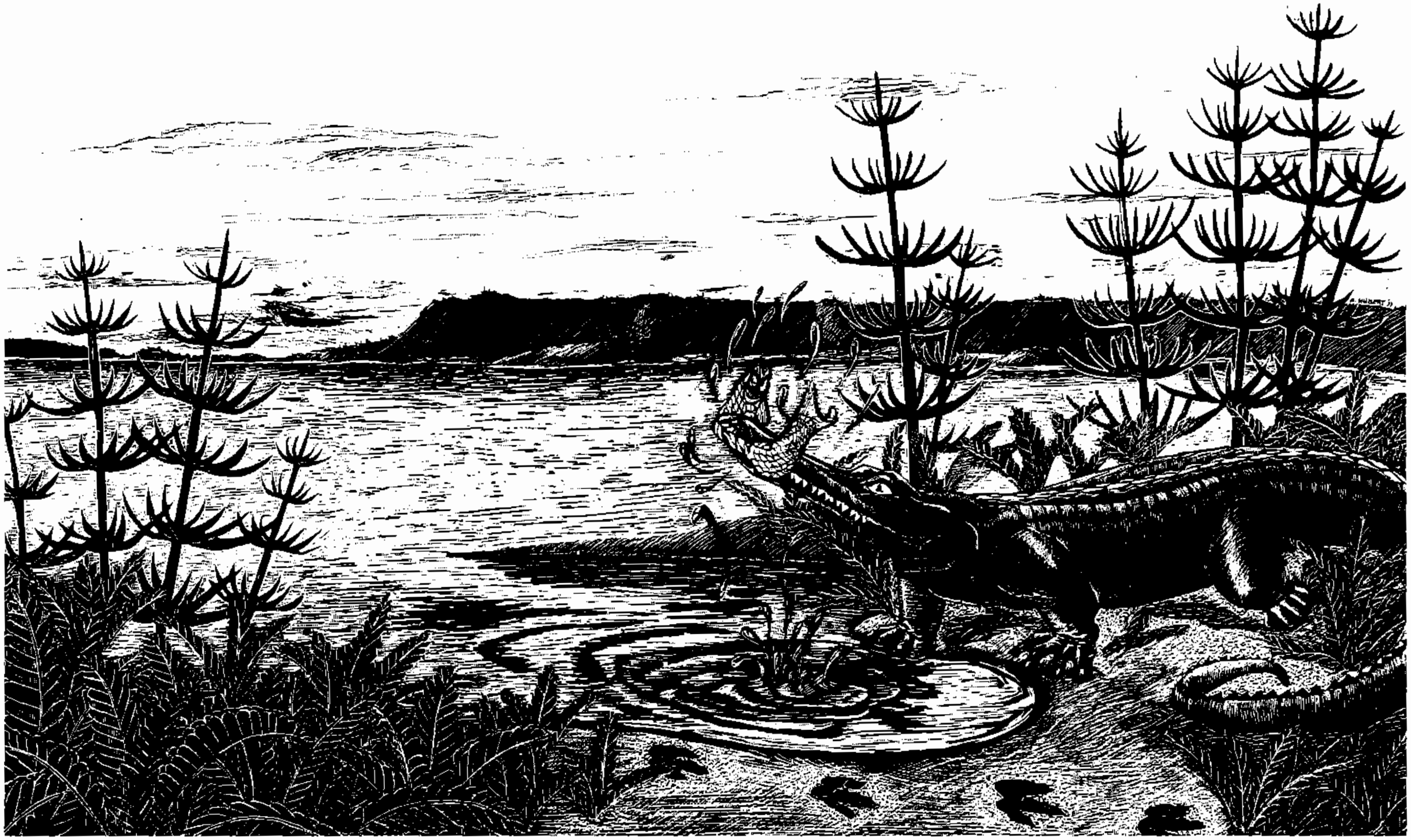


Fig. 40. On the shore of a large East Berlin lake. In the early morning, the 7-m phytosaur Rutiodon snatches a Semionotus from the shallows. The tall horsetail Equisetum and cycad Otozamites thrive in the wet mud of the lake strand. Stands of the conifer Araucarioxylon tower 60 m high along the distant horizon on sandy soils of the well-drained uplands. The dinosaur Eubrontes passed this way the previous evening.

Paleoslopes

The paleoslope directions of the lake floors were mapped using slump sheets that occur in the lacustrine sequences throughout the Hartford Basin. Similar slump sheets with recumbent folds are found in modern lakes near active faults; some of the slump sheets have been correlated with specific historical earthquakes (Sims, 1973, p. 163). The outcrops with slump horizons in the East Berlin Formation are within 20 km of the fault-bounded eastern escarpment, suggesting that earthquake shocks generated the slides.

At stop 7, the paleoslope of the lowest perennial lake sequence is to the west, as evidenced by data from slump sheets at section 5 (Fig. 35). At other localities in the Hartford Basin where paleoslope data are available for several lake sequences at one outcrop, the paleoslope azimuths are nearly in the same direction (Hubert _et al_., 1976, p. 1286). There is also a striking consistency of paleoslope directions to the southwest for perennial lake floors in Massachusetts and central Connecticut (Fig. 41).

Paleocurrents

The vector mean of the paleocurrents for the gray mudstone and sandstone of the perennial lake sequences at stop 7 were obtained by combining all available readings. The paleocurrents flowed northeast. This vector mean and the vector means for twelve other outcrops in the Hartford Basin are statistically significant at greater than the 95 percent level when tested by the Rayleigh statistic, except for one outcrop with only two readings (Fig. 41). The paleocurrent readings are mostly from the interbedded gray mudstone and sandstone near the top and bottom of the lake cycles and thus reflect the more shallow, nearshore parts of the lakes.

Figure 41 shows that the paleocurrents in the perennial lakes dominantly flowed to the southeast, except locally in central Connecticut where they flowed northeast. We interpret the regional pattern of southeast paleocurrents as due to paleowinds that persistently blew from the northwest over the lakes (Figs. 41, 42). The winds generated wave-driven currents that flowed southeast across the southwest-sloping lake floors.

At stop 7, and at Rocky Hill and Dinosaur State Park (sections 6, 7 on Fig. 35), the waves traveled northeast, up the southwest-sloping floors of the lakes. Most of the readings are from shallow-water nearshore sandstone at the top and bottom of the lake sequences. Wave refraction evidently oriented wave crests parallel to local northwest-southeast trends of the lake shores (Fig. 43).

The paleowinds were dominantly from the northwest both for successive lake cycles at any one outcrop and also for the 108 km from Massachusetts to Connecticut (Figs. 35, 41). The fetch of the lakes is unknown but was probably much greater than the minimum 20 km demonstrated by the outcrops in Connecticut. With a fetch of 20 km, a force 8 wind blowing 19.5 m per second (45 mi per hour) will stir the bottom at a depth of 22 m.

There are no bimodal paleocurrent patterns at the outcrop or regional level interpretable as the record of contrasting summer and winter paleowinds. The paleocurrents at each outcrop tend to have a unimodal distribution. The variances range from 382 to 3,492, averaging 1,551.

Interpretation of regional paleowind patterns for the Early Jurassic is difficult because data are limited and widely separated. In pre-continental drift orientation, the East Berlin paleowinds dominantly blew from the northwest. The partly eolian sandstones of Early Jurassic age in the western United States, including the Nugget, Navajo, and Aztec, also

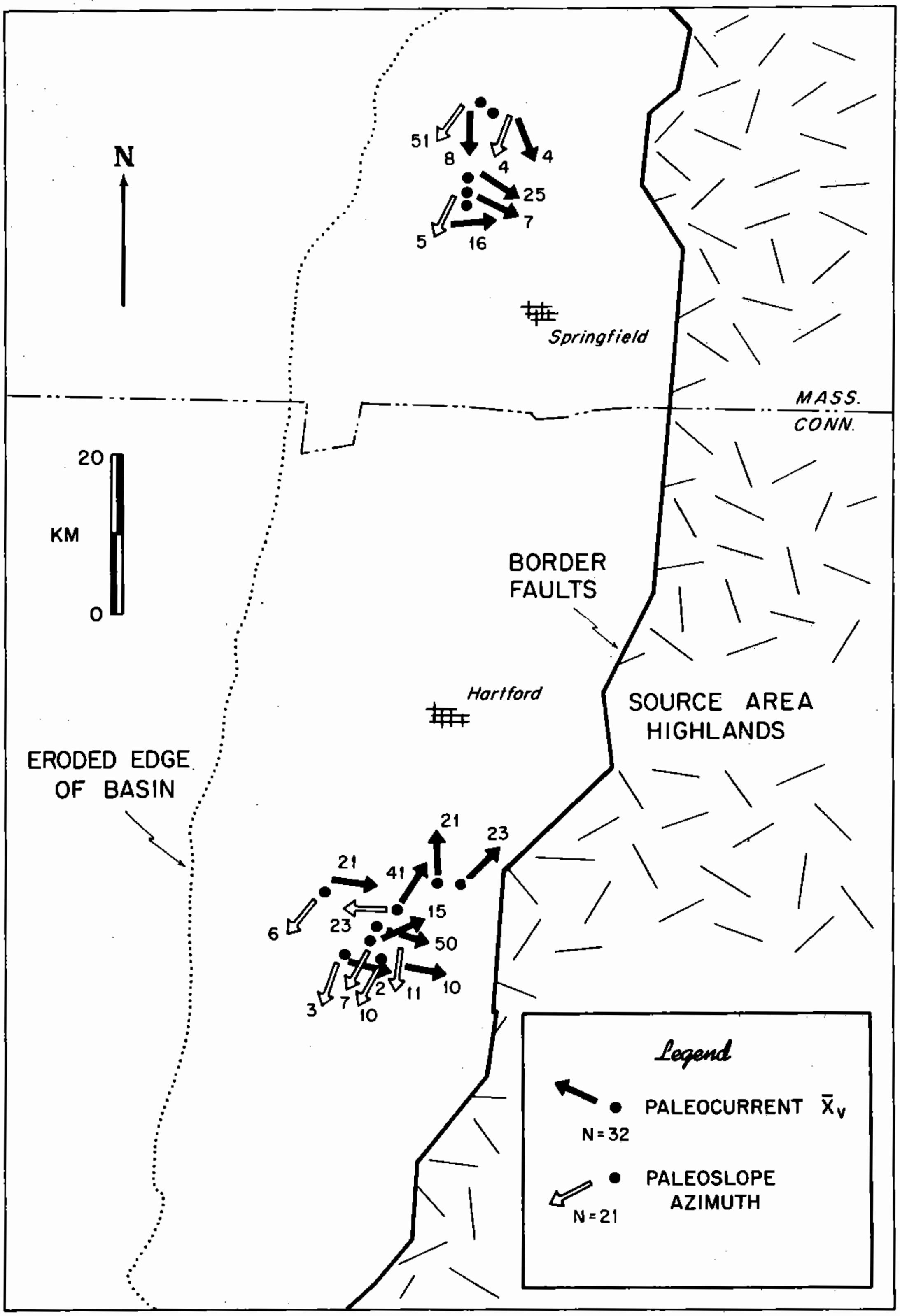


Fig. 41. Paleocurrents and paleoslopes for the black shale and gray mudstone and sandstone deposited in perennial lakes of the East Berlin Formation. Stop 7 is the dot with 41 paleocurrent readings. From Hubert <u>et</u> <u>al</u>., 1976, Fig. 8.

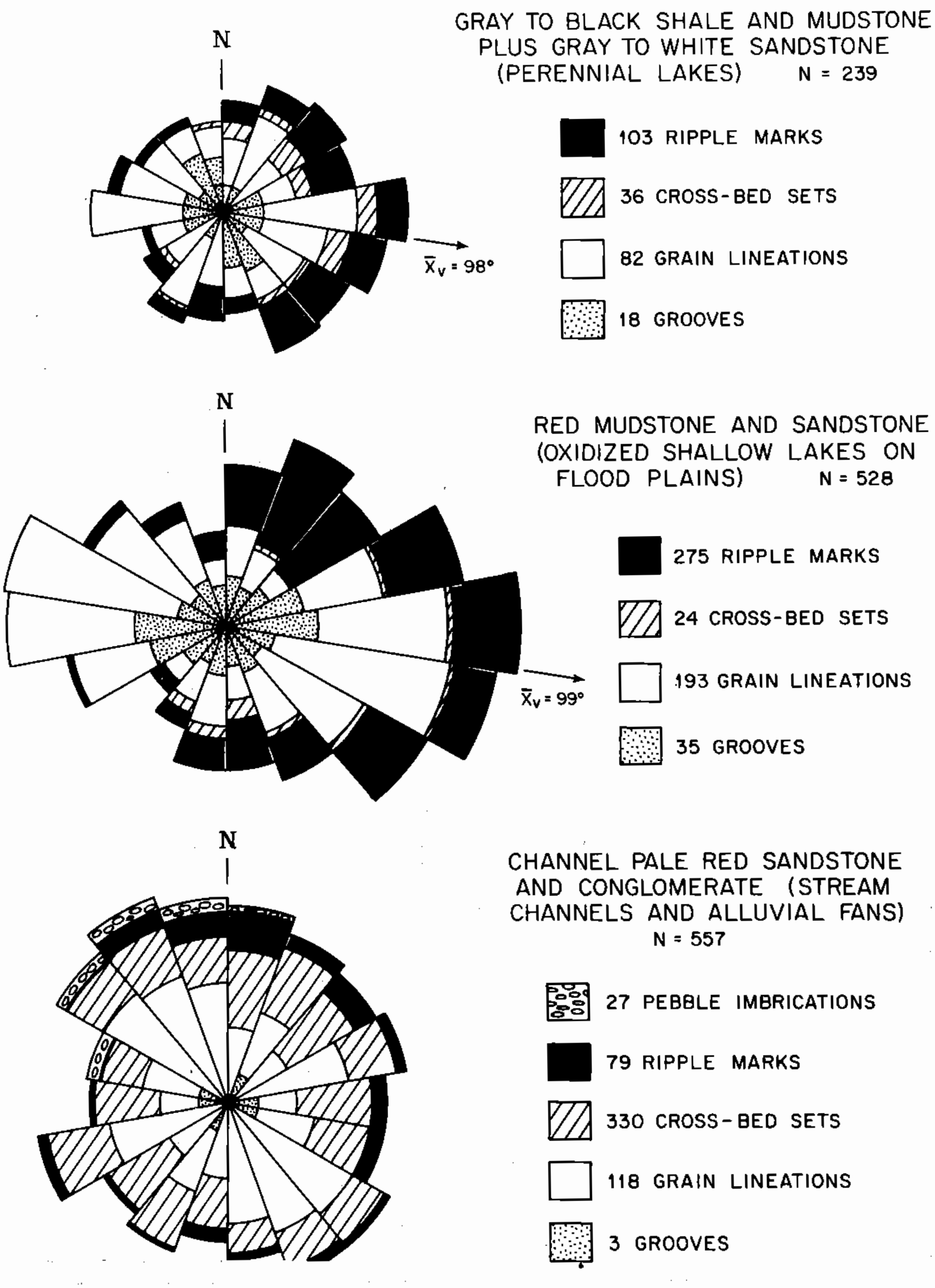


Fig. 42. Summary of paleocurrent data for the East Berlin Formation. No vector mean is calculated for the fluvial sandstone and conglomerate because of the absence of a strong mode in the data. From Hubert et al., 1976, Fig. 6.

record paleowinds from the northwest in the reconstructed continent (Poole, 1963, p. 402; Stanley et al., 1971, p. 13). The northwest paleowinds do not fit a simple system of northeast trade winds that might be anticipated at 15 degrees north paleolatitude. This is so even though the warmer, more uniform climate of the Early Jurassic widened and strengthened the trade wind belt as compared to today. Local mountainous relief on the rifting continental landmass probably strongly influenced the planetary surface winds.

Origin of Laminated Dolomite-Black Shale or Gray Mudstone Couplets

Couplets of dolomite and black shale or gray mudstone are common in the perennial lake sequences of stop 7 (Figs. 36, 38). The couplets reflect rhythmic alternation of two contrasting sets of environmental conditions. We infer that each carbonate lamina was precipitated during a dry period when river discharge was low and the lakes received little terrigenous sediment and humus.

The problem of interpretation is whether or not the couplets are varves. The varve hypothesis is suggested by the large size of the lakes, measured in thousands of square kilometers, the inferred depths of tens of meters, and the less than 1 mm thickness of most of the laminae. In the varve model, each carbonate lamina was precipitated during the dry season of a tropical wet-dry climate, lasting perhaps two to three months. The Hartford Basin at this time was in the tropics at about 15 degrees north latitude (Van Houten, 1977, p. 93). Counts of couplets indicate a minimum sedimentation rate of 35 to 45 cm of post-compaction sedimentary rock per 1,000 years. Lacustrine dolomite-black shale couplets in the Shuttle Meadow Formation have been cited as varves by Krynine (1950, p. 161), Cornet et al., (1973, p. 1245),

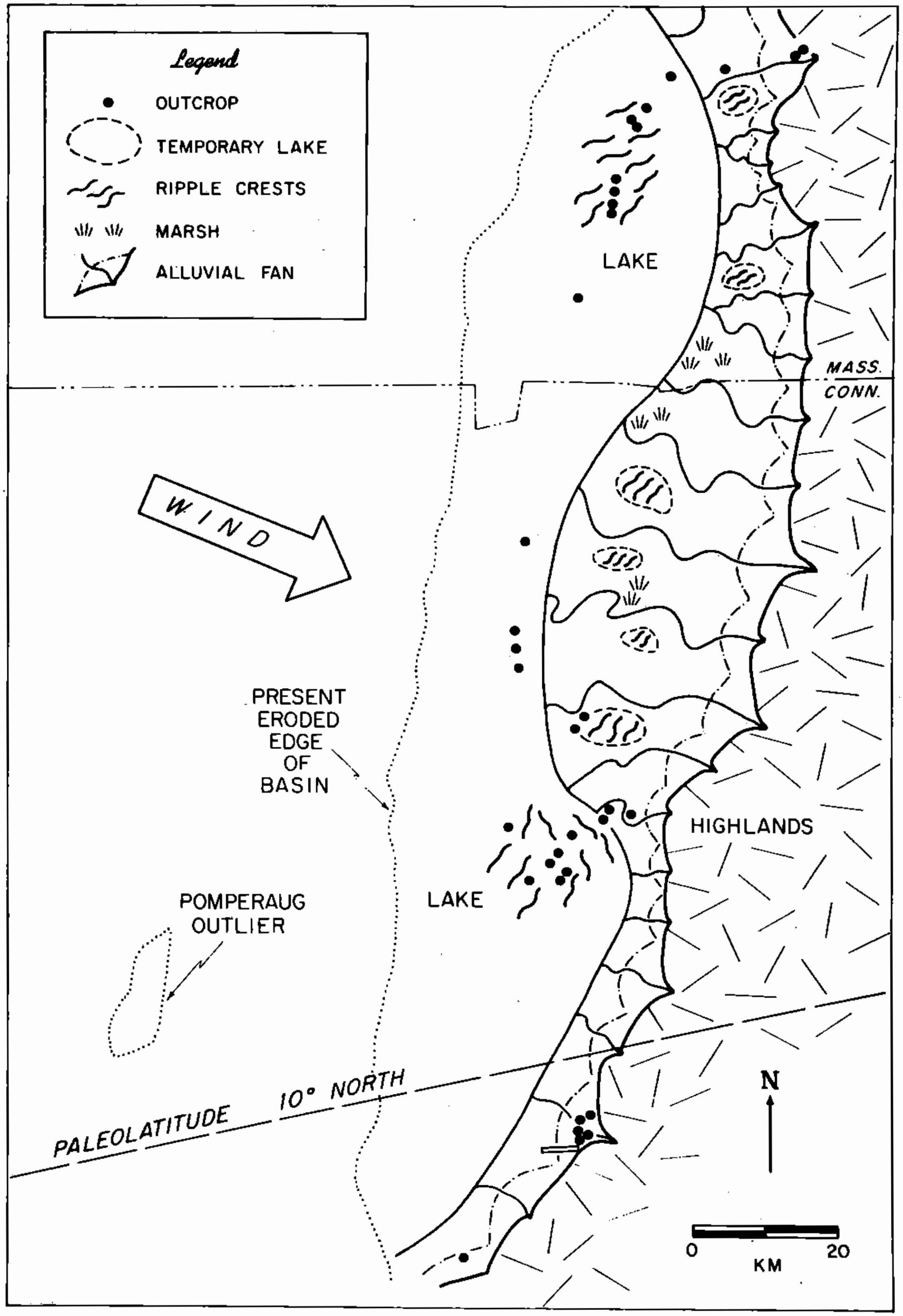


Fig. 43. East Berlin paleogeography at a time of a large perennial lake. It locally encroached on the alluvial fans as evidenced by the sequence in the Sugarloaf aqueduct tunnel (rectangle). From Hubert et al., 1976, Fig. 9.

and Cornet and Traverse (1975, p. 30).

The alternative hypothesis is that the dolomite laminae, or some of them, are not annual, but formed over several consecutive dry years when the rainy season was very short and little detritus reached the lakes. This idea gains support from the observation that the dolomite laminae are poorly defined in places, particularly in the gray mudstone, contain terrigenous clay, and are thicker than 1 mm, suggesting a combination of precipitation and detrital processes. Perhaps some laminae and thin layers of terrigenous clayey dolomite, especially bioturbated layers near the top and bottom of the lake cycles, formed over a span of years in shallow water at or just above the thermocline. Without an independent check on the seasonal origin of the carbonate laminae, such as correlation with tree rings in fossil logs, the rock record unfortunately does not permit the distinction between an annual dry season and a span of drought years. The dolomite laminae in similar rocks ("cementstone") in Carboniferous lacustrine sequences of Nova Scotia have been explained by carbonate precipitation during lake contraction in response to prolonged drought (Belt et al., 1967, p. 720).

Dolomite laminae are not uniformly present in the beds of black shale but form bundles (Fig. 36) that may reflect times when the surface waters were relatively more concentrated or the dry season of longer duration. Today in tropical wet-dry climates, a change to less annual precipitation is commonly accompanied by lengthening of the dry season.

Couplets of gray to brown, kerogen-rich oil shale-dolomite in the Eocene Green River Formation of Lkae Gosiute in Wyoming (Surdam and Wolfbauer, 1975, p. 388) and Lake Unita in Colorado (Lundell and Surdam, 1975, p. 495) have been explained by a playa-lake model. The playa-lake model is not appropriate for the kerogen-bearing black shale of the East Berlin Formation because of the absence of typical playa features. Missing are flat-pebble conglomerate of kerogen-rich shale derived by breakup of subaerial mudcracks, and interbeds of evaporite minerals, such as trona, nahcolite, halite, and Magadi-type chert. The absence of evaporite beds in the East Berlin Formation is especially striking and implies that residual brines did not develop with the inevitable filling in and drying up of the lakes.

Whether a varve or not, each nearly pure carbonate lamina in the East Berlin lakes evidently formed during a dry period (Fig. 44). At this time there would be a combination of tropical heat, intense evaporation, plant photosynthesis, lack of appreciable rainfall, decrease in sediment supply, and drop in lake level. The elevated surface temperature, high pH, and increased ion concentrations of the surface water would favor precipitation of calcium carbonate. The thickness of a carbonate lamina would vary due to fluctuations in amount of precipitated carbonate and dissolution in the bottom water enriched in CO_2 from decaying plant material.

Mg-calcite was most likely the major carbonate precipitated because of the ease with which the carbonate was dolomitized, leaving no relict calcium carbonate visible in 35 thin sections stained with alizarin red-S. In modern lakes, a Mg/Ca ratio of 2 is adequate to favor precipitation of Mg-calcite with up to 12 mol percent $MgCO_3$ over calcite (Müller _et al._, 1972, p. 161). Calcite and aragonite may have been precipitated at times in some of the lakes.

Rapid dolomitization evidently occurred during burial diagenesis in pore waters with Mg/Ca ratios over 7 and of fairly low salinity, a process known to convert Mg-calcite to dolomite in modern lakes (Müller

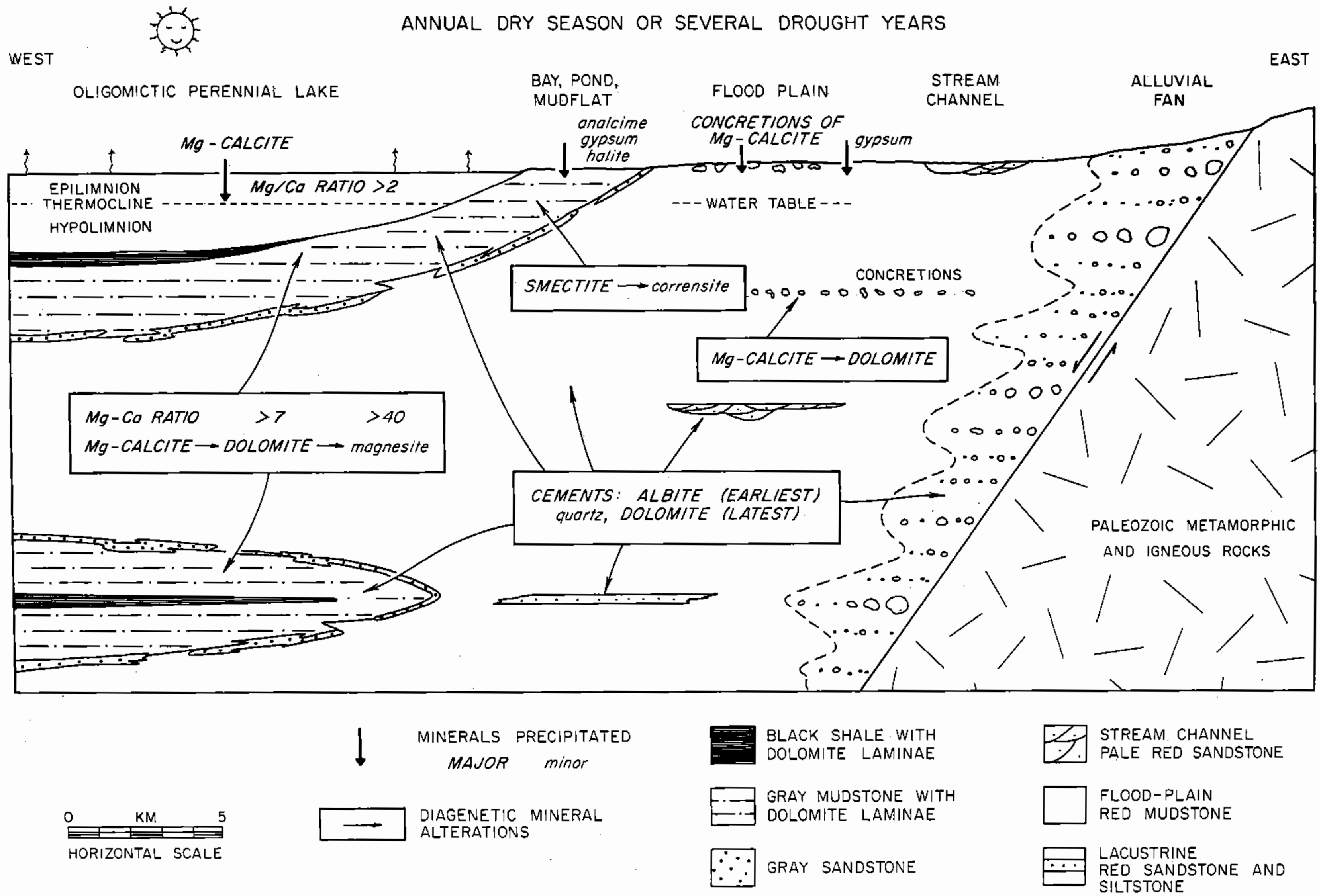


Fig. 44. Diagrammatic cross section of the valley showing deposition of carbonate laminae in an East Berlin lake during an annual dry season or a drought of several years duration. Diagenetic processes are shown in the rectangles.

et al., 1972, p. 163; Folk and Land, 1975, p. 63). Precipitation of calcium carbonate from lake water raises the Mg/Ca ratio so that it is enriched in Mg compared to the inflowing river water. The higher Mg/Ca ratios favor dolomitization of the calcium carbonate in the interstitial pore water during burial diagenesis. Traces of magnesite in black shale of the East Berlin Formation reflect conversion of dolomite when the Mg/Ca ratio exceeded 40, by analogy with modern lakes (Müller *et al.*, 1972). Mg-rich pore water is also implied by variable amounts of corrensite (regularly interstratified chlorite-smectite) found in some of the gray mudstones. The corrensite evidently formed by post-burial conversion of detrital smectite (April, R.H., pers. comm., 1977). The original calcite of the rare ostracod shells is now completely dolomitized.

Some of the gray mudstones have traces of analcime and gypsum plus molds of halite and possibly glauberite. The combined mineral assemblage of the black shale and gray mudstone suggests alkaline, hard water lakes with abundant Mg^{++}, Ca^{++}, and Na^{+} cations and HCO_3^{-} and $SO_4^{=}$ anions. The inferred high Mg/Ca ratios seem reasonable by analogy with the 5.5 to 7.3 Mg/Ca ratios in surface waters of lakes Tanganyika, Albert, Edward, and Kivu in the Western Rift Valley of East Africa (Beadle, 1974, p. 50).

In the varve model, each black or gray lamina of kerogen-bearing terrigenous clay accumulated during the rainy season, when swollen rivers brought detrital sediment and humus to the lakes. This is also the time of greatest nutrient availability and lake productivity, yielding the annual maximum of autochthonous organic matter. The thicknesses of the laminae vary, reflecting magnitude of river discharge, availability of detritus, and distance from entering rivers. The absence of grasses on the highland slopes and valleys would promote high rates of runoff, resulting in flash floods and serve erosion of unconsolidated sediment when the first storms announced the start of the rainy season. The initial storms in the wet-dry tropics commonly are violent cloudbursts.

Alternating fine laminae of carbonate and terrigenous clay are best preserved from burrowing organisms in oligomictic and meromictic lakes (Ludlam, 1969, p. 849). The laminae tend to be preserved below the thermocline where the anerobic hypolimnion prevents destruction by burrowing organisms. With a fetch of more than 40 km, the thermoclines in the East Berlin lakes would be a few tens of meters deep in response to wind mixing. Tropical lowland lakes, unless a few hundreds of meters deep, are mostly oligomictic where the thermally stratified water mixes at infrequent intervals. Mixing is caused by very high winds accompanying an unusually severe storm along a cold front, or by warm steady winds blowing over the lake in the dry season (Beadle, 1974, p. 73). The wind causes rapid evaporation and cooling of the surface water so that it is able to sink and mix with the bottom water, which in the tropics is commonly only 2 to 3°C cooler than the surface water. After mixing, the tropical heat soon reestablishes a pronounced thermocline and the hypolimnion in a few weeks is again depleted in oxygen. The brief, infrequent intervals when the bottom water is oxygenated are inadequate for a bottom fauna to become established. The East Berlin lakes were perhaps at times meromictic if deeper than a few hundred meters.

The East Berlin black shale accumulated in stagnant, anerobic bottom water rich in hydrogen sulphide, as shown by the relatively high organic content, abundant fossil fish, authigenic pyrite, and absence of burrows and fossils of scavenging animals. The fish skeletons commonly are articulated and restricted to certain laminae, suggesting fish kills caused by overturning of stratified water (McDonald, 1975, p. 107). The kerogen-bearing black shale contains about 1 percent organic matter, including aromatic and saturated hydrocarbons (Lawlor _et al._, 1967, p. 128).

Size and Depth of the Perennial Lakes

The minimum size of one of the lakes can be measured using the areal extent of the bed of black shale at stop 7 that is 28 m below the basal lava flow of the Hampden Basalt (Fig. 35). This black shale also crops out in northern Connecticut southeast of Tariffville 30 m below the Hampden Basalt (Davis and Loper, 1891, p. 427). During construction of the Sugarloaf aqueduct tunnel in southern Connecticut at the northeast end of Lake Gaillard, black shale was encountered 35 m below the Hampden Basalt (Thorpe, 1929, p. 281). Near Long Island Sound at East Haven, the black shale crops out at the south end of Lake Saltonstall, 30 m below the Hampden Basalt, in a fossil fish locality (Davis and Loper, 1891, p. 427).

The black shale is thus a continuous rock body for at least 108 km in a north-south direction, and 20 km east-west. The minimum size of the lake is 2,160 km^2. The lake extended west of the outcrop belt as implied by the west and southwest dip directions of the lake floors (Fig. 41) and by the paleocurrents to the west, southwest, and northwest recorded by river-channel sandstone in the westernmost outcrops (Fig. 47). Furthermore, the Shuttle Meadow Formation (?) in the Pomperaug Outlier contains beds of black shale with fish fossils (Hobbs, 1901, p. 55; Scott, 1974, p. 34). If the East Berlin lake reached the Pomperaug Outlier, as seems likely, it was at least 44 km wide, with an area of 4,717 km^2. This is slightly larger than Great Salt Lake. The lake was larger than 5,000 km^2 because it must have extended beyond the existing outcrops south of East Haven, Connecticut, north of Mount Tom in Massachusetts, and west of the Pomperaug Outlier.

The depth of the lake can be roughly estimated using an assumed gradient for the lake floor. A gradient of one-quarter of one degree seems a modest value, considering the slope implied by the numerous slump horizons and transgression of the lake over alluvial-plain sediments spread away from an active fault scarp. A one-quarter of one degree gradient over the minimum 20 km of southwest-sloping lake floor exposed in central Connecticut implies a depth of 80 m. This value is arbitrary, but suggests depths conservatively estimated in tens of meters.

Each sequence of black shale and gray mudstone and sandstone records the formation and expansion of a perennial lake followed by its contraction and disappearance. In the tectonic setting of the rift valley, the major control of the cycles seems to have been climatic fluctuations on the order of some tens of thousands of years due to shifting of the transitional boundary between the tropical humid wet-dry and tropical semi-arid climatic belts.

It is not known whether the perennial lakes were closed or through-flowing with spillways, because the sequences of gray mudstone and black shale are truncated by Long Island Sound. Also the lakes may have varied from open exit to closed. The available evidence suggests they commonly were closed, with waters that were alkaline, of variable salinity, and with substantial dissolved Mg^{++}, Ca^{++}, Na^{+}, HCO_3^{-} and $SO_4^{=}$. Especially suggestive are the large volume of carbonate precipitated, extensive dolomitization, traces of analcime, gypsum, and halite, abundant authigenic albite and dolomite cements in the sandstones, absence of fragments of gastropods and pelecypods and scarcity of ostracods in thin sections, and the presence of the lacustrine symmetrical cycles. Furthermore, the dolomite concretions in the gray mudstone show stable isotope compositions for C and O that fall in the saline range (Mahoney, 1974, p. 66-68). She concluded (p. 86) that the data are compatible with an interpretation of repeated flooding and extreme evaporation of pore solutions.

The average boron content of the less than 2 μm fraction of 17 gray and black mudstones from the East Berlin Formation is 225 ppm, varying from 137 to 284 (Wakeland, M.E., Jr., 1976, pers. comm.). The abundant ions of sulphate, magnesium, and boron in the waters of the lakes evidently prevented colonization by gastropods, pelecypods, and decapod crustaceans. In modern lakes these three ions have a strong adverse effect on the invertebrate fauna.

Some of the dolomitic gray mudstones that enclose the deeper water black shale contain traces of analcime and gypsum, plus molds of halite and possible glauberite. These minerals may record concentration of ions in shallow bays or isolated pools during falling lake levels. Prolonged periods of drought may at times have caused lake levels to fall below a spillway, if the lake was not already closed. With no outlet, a drought of even a few years will cause a lake strand to withdraw many kilometers. The absence of beds of evaporite minerals suggests that the main lake bodies did not go to dryness and probably were never hypersaline, unlike the large playa lakes with brines that generated analcime-rich mudstone at the top of the lacustrine asymmetrical cycles in the Upper Triassic Lockatong Formation in New Jersey (Van Houten, 1964, p. 509).

Burial Diagenesis

During burial diagenesis, the lake and river sands were everywhere cemented by large amounts of pure albite overgrowths on detrital plagioclase (Fig. 45). The albite overgrowths are nearly pure, varying from 99.4 to 100.0 percent albite molecule by X-ray fluorescence analysis. The petrography of the albite cement is described by Heald (1956, p. 1148).

The sequence of precipitation of cements was quartz, then albite, followed by dolomite, commonly slightly ferroan, and lastly calcite. The calcite is confined to faults and joints and is relatively rare. The pore water of the sands contained substantial dissolved Na^{+}, Ca^{++}, and Mg^{++}, because albite cement comprises 10 percent by volume of the sandstones and dolomite 13 percent. The source of these ions was the closed-exit lakes with alkaline, hard water that several times covered most of the floor of the rift valley. Lake water evidently entered the pores of the channel sands by percolation through the valley alluvium in response to the regional hydraulic gradient during burial compaction.

Plagioclase grains dominate over K-feldspar in a 7:1 ratio in the arkoses, suggesting that the sodium in the lakes was largely derived from weathering of plagioclase in the high-grade metamorphic rocks of the highlands along the rift valley. The plagioclase grains average calcic albite in composition, varying from 71 to 100 percent albite molecule by X-ray fluorescence analysis.

Lacustrine Redbeds

Thin, even beds of red very fine sandstone and coarse siltstone are interbedded with the floodplain red mudstone. The tabular bodies contain abundant ripple cross-lamination and vary in thickness from a few centi-

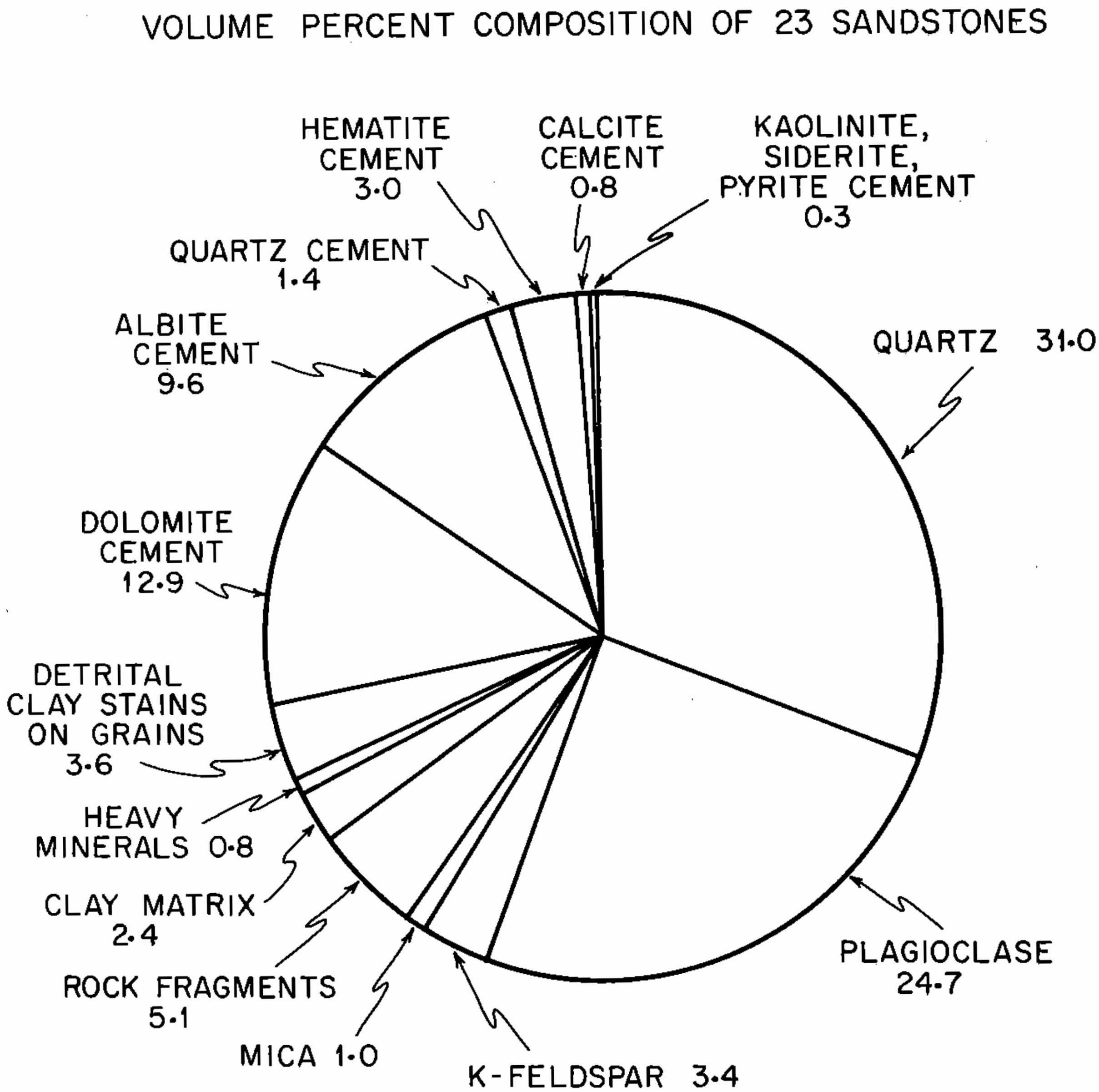


Fig. 45. Composition of 23 fluvial and lacustrine sandstones in the East Berlin Formation by volume percent determined in petrographic modal analyses.

meters to a meter, with a modal thickness of 20 to 30 cm. The areal extent of the sandstone and siltstone bodies commonly is less than 2 km^2 and they are enclosed within floodplain red mudstone and stream-channel pale-red sandstone (Fig. 35). The lower contacts of some of the sandstone and siltstone bodies are planar surfaces that extend for tens of meters across the outcrops. We interpret the surfaces as the initial lake floors cut into floodplain mud. Some of the lacustrine bodies show flaser bedding that grades laterally and vertically into starved ripples, making difficult the determination of the boundaries of the lake sequences. Krynine (1950, p. 60) and Sanders (1968, p. 289) also viewed these redbeds as having accumulated in shallow oxidized lakes.

Layers of dolomite nodules are common in the red mudstone at stop 7 (Fig. 35). The nodules are of early diagenetic origin because mud laminae drape over some of them, others are deformed in slump horizons, and some are reworked as intraformation pebbles in fluvial sandstone. Many of the layers of carbonate nodules seem to be the result of precipitation of carbonate in the pore water of the sediment as shallow lakes evaporated to dryness. At some horizons, clusters of gypsum crystals grew in the mud perpendicular to the lamination.

Some of the lakes probably persisted for many seasons, but the red color of the sandstone and siltstone implies destruction of organic matter at oxidizing shallow depths and probably lack of permanence. The lakes lay in low areas in the flood basin and were replenished during river flooding.

Grain lineation, grooves, and ripple crests were used to map paleocurrents. At the outcrop level, the frequency distributions of paleocurrent azimuths tend to be unimodal, with variances from 43 to 6,453, averaging 1,131. At each of the 21 outcrops the vector mean is statistically significant when tested by the Rayleigh statistic, except for the type section which has only eleven readings.

The paleocurrents that deposited the lacustrine sand and silt consistently flowed southeast at 18 outcrops in Massachusetts and Connecticut, and to the northeast at three outcrops in central Connecticut, including stop 7 (Figs. 35, 46). We interpret the paleocurrents as due to dominant northwest winds that blew over the surfaces of the shallow lakes, generating waves that flowed to the southeast. In central Connecticut, the paleocurrents flowed northeast, east, and southeast, whereas the paleoslopes of the floodplains were to the southwest. Wave refraction evidently oriented ripple crests parallel to arcuate lake shores, concave to the northwest.

River-channel Sandstone

Sandstone bodies with erosional lower surfaces occur throughout the floodplain sequences of red mudstone at stop 7. The lensing channel sandstones commonly are about 0.5 m thick. The pale-red sandstones contain intraformational pebbles of red mudstone. A few of the thicker sandstones at other outcrops range up to 2 m thick and are interpreted to be river point-bar sequences because they fine upwards from medium sandstone with festoon cross-beds to fine sandstone and siltstone with interstratified plane beds and ripples. The thinner sandstone bodies are horizontally laminated with a few festoon or planar cross-beds, but lack a fining-upward sequence.

The paleocurrent vector means of 8 of the 9 outcrops are significant at the 95 percent level when tested by the Rayleigh statistic, demonstrating a preferred orientation. The exception is an outcrop with only 5

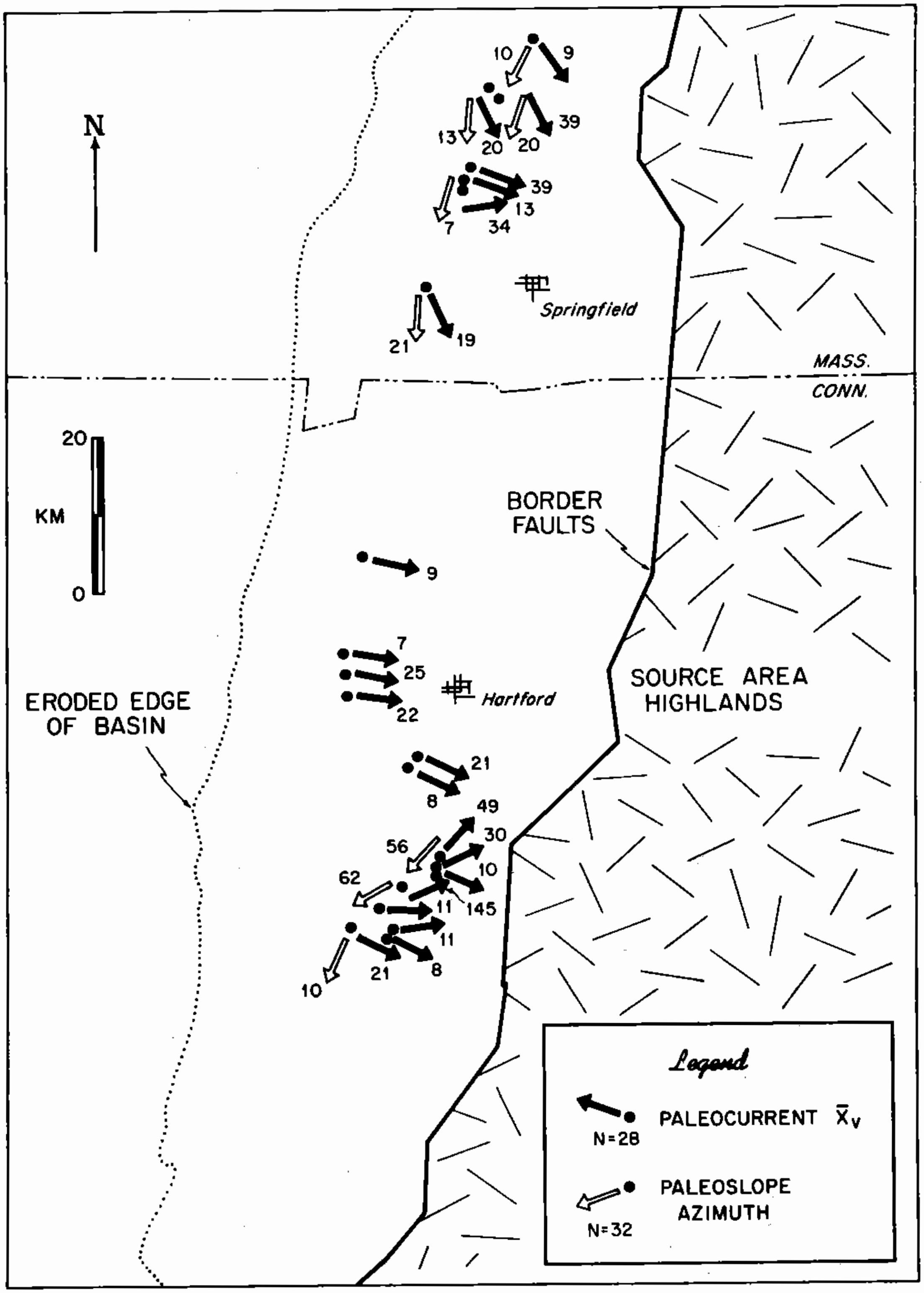


Fig. 46. Paleocurrents for red sandstone and siltstone deposited in oxidized shallow lakes on floodplains of the East Berlin Formation. Paleoslopes for floodplain red mudstone with thin sandstone layers show the paleoslope of the valley floor. Stop 7 is the dot with 49 paleocurrent readings. From Hubert <u>et al</u>., 1976, Fig. 5.

readings. The variances within outcrops are large, 1,094 to 7,544, averaging 3,935.

The outcrop vector means for paleocurrents in the channel sandstones generally radiate radiate away from the eastern escarpment (Fig. 47). The rivers flowed northeast at stop 7 and Rocky Hill in central Connecticut; meandering rivers are suggested at these localities by the large scatter of azimuth directions.

In general, the rivers meandered across the valley floor, as evidenced by the interbedded red mudstone, fining-upward sandstone sequences, radiating paleocurrent pattern of outcrop vector means, and large variances within outcrops.

Floodplain Red Mudstone

At stop 7, beds of mostly horizontally laminated, grayish-red mudstone form sequences up to 4 m thick, most being 0.5 to 1 m (Fig. 35). Mudcracks are common on the bedding surfaces, and dinosaur tracks and raindrop impressions are occasionally seen. The mudstones are interpreted as floodplain deposits that graded laterally into lacustrine mudflats. Some horizontally laminated, thin beds that grade from sandstone to mudstone evidently reflect settling of sediment from swirling floodwater. Grayish-brown to dusky-brown mudstone may have accumulated in organic-rich marshes.

Origin of the Color of the Redbeds

Introduction

The origin of the hematite pigment in the redbeds of the Connecticut Valley has long been controversial. Russell (1889, p. 15) thought that the red colors imply transportation of hematite-stained particles from lateritic soils that mantled hot, wet highlands along the rift valley.

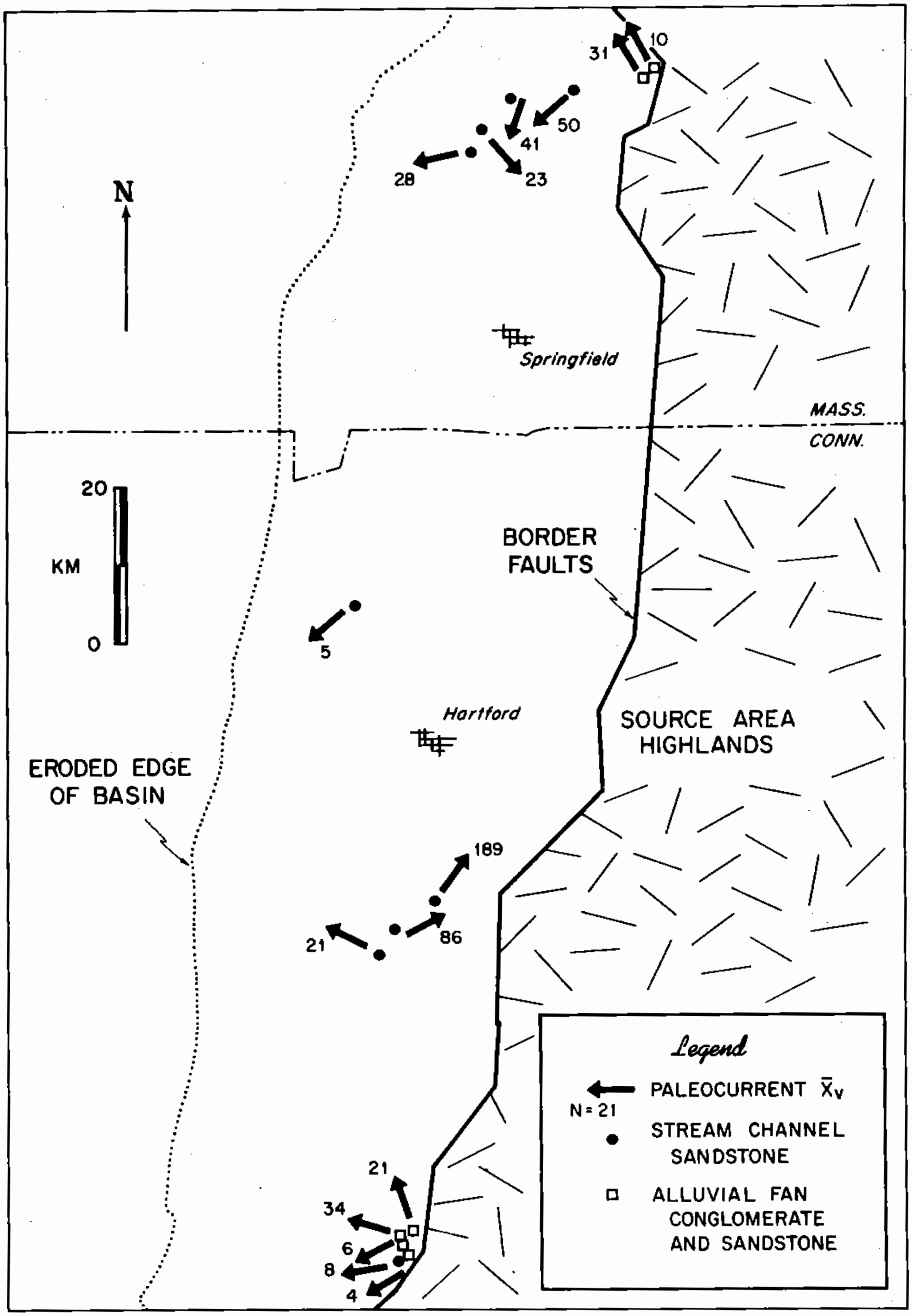


Fig. 47. Paleocurrents for sandstone and conglomerate deposited in river channels and alluvial fans of the East Berlin Formation. Stop 7 is the dot with 189 paleocurrent readings. From Hubert et al., 1976, Fig. 4.

Impressed by fresh feldspar grains, abundant mudcrack horizons, and local molds of halite, hypsum, and glauberite, Barrell (1915, p. 30) preferred semiaridity with little chemical weathering. Krynine (1950, p. 180) tried to reconcile humidity and aridity in a tropical regime where laterites formed under at least 1250 mm of annual precipitation, but with a brief dry season. His central idea was that highland streams incised V-shaped canyons through laterites to erode fresh rock, the resulting alluvium being a red mixture of weathered, hematite-stained particles and fresh detritus. Dilution of red soil with fresh detritus, however, invariably yields brown or yellow-brown, not red, alluvium, even in areas of thickest laterites (Van Houten, 1961, p. 111; Walker, 1967a, p. 917). The paleoclimate and origin of the red color of a specific formation must be determined by independent evidence.

Redbeds can form wherever sediment accumulates in an oxidizing environment because the surfaces of sand and mud particles in almost every climate are stained brown or yellow brown by hydrated iron oxides, collectively called limonite (Walker, 1967a, 1976; Van Houten, 1961, 1972, 1973). With aging over some tens of thousands of years in oxidizing, alkaline interstitial water, the limonite converts with dehydration to hematite. In tropical climates that range from humid to arid, the volume of limonite stains on the particles is more than adequate to color the corresponding sandstone and mudstone various shades of red.

Fluvial strata can also become progressively reddened over tens of millions of years by hematite pigment generated by post-depositional dissolution of Fe-silicate grains, especially amphibole, pyroxene, epidote, chlorite, and biotite (Walker, 1967b, 1976). Dissolution of the grains by hydrolysis releases iron, which in oxidizing, alkaline pore water is precipitated as hematite, or perhaps as a red ferric oxide, which upon aging converts to hematite.

Colors in the East Berlin Formation

Redbeds comprise 65 percent of the measured sections of the East Berlin Formation. Fifty-two percent is grayish-red mudstone (Munsell Color Chart 5R 4/2) that accumulated on floodplains. Pale-reddish brown (10R 5/4) and grayish-red (10R 4/2) very fine sandstone and coarse siltstone that were deposited in shallow, oxidized lakes constitute 10 percent. Meandering streams deposited pale-red (10R 6/2) and pale-yellowish brown (10YR 6/2) channel-sandstone that forms the remaining 3 percent.

Large perennial lakes left a record of drab-colored rocks that total 35 percent of the formation. The deeper water, kerogen-bearing, pyritic shale is grayish black (N6) and dark gray (N3), whereas coarser mudstone closer to the lake strands is medium dark gray (N4) and medium gray (N5). Nearshore and strandline sandstones are mostly medium light gray (N6) and light gray (N7). Irregularly shaped patches and streaks in some of these gray sandstones are altered to yellowish brown (10YR 6/2), light olive gray (5Y 6/1), or grayish orange pink (5YR 7/2).

Genesis of Hematite Pigment

In the fluvial pale-red sandstones, rims of hematite-stained clay are present on plagioclase and quartz grains beneath albite and quartz overgrowths (Fig. 48). Brown and yellow-brown limonite evidently stained the surfaces of the sand and mud particles transported by the streams, and with aging the limonite converted to hematite.

Hematite pigment in the red sandstone and mudstone of the East Berlin Formation is authigenic, generated over millions of years by four post-depositional processes (Hubert and Reed, 1968, in press).

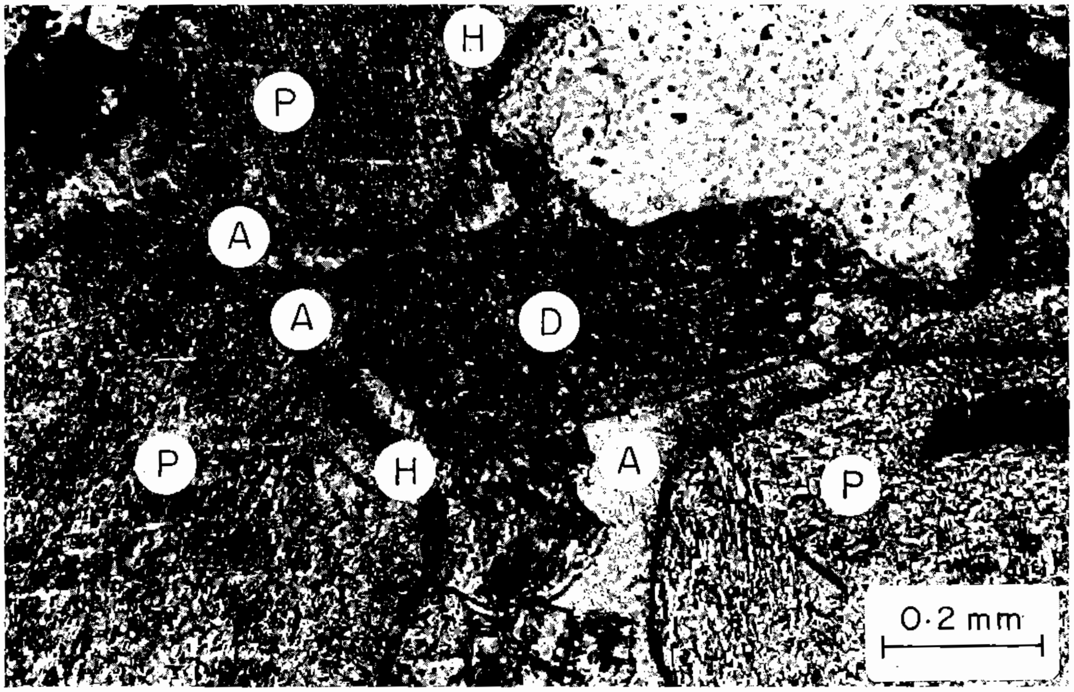


Fig. 48. Hematite-stained (H) clay coats detrital grains in a fluvial pale-red sandstone in the East Berlin Formation. Albite overgrowths (A) on plagioclase grains (P) are overlain by younger dolomite cement (D) which appears dark due to potassium ferricyanide staining. Plain light.

Fig. 49. Authigenic hematite alteration (black) on biotite in a biotite-quartz grain in a fluvial pale-red sandstone in the East Berlin Formation. The hematite is the result of intrastratal solution; hematite was also precipitated in intergranular pores. Plain light.

1. Brown and yellow-brown limonite surface stains on sand and mud particles that accumulated in the oxidizing environments of stream channels, floodplains, and shallow lakes were converted to hematite with aging. This was the major source of hematite pigment in floodplain grayish-red mudstone.

2. Intrastratal solution of Fe-silicate grains, especially pyroxene, amphibole, epidote, chlorite, and biotite, was pervasive in the fluvial and lacustrine sandstones, generating about 3 percent by volume of hematite (Fig. 49). The proportion of pyroxene-amphibole-epidote grains among the nonmicaceous nonopaque heavy minerals was reduced to 5 percent, and many of the grains show hematite surface alteration and partially dissolved terminations. The original nonopaque assemblage is largely preserved in impermeable dolomite concretions and in fluvial and lacustrine mudstone where it averages 32 percent pyroxene-amphibole-epidote. Dissolution of Fe-silicate grains by hydrolysis became an important process after partial filling of intergranular pores by quartz, albite, and dolomite cements. Processes (1) and (2) were volumetrically the most important in generating hematite pigment.

3. In all the redbeds, the abundant magnetite grains are oxidized to hematite pseudomorphs, identified in refelcted light with polished thin sections. The relatively small proportion of ilmenite grains in the detritus is oxidized to hematite-rutile intergrowths.

4. An additional amount of iron was released into the pore water of the sandstones as dolomite cement preferentially attacked and replaced grains of pyroxene, amphibole, epidote, chlorite, and biotite. All stages of replacement are present, including dolomite pseudomorphs after Fe-silicate grains. The iron released to the pore water was available to form a considerable amount of hematite pigment.

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ROAD LOG

The trip starts from the Science Building at Wesleyan University, Middletown, Connecticut. The Science Building is on Church Street across from Olin LIbrary. The bus is parked facing north on Pine Street at the side of the Science Building.

MILES

0.0	0.0	Turn right on Church Street.
0.2	0.2	Turn left on High Street.
0.7	0.5	Turn right at the traffic lights on Washington Street (route 66) and go downhill toward the Connecticut River. The hills in the distance are crystalline rocks of Lower Paleozoic age east of the Mesozoic border fault.
1.1	0.4	Turn right (south) on route 9 parallel to the Connecticut River.
1.6	0.5	Turn right on route 17.
2.1	0.5	Proceed around the traffic circle and continue south on route 17.
2.9	0.8	Sandstone of the Portland Formation on the right.
4.3	1.4	Durham Historical District with lovely old homes.
5.5	1.2	Take the right fork and continue on route 17. Escarpment on left follows the Mesozoic border fault that separates crystalline rocks on the east from the Portland Formation on the west.
5.7	0.2	Turn left on route 77.

5.8	0.1	STOP 1. PORTLAND FORMATION, DURHAM.
		Park at two-story yellow house inside the Y-junction of routes 17 and 77. This is the first house on the right on route 77. At the Y-junction, walk east across the road and plunge into the woods. The rock ledge is about 30 m east of the road.
		Leave the parking area on the west side and turn left (south) on route 17.
9.9	4.1	The ridge on the right is the Holyoke Basalt.
13.6	3.7	Turn right on route 22.
13.9	0.3	Junction with route 150. Take left fork, following route 22.
17.3	3.4	Turn right on route 5
18.1	0.8	Proceed through underpass; turn left onto I-91 and go south.
20.0	1.9	Leave I-91 at Exit 10 (route 40; Mt. Carmel, Hamden).
22.1	2.1	STOP 2. NEW HAVEN ARKOSE, NORTH HAVEN.
		Stop in breakdown lane of route 40. The illustrations for stop 2 refer to the north side of the roadcut. Proceed west on route 40.
22.9	0.8	Turn left (south) on route 10.
23.1	0.2	Turn around in parking lot of Our Lady of Mt. Carmel Church on left side of street and proceed north on route 10.
23.3	0.2	Turn right on route 40.
25.7	2.4	Take I-91 toward Hartford. As you enter I-91, the intrusive basalt of Sleeping Giant is ahead on the left. His head is on the west and feet on the east.

28.9	3.2	The roadcut on the right is New Haven Arkose with caliche horizons.
30.9	2.0	The roadcuts on both sides of I-91 are New Haven Arkose with caliche horizons.
32.9	2.0	New Haven Arkose on both sides of I-91.
36.0	3.1	New Haven Arkose along the center strip of I-91.
36.3	0.3	On the left are the Hanging Hills of Meriden (Holyoke Basalt). The TV installations are on West Peak.
37.2	0.9	The cliffs on the right are Holyoke Basalt.
38.2	1.0	Take exit 17 to the right following the sign to "route 66 west".
38.9	0.7	Take left fork toward route 66 west.
40.2	1.3	Go right toward route 66 west.
40.5	0.3	Junction with route 66 west.
41.0	0.5	New Haven Arkose on right in roadcut.
41.4	0.4	Leave route 66 at exit 6. Stop 3 is behind the G. Fox Store seen on the north side of route 66.
41.7	0.3	Turn left into Meriden Square.
41.8	0.1	Turn left and drive around the perimeter of the parking lot to behind the G. Fox store, which is at the left end of the shopping center.
42.2	0.4	STOP 3. TALCOTT BASALT, MERIDEN. The outcrop is directly behind the G. Fox store. Proceed to northwest corner of parking lot. Turn left at stop sign and go up the hill. Holyoke Basalt is on the skyline.
42.7	0.5	Turn right at traffic lights on route 71 (Capitol Avenue).

43.1	0.4	Holyoke Basalt is on the right and also ahead on left at bend in road.
45.7	2.6	Go left on Butler Street (note that there is no street sign at this end of Butler Street).
45.9	0.2	Turn left on Park Drive.
46.8	0.9	Cross first bridge; turn right and proceed over second bridge onto Percival Park Road. Drive carefully because this road is very narrow. You are passing the north end of Merimere Reservoir.
47.4	0.6	You are driving up the dip slope of the Holyoke Basalt.
48.2	0.8	Turn left on road to East Peak. The right fork leads to West Peak.
48.6	0.4	STOP 4. EAST PEAK OF THE HANGING HILLS OF MERIDEN. Park in parking lot and ascend to top of stone tower. Leave parking lot by exit <u>adjacent</u> to stone tower in order to follow one-way loop. Return down Percival Park Road.
49.1	0.5	Drive past road on left that goes to West Peak.
50.3	1.2	Go past end of Merimere Reservoir.
50.5	0.2	Cross bridge and turn right on Reservoir Avenue.
50.9	0.4	Holyoke Basalt forms cliffs across lake on right.
51.3	0.4	View of stone tower on right.
51.8	0.5	Go through underpass below route 66. Turn right at second road into parking lot of Hubbard Park, Meriden. LUNCH STOP. Leave parking lot by turning left on unnamed park road. Drive past Mirror Lake on the right.

52.6	0.8	Turn right at traffic lights on west Main Street.
53.3	0.7	Sign for junction with route 66.
53.5	0.2	Cross bridge and immediately past traffic lights turn left and park on unfinished access road.
		STOP 5. NEW HAVEN ARKOSE ALONG ROUTE 66, MERIDEN
		Proceed west on route 66, following sign to I-84.
55.9	2.4	Turn right at traffic lights on road leading to route 10.
56.0	0.1	Turn right on route 10.
56.6	0.6	Turn left on access road to I-84 and go east on I-84.
58.4	1.8	Hills on left are Lower Paleozoic crystalline rocks west of the Mesozoic redbeds.
59.9	1.5	New Haven Arkose with caliche horizons adjacent to and under bridge.
61.6	1.7	New Haven Arkose on both sides of I-84.
62.9	1.3	TV towers on Holyoke Basalt seen to the north.
63.2	0.3	Leave I-84 at Exit 34 leading to route 66 west.
63.4	0.2	Turn left at stop sign. Holyoke Basalt to right on skyline.
63.9	0.5	Turn right on route 72. Go past Getty Gas station and park on south side of road in pull-off just before entrance sign to Holiday Inn.
		STOP 6. SHUTTLE MEADOW FORMATION, PLAINVILLE.
		The quarry is on the north side of the road. Please do <u>not</u> climb the rock faces. This is a working quarry and the blocks are loose. Do not examine rock face behind the Getty Gas station because the owner does not allow visitors and has two guard dogs. The Tomasso trap rock

		quarry in the Holyoke Basalt is on the south side of I-84. Proceed east on route 72.
65.5	1.6	Turn right to follow route 72 at the Texaco station. Go over the bridge and continue on route 72 east.
70.6	5.1	Bridge over routes 5/15.
70.7	0.1	Roadcuts with the type section of the East Berlin Formation (Lehmann, 1959, p. 16-21). The measured section (section 1 on Fig. 35) shows symmetrical lake cycles and river-channel sandstone and floodplain red mudstone. The contact with the overlying Hampden Basalt is especially well exposed. The measured section of the 33 m of exposed basalt details 8 lava flows (Chapman, 1965, Fig. 12). Watch out for high-speed cars - this is a major east-west road.
74.0	3.3	Underpass beneath I-91 with Hampden Basalt on left.
74.6	0.6	Turn left at traffic lights on Coles Road (route 217).
75.3	0.7	Turn left on North Road. Just ahead is underpass beneath I-91.
76.1	0.8	Junction with Pasco Hill Road. Drive straight ahead on road marked "dead end."
76.7	0.6	Turn around at dead end of road.
77.0	0.3	Park on right just before bridge over brook. Trail to stop 7 begins on left (east) side of road about 10 m beyond (south) of bridge. Follow trail to large roadcuts in unfinished access lanes to I-91.
		STOP 7. EAST BERLIN FORMATION, CROMWELL.
		Proceed south again on "dead end" road.

77.1	0.1	On left are red sandstone and mudstone in the East Berlin Formation.
77.4	0.3	Junction with Pasco Hill Road. Proceed directly ahead on North Road.
78.2	0.8	Turn right on Coles Road
78.8	0.6	Turn left at traffic lights on route 72 east.
79.5	0.7	At traffic lights proceed straight ahead following sign to routes 3 and 9. Do not follow route 72, which turns right.
79.8	0.3	Turn right (south) on route 9.
83.4	3.6	Turn right at sign for "route 66 west and Wesleyan University."
		Proceed up the street to traffic lights. Go straight ahead on route 66 west.
83.9	0.5	Turn left on High Street at sign to Wesleyan University.
84.4	0.5	Turn right on Church Street. Pass Science Building on left and Olin Library on right.
84.8	0.4	Turn left on Pine Street and park beside the Science Building.
		End of field trip.