DEPOSITIONAL ENVIRONMENT OF THE ONEONTA FORMATION (CATSKILL FACIES), NEAR UNADILLA, NEW YORK

JOHN S. BRIDGE AND LORRAINE E. DINGMAN

Department of Geological Sciences
State University of New York
Binghamton, New York 13901

INTRODUCTION

A large roadside exposure on Interstate 88 near Unadilla (Fig. 1) offers a unique opportunity for detailed study of both vertical and lateral lithofacies variations in the Upper Devonian Catskill facies. The rocks are assignable to the Oneonta Formation (Fletcher, 1963; Rickard, 1975), and the postulated early-middle Frasnian age (*Polygnathus asymmetricus* conodont zone) of this formation is supported independently by the miospore studies of J. B. Richardson (pers. comm.). This paper presents a detailed description and interpretation of the lithofacies, and a brief comparison with other published studies of Catskill-facies sedimentology. For convenience, the rocks were separated into two main facies associations which can usually be clearly distinguished in the field.

THE SEDIMENTARY LITHOFACIES

Facies Association 1

This association comprises mainly fine- to very fine-grained, moderately to well-sorted sandstones, medium-dark to medium-light gray (N4-N6) in color (Goddard *et al.*, 1970), and muddy sublitharenite (McBride, 1963) in composition. Gravel-sized intraformational sandstone, siltstone, and mudstone fragments are common, locally concentrated as breccias above erosion surfaces. These breccias also contain plant remains, fish fragments, and reworked calcareous concretions. In color, they range from dark yellow orange (10YR6/6) to medium-dark gray (N3), depending on the type of fragments present. Discontinuous layers of medium-dark to olive gray (5Y4/1) siltstone and mudstone also occur locally.

The dominant internal structures are small- and large-scale trough cross-stratification, and horizontal stratification. Small-scale trough cross-stratification is usually the climbing type (Type A of Jopling and Walker, 1968); set thickness is about 1 to 3 cm, and trough width, 3 to 6 cm. Thickness of the large-scale cross-bed sets lies in the range 5 to 30 cm and trough widths are 10 cm to 3m. Horizontal stratification (which may actually have a slight inclination) is comprised of mm-scale laminae and is typically associated with parting lineation. Commonly interbedded with horizontal stratification are large-scale nearly planar cross-bed sets.

The spatial organization of texture and internal structure within the four exposed sandstone bodies of facies association 1 is complex, but systematic. Major erosion surfaces with lateral extent on the order of hundreds of meters underlie each sandstone body, but may also occur within a sandstone body, imparting a 'multistory' character (Fig. 2). Immediately overlying beds are usually intraformational breccias, with clasts up to decimeters across set in a fine to very fine sandstone matrix.

Within each sandstone body it is usually possible to recognize bedsets (sedimentation units) up to 1 meter thick, which can be traced laterally for many tens of meters (Fig. 3). Commonly, the base of a bedset is a relatively minor erosion surface with intraformational breccia, variably developed along the bedset. For instance, the breccias lower in a sandstone body are typically better developed and have larger clasts (up to 40cm). The top of each bedset may fine locally to siltstone or mudstone; sandstone bedsets overlying the thicker occurrences of fines (Fig. 3) may display load-casted as well as eroded bases. Sedimentary structures may also vary vertically within a bedset, for example horizontal stratification may be overlain by small-scale cross-stratification and finally capped by ripple marks (Fig. 3). Figure 3 further shows that the thickness, texture, and internal structure of any bedset may vary laterally as well as vertically.

Bedsets may be inclined up to about 10° with respect to the base of a sandstone body (i.e. epsilon cross-stratification, Allen, 1963) or occur as concave upwards channel-fills, evident in sections transverse to paleocurrents (Fig. 2, body 1 and 2). Where paleocurrents essentially parallel the outcrop, bedsets are broadly parallel to the basal erosion surface (e.g. body 3). As explained below the orientation of the bedsets may vary within <u>and</u> between the different stories of a sandstone body.

Within each story of a sandstone body is large-scale vertical and lateral variation in bedset orientation, thickness, texture, and internal structure (Figs. 2 and 3). In some instances, there is an overall fining-upwards tendency associated with a systematic variation in internal structure; however, this is not ubiquitous. The uppermost story of sandstone body 1 (Figs. 2 and 3) shows epsilon cross-stratification becoming steeper to the northeast, and finally changing to a channel fill; there is an associated change of vertical facies sequence. The immediately adjacent channel fill is also complex, with more fines in the base of the channel than higher up.

Paleocurrents are consistently unidirectional for a given story, and the range of values is less than 30° (Figs. 2 and 3). Mean directions <u>between</u> different stories may be distinctly different, thereby assisting in the recognition of separate stories in each sandstone body (c.f. Puigdefabregas and Van Vliet, 1978).

In the top meter of each sandstone body (and in the overlying facies association 2) are abundant siltstone casts of <u>in situ</u> and transported plant roots and stems, about 1 to 3 cm across, and up to 30 cm long (Fig. 5 E, F).

<u>Facies Association 2</u>

Facies association 2 comprises complex interbedding of very fine grained, moderately to poorly sorted sandstones, siltstones, and mudstones.

The color of sandstones varies from medium gray (N5) to grayish red (10R4/2), with siltstones and mudstones ranging from medium-dark gray (N4) to grayish red (10R4/2).

The beds are arranged in sets, centimeters to decimeters thick; each fines upwards. Bases of the bedsets are erosional with relief of up to 5 cm and common intraformational rock fragments up to 10 cm across. The fragments commonly match immediately underlying beds. Flow-parallel furrows (about 5 cm deep and 10 cm across) and load casts are locally associated with erosional bases.

Typical vertical variation in texture and sedimentary structure of facies association 2 is summarized in Figure 4. Large-scale cross-stratification is most commonly of the trough-type, set thicknesses 5 to 20 cm thick, and troughs 10 to 100 cm across. Small-scale cross-stratification (sets 0.5 to 3 cm thick) may be trough (widths 1 to 5 cm) or planar type; troughs are most common. Climbing types correspond to type A of Jopling and Walker (1968). Asymmetrical and symmetrical ripple marks up to a centimeter high and a few centimeters long are common on bedding surfaces (Fig. 5G). Desiccation cracks are ubiquitous, and rare raindrop imprints can be seen (Fig. 5 C,D,G).

Large vertical (and partly horizontal) burrows with roughly circular sections up to 10 cm across occur mainly in siltstones and mudstones. They are commonly filled with relatively coarse sediment and mud chips. Two varieties of surface trails (as yet unidentified) have also been found (Fig. 5 A,B).

Drifted and _in situ_ plant remains, siltstone casts of stems and rootlets, are common in this facies association. The casts and immediately surrounding rock may display a local color change to greenish gray (5G6/1). Concretions, moderate to dark yellowish brown in color (5YR3/4 to 10YR4/2), occur in very fine sandstones and siltstones and are commonly associated with large branching rootlet traces (Fig. 5F). They are irregularly globular, up to 3 cm across, and are similar to type A of Allen (1974).

The bedsets may occur as laterally extensive sheets (tens or hundreds of meters) or filling broad channels (Fig. 2). The channels are typically asymmetrical in section, 3 to 30m across and 0.3 to 1m in maximum depth. Thickness, texture and internal structures of the bedsets change laterally in both sheets and channels (Fig. 3). It appears that the coarser grained representatives of this association occur directly on top of facies association 1 (Fig. 3), and in sandstone body 2 there is lateral transition into association 1. Also there are two examples of small, isolated coarse-grained channel fills, one of which has an asymmetrical section and adjacent epsilon cross-stratified sandstone (Figs. 2 and 3).

Paleocurrents from large-scale cross-stratification, parting lineation, and channels indicate unidirectional flow, subparallel to those from facies association 1 (Fig. 3). Directional variation between bedsets of up to 30° is common, but small-scale cross-stratification and asymmetrical ripple marks are more variable. Crestlines of symmetrical ripples marks show no systematic orientation (Fig. 3).

INTERPRETATION OF LITHOFACIES

Facies Association 1

The sedimentary characteristics of facies association 1 strongly suggest deposition in river channels migrating laterally across alluvial plains. The major laterally extensive erosion surfaces are ascribed to lateral migration of erosion-dominated areas deep in river channels. It is well known that slumped fragments of alluvial-plain sediment from adjacent retreating cut-banks accumulate as lag gravels in these channel deeps. Accordingly, the overlying sandstones and breccias represent the deposits of laterally migrating channel bars or coarse-grained channel fills.

The large-scale vertical sequence of texture and internal structure in each sandstone story has parallels in some modern single-channel streams with sinuous talwegs (e.g. Harms et al., 1963; Davies, 1966; Sarkar and Basumallick, 1968; Bernard et al., 1970; Shelton and Noble, 1974). Large- and small-scale trough cross-stratification record the downstream movement of three-dimensional dunes and ripples, whereas horizontal stratification was deposited on upper-stage plane beds. Climbing ripples and upper-stage plane beds generally indicate the presence of significant suspended load as well as bedload transport within the channels. The vertical juxtaposition of bed configurations and textures reflects their original spatial distribution on the laterally migrating inclined bar surface, in response to locally variable velocity, depth, and slope. It is worth remembering that the facies sequence predicted by the well-known fining-upwards model (e.g. Allen, 1970) will only be present in the downstream part of channel bars (Jackson, 1976; Bridge, 1978).

Perhaps the strongest support for lateral deposition on a channel bar comes from epsilon cross-stratification, where the low-angle stratification surfaces dipping approximately normal to local paleocurrent direction represent ancient bar surfaces. Small-scale vertical facies variation within the individual bedsets that defines the epsilon cross-stratification can also be seen in modern channel-bar deposits (see references above). The upward decrease in grain size and change in internal structure in each bedset probably records deposition during falling flow stages. Desicatted low-flow deposits may be incorporated as intraformational fragments in the deposits of an ensuing flood. The restriction of the in-situ plants to topographically high areas of the bars, and the relative lack of finest sediment grades within the lower parts, suggests that the rivers were perennial.

The channel fills with asymmetrical cross-sections have well-exposed cut banks, supporting an interpretation as sinuous, laterally migrating channels. The sediment in the fills is relatively coarse, and the spatial facies variations within the fills are complex (see later discussion).

The multistory character of the sandstone bodies is expected when lateral bar migration is combined with net floodplain aggradation (Bluck, 1971; Bridge, 1975; Bridge and Leeder, 1979). Each story is a single bar deposit which has been overriden and eroded by a different channel segment and associated bar. The unidirectional paleocurrents in a single story are consistent with channel deposition; however, if the migrating channels had any curvature, it is reasonable to expect different mean paleocurrents from superimposed stories (Figs. 2 and 3).

Sandstone body 1 (Figs. 2 and 3) records evidence of particularly complex channel behaviour. Channel fill 1A is probably associated with the immediately adjacent bar deposit, by virtue of a common basal erosion surface and similar paleocurrent direction; however there is evidence of subsequent erosion of the bar deposits low in the channel. These bar deposits form the cut bank of channel 1B, thereby predating it. Both channels may have existed simultaneously for some time, however, being separated by the earlier bar deposits of channel 1A. The steepening of the epsilon cross-bed sets and change in facies as channel 1B migrated laterally is a result of change in hydraulics and channel geometry. This could result from either (a) observation of the same bar at different longitudinal positions in a given river bend (e.g. Bridge, 1978), or (b) changes in plan form and hydraulics during channel migration, such as increase in sinuosity and decrease in mean slope and velocity. The laterally extensive erosion surface (overlain by sandstone) that truncates the top of the epsilon sets appears to be associated with the last of the coarse-grained fill of channel 1B. Finer grained beds overlying this fill are laterally equivalent to beds only a few meters from the base of channel fill 1A, implying that channel 1A was at least partly open once 1B had filled up. In fact, Figure 2 shows evidence that a substantially shallower channel 1A was actively migrating in the final stages of filling.

A possible sequence of events to explain these complicated facies patterns is:

(a) lateral migration of channel 1A,
(b) chute cut-off and formation of channel 1B,
(c) lateral migration of channel 1B, while channel 1A was filling and partially eroding its previous bar deposits,
(d) gradual filling of channels 1A and 1B; the relatively coarse fills imply that the channels still carried an appreciable discharge, and
(e) diversion of more discharge back into channel 1A, with some renewed bank erosion and lateral deposition during the final filling stage.

Sandstone body 4 (Fig. 2) also shows lateral and vertical transition to a major channel fill, the exposed top parts of which are filled with bedsets of facies association 2.

Facies Association 2

Each bedset is interpreted as the deposit of a single flood on a floodplain. Lower bounding surfaces record some erosion of previously desiccated deposits prior to deposition. Large- and small-scale cross-stratification and horizontal stratification were produced by bed-load deposition of sand moving as dunes, ripples, or an upper-stage plane bed. However, textures and internal structures indicate substantial transfer of suspended sediment to the bed during deposition. The vertical sequence in a bedset records waning overbank flood velocities, with the siltstone and mudstones representing purely suspended-load deposition of the final flood stages. Wind action on ponded areas is recorded by wave ripples, and the raindrop impressions and abundant desiccation cracks indicate subsequent exposure of the floodplain surface. Concretions represent the calcareous soil horizons common beneath floodplains in arid and semi-arid climates (see Allen, 1974, for summary).

Thicknesses of these postulated flood deposits are consistent with observations from modern rivers (e.g. Bridge and Leeder, 1979). The coarser and thicker channel-filling facies are similar to modern crevasse-channel and splay deposits (e.g. Kruit, 1955; Coleman, 1969; Singh, 1972). The two isolated coarse-grained channel fills immediately below sandstone body 2 are specifically interpreted as crevasse channels that were probably open during the deposition of sandstone body 1. The more sheet-like facies closely resemble modern levee deposits (e.g. Fisk, 1947; Singh, 1972; Ray, 1976), although some of these bedsets may have occurred during the final stages of filling of major channels (Figs. 2 and 3). The locally increased dip of bedsets to the west of sandstone body 4, and their sheet-like geometry, reflects the topographic dip of a levee away from a major channel. The finest bedsets have parallels in the flood-basin deposits of modern floodplains (Jahns, 1947; Allen, 1965).

The most abundant _in situ_ and drifted plant remains occur in facies that are interpreted here as upper channel bar, levee, and crevasse deposits; that is, topographically high areas immediately adjacent to major channels. It appears that during floods much plant debris was buried rapidly before the fragments could be transported into the flood-basin (see also Bridge _et al_, 1980). The restriction of the major flora to areas near channels may be due to the presence of a locally high water table near perennial streams in a semi-arid climate. The close association of many concretions with _in situ_ plant roots suggests that concentration of carbonate in the groundwater is associated with loss of water to the atmosphere through plants. The color difference between red overbank and gray channel deposits is most probably due to differences in the oxygenation of the groundwater during early diagenesis.

The origin of the trace fossils cannot be ascertained conclusively due to lack of preservation of organisms responsible. The larger burrows are similar to those interpreted elsewhere as dipnoan aestivation burrows (Woodrow, 1968) or maybe even arthropod burrows (Rolfe, 1980). Arthropods were probably responsible for surface traces shown in Figure 5A and B.

Sandstone body 2 (Fig. 2) shows a lateral transition from a postulated crevasse-splay deposit to a single major channel with evidence of lateral accretion; the channel then became progressively shallower. This behaviour is consistent with the known tendency of streams to change course during major floods by diversion of discharge through crevasse channels (i.e. avulsion). In this case, the diverted channel did not develop a major channel belt, as represented by the other sandstone bodies. The spatial organization of facies associations 1 and 2 can, however, be explained by avulsion of major channels whilst net floodplain aggradation proceeds (Allen, 1965, 1974; Bridge and Leeder, 1979). Many authors suggest that the presence of appreciable fine floodplain sediment is a result of a degree of channel belt stability present only in meandering streams. Bridge and Leeder (1979) show theoretically that the relative proportion of channel and overbank deposits in an alluvial succession is unlikely to be diagnostic of channel planform. It is the detailed nature of the channel sandstone bodies that supports the single curved-channel interpretation here.

COMPARISON WITH OTHER PUBLISHED INTERPRETATIONS

The foregoing discussion agrees with the generally accepted view that the Catskill facies is alluvial in origin (e.g. Shepps, 1963; Rickard, 1975). The Oneonta Formation in this region has been more specifically interpreted as meandering-river deposits of a lowland alluvial plain (Woodrow and Fletcher, 1967; Johnson and Friedman, 1969). Although many of the sedimentary features described here have been recognized in previous detailed studies of Catskill facies (e.g. Allen and Friend, 1968; Johnson and Friedman, 1969; Allen, 1970) the observations at the Unadilla outcrop have allowed a much more refined and unambiguous interpretation than has been possible to date. Paleoclimatic implications of this study concur with Woodrow et al.'s (1973) reconstruction of the Upper Devonian paleogeography of this area.

ACKNOWLEDGEMENTS

We are most grateful to Paul Enos, Jim Sorauf, and Dick Beerbower for reviewing the manuscript and to Olga Kurty for drafting the diagrams.

REFERENCES CITED

Allen, J.R.L., 1963, The classification of cross-stratified units, with notes on their origin: Sedimentology, v. 2, p. 93-114.

____________, 1965, A review of the origin and characteristics of Recent alluvial sediments: Sedimentology, v. 5, p. 89-191.

____________, 1970, Studies in fluviatile sedimentation: a comparison of fining-upwards cyclothems with special references to coarse-member composition and interpretation: Jour. Sed. Petrology, v. 40, p. 298-323.

____________, 1974, Studies in fluviatile sedimentation: implications of pedogenic carbonate units, Lower Old Red Sandstone, Anglo-Welsh outcrop: Geol. Jour., v. 9, p. 181-208.

Allen, J.R.L., and Friend, P. F., 1968, Deposition of the Catskill facies, Appalachian region: with notes on some other Old Red Sandstone basins, in Klein, G.deV., ed., Late Paleozoic and Mesozoic continental sedimentation, northeastern North America: Geol. Soc. America Spec. Pap. 106, p. 21-74.

Bernard, H. A., Major, C. F., Parrott, B. S., and LeBlanc, R. J., 1970, Recent sediments of southeast Texas: Univ. Texas, Austin, Bureau Econ. Geol. Guidebook 11.

Bluck, B. J., 1971, Sedimentation in the meandering River Endrick: Scott. Jour. Geol., v. 7, p. 93-138.

Bridge, J. S., 1975, Computer simulation of sedimentation in meandering streams: Sedimentology, v. 22, p. 3-43.

____________, 1978, Palaeohydraulic interpretation using mathematical models of contemporary flow and sedimentation in meandering channels, in Miall, A. D., ed., Fluvial Sedimentology: Can. Soc. Petrol. Geol. Mem. 5, p. 723-742.

Bridge, J. S., and Leeder, M. R., 1979, A simulation model of alluvial stratigraphy: Sedimentology, v. 26, p. 617-644.

Bridge, J. S., Van Veen, P. M., and Matten, L. C., 1980, Aspects of the sedimentology, palynology and palaeobotany of the Upper Devonian of southern Kerry Head, Co. Kerry, Ireland: Geol. Jour., v. 15, p. 143-170.

Coleman, J. M., 1969, Brahmaputra River: Channel processes and sedimentation: Sediment. Geol., v. 3, p. 129-239.

Davies, D. K., 1966, Sedimentary structures and subfacies of a Mississippi River point bar: Jour. Geol., v. 74, p. 234-239.

Fisk, H. N., 1947, Fine-grained alluvial deposits and their effects on Mississippi River activity: Mississippi River Commission, Vicksburg, Miss., U.S.A., 82p.

Fletcher, F. W., 1963, Regional stratigraphy of Middle and Upper Devonian non-marine rocks in southeastern New York, in Shepps, V. C., ed., Symposium on Middle and Upper Devonian stratigraphy of Pennsylvania and adjacent states: Penn. Geol. Survey Bull. G-39, p. 25-41.

Goddard, E. N., Trask, P. D., Deford, R. K., Rove, O. N., Singewald, J. T., and Overbeck, R. M., 1970, Rock-color chart: Geol. Soc. America.

Harms, J. C., McKenzie, D. B., and McCubbin, D. G., 1963, Stratification in modern sands of the Red River, Louisiana: Jour. Geol., v. 71, p. 566-580.

Jackson, R. G., 1976, Depositional model of point bars in the lower Wabash River: Jour. Sed. Petrology, v. 46, p. 579-594.

Jahns, R. H., 1947, Geologic features of the Connecticut Valley, Massachusetts, as related to recent floods: U. S. Geol. Surv. Water Supply Paper 996, 158p.

Johnson, K. G. and Friedman, G. M., 1969, The Tully clastic correlatives (Upper Devonian) of New York State: a model for recognition of alluvial, dune (?), tidal, nearshore (bar and lagoon), and off-shore sedimentary environments in a tectonic delta complex: Jour. Sed. Petrology, v. 39, p. 451-485.

Jopling, A. V. and Walker, R. G., 1968, Morphology and origins of ripple-drift cross-lamination, with examples from the Pleistocene of Massachusetts: Jour. Sed. Petrology, v. 38, p. 971-984.

Kruit, C. C., 1955, Sediments of the Rhone delta. Grain size and microfauna: Verhandel. Koninkl. Ned. Geol. Mijnbouwk. Genoot., Geol. Ser. 15, p. 357-514.

McBridge, E. F., 1963, A classification of common sandstones: Jour. Sed. Petrology, v. 33, p. 664-669.

Puigdefabregas, C. and Van Vliet, A., 1978, Meandering stream deposits from the Tertiary of the southern Pyrenees, in Miall, A. D., ed., Fluvial sedimentology: Can. Soc. Petrol. Geol. Mem. 5, p. 469-485.

Ray, P. K., 1976, Structure and sedimentological history of the overbank deposits of a Mississippi River point bar: Jour. Sed. Petrology, v. 46, p. 788-801.

Rickard, L. V., 1975, Correlation of the Silurian and Devonian rocks in New York State: New York State Museum and Science Service, Map and Chart Series No. 24, 16p.

Rolfe, W. D. I., 1980, Early invertebrate terrestrial faunas, in Panchen, A. L., ed., The terrestrial environment and the origin of land vertebrates: London, Academic Press, p. 117-157.

Sarkar, S. K. and Basumallick, S., 1968, Morphology, structure, and evolution of a channel island in the Barakar River, Barakar, West Bengal: Jour. Sed. Petrology, v. 38, p. 747-754.

Shelton, J. W. and Noble, R. L., 1974, Depositional features of braided-meandering stream: Am. Assoc. Petrol. Geol. Bull., v. 58, p. 742-749.

Shepps, V. C., ed., 1963, Symposium on Middle and Upper Devonian stratigraphy of Pennsylvania and adjacent states: Penn. Geol. Survey Bull. G-39.

Singh, I. B., 1972, On the bedding in the natural-levee and the point-bar deposits of the Gomti River, Uttar Pradesh, India: Sediment. Geol., v. 7, p. 309-317.

Woodrow, D. L., 1968, Devonian Dipnoan aestivation cylinders (abs.): Geol. Soc. America Spec. Pap. 121, p. 383-384.

Woodrow, D. L., and Fletcher, F. W., 1967, Late Devonian paleogeography in southeastern New York and northeastern Pennsylvania, in Oswald, D. H., ed., International Symposium on the Devonian System, Calgary, p. 1327-1334.

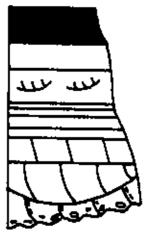
Woodrow, D. L., and Ahrnsbrak, W. F., 1973, Paleogeography and paleoclimate at the depositional sites of the Devonian Catskill and Old Red Facies: Bull. Geol. Soc. Am., v. 84, p. 3051-3064.

ROAD LOG

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Trip starts at Bartle-Drive (main) exit of SUNY-Binghamton campus. Turn east on to NY 434 (Vestal Parkway) towards Binghamton.
3.0	3.0	Bear to the right off NY 434 immediately after crossing the State Street Bridge in Binghamton. At this point, follow the road signs to NY 17 and I 81, but subsequently stay in the lane for NY 7N and I 88 towards Oneonta.
9.0	6.0	Chenango Bridge; NY 7N becomes I 88. Continue on I 88.
45.0	36.0	Stop at roadside exposure on right-hand (south) side of I 88 just past Unadilla exit.

STOP. ONEONTA FORMATION, CATSKILL FACIES.

Key to Figures 2-4



Siltstone or mudstone
Indistinct bedding
Small-scale cross-stratified sandstone to silty sandstone
'Horizontal' stratified sandstone or silty sandstone
Large-scale planar cross-stratified sandstone
Large-scale trough cross-stratified sandstone
Scoured surface overlain by intra-formational breccia

Epsilon cross-stratification
Asymmetrical ripple marks
Symmetrical ripple marks
Mean paleocurrent azimuths (N to top of page) derived from:
a) Channels
b) Large-scale cross-stratification and parting lineation
c) Small-scale cross-stratification or asymmetrical ripple marks

Load casts and flame structures
Raindrop imprints
Desiccation cracks
Concretions
Burrows
Tracks and trails
Drifted and insitu plant remains (various sizes)
Fish bone fragments

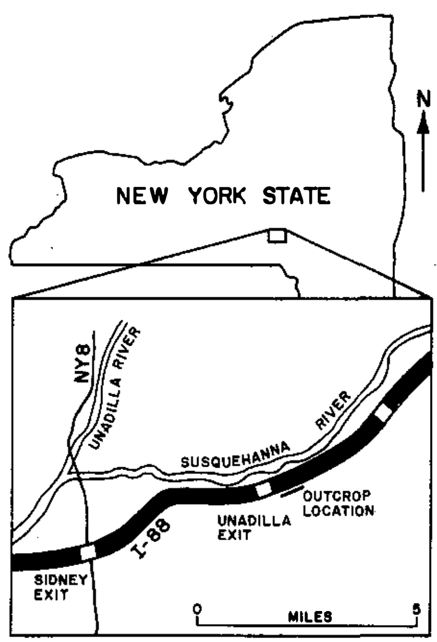
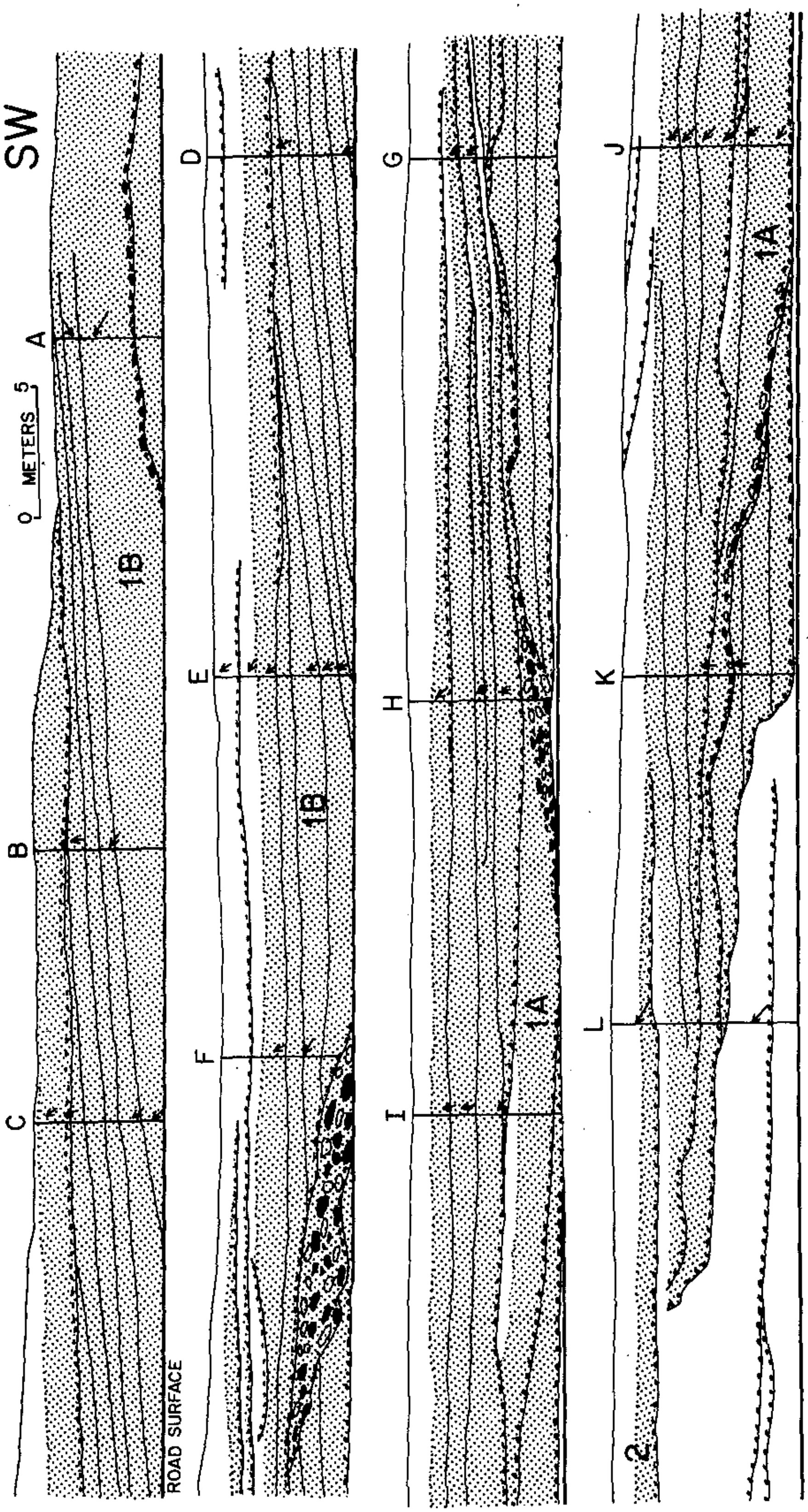
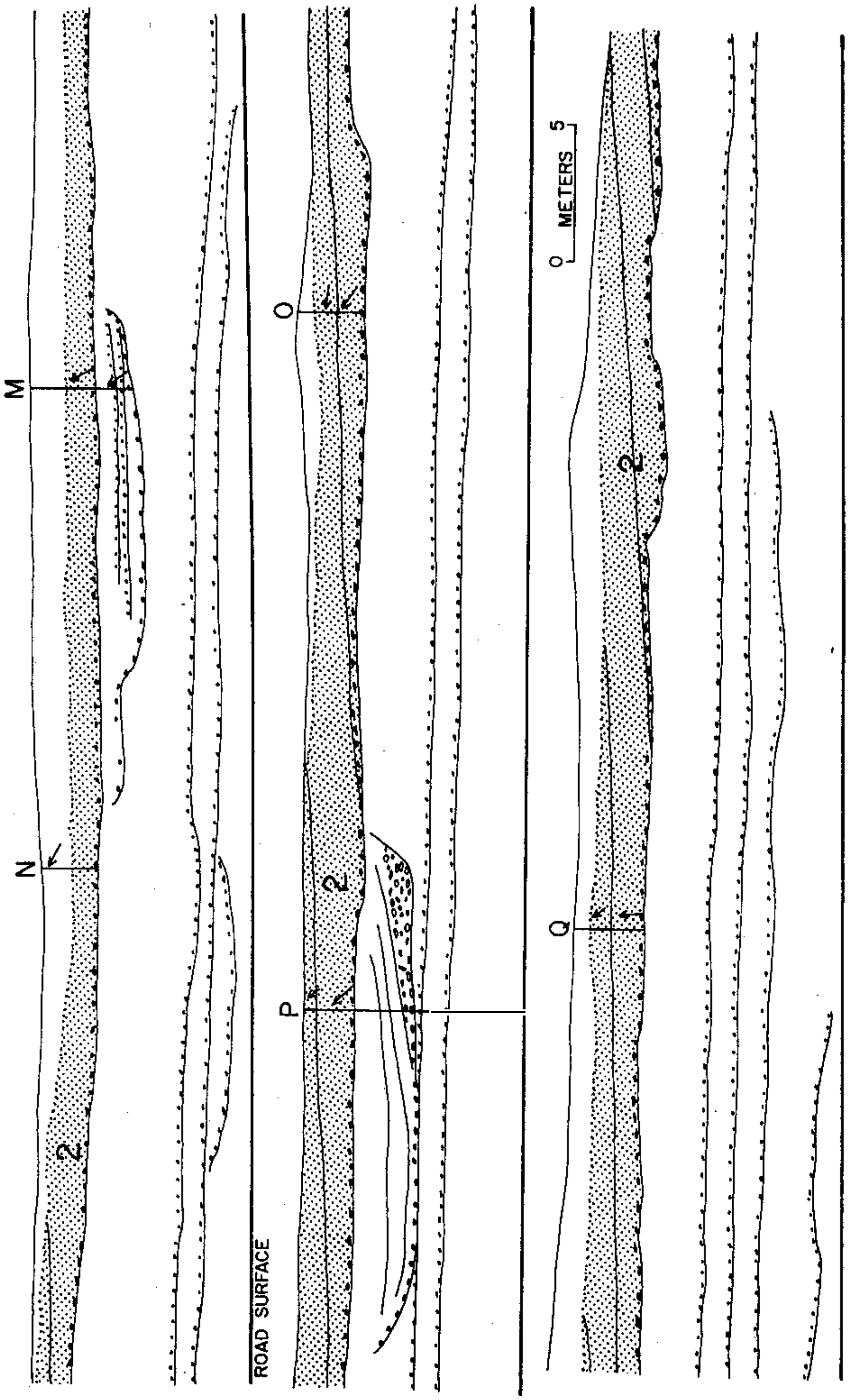
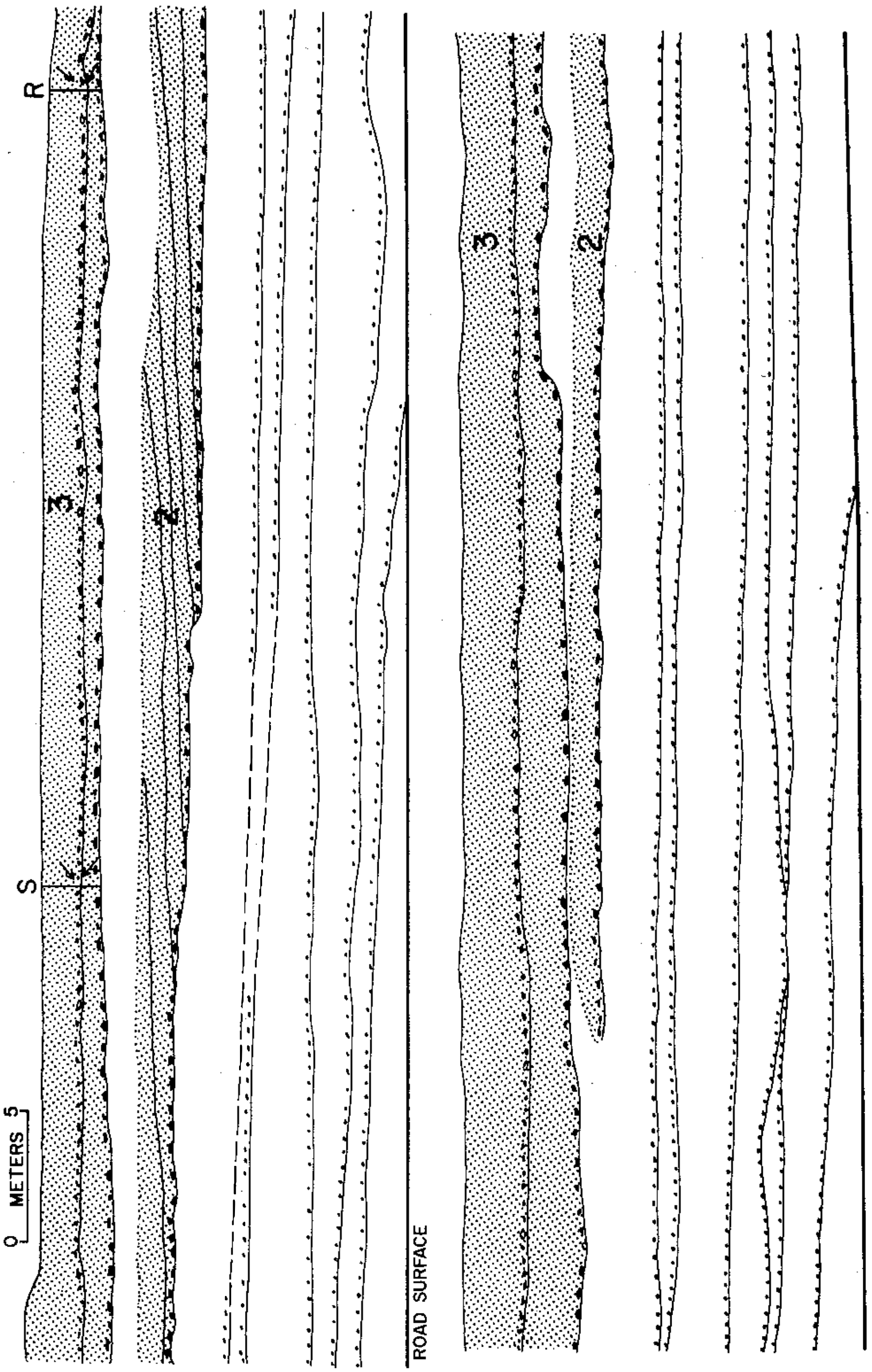


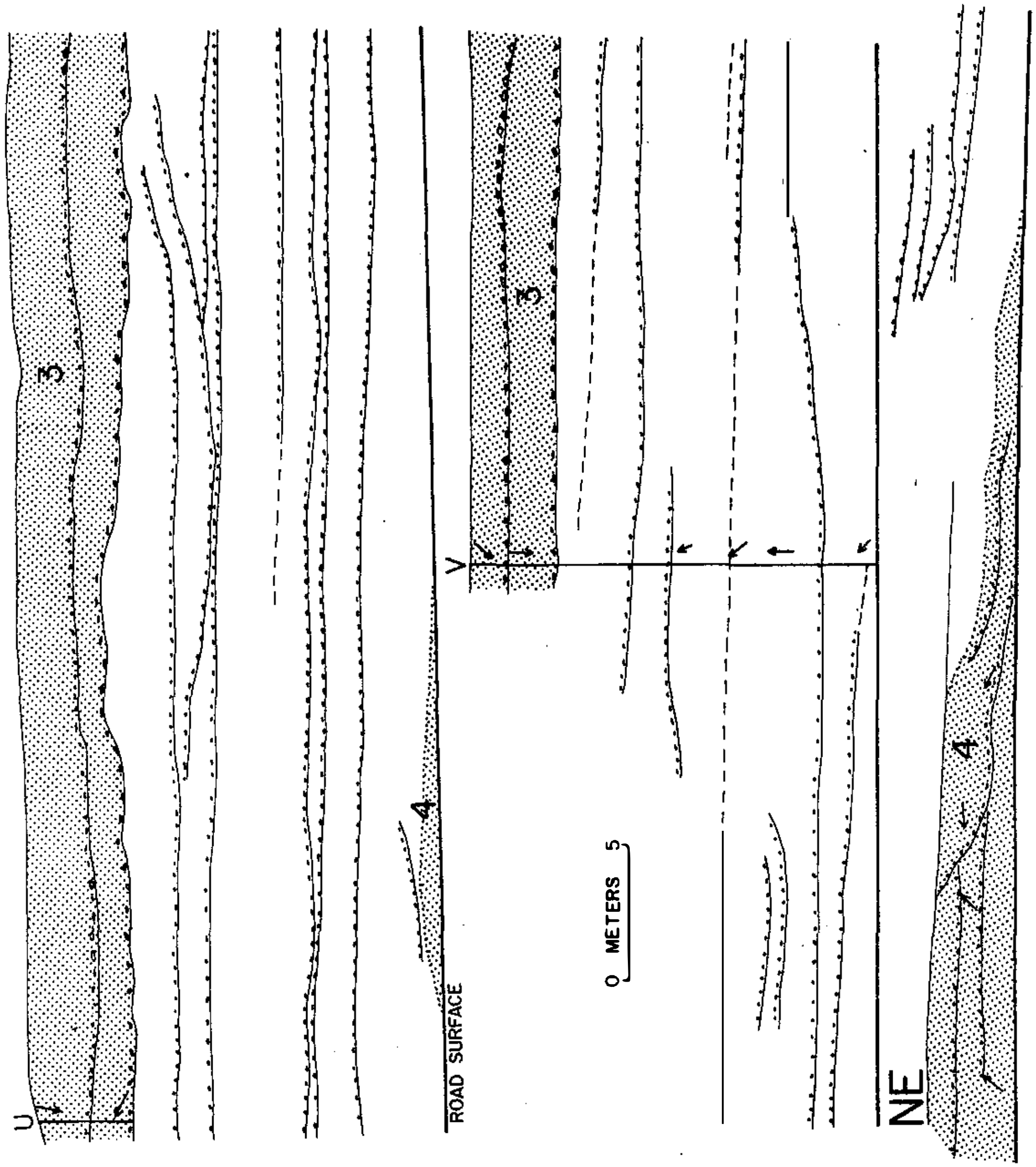
FIG. 1 - Location map of outcrop.

FIG. 2 - Scale diagram of outcrop studied, showing major bedding features and facies associations. The diagram represents a continuous section, starting at the southwestern edge of the outcrop (nearest Binghamton) and finishing at the northeastern edge. The vertical scale is not strictly correct for the upper part of the outcrop because of some distortion in the photographs from which the diagram was drawn; however true vertical thicknesses can be obtained from detailed measured sections (Fig. 3), the positions of which are marked by letters. Stippled areas are sandstone bodies of facies association 1, with some important interbedded shales unstippled. Facies association 2 is unstippled, except for parts of sand body 2 that are transitional with facies association 1. Only representative bedding surfaces are marked, the size of the overlying ornament reflecting degree of erosion and development of intraformational breccia. Paleocurrent azimuths (north to top of page) are shown; the azimuth of the outcrop is approximately N50^{o}E. Individual sandstone bodies are numbered for reference from text.


SW
0 METERS 5
A
B
C
D
E
F
G
H
I
J
K
L
1B
1B
1A
1A
2
ROAD SURFACE


M
N
2
ROAD SURFACE
O
P
2
0 METERS 5
2
Q


R
S
3
2
3
2
0 METERS 5
ROAD SURFACE


U
3
4
ROAD SURFACE
V
3
0 METERS 5
4
NE

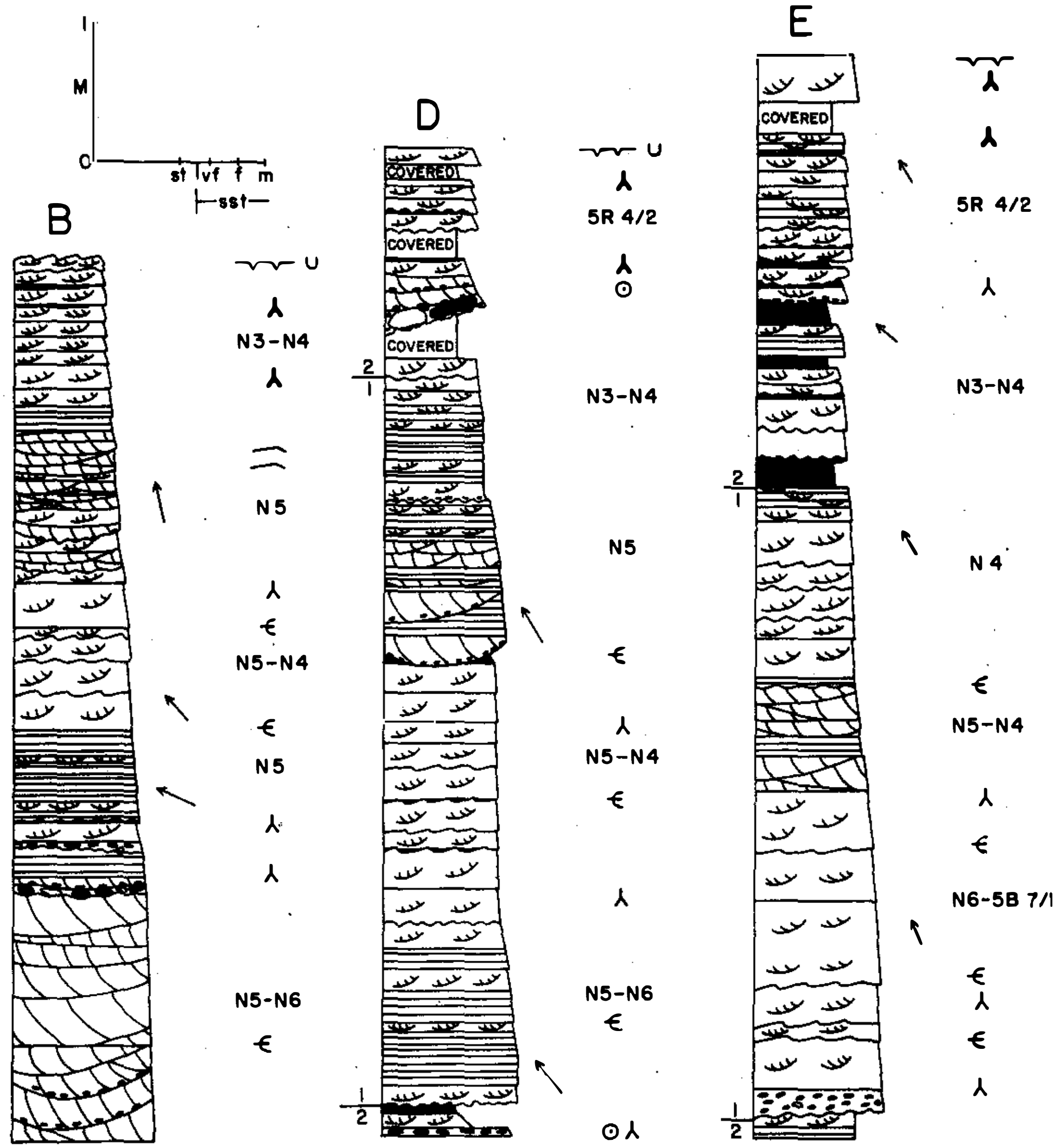
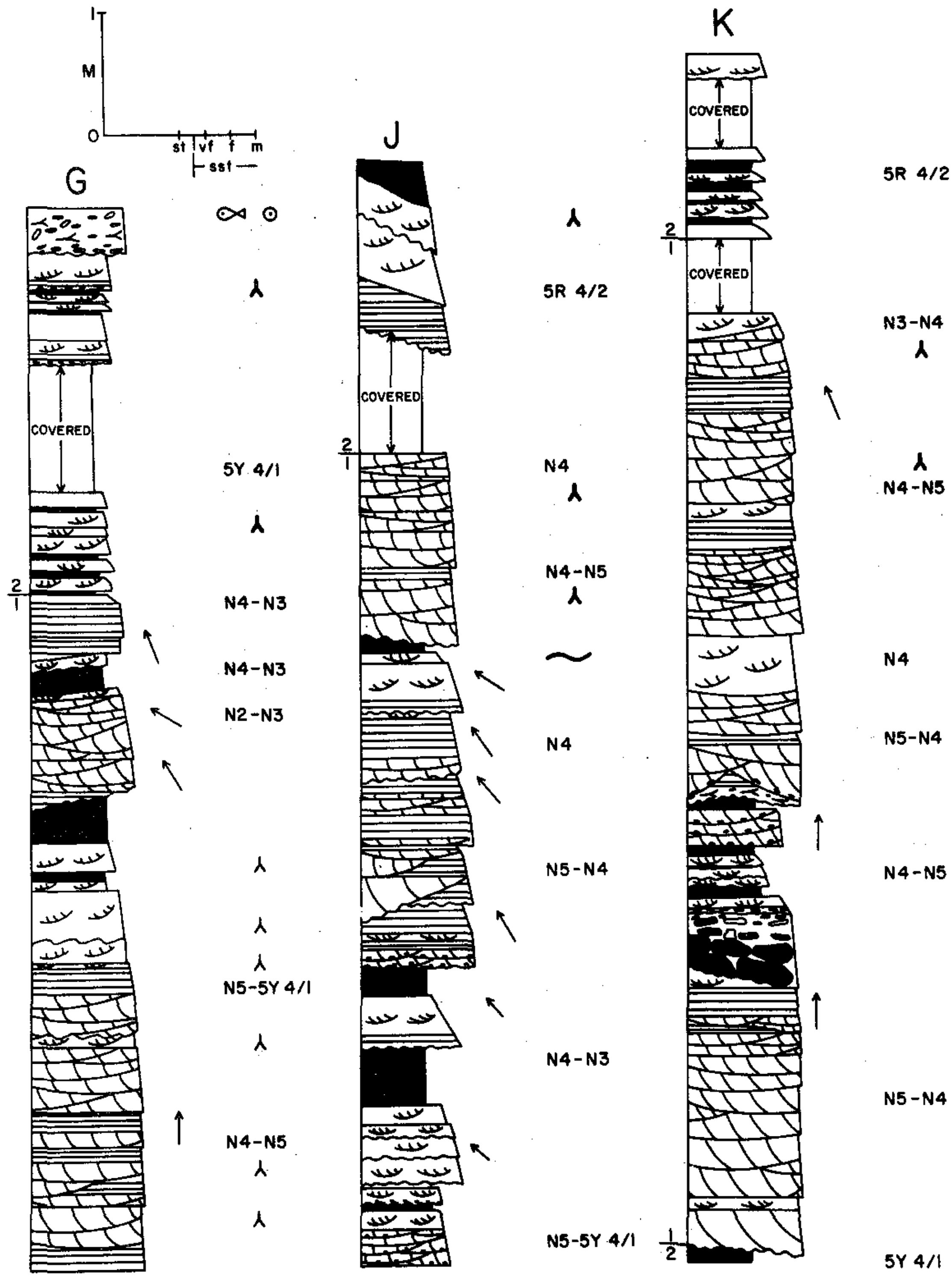
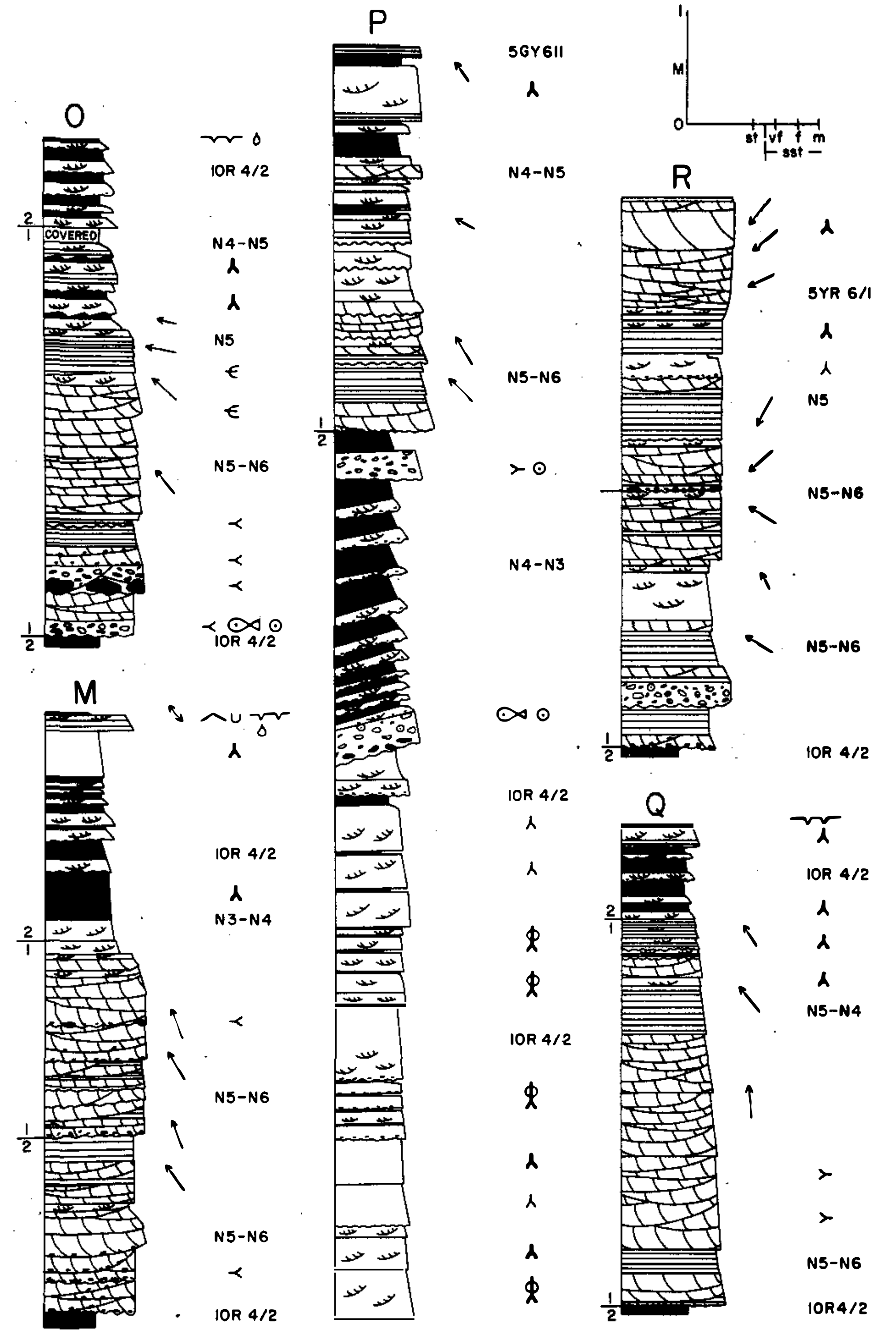


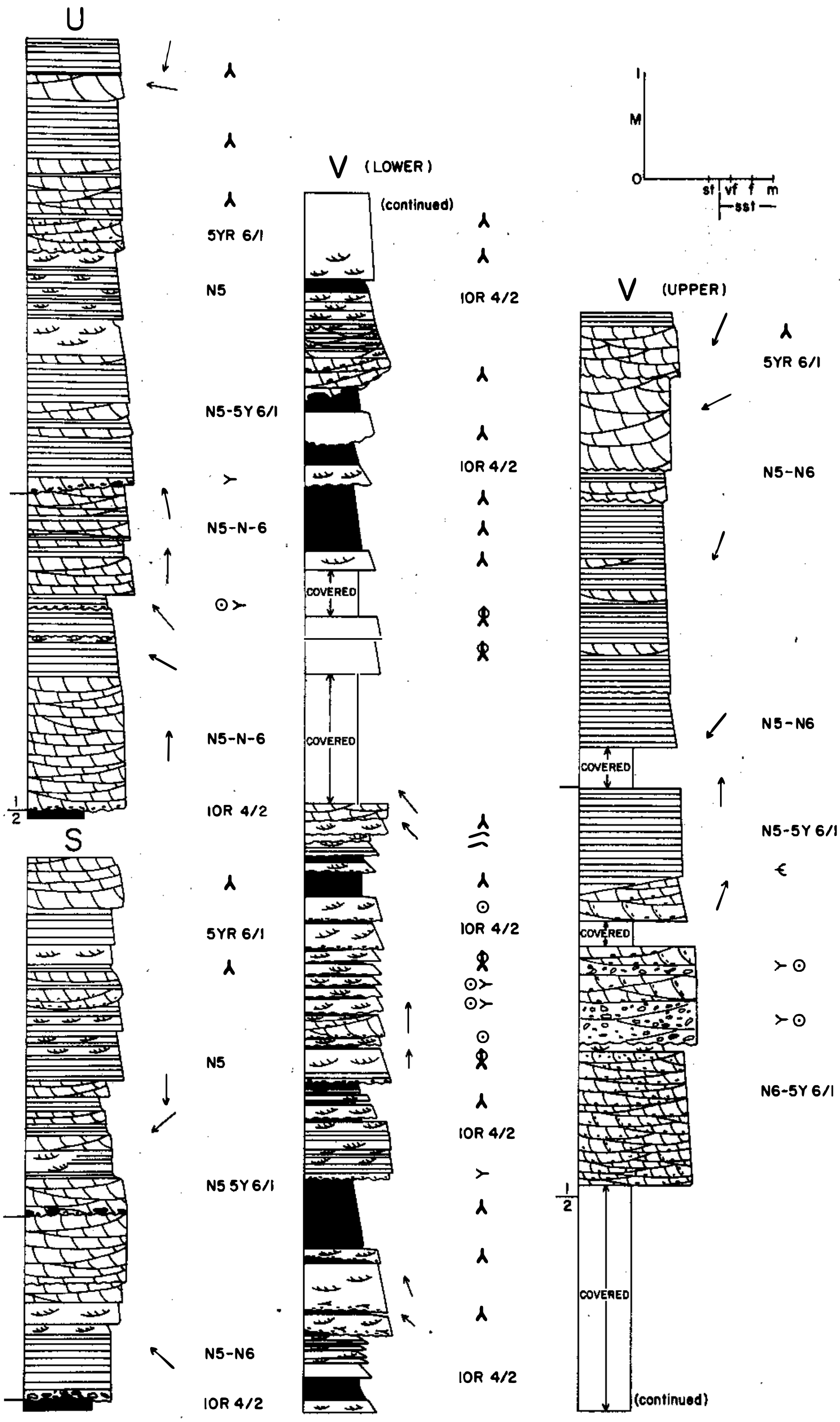
FIG. 3 - Selected detailed vertical sections, as shown on Fig. 2.
See legend. Paleocurrents (north to top of page) and additional sedimentary properties are shown to right of each graphic log. Facies association given to left of each log.



1
M
0
st
vf
f
m
sst
G
J
K
COVERED
COVERED
COVERED
COVERED
2
1
5Y 4/1
N4-N3
N4-N3
N2-N3
N5-5Y 4/1
N4-N5
5R 4/2
N4
N4-N5
N4
N5-N4
N4-N3
N5-5Y 4/1
5R 4/2
N3-N4
N4-N5
N4
N5-N4
N4-N5
N5-N4
5Y 4/1



O
10R 4/2
2
1
COVERED
N4-N5
N5
N5-N6
1
2
10R 4/2
M
10R 4/2
N3-N4
2
1
N5-N6
1
2
N5-N6
10R 4/2
P
5GY6/1
N4-N5
N5-N6
1
2
N4-N3
10R 4/2
10R 4/2
1
M
0
st
vf
f
m
sst
R
5YR 6/1
N5
N5-N6
N5-N6
1
2
10R 4/2
Q
10R 4/2
2
1
N5-N4
N5-N6
1
2
10R4/2


U
5YR 6/1
N5
N5-5Y 6/1
N5-N-6
N5-N-6
1/2
10R 4/2
S
5YR 6/1
N5
N5 5Y 6/1
N5-N6
10R 4/2
V (LOWER)
(continued)
10R 4/2
10R 4/2
COVERED
COVERED
10R 4/2
10R 4/2
10R 4/2
V (UPPER)
5YR 6/1
N5-N6
N5-N6
COVERED
N5-5Y 6/1
COVERED
N6-5Y 6/1
1/2
COVERED
(continued)
1
M
0
st
vf
f
m
sst

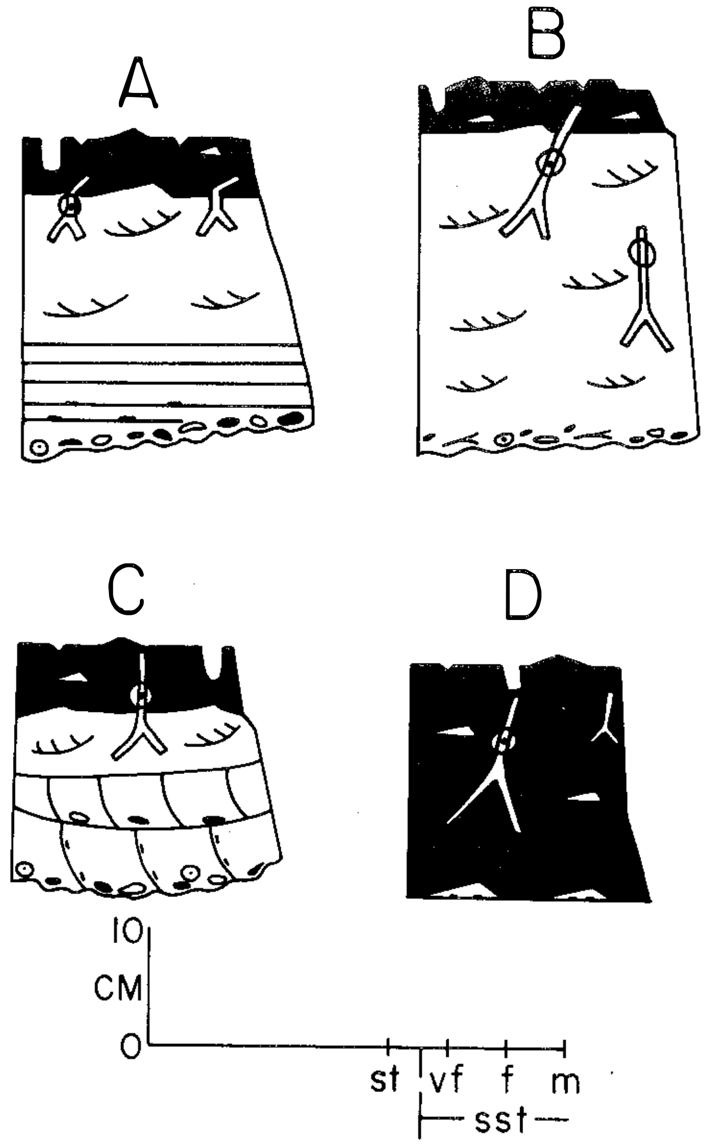


FIG. 4 - Typical small-scale facies sequences comprising facies association 2. See legend.

FIG. 5 - A) Top view of surface trace (facies association 2). B) Basal views of arthropod? trail (f.a.1, channel-fill 1A). C) Raindrop imprints, top surface (f.a.2). D) Large vertical burrows and desiccation cracks (f.a.2). E) Siltstone case of drifted plants on ripple-marked surface (f.a.1, top of 1B). F) Calcareous concretions associated with in situ plant roots (f.a.2). G) Desiccation cracks and symmetrical ripple marks (f.a.2). Black board scale is 10 cm in all photos.