SEDIMENTOLOGY AND REGIONAL IMPLICATIONS OF FLUVIAL QUARTZOSE SANDSTONES OF THE LEE FORMATION, CENTRAL APPALACHIAN BASIN

by

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(ABSTRACT)

Sedimentological analyses, including detailed facies characterization and lateral profiling, demonstrate deposition in a bedload-dominated fluvial system for the quartzose sandstones of Lee Formation. Internal (architectural) elements of the sandstones consist primarily of truncated channel-fill sequences. Individual channel elements, up to 20 meters thick, contain a complex hierarchy of bedform deposits. The principal internal component of channels were downstream-accreting (mid-channel?) macroforms; channel elements frequently contain deposits of more than one macroform. Reconstruction of the macroforms reveals accretion primarily by superposed bedforms that migrated down a low-angle front. Steeper, giant foresets, transitional along flow with the low-angle facies, indicate that the macroform episodically developed a steep slipface. Uppermost channels within the Rockcastle Member contain macroform elements with components of lateral accretion, interpreted as deposits of alternate bank-attached macroforms. Also recognized within channel-fill deposits are minor-channel, sandy-bedform, gravity-flow (attributed to bank slumping), and channel-bottom elements. The latter element is contained within a facies sequence that suggests rising- to flood- to waning-stage deposition. In general, deposition was probably during relatively high stage; little evidence of low-stage flow was recognized. Subordinate fine-grained facies are interpreted as levee and overbank deposits.

Strongly unimodal paleocurrents, lack of facies that suggest low-stage reworking and paucity of lateral-accretion features indicate deposition in a single-channel, low-sinuousity, system (i.e., a low braiding index). Fluvial architecture similar to that found in the Lee Formation has been previously explained by deposition in multi-channel, braided-river systems. However, the internal architecture of sandstone members is also consistent with a single-channel origin. Individual channels were temporarily confined, during which time the passage of several macroforms aggraded the channel. Position of the channel in the alluvial plain was largely controlled by avulsion of the river from fully aggraded channel belts to other areas of the plain. Calculations reveal that avulsion of a single-channel system across a wide alluvial plain is a plausible mechanism for building the sheet-like sandstone bodies of the Lee Formation. Spatial arrangement of individual sandstone members of the Lee Formation was probably controlled by tectonic processes. Episodic thrust-loading in the orogenic belt to the east and subsequent flexure of the crust in the foreland basin caused a step-wise progression of the river system towards the west.

Petrographic, sedimentologic and stratigraphic data indicate that source area and climate functioned as the primary controls on the mature composition of sandstones in the Lee Formation. Source areas were composed primarily of quartz-rich sedimentary rocks and were located chiefly to the northeast/north. A east/southeast source area supplied subordinate and low-grade metamorphic rock fragments. Intense weathering, associated with humid tropical climates, acted upon the detritus throughout the sedimentation cycle. Less important controls on composition were tectonics and transport/depositional processes that extended exposure of the sediments to the severe climatic conditions. Quartzose sandstones of the Lee Formation reflect lower rates of tectonic subsidence and greater recycling of sand-sized grains during transportation and temporary deposition on the alluvial plain, relative to lithic time equivalents to the east.

ACKNOWLEDGEMENTS

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Ken Eriksson, my advisor, has demonstrated remarkable patience and provided encouragement in guiding me through my dissertation. I would also like to thank my other committee members, Richard Bambach, Lynn Glover, Fred Read and Steve Scheckler for their help and support. Departmental employees have been a great help: Dean, Belinda, Karen, Linda, Carolyn, Pat and Ethel have all greatly assisted me and often went out of their way to lend a hand. Special thanks to Llyn Sharp, who taught me the photographic ropes, and is a good and very supportive friend. My other best friends at Tech have been fellow graduate students, who are, fortunately, too numerous to name. Ed Simpson, Charles Harris and Chis Fedo have provided me with technical expertise, scientific insight, and friendship during our tenures as officemates.

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ABSTRACT

Lateral profiling, including bounding surface and architectural-element analyses, are utilized to demonstrate a fluvial origin for the quartzose sandstones of Lee Formation. Internal architecture of the sandstones consists primarily of multistorey and multilateral channel bodies. Individual channel elements, up to 20 meters thick, contain a complex hierarchy of bedform deposits. The primary depositional element within the channels were downstream-accreting (mid-channel?) macroforms; channel elements frequently contain deposits of more than one macroform. Reconstruction of the macroforms reveals accretion primarily by superposed bedforms that migrated down a low-angle front. Steeper, giant foresets, transitional along flow with the low-angle facies, indicate that the macroform episodically developed a steep slipface. The uppermost channels within the Rockcastle Member contain macroform elements with components of lateral accretion, interpreted as deposits of alternate bank-attached macroforms. Other depositional elements recognized within channel-fill deposits are minor-channel, sandy-bedform, gravity-flow (attributed to bank slumping), and channel-bottom elements. The latter element is contained within a facies sequence that suggests rising- to flood- to waning-stage deposition. In general, deposition probably occurred during relatively high stage; little evidence of low-stage flow was recognized. Subordinate fine-grained facies are interpreted as levee and overbank deposits.

The majority of the Lee Formation was deposited in deep, low-sinuousity, bedload-dominated fluvial channels. Strongly unimodal paleocurrents, lack of facies that suggest low-stage reworking and paucity of lateral-accretion features indicate deposition in a single-

channel system (i.e., a low braiding index). Fluvial architecture similar to that found in sandstone members of the Lee Formation has been previously explained by deposition in multi-channel, braided-river systems. However, the internal architecture of sandstone members is also consistent with a single-channel origin. Individual channels were temporarily confined, during which time the passage of several macroforms aggraded the channel. Position of the channel in the alluvial plain was largely controlled by avulsion of the river from fully aggraded channel belts to other areas of the plain. Calculations reveal that avulsion of a single-channel system across a wide alluvial plain is a plausible mechanism for building the sheet-like sandstone bodies of the Lee Formation. Allocyclic controls (i.e., tectonism and eustatic sea-level changes) on avulsion processes, although likely to have occurred, cannot be ascertained. The spatial arrangement of individual sandstone members of the Lee Formation was probably controlled by tectonic processes. Episodic thrust-loading in the orogenic belt to the east and subsequent flexure of the crust in the foreland basin appears to have caused a step-wise progression of the river system towards the west.

INTRODUCTION

Various depositional environments have been proposed for the Lee Formation in the central Appalachian Basin (Table 1). Controversy endures because: 1) Lee Formation sandstones are located stratigraphically between marine deposits (Newman and Pennington Formations) and coal-bearing fluvial-deltaic deposits (Breathitt Formation); 2) stratigraphic relationships and correlations are often tenuous; 3) the sandstones, relative to those of the Pennington and Breathitt Formations, are chemically and texturally mature; and 4) the sandstones are non-fossiliferous. Embroiled in the controversy is the nature of Mississippian-Pennsylvanian systemic boundary. In general, proponents of a nearshore

Table 1. Depositional environments proposed for the quartz arenite units of the Lee Formation in Kentucky, Virginia and northern Tennessee.

Authors	Depositional Environments	Methods	Location[1]	Members[2]
Wanless (1946)	Subareal	Outcrop	C,P,E	L,M,B,R,C[3]
Mitchum (1954)	Nearshore marine	Outcrop, subsurface	C,P,E	L,M,B,R,C
Potter and Seiver (1956)	Fluvial	Paleocurrents	E	R,C[3]
Bergenback and Wilson (1961)	Non-marine	Subsurface[4]	E	R,C
Bowen (1963)	Nearshore marine	Subsurface	SW	L,M,B
Englund (1968)	Beach to swamp	Subsurface, outcrop	P,C	L,M,B,R,C
Ferm, et al. (1971)	Barrier beach	Outcrop	E	C(?)
Miller (1974)	Tidal delta, barrier bar	Subsurface	SW	L,M,B
Horne, et al. (1974)	Barrier beach	Outcrop	E	C(?)
BeMent (1976)	Shoreface, tidal, fluvial	Outcrop	P,C	L,M,B
Ellsworth (1976)	Deltaic	Outcrop(?)	E	C
Short (1978)	Fluvial	Outcrop, subsurface	E	C
Hester and Taylor (1977, 1981)	Braided-meandering fluvial	Outcrop	E	C
Hayes and Conner (1982)	Barrier island	Quadrangle mapping	E	C
Rice (1984)	Fluvial	Mapping synthesis[4]	E	C,R
Rice (1985)	Fluvial	Subsurface	SW	L,M,B
Englund et al. (1986)	Barrier bar	Subsurface, outcrop	SW	L,M,B
Rice and Schwietering (1988)	Fluvial	Subsurface[4]	E,P	R,C?
Chesnut (1988)	Fluvial	Subsurface	E,P,C	L,M,B,R,C[5]
Cecil and Englund (1989)	Tidal delta/straits	Outcrop	SW	?
Wnuk and Maberry (1990)	Longshore bar	Outcrop	P	M
Chesnut and Cobb (in press)	Fluvial	Outcrop	P	M

1 Study locations (Figure 2): **E** - Eastern Kentucky Cumberland Plateau **P** - Pine Mountain **C** - Cumberland Mountain **SW** - Southwest Virginia.

2 Members studied (Figure 3): **L** - Lower Lee Formation of SW Virginia (below Middlesboro Member) **M** - Middlesboro **B** - Bee Rock **R** - Rockcastle **C** - Corbin

3 Current stratigraphic nomenclature not used and/or undifferentiated.

4 Study of Lee and underlying units, documenting paleovalleys.

5 Redefined stratigraphy (See text and Figure 3).

marine environment favor a conformable relationship, whereas supporters of a fluvial interpretation argue for a major unconformity (Fig. 1).

The Lee-Newman barrier-shoreline model (Fig. 1a; Horne *et al.*, 1978; Ferm and Horne, 1979) has created an often active debate. This model contends that the Lee Formation represents reworked, fluvial lithic sands deposited in barrier-island environments adjacent to an extensive, orogen-sourced deltaic system. In the barrier-shoreline model, formations are interpreted as components of a regional time-transgressive facies continuum, such that Mississippian marine rocks were deposited alongside Pennsylvanian fluvial units. The Mississippian-Pennsylvanian unconformity is interpreted as a discontinuous series of local erosional surfaces.

Alternatively, several workers (Table 1 and review in Ettensohn, 1980) have suggested that the Lee Formation was deposited by a fluvial system on an extensively eroded continental land mass (Fig. 1b). Recent detailed stratigraphic work in eastern Kentucky (Rice, 1984; Chesnut, 1988; Rice and Schwietering, 1988) has supported an unconformable relationship. However, because previous studies of the Lee Formation concentrated primarily on regional stratigraphy and not detailed sedimentological modeling, Shepherd *et al.* (1986) argue that a fluvial interpretation remains unfounded.

This study provides the sedimentological data necessary to better understand the depositional system of the Lee Formation. Outcrops of the Lee Formation provide a unique opportunity to examine many levels of fluvial deposition. Few previous studies (*cf.* Miall and Turner-Peterson, 1989) have combined detailed reconstruction of macroforms with analysis of large-scale fluvial architecture. Lateral profiling, including bounding surface and architectural-element analysis, was utilized taking advantage of extensive outcrops of the Lee Formation in the study area (Fig. 2). This technique is essential for detailed analysis of channelized systems where significant facies changes occur over short lateral

Figure 1. Depositional environments proposed for the Lee Formation: a) barrier-shoreline model; and b) fluvial model. Note absence of unconformity in (a).

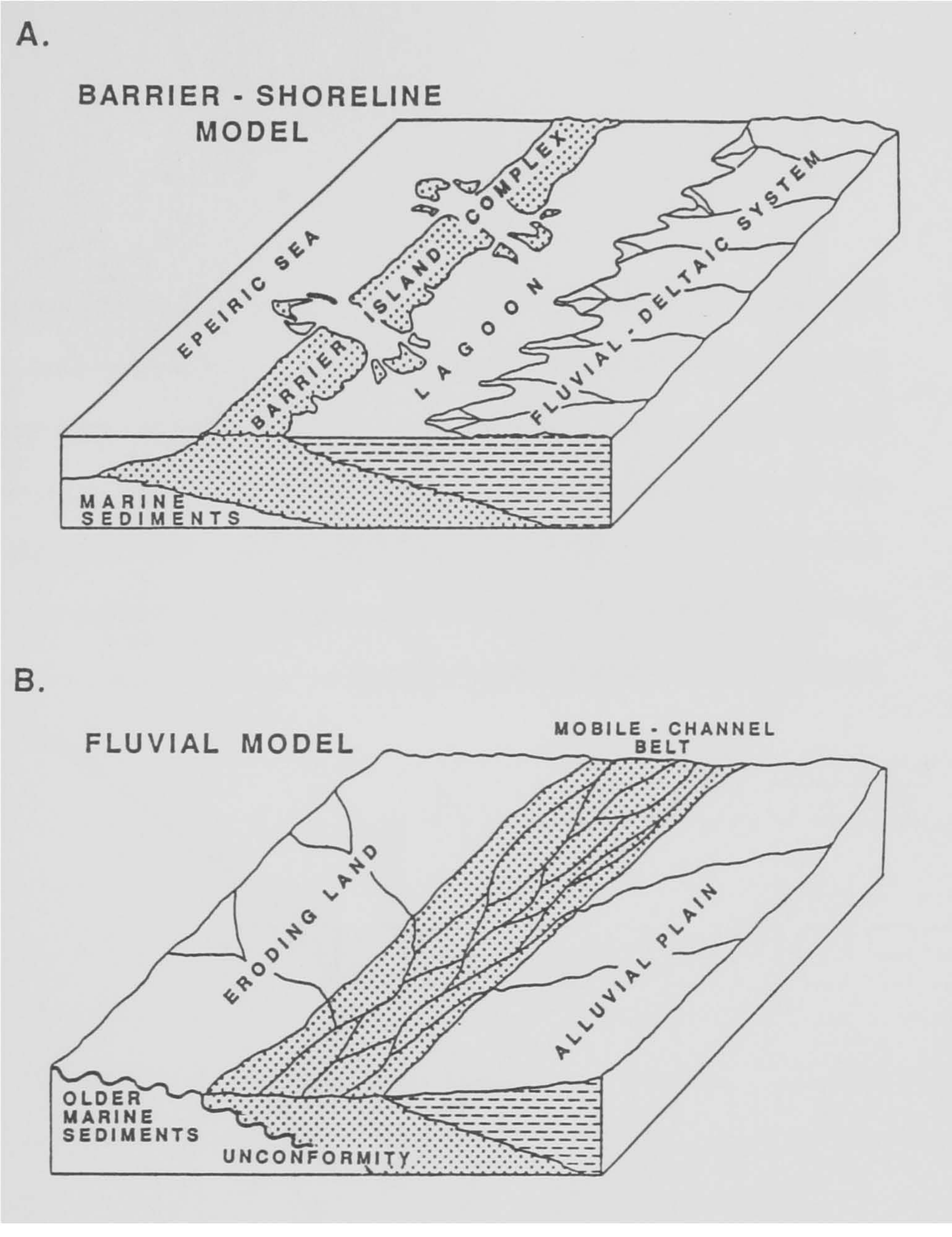
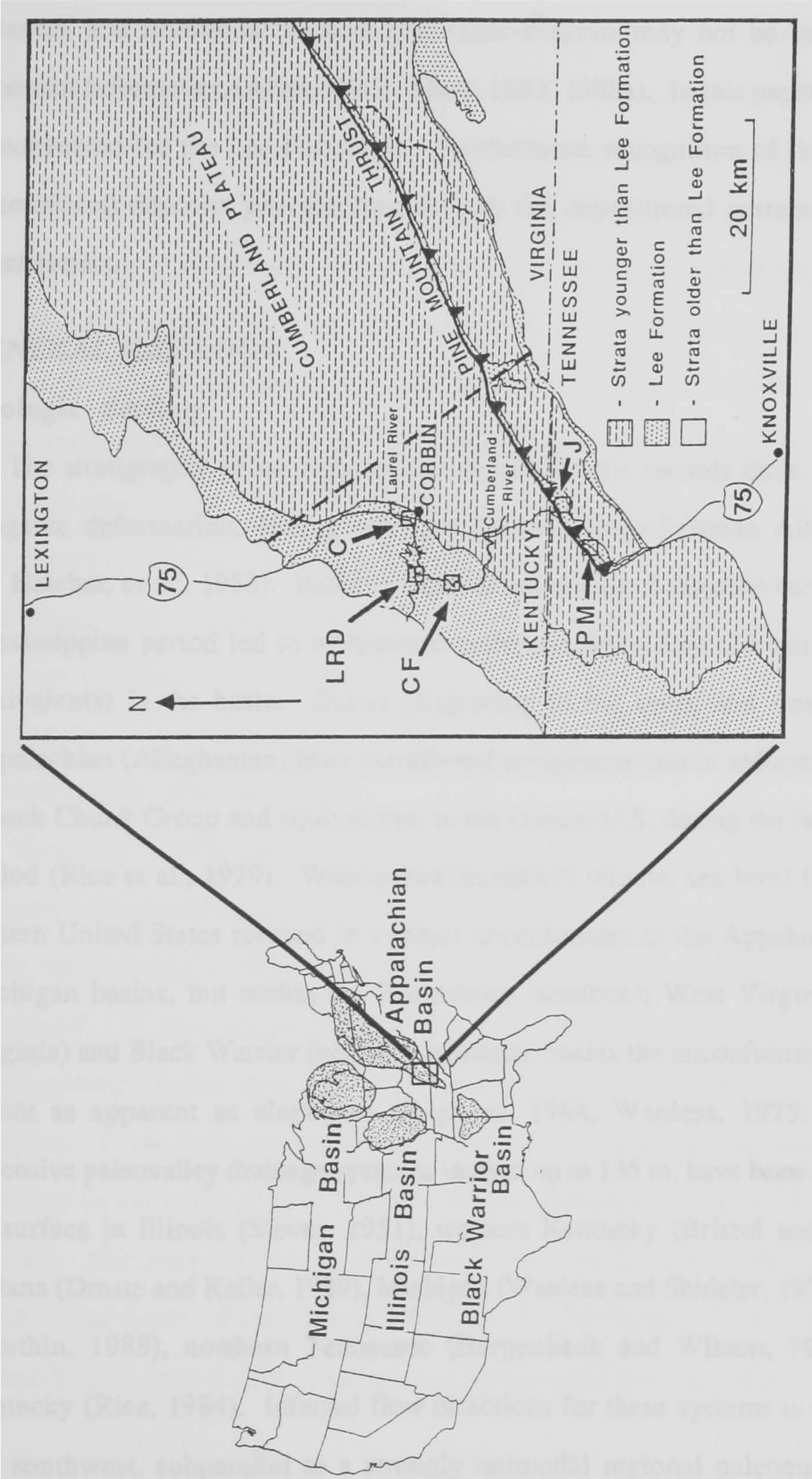

A.
BARRIER - SHORELINE
MODEL
EPEIRIC SEA
BARRIER ISLAND COMPLEX
LAGOON
FLUVIAL - DELTAIC SYSTEM
MARINE
SEDIMENTS
B.
FLUVIAL MODEL
MOBILE - CHANNEL
BELT
ERODING LAND
ALLUVIAL PLAIN
OLDER
MARINE
SEDIMENTS
UNCONFORMITY

Figure 2. Maps showing eastern United States sedimentary basins (left) and general geologic map of study area (right). From NW to SE the three main outcrop belts for the Lee Formation are the Cumberland Escarpment, Pine Mountain and Cumberland Mountain. Study sites as follows: C - Corbin area; LRD - Laurel River Dam; CF - Cumberland Falls; PM - Pine Mountain (I-75); J - Route 25W, south of Jellico, Tennessee. Dashed line shows location of schematic cross-section (Fig. 3).



LEXINGTON
75
CUMBERLAND PLATEAU
PINE MOUNTAIN THRUST
Laurel River
CORBIN
Cumberland River
C
LRD
CF
N
KENTUCKY
VIRGINIA
TENNESSEE
PM
J
75
KNOXVILLE
- Strata younger than Lee Formation
- Lee Formation
- Strata older than Lee Formation
20 km
Michigan Basin
Illinois Basin
Black Warrior Basin
Appalachian Basin

distances and traditional vertical sequences analysis may not be sufficient to record important information (Allen, 1983; Miall, 1985, 1988a). In this paper a fluvial origin is demonstrated for the Lee Formation. Furthermore, recognition of large-scale bedform (macroform) deposits provides insight into the depositional system, aiding in model development.

GENERAL GEOLOGY

Geologic Setting

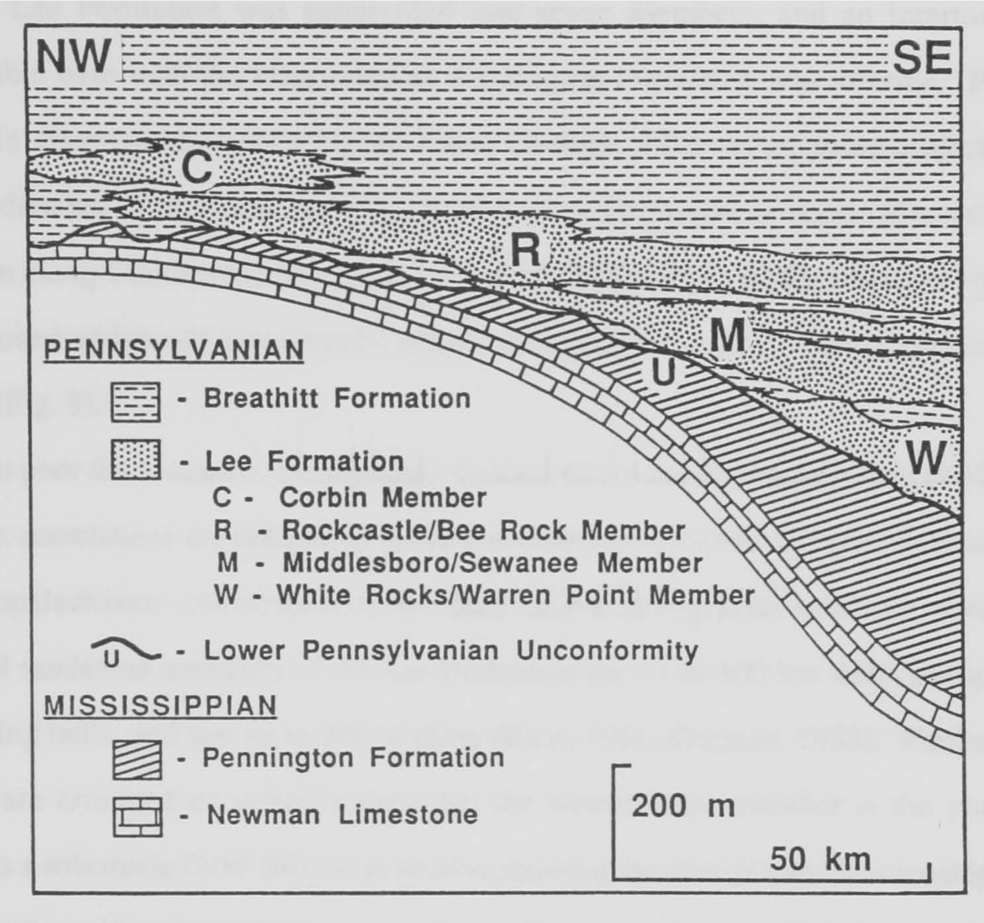
The stratigraphy of the Appalachian foreland basin records three major episodes of orogenic deformation, ending with the Pennsylvanian-Permian Alleghanian episode (*cf.* Hatcher, *et al.*, 1983). Relative tectonic quiescence during the early part of the Late Mississippian period led to widespread carbonate deposition (Newman Limestone and equivalents) in the basin. Deltas prograding to the south and west from the rising Appalachian (Alleghanian) front contributed terrigenous clastic sediments (Pennington or Mauch Chunk Group and equivalents) to the eastern U.S. during the latest Mississippian period (Rice et al., 1979). Widespread (eustatic?) relative sea level fall throughout the eastern United States resulted in a major unconformity in the Appalachian, Illinois and Michigan basins, but within the Pocahontas (southeast West Virginia and southwest Virginia) and Black Warrior (northern Alabama) basins the unconformity is not present or is not as apparent as elsewhere (Englund, 1964; Wanless, 1975; Chesnut, 1988). Extensive paleovalley drainage systems, incised up to 135 m, have been documented in the subsurface in Illinois (Siever, 1951), western Kentucky (Bristol and Howard, 1974), Indiana (Droste and Keller, 1989), Michigan (Wanless and Shideler, 1975), West Virginia (Beuthin, 1988), northern Tennessee (Bergenback and Wilson, 1961), and eastern Kentucky (Rice, 1984). Inferred flow directions for these systems is towards the south and southwest, subparallel to a strongly unimodal regional paleocurrent direction as observed in surface exposures of strata within and overlying the valleys. The nature of the unconformity in the central Appalachian basin is such that progressively older rocks are truncated to the north and northwest, supporting a south-dipping paleoslope (Rice and Schwietering, 1988; Chesnut, 1988; Fig. 3). Early Pennsylvanian (Morrowan Series) strata, primarily Lee Formation and lithologic equivalents, overlie the unconformity.

The Lee Formation thickens to the southeast, and its maximum thickness of more than 500 m (includes intervening Breathitt Shales; Rice, 1984) coincides with the erosional eastern outcrop boundary. As a result of Alleghanian deformation and subsequent erosion there are three main outcrop belts of the Lee Formation (Fig. 2). Cumberland Mountain, the eastern-most belt, and Pine Mountain, the central belt, are part of the Cumberland overthrust sheet. Tectonic transport is approximately 18 kilometers (northwest) at the southern terminus of the sheet (Englund, 1968). The western-most outcrop belt is the Cumberland Plateau escarpment, where the Lee Formation is tectonically undisturbed.

Stratigraphic Relationships

Campbell (1893) assigned the name Lee Formation to Carboniferous quartzose conglomerates forming the crest of Cumberland Mountain in Lee County, Virginia. Later work by Campbell (1898a, b) correlated the Lee Formation with similar rocks in eastern Kentucky. In fact, all along the western margin of the Appalachian Basin, several lithologically similar units have been correlated with the Lee Formation in southwest Virginia. These units are easily differentiated from surrounding units by their relative maturity (commonly > 90% quartz) and the presence of white, well-rounded quartz pebbles. Although not directly correlatable with the Lee-type rocks of the Appalachian basin, other Lower Pennsylvanian units in the Michigan, Illinois, and Black Warrior Basins (Fig. 2) have similar lithologic and stratigraphic characteristics.

Figure 3. Schematic stratigraphic cross-section from near Corbin, Ky. to southeast of Pine Mountain (Fig. 2). After Rice (1984) and Chesnut (1988).


NW
SE
C
R
M
U
W
PENNSYLVANIAN
- Breathitt Formation
- Lee Formation
C - Corbin Member
R - Rockcastle/Bee Rock Member
M - Middlesboro/Sewanee Member
W - White Rocks/Warren Point Member
U - Lower Pennsylvanian Unconformity
MISSISSIPPIAN
- Pennington Formation
- Newman Limestone
200 m
50 km

A type section of the Lee Formation was not proposed until Englund (1964) described a section within the Cumberland Gap National Park (Cumberland Mountain outcrop belt). Here the Lee Formation was subdivided into seven members, and an intertonguing, conformable base with the Mississippian Pennington Formation was inferred. Englund (1968) and Rice (summary of a United States Geological Survey mapping project; 1984) provided detailed stratigraphic correlations of the Lee Formation in eastern Kentucky with the section along Cumberland Mountain. Chesnut (1988) further refined these correlations, and proposed additional correlations with Lee-equivalent units in Tennessee and West Virginia (Fig. 3).

Due to poor fossil control, stratigraphic discontinuities and lateral and vertical lithologic variations, correlations are difficult to make and often are untenable in Pennsylvanian strata of the Appalachians. Nevertheless, the most recent stratigraphic analyses reveal that individual sandstone members of the Lee Formation are 17 to 100 km wide, occupy SW-NE trending belts, and are up to 200 m thick (Rice, 1984; Chesnut, 1988). Furthermore, the belts are arranged *en echelon* such that the westernmost member is the youngest. Figure 3 is a schematic (NW-SE) cross-section showing the stratigraphic relationship of the Lee Formation with surrounding strata, utilizing the nomenclature of both Rice (1984) and Chesnut (1988). Overlying the Lee Formation is the fluvial-deltaic Breathitt Formation (and equivalents) which contain considerably more fine-grained sediment than the Lee Formation and whose channel sandstones contain abundant lithic fragments (Ferm and Horne, 1979). Current stratigraphic terminology (Rice, 1986; Chesnut, 1988), includes the fine-grained units between sandstone members of the Lee Formation as part of the Breathitt Formation (referred to as the lower tongue of the Breathitt Formation). Whether these rocks represent fine-grained facies of the Lee system, or periods of increased progradation of the Breathitt system is not presently understood.

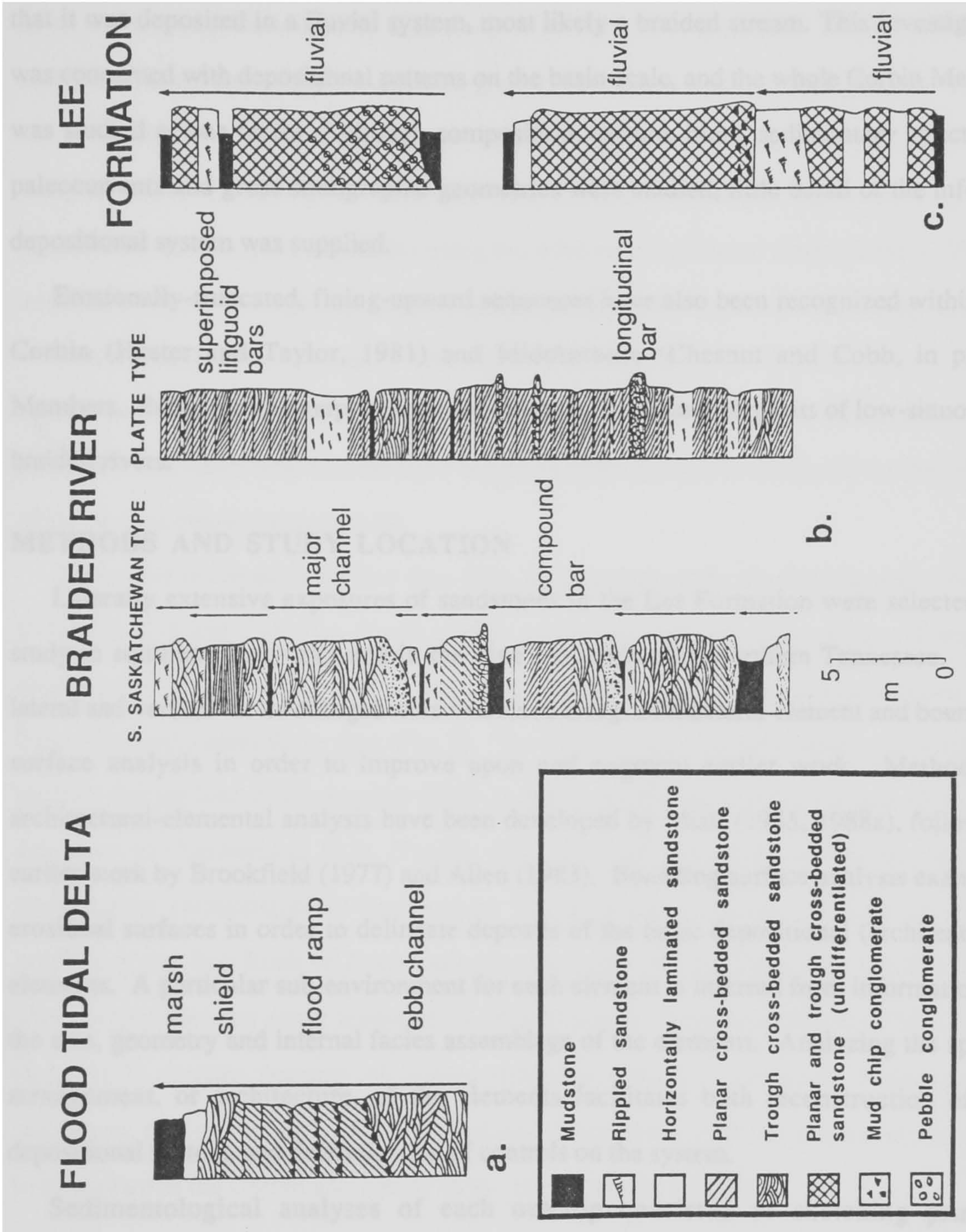
Previous Sedimentological Work

Many workers have studied the Lee Formation and equivalents in the central Appalachian Basin (Table 1), but most studies have focused on regional-scale stratigraphic relationships. Few detailed sedimentological studies of the Lee Formation have been published. Of the studies compiled by Ferm and Horne (1979), only that by Hobday and Horne (1977) discusses details of the orthoquartzite barrier sandstones. Several authors have inferred a fluvial environment for the Lee Formation, but none has presented a refined model of the implicit fluvial environment. The studies of BeMent (1976) and Short (1978) are the most detailed fluvial analyses.

Expanding on earlier work in Alabama (Hobday, 1969), Hobday and Horne (1977) interpreted orthoquartzite sandstones (details of stratigraphic position not discussed) in West Virginia to be largely deposited in tidal deltas and inlets. Fining-upward, channel-fill deposits that display an upward decrease in set size and increasing paleocurrent dispersion, were interpreted as tidal channel deposits. The rocks differed from the Alabama units in that typical beach foreshore structures were not recognized in West Virginia.

BeMent (1976) inferred the Middlesboro and Bee Rock Members of the Lee Formation to be the product of braided streams, and that sandstones below the Middlesboro Member were deposited in tidally influenced nearshore marine environments. BeMent (1976) based his interpretation on an analysis of composition, texture, biota, sedimentary structures, paleocurrents and gross stratigraphic geometries. Several paleohydraulic parameters were calculated, primarily from equations that utilized channel depths that were estimated from the thicknesses of repetitive, truncated, fining-upward sequences. Few sedimentologic details of the studied outcrops were supplied (Fig. 4), probably because at the time of the study relatively little was known of the details within many sedimentary systems, including braided streams.

Figure 4. Comparison of vertical sequences with similar characteristics: a) Tidal delta (modified from Boothroyd, 1985); b) Braided rivers (modified from Miall, 1978); c) Measured section through Lee Formation (modified from BeMent, 1976). In each sequence major depositional components are identified and fining-upward sequences indicated (arrows). Note: In BeMent's study individual beds (20-200 cm) are not differentiated.


FLOOD TIDAL DELTA
marsh
shield
flood ramp
ebb channel
a.
BRAIDED RIVER
S. SASKATCHEWAN TYPE
PLATTE TYPE
major channel
compound bar
superimposed linguoid bars
longitudinal bar
5
m
0
b.
LEE FORMATION
fluvial
fluvial
fluvial
c.
Mudstone
Rippled sandstone
Horizontally laminated sandstone
Planar cross-bedded sandstone
Trough cross-bedded sandstone
Planar and trough cross-bedded sandstone (undifferentiated)
Mud chip conglomerate
Pebble conglomerate

In northeastern Kentucky, Short (1978) examined the Corbin Member and concluded that it was deposited in a fluvial system, most likely a braided stream. This investigation was concerned with depositional patterns on the basin-scale, and the whole Corbin Member was studied as one facies. Although composition, texture, biota, sedimentary structures, paleocurrents and gross stratigraphic geometries were studied, little detail of the inferred depositional system was supplied.

Erosionally-truncated, fining-upward sequences have also been recognized within the Corbin (Hester and Taylor, 1981) and Middlesboro (Chesnut and Cobb, in press) Members. Both studies interpreted the sequences to be channel deposits of low-sinuousity braided rivers.

METHODS AND STUDY LOCATION

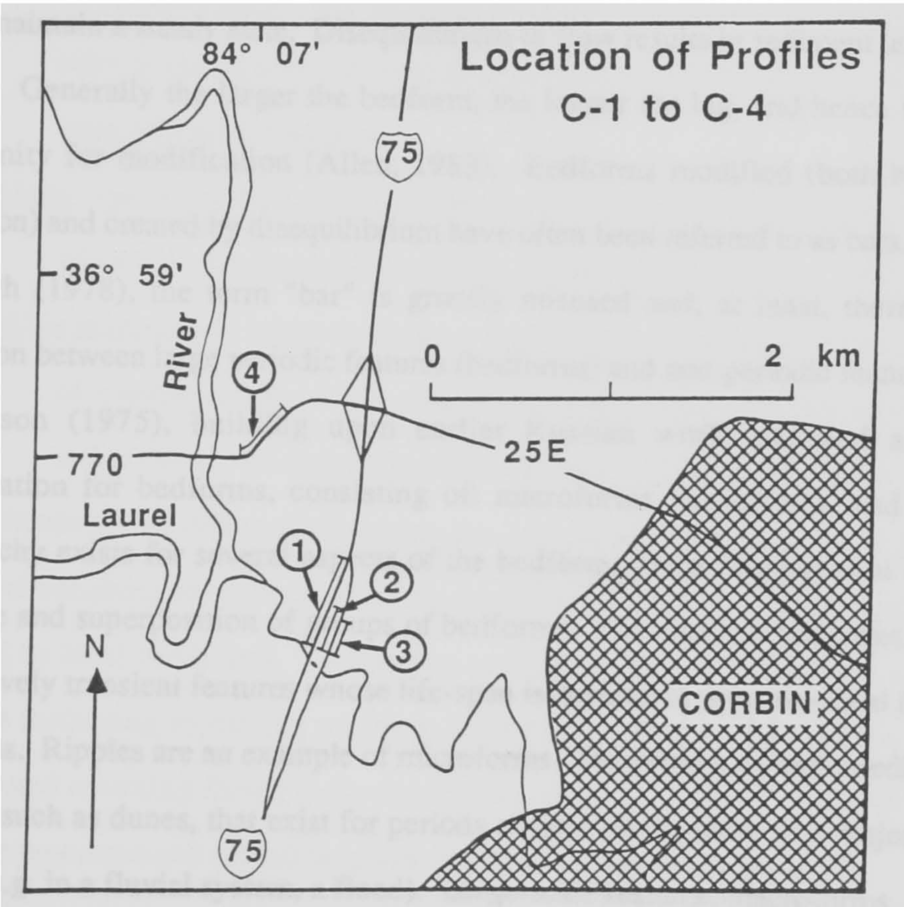
Laterally extensive exposures of sandstones in the Lee Formation were selected for study in southeast Kentucky and in the Pine Mountain area, northern Tennessee. Both lateral and vertical facies changes were examined using architectural-element and bounding surface analysis in order to improve upon and augment earlier work. Methods of architectural-elemental analysis have been developed by Miall (1985, 1988a), following earlier work by Brookfield (1977) and Allen (1983). Bounding surface analysis examines erosional surfaces in order to delineate deposits of the basic depositional (architectural) elements. A particular sub-environment for each element is inferred from information on the size, geometry and internal facies assemblage of the elements. Analyzing the spatial arrangement, or architecture, of the elements facilitates both reconstruction of the depositional system, and determination of controls on the system.

Sedimentological analyses of each outcrop consisted of sketching primary sedimentological features, and measuring grain size and paleocurrent directions. Centimeter-scale features were recorded on transparent overlays of field photographs (8 by

10 inch), each normally covering less than 10 m (lateral) of an outcrop face. The lower 5 m of the outcrop were closely examined with the aid of a ladder. Where necessary, additional information from the remaining upper part of the outcrops was obtained with the aid of binoculars. Photomosaics of the entire length of each exposure were constructed, using the methods outlined in Appendix A, to minimize distortion. Line drawings were prepared for each exposure (profile) using the information gathered from the field. Thus, a largely distortion-free representation of each exposure was obtained, documenting lateral as well as vertical facies changes. Because of lateral extent, each profile could not be reproduced in a continuous manner without loss of detail. Profiles are sectioned and presented on a sheet such that the right end of a section (panel) is continued on the left end of the section below (Appendix B).

Data were collected from five sites (Fig. 2). Outcrops of the Corbin Member (Profiles C-1, C-2, C-3 and C-4; Fig. 5) are readily accessible, laterally extensive, and oriented nearly parallel to paleoflow. The Rockcastle Member was studied at Cumberland Falls State Park and Laurel River dam (Fig. 2). The latter location contains relatively accessible profiles that are both nearly parallel (LRD-1) and perpendicular (LRD-2) to paleoflow. The exposure at Cumberland Falls (Profile CF), although not easily accessible, revealed architectural relationships of higher-order (larger-scale) elements. Additional, shorter profiles of the Bee Rock or Rockcastle Member near Jellico, Tennessee (Profile J) and Middlesboro Member along Interstate 75, Pine Mountain Tennessee (Profiles PM-1, PM-2, and PM-3) contain features of sedimentological interest. Complete fining-upward sequences, recognized only at the gradational upper contacts of the sandstone members, were studied at the Corbin and Laurel River dam locations (Fig. 2).

Figure 5. Detailed map of the Corbin study area (Fig. 2), showing locations of Profiles C-1, C-2, C-3 and C-4.


84° 07'
Location of Profiles
C-1 to C-4
75
36° 59'
River
0
2 km
4
770
25E
Laurel
1
2
3
N
CORBIN
75

BEDFORM TERMINOLOGY

The plethora of bedform nomenclature in sedimentology necessitates definition of the terminology utilized in this paper. Studies of river systems, because of channelization and frequently unsteady flow conditions, have been especially susceptible to and hampered by copious terminology (Ashley, 1990). Ideally, bedforms represent a steady-state equilibrium between fluid and sediment transport, but, by their very nature, fluvial systems do not maintain a steady state. Disequilibrium of flow results in sediment lag or prolonged storage. Generally the larger the bedform, the longer the lag, and hence the greater the opportunity for modification (Allen, 1983). Bedforms modified (both by erosion and deposition) and created by disequilibrium have often been referred to as bars. As discussed by Smith (1978), the term "bar" is grossly misused and, at least, there should be a distinction between large periodic features (bedforms) and non-periodic features (bars).

Jackson (1975), building upon earlier Russian work, outlined a hierarchical classification for bedforms, consisting of: microforms, mesoforms, and macroforms. A hierarchy exists for several aspects of the bedforms, including physical scale, time of existence and superposition of groups of bedforms. The smallest bedforms, microforms, are relatively transient features whose life-span is dependent upon the local fluid dynamic conditions. Ripples are an example of microforms. Mesoforms are intermediate-scale bed features, such as dunes, that exist for periods of time on the order of a major event in the system (e.g. in a fluvial system, a flood). Large-scale features, macroforms, are governed by the geomorphological regime of the system and may endure many major events in the system. Point bars are an example of a macroform. Superposition of bedforms upon another bedform of equal or greater size is common for both equilibrium and non-equilibrium flow conditions. Unfortunately, in his classification, Jackson (1975) included many non-periodic bed configurations.

Using a similar hierarchical nomenclature, Ashley (Chairperson, symposium on classification of large-scale flow-transverse bedforms; 1990) defined four types of sediment storage bodies in fluvial systems:

1) Bedforms - periodic, relatively dynamic and short-lived sediment storage bodies that occur *within the channel* (microforms and mesoforms of Jackson, 1975).
2) Channel form - also periodic, but generally having a slow response to changes in hydraulic regime, and an order of magnitude larger than bedforms. They are *part of the channel* (macroforms of Jackson, 1975) and are often superposed with migrating bedforms.
3) Unit bars - quasi-periodic or solitary bodies that occur *in the channel*, but are products of local hydraulic conditions (also mesoforms of Jackson, 1975).
4) Braid bar complexes - relatively large, quasi-periodic or solitary bodies that can consist of amalgamations of the other three types. They may persist for many years and are likely the result of intricate episodes of deposition and erosion.

The fundamental difference between the two types of periodic forms is that channel forms (macroforms) are controlled by the (bankfull) channel discharge, and bedforms are governed by local conditions. Symposium participants concluded that the wide variety of bedforms (reflecting modifying processes such as channelization, fluctuating water levels, unsteady and reversing flows) in fluvial and nearshore marine environments are similar in formative processes and should be given one name: 'dune' (Ashley, 1990).

The term dune is utilized in this paper for periodic, intermediate-scale, bed configurations. The term will be modified by size and crest shape descriptors. Dunes greater than a meter in height are referred to as large-scale and those less than a meter as small-scale. Dunes with laterally continuous, flow-transverse, crests are considered 2-dimensional (2-D), whereas if morphology varies significantly in the flow direction, they are 3-dimensional (3-D). Set bounding surfaces created by 2-D dunes are flat, whereas those of 3-D dunes are trough shaped. In addition, 2-D dune crests may be either sinuous or straight. A secondary modifier will be used to indicate the presence (compound) or absence (simple) of superposed bedforms. Bedforms of less than 10 cm height are referred to as ripples and channel-scale, periodic bed configurations are termed macroforms.

FACIES DESCRIPTIONS

The Lee Formation sandstones are primarily quartzose, but commonly contain 5 to 10% (sometimes up to 20%) labile components, chiefly mica and low-grade metamorphic rock fragments. Texturally, they are fine to coarse grained (mostly medium grained) and are subangular to rounded with generally good sorting. Quartz pebbles (up to 4 cm, most $\leq$ 1 cm) are common and some conglomerate beds are found, especially near the base of sandstone members. The Rockcastle Member (Fig. 3) is coarser than the other members and is informally divided into two units because of a regional, distinctive vertical change in texture from a lower conglomeratic to an upper pebbly sandstone.

Shale and mudstone facies comprise only a few percent of the Lee Formation. This is because individual members are defined by lithology and any significant intervening fine-grained units are assigned to the Breathitt Formation. However, minor lenses of shales and mudstones are found within the sandstones at the tops of erosionally based, fining-upward sequences, that are typically truncated by subsequent sequences. Most information on fine-grained sediment in the Lee Formation was garnered from the gradational contacts between the Rockcastle and Corbin Members and the overlying Breathitt Formation.

Several distinct facies are recognized within the Lee Formation (Table 2). Cross-bedded sandstones dominate, accounting for more than 90 percent of the Lee Formation. Massive sandstones are interbedded within cross-bedded facies and are estimated to constitute less than 10 percent of sandstones in the Lee Formation. Fine-grained facies are

Table 2. Facies characteristics.

Cross-bedded facies

Facies	Description	Dimensions	Geometry	Interpretation
Conglomerate and conglomeratic sandstone Fig. 6 S_{cg}	Poorly sorted, medium-grained sandstone to conglomerate. May contain rounded mudclasts (up to 20 cm) and quartz pebbles ($\leq$ 2 cm), and plant fragments ($\leq$ 1 m). Poorly defined cross beds. Common facies.	0 to 3 m thick (normally < 1 m), discontinuous over 100s' of m.	Pinches and swells irregularly along outcrop.	Coarse-grained, channel-bottom material. Some deposits of 3-D dunes.
Tangential cross beds Figs. 7 and 8 S_t	Fine- to medium-grained sandstone. Occurs at 3 scales: Small (s), medium (m) and large (l) Foresets concave-up, maximum dip $\leq$ 22°, some overturned. Reactivation surfaces common. Toesets tangential to planar erosional base, and contain poorly preserved S_r (rare). Bounding surfaces to sets and cosets horizontal to gently inclined. Abundant facies	Thickness 10-30 cm (s), 30-60 cm (m), and 60 cm to 1.5 m (l). Width ~10 times, and length > 100 times set thickness.	Wedge- to tabular-shaped sets.	2-D dunes with slightly sinuous crests.
Subfacies				
Long-bottomset cross beds Fig. 7b S_{lb}	Facies S_t with 1 to 5 m long bottomsets sub-parallel to lower erosional surface. Within fine-grained bottomsets are < 1 cm thick laminations of plant- and mica-rich material that may extend 10 cm up toesets. Reactivation surfaces abundant. Common facies.	Similar to large-scale S_t facies.	Tabular -to wedge-shaped sets.	2-D dunes with sinuous-crests. High-flow deposits.
Sigmoidal cross beds Fig. 9 S_s	Topset preserved S_t facies. Convex-up top and flat base. Minor internal erosional surfaces. 10 cm thick massive bed overlies sets. Rare facies.	Sets 1-2 m thick, 10 m long and probably $\geq$10 m wide.	Lenticular shaped sets.	Lower- to upper-flow regime transitional bedform.
Asymmetric troughs S_a	Meters long sets of facies S_t (m), gradational into 1-2 m wide erosional trough, oriented orthogonal to oblique to S_t. Rare facies.	20 to 60 cm thick, width probably 1-2 m, length > 10 m.	Wedge- to lenticular-shaped beds.	Medium-scale, 3-D dunes, shallow-water feature.
Preserved bedform Fig. 10 S_b	Undulating top of sandstone bed and internal cross bedding reveals train of 5 preserved climbing bedforms. Flat to undulatory base on facies S_r and M. Sharply overlain by thin silty-mudstone bed. Little modification of top.	Vertically, crest to trough is 1 m. Wavelength is ~ 18 m, width of bedform unknown.	Asymmetric bedform.	Sinuous-crested, 2-D dune. Formed during gradually increasing flow and preserved as a result of rapid decrease in flow.
Planar-tabular cross beds Fig. 11 S_p	Fine- to coarse-grained, often pebbly, sandstone. Planar to slightly tangential foresets dip ~25°. No reactivation surfaces. Overlies thin, weakly sub-horizontally laminated bed. Abundant facies.	0.5 to 2.5 m thick. Sets are >100 m long and at least 10s' of m wide.	Tabular- to wedge-shaped sets.	2-D dunes with straight-crests.
Compound cross beds Figs. 12a & b S_c	Fine- to coarse-grained sandstone. Compound foresets dip < 15° in general flow direction and contain intrasets of facies S_t or S_p, their foresets dip sub-parallel to flow direction. Rarely, dip directions of intraset and compound foresets are orthogonal to each other. Bounding surfaces to compound sets dip 10° ; sets have flat to undulatory erosional bases and flat to convex-up tops. Common facies.	Thickness 1 to 5 m, most < 2m. Beds are > 100 m long and at least 10s' of m wide.	Wedge- to tabular-shaped sets.	Compound dunes. Smaller-scale, 2-D dunes with striaight to sinuous crests superposed on larger-scale 2-D dunes with straight to sinuous crests.
Subfacies				
Topset-preserved cross strata Fig. 12c & d S_{tp}	Similar to S_c facies but intraset foreset in compound foresets often low-angle. Topsets and foresets dip oblique (up to 60°) to each other. Topset intrasets are facies S_t (s); their foresets dip in direction oblique to dip direction of low-angle intrasets. Compound foresets often found in slight depressions. Rare facies.	1 to 3 m thick and up to 30 m long. Probably 10s' of m wide.	Wedge-shaped sets.	Lateral accretion deposits of dunes or lobate mesoform.
Trough cross beds Fig. 13 S_{tr}	Fine- to coarse-grained sandstone. Highly erosive, concave-up base, low-angle foresets. Relatively rare facies.	Extend ~1-2 m in all directions. Set 5 m wide in facies S_{cg}.	Lenticular shaped sets.	3-D dunes.
Giant cross beds Fig. 14 S_g	Medium- to coarse-grained sandstone. Very large cross beds with rare intrasets. Planar- or tangential-shaped foresets. Rare facies.	4 to 6 m thick sets. Width and length > 10s' of m.	Tabular- to wedge-shaped sets.	Macroform and large-scale dune deposits.

Table 2 continued. Facies characteristics.

Facies	Description	Dimensions	Geometry	Interpretation
Massive facies				
Massive sheet-like sandstone Fig. 15 S_{ms}	Same composition and texture as surrounding cross-bedded facies, but quartz granules (no mudclasts or plant debris) at base. Undulatory, erosive bases and flat, sub-horizontal tops. Basal scours ~ 90° to regional flow. Very poorly defined, flat to slightly concave-up internal laminations. Basal laminations sub-parallel to margin. Rare facies.	3 to 10 m thick, extend for more than 100 m along outcrop face, can be traced > 20 m transverse to face.	Sheet to lenticular shaped.	Gravity flow deposit generated by channel-bank slump and/or as a flood-stage (major) channel-bottom deposit.
Massive channel-form sandstone Fig. 16 S_{mc}	Internally similar to facies S_{ms}, but external shape is steep-sided, slightly asymmetric channel-form. Upflow end is less well-defined, lateral-accretion surfaces in one case. Tops widen rapidly giving form 'wings'. Rare facies.	2.5 to 5 m thick, width less than 10 m, other dimension not known.	Channel-form, possibly ribbon-shaped.	Rapidly filled secondary (minor) channel.
Fine-grained facies				
Rippled sandstone S_r	Fine-grained sandstone with cross laminations. Abundant disseminated plant debris. Coarsening-up beds (of S_r) at top of Corbin Member contain unidirectional trough- and (rare) planar-laminated sets. Some climbing sets. Bioturbation rare. Rare facies.	Individual sets $\leq$1 cm, $\leq$ 30 cm long and < 5 cm wide. Cosets are $\leq$ 2 m thick, at least 10 m long and several m wide.	Sets lenticular, beds are tabular- to lenticular-shaped.	2-D and 3-D current ripples.
Mudstone M	Silty dark-grey mudstone with weak horizontal laminations. Mostly thin lenses associated with erosional surfaces. Extensive beds at tops of members. Plant fossils only. Rare facies.	Lenses $\leq$ 30 cm thick and $\leq$ few m long and wide. Beds capping members 10 cm to > 1 m thick, and extend laterally 10s' to 100s'(?) m.	Sheet- to lenticular-shaped beds.	Vertical accretion of fine-grained sediment.

a relatively minor component. Bedding contacts, or bounding surfaces, of various scales occur within the sandstones. Minor bounding surfaces that separate cross-bed sets and cosets are described in this section. Major surfaces, important in the understanding of large-scale depositional features, are discussed in the lateral profiling section.

Cross-bedded Sandstones

Several types of cross beds (Table 2) constitute a continuum from planar to tangential to trough shaped. Both simple and compound types of cross beds are recognized. Simple tangential cross beds comprise the bulk of the sandstones.

Conglomerate and conglomeratic sandstone (S_{cg}). Poorly-sorted, medium- to coarse-grained, conglomeratic sandstones and conglomerates are found immediately above extensive erosional surfaces in the Lee Formation. Coarsest at the base, conglomeratic sandstones contain abundant subangular to rounded quartz pebbles (up to 2 cm in diameter); grey, laminated, subangular to well-rounded mudclasts (up to 20 cm size); and impressions of plant fragments (up to 30 cm in diameter) more than 1 m long. Poorly defined trough and planar cross beds are the only structures contained within the conglomerates and sandstones. This facies is considered part of the cross-bedded facies because of its close spatial association to cross-bedded sandstones. Both conglomerates and conglomeratic sandstones grade upwards into cross-bedded sandstones. Lower erosional surfaces are undulatory, effecting a pinch and swell geometry (0 to 3 m thickness). This facies is not always present or apparent over the entire length of major erosional surfaces (Fig. 6). In the lower Rockcastle Member, common conglomerate beds are composed almost entirely of quartz pebbles. Conglomerates are best developed above local concave-up, scoured depressions (1 to 10 m local relief) along erosional surfaces (Fig. 6). Clasts, particularly plant debris, are concentrated along the down-flow side (as recognized from overlying cross-bed orientations) of the local scours.

Figure 6. Erosional surface (arrows) in Corbin Member (Profile C-1). Note change in relief of surface and character of overlying sediment over a short lateral distance. Conglomeratic sandstone (facies S_C), with abundant plant fossils, is concentrated above the downflow side of local scour (left). Scale is 15 cm tall.

Tangential cross beds (S_t). This is the most common type of cross bed and consists of 10 to 100 cm-thick tangential sets. Three distinct occurrences of this cross-bed type are recognized on the basis of set thickness. Small-scale sets (10 to 30 cm thick) are present within 1 to 1.5 m-thick cosets of downflow-dipping sets (see compound cross-bed facies below). Medium-scale sets (30 to 70 cm) are isolated or are grouped into sub-horizontal cosets. Large-scale sets (70 to 150 cm thick) occur mainly above major erosional surfaces and are horizontal to gently inclined (< 10°) . In all three types, foresets are tangential to erosive, lower bounding surfaces. Sets are wedge to tabular shaped. Large-scale sets are typically continuous parallel to flow for more than 100 m. In rare bedding-plane exposures, large-scale, 2-D bedforms are sinuous-shaped; their crests extend laterally for more than 10 m (Fig. 7a).

Where preserved, the upper foresets of tangential cross beds are nearly planar with a maximum dip of 17° to 22°. Foresets grade into curved toesets that dip a few degrees and sometimes pass into nearly horizontal bottomsets. Minor internal discontinuities, or reactivation surfaces, are common within tangential cross-bed sets; these are best developed and most frequent in sets containing markedly concave-up cross beds.

Soft-sediment deformation is common in this facies, primarily as oversteepened and overturned foresets (Fig. 8). Overturned sets are up to 70 cm thick. Most occurrences are isolated or incorporate a few contiguous sets, but in the Bee Rock Member (Profile J) deformation extends through several meters of stacked sets (Fig. 8). Highly distorted, disharmonically-folded cross beds are also contained in these sets.

Four subfacies of tangential cross beds are recognized in the Lee Formation (Table 2). In some large-scale sets, well-developed bottomsets are continuous for several meters (Fig. 7b; subfacies S_{tb}). Cursory examination could easily lead to misinterpretation of bottomsets as low-angle to sub-horizontal planar laminations. This problem is

Figure 7. a) Exhumed 2-D, sinuous-crested dunes in Rockcastle Member (located at dam spillway, 1 km southwest of Profile LRD-1). Cross-bed set along center of photograph is about 1 m thick. b) Tangential cross beds with long bottomsets (facies S_{lb}) in Corbin Member (Profile C-2). Scale increments = 10 cm.

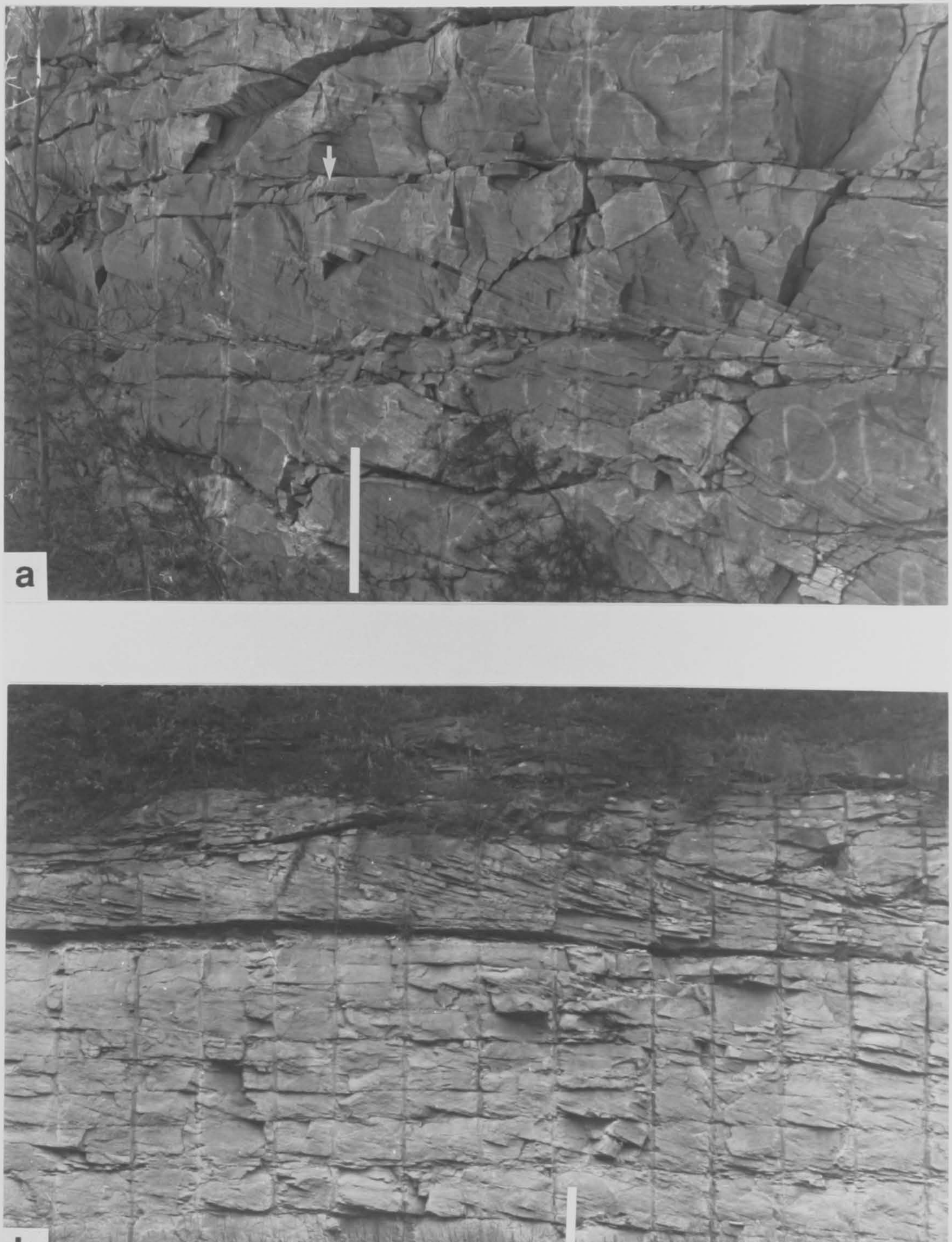

a
b

Figure 8. Overturned foresets of tangential cross beds (facies S_t) in Bee Rock Member (Profile J). Bar scale = 1 m.

compounded by the finer grain size of the bottomsets, which often weather differently from the overlying but contiguous foresets. In addition, subsequent bedforms may erode the foreset component of the cross beds, leaving only the bottomsets. Interbedded with sandstone in the bottomsets are thin laminae of micaceous and organic material that sometimes extend 10 cm up the foresets. Reactivation surfaces that truncate cross strata at a low angle are common in this subfacies. Minor changes in paleocurrent orientation are discernable across these surfaces. Some bottomsets contain what appear to be ripples, but preservation is never good enough to substantiate this possibility.

The sigmoidal, cross-bedded sandstones subfacies (S_S) consists of topset-preserved, simple tangential sets, about 1 m thick (Fig. 9a). Cross beds contain only minor internal erosion surfaces. The set is continuous downflow for about 10 m, before the convex-up upper, and flat lower bounding surfaces converge. The sigmoidal cross beds are transitional into tangential cross beds in the upflow direction. Immediately overlying the sigmoidal set is a 10 cm thick, massive to weakly laminated sandstone bed that conforms to the upper set boundary. The only occurrence of this subfacies is in the Middlesboro Member at Pine Mountain (Profile PM-147E; Fig. 9a). However, an isolated lenticular cross-bed set, similar to the sigmoidal cross-bed set in internal and external geometry, is located immediately above a major erosional surface in the Corbin Member (Profile C-4; Fig. 9b). A flat, lower bounding surface and convex-up upper surface outline a nearly symmetrical morphology, nearly 1.5 m high and 20 m long. This shape resembles the 'lenticular cross-set (preserved bedform)' features described by Røe (1987). Internal cross-stratification is concave-up, or tangential, with extensive bottomsets that can be traced for several meters in this oblique section. A 10 to 20 cm-thick, massive to parallel-laminated, sandstone bed appears to truncate the stoss-side and crest of the bedform and

Figure 9. a) Sigmoidal cross-bedded sandstone (facies S_S) in Middlesboro Member (Profile 147W). Note uniform nature of foresets in flow direction and 10 centimeter thick massive to weakly planar bed (arrow) overlying the cross-bed set. b) Lenticular cross-bed set (preserved bedform) overlying erosional surface in Corbin Member (Profile C-2). Note near-symmetrical shape. Bar scales = 1 m.


a
b

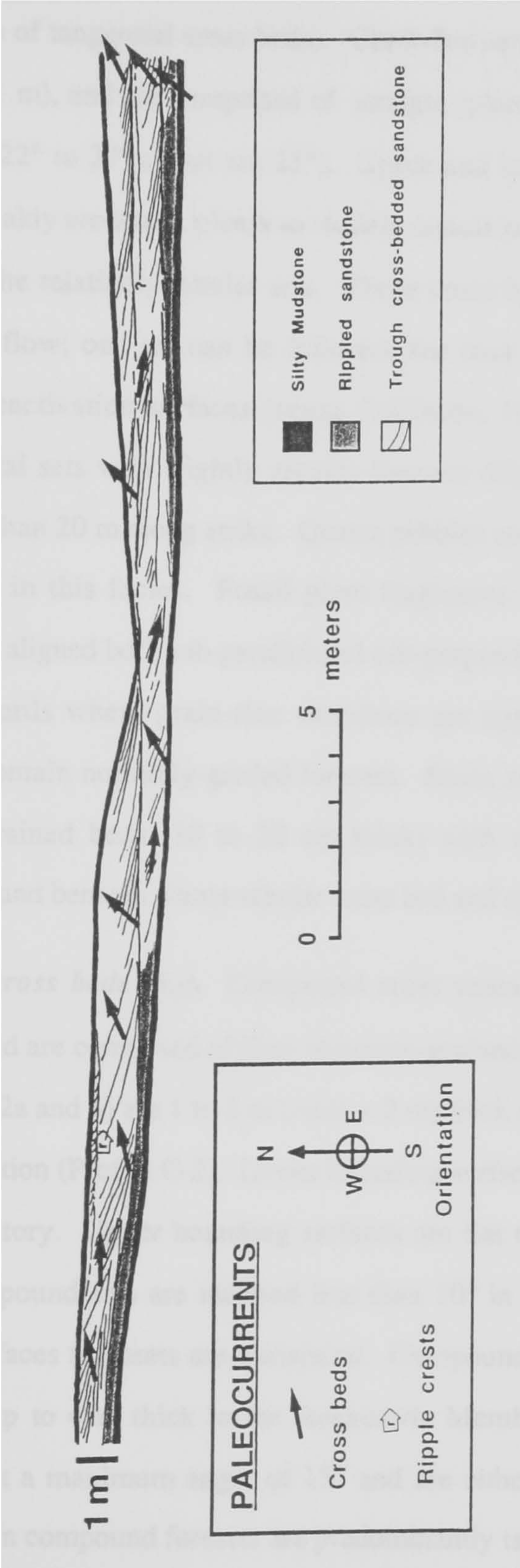
merge with the lee-side foresets. Cross beds on the downflow end of the set are sigmoidal in appearance.

Asymmetric-filled troughs (subfacies S_a) consist of two components: a medium-scale, tabular-tangential cross-bed set that grades in the downcurrent direction of sediment transport into a concave-up, asymmetric-filled, 1 to 2 m wide trough. The trough is significantly more erosive than the tangential cross-bed set (Profile C-4). No significant erosional surface is developed between the two types of foresets. Paleocurrent data indicates that the trough is highly oblique to the tangential foresets.

The preserved bedform subfacies (S_b) is recognized in the upper part of the Corbin Member (Fig. 10). An undulating upper surface of a 1 to 2 m-thick sandstone bed exhibits five distinctly recognizable crests with consistent asymmetry. Internal, tangential cross stratification conforms with the steeper side, indicating fully preserved, large-scale bedforms. Crest heights of about 1 m and an average bedform wavelength of 18 m are accurately estimated for the bedform, because the outcrop is nearly parallel to the paleoflow direction. The lengths and morphologies of the crests are not known as crestal exposure is limited. The bedforms have a flattened shape and the crests appear rounded. On only one crest, where cm-scale 3-D ripples are found, is there evidence of reworking. Foreset dips range from 10° to 20°, most are nearly 20°. Cross-bed set bounding surfaces dip upcurrent (Fig. 10). The sandstone bed is sharply overlain by a 10 cm-thick, silty-mudstone bed. The base of the sandstone bed is flat to undulatory and overlies either rippled sandstone or silty-mudstone. The rippled sandstones are often located upcurrent of the cross beds (Fig. 10).

Planar-tabular cross beds (S_p). Planar-tabular cross beds are relatively common and consist of medium- to coarse-grained sandstones. The coarser-grained Rockcastle Member is composed of a greater percentage of planar-tabular cross beds than the other members (to

Figure 10. Line drawing of preserved bedform (facies S_b) at top of Corbin Member (located 1/2 km west of Profile C-4).

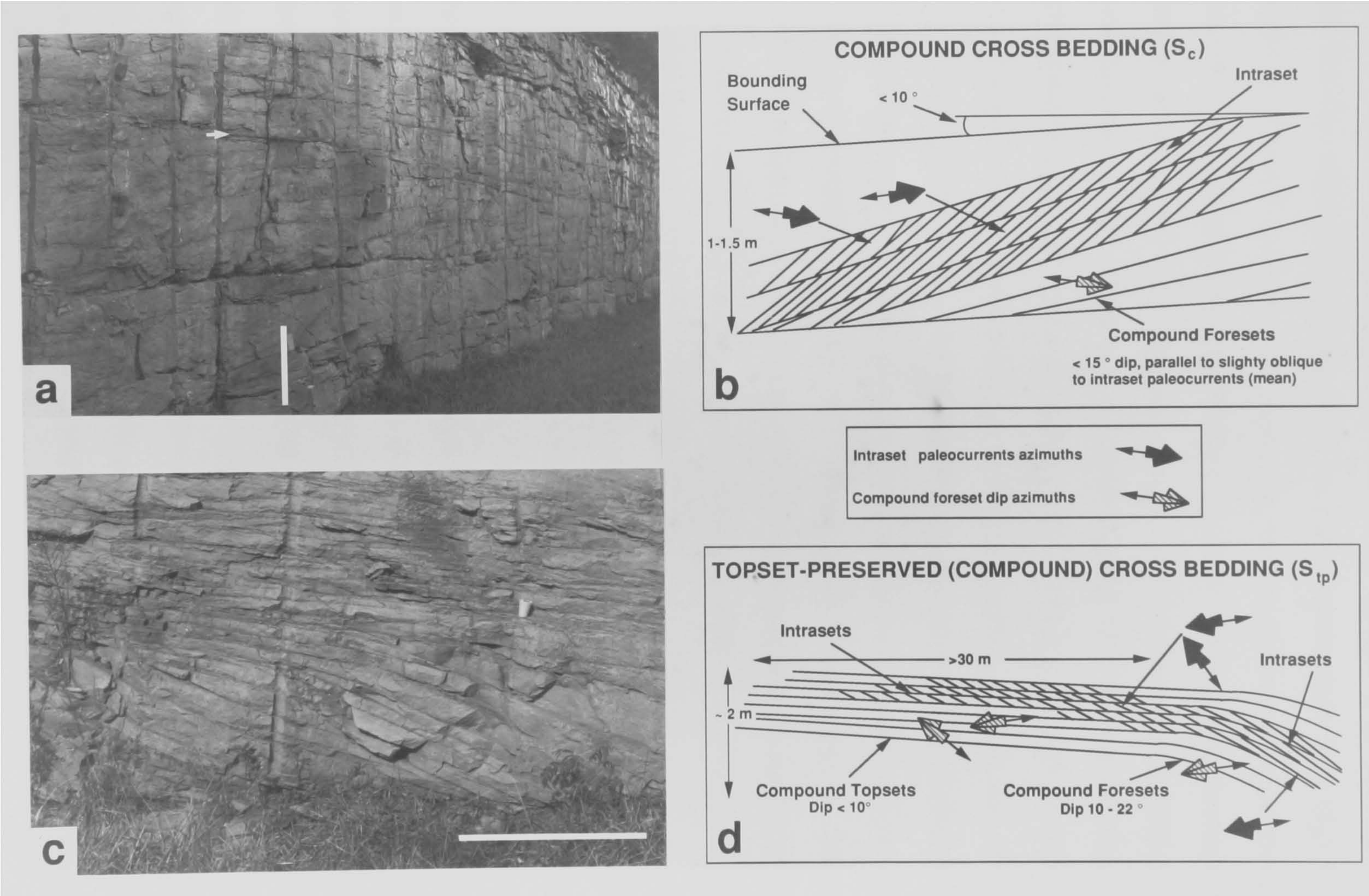

PALEOCURRENTS
Cross beds
Ripple crests
N
W
E
S
Orientation
1 m
0
5
meters
Silty Mudstone
Rippled sandstone
Trough cross-bedded sandstone

the exclusion of tangential cross beds). Cross-bed sets are 30 cm to more than 2 m thick (commonly 1 m), and are comprised of straight (planar with minor toesets) foresets that are inclined 22° to 27° (most are 25°). Upper and lower bounding surfaces to sets and cosets are weakly erosional, planar to slightly undulatory, resulting in local thickening and thinning of the relatively tabular sets. These cross beds are remarkably uniform in the direction of flow; one set can be followed for over 200 m (Profile C-1) without any significant reactivation surfaces (*sensu* Collinson, 1970a). A bedding-plane exposure reveals several sets with slightly arcuate foresets (Fig. 11). Individual foresets can be traced more than 20 m along strike. Quartz pebbles and mud chips (up to 1 cm in length) are common in this facies. Fossil plant fragments (up to 50 cm long and 30 cm in diameter) are aligned both sub-parallel and sub-perpendicular to paleoflow. Cross-bed sets coarsen upwards where grain-size variations are apparent. Planar-tabular cross beds commonly contain normally-graded foresets. Some sets contain a quartz-pebble veneer top. Fine-grained beds (10 to 20 cm thick) with weak sub-horizontal laminae, are sometimes found beneath planar-tabular cross bed and appear to define bottomsets.

Compound cross beds (S_c). Compound cross strata are common features in the Lee sandstones and are composed of fine- to coarse-grained sandstones. Compound cross-bed cosets (Fig. 12a and b) are 1 to 5 m (most < 2 m) thick and extend for more than 100 m in the flow direction (Profile C-2). Lower bounding surfaces of compound sets are erosional, flat to undulatory. Upper bounding surfaces are flat to convex up. Bounding surfaces between compound sets are inclined less than 10° in the downflow direction, whereas bounding surfaces to cosets are horizontal. Compound sets are generally 1-1.5 m thick, but may be up to 4 m thick in the Rockcastle Member (Profile LRD-2). Compound foresets dip at a maximum angle of 15° and are either planar or concave-up in shape. Intrasets within compound foresets are predominantly tangential cross beds (S_t), generally

Figure 11. Planar-tangential cross-bedded sandstone (facies S_p) in Rockcastle Member (Cumberland Falls State Park). Foresets can be traced along bedding-plane exposure. Note arcuate shape to foresets. Scale increments = 10 cm.

Figure 12. a) Compound cross-bedded sandstone (facies S_c) in Corbin Member (Profile C-2). Set truncated by third-order bounding surface (arrow). b) Schematic drawing of compound cross bed set (S_c). c) Topset-preserved, compound cross-bed set (S_{tp}) in Corbin Member (Profile C-4). d) Schematic drawing of facies S_{tp}. Bar scales = 1 m



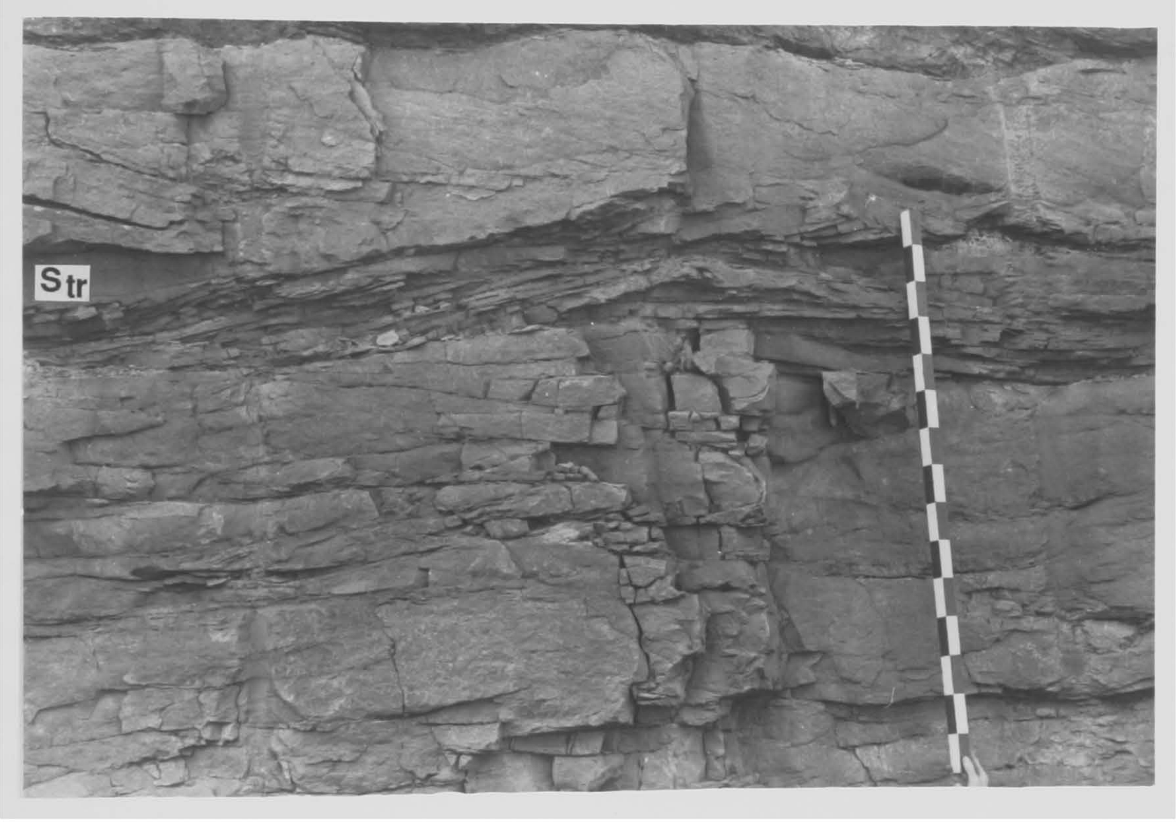
a
COMPOUND CROSS BEDDING (S_c)
Bounding Surface
< 10 °
Intraset
1-1.5 m
Compound Foresets
< 15 ° dip, parallel to slighty oblique to intraset paleocurrents (mean)
b
Intraset paleocurrents azimuths
Compound foreset dip azimuths
c
TOPSET-PRESERVED (COMPOUND) CROSS BEDDING (S_{tp})
Intrasets
>30 m
Intrasets
~ 2 m
Compound Topsets
Dip < 10°
Compound Foresets
Dip 10 - 22 °
d

less than 30 cm thick, and contain numerous reactivation surfaces (Fig. 12b). In the Rockcastle Member, intrasets in compound sets are planar-tabular cross beds (S_p), less than 1 m thick (Profiles CF and LRD-2). In most cases, foresets of intrasets and compound foresets are inclined sub-parallel (Fig. 12b, Profile C-2). Rarely, paleocurrent directions of intraset foresets are orthogonal to the dip direction of compound foresets (Profile LRD-1).

Topset-preserved compound cross beds (subfacies S_{tp}; Fig. 12c) are a variant of facies S_c. Dip directions of compound topsets are up to 60° different from the compound foresets (Fig. 12d). Topsets, which in some cases can be followed upflow for more than 30 m, contain well developed intrasets. Dips of intraset foresets decrease in a downcurrent direction and, as the compound topsets pass into compound foresets, become sub-parallel to the enclosing compound foresets (Fig. 12d). Intraset paleocurrents are either oblique (Profile C-4), or parallel (Profile PM-151) to the compound topsets and are parallel to the compound foresets (Fig. 12d). Often, topset-preserved, compound cross beds are associated with depressions in the bedding surface (Profiles C-4 and PM-151).

Trough cross beds (S_{tr}). Trough cross beds are not a particularly common facies in the Lee Formation. This facies was only recognized in sections oblique to or nearly perpendicular to inferred flow directions. Two types of trough cross beds are distinguished; one is found above, and the other below extensive erosional surfaces. In both types, upper and lower bounding surfaces of the troughs are erosional. Trough cross beds above erosional surfaces are contained within conglomeratic sandstones. These troughs are broad, low-angle features that may be up to 1 m thick by 5 m wide (Profile C-4). They are manifest by pebble and mudclast stringers in otherwise relatively homogeneous, fine- to medium-grained sandstone. Laminations within the troughs are sub-parallel to the stringers. Trough cross strata sets beneath erosional surfaces are 50 to 100 cm thick and less than 2 m wide (Fig. 13; Profile C-1b). These troughs are developed

Figure 13. Trough cross-bedded sandstone (facies S_{tr}) in Corbin Member (Profile C-2). Scale increments = 10 cm.


Str

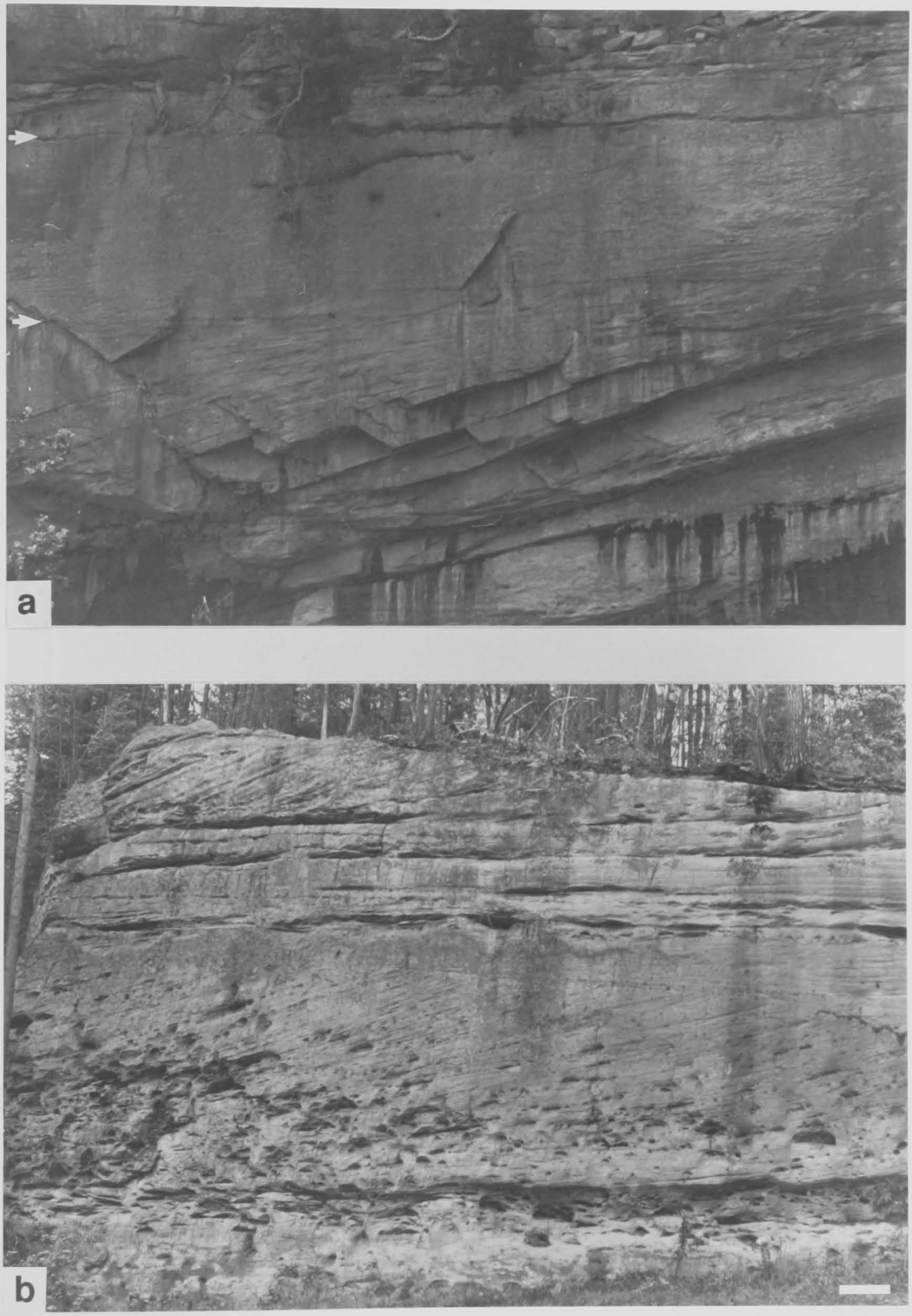
in fine- to medium-grained sandstone and form a laterally continuous belt (coset) up to 2 m thick.

Giant cross beds (S_g). Several examples of giant cross beds were observed in the Lee Formation, but most were inaccessible for detailed analysis. Three examples in the Rockcastle Member are discussed here (Fig. 14). One set, about 6 m thick, can be traced in the direction of flow for more than 60 m and is transitional into low-angle, downflow-dipping sets (S_c; Profile CF). A similar sequence of downflow-dipping sets (S_c) overlie the giant cross beds (Fig. 14a). The giant cross beds dip at greater than 20° and are slightly convex-up. Rare intrasets are recognized between the steeper foresets. Another planar giant, planar set (Fig. 14b; see Chesnut, 1989) is more than 6 m thick and is continuous for more than 25 m parallel and 50 m perpendicular to paleoflow direction. Both of these giant cross bed sets are within the lower conglomeratic sandstone of the Rockcastle Member. Giant cross beds present in the upper Rockcastle Member (Profile LRD-1) are up to 3.5 m thick and are continuous for more than 30 m in the flow direction. The cross beds have a maximum dip of 20° and are tangential to a basal erosional surface. Overlying low-angle beds (S_c) both truncate and are continuous with the giant tangential set, suggesting coeval deposition. Near the base of the giant cross beds are rare intrasets of small-scale tangential cross beds. The giant foresets contain three low-angle internal erosional surfaces.

Massive Sandstones

Sandstone units devoid of sedimentary features are relatively common in the Lee Formation. Many examples of featureless, or massive, sandstones are probably caused by either diagenesis or weathering. There are, however, several explicit occurrences of massive sandstones with distinct geometries interbedded within the cross-bedded facies. Massive sandstone are erosionally based. Two geometries of massive sandstone are recognized: sheet-like bodies and channel forms.

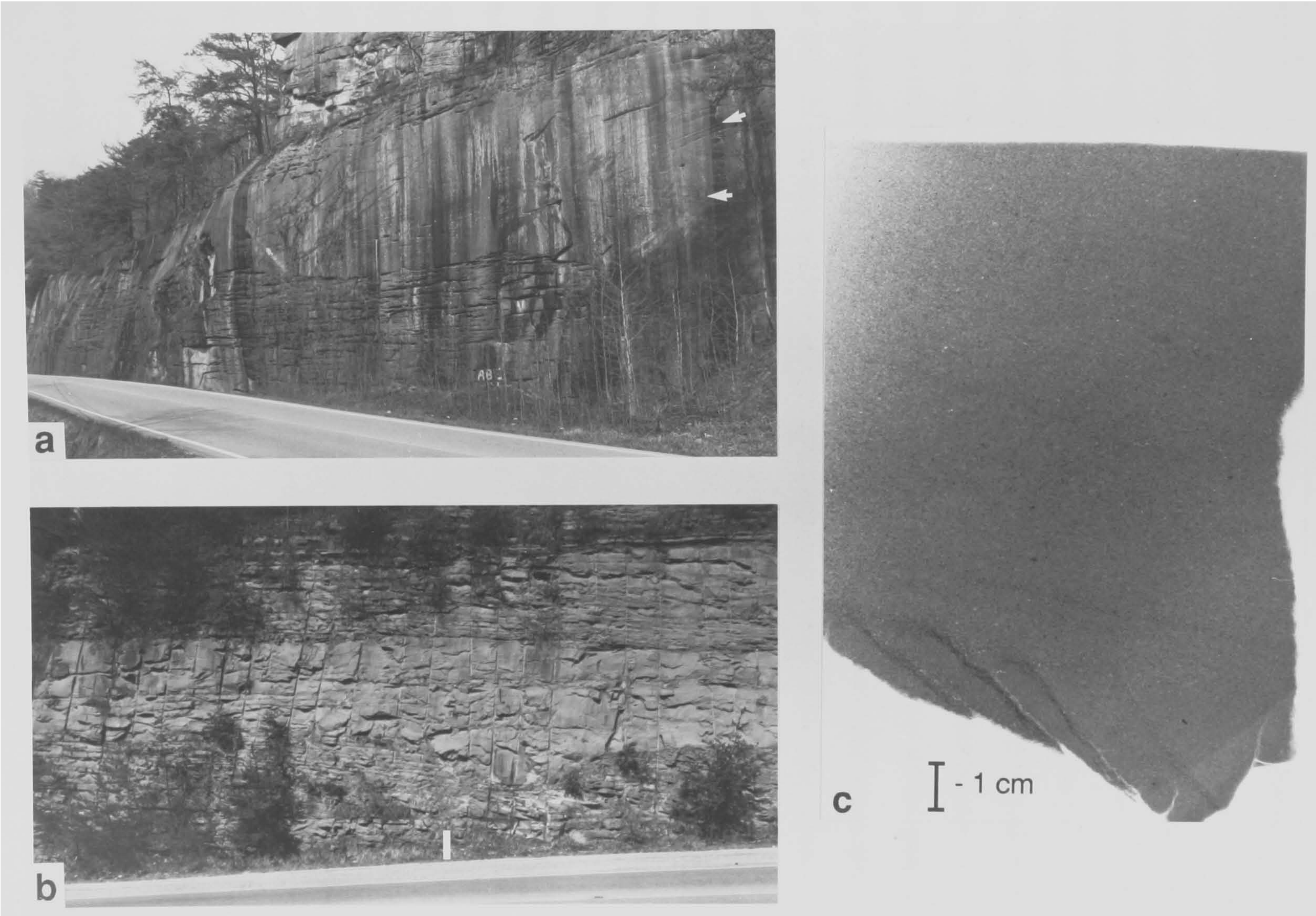
Figure 14. a) Close-up photograph of Rockcastle Member outcrop in Profile CF (box in Profile CF). Giant cross-bed set (facies S_g) about 6 m thick. Paleoflow towards left (southwest); arrows mark erosional surfaces. Sequence between upper two erosional surfaces consists of basal, planar-tabular set, overlain by downflow-dipping low-angle sets. b) Six meter-thick cross-bed set (facies S_g) in the Rockcastle Member (located along Rt. 80, Ky., 25 km north of LRD location, Fig. 2). Bar scale = 1 m.


a
b

Sheet-like bodies (S_{ms}). Relatively featureless, sheet-like sandstone bodies were found in the Corbin (Profile C-4) and Bee Rock (Profile J) Members (Fig. 15a and b). Both bodies have undulating erosive bases and flat, horizontal to convex-up tops, but are of different scales. Sheet-like massive sandstones are similar in composition and texture to surrounding cross-bedded facies, except that they contain basal concentrations of subangular to rounded quartz and mudclast granules. These granules are most frequently concentrated in concave-up basal scours. Nearly symmetrical basal scours indicate a flow direction highly oblique to paleocurrent direction obtained from surrounding cross beds; steep basal-scour margins in the Corbin trend almost perpendicular to paleoflow.

The sheet-like, massive sandstone body in the Corbin Member is up to 5 m thick, more than 100 m wide along the outcrop face and extends more than 20 m perpendicular to the outcrop (across the road). The massive body thins to a few meters thick in the down- and across-flow directions (as determined by paleocurrents of cross beds above and below the massive sandstone). At the northeastern (downflow) end of the sheet-like body, the upper surface twice steps down abruptly, each by approximately a meter. Internally, the massive sandstone is featureless except for numerous extensive, sub-horizontal surfaces, up to 10 m long (Profile C-4). These surfaces are flat, but contain mm-scale undulations. X-ray examination of a sample from the base of the massive facies reveals faint laminations sub-parallel to the margin (Fig. 15c). Similar examination of the massive facies well above the base, including specimens containing the sub-horizontal surfaces, did not reveal any features. In the poorly-exposed outcrop across from Profile C-4, the massive sandbody contains internal undulating surfaces. Inclinations of the crests and swales are approximately perpendicular to paleoflow. Freshly broken surfaces of this facies reveal cm-scale, round, patchy, brown-grey spots. Thin section examination of these features indicates that they consist of an interstitial organic residue concentrated in grains of

Figure 15. a) Sheet-like, massive sandstone unit (facies S_{ms}) in Bee Rock Member (Profile J). Note faint internal trough-shaped laminations (arrows). Letter 'A' is 0.5 m tall. b) Sheet-like massive sandstone bed (facies S_{ms}) in Corbin Member (Profile C-4). Bar scale = 1 m. c) Print from x-ray negative of contact between massive facies and cross-bedded facies in (b), showing faint basal laminations in facies S_{ms}.


a
b
c
- 1 cm

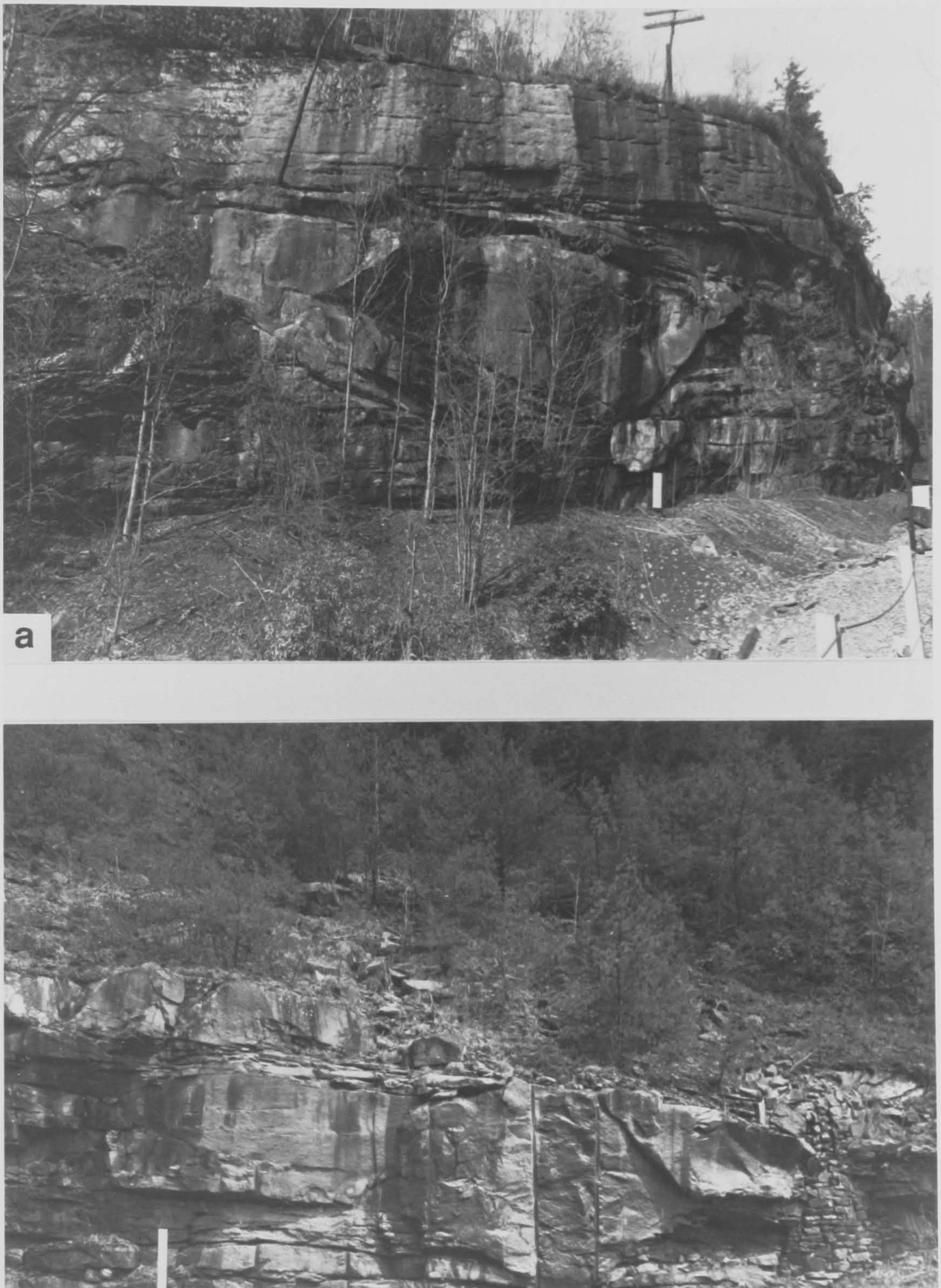
reworked low-grade metamorphics. Geochemical tests indicate that the mottling is not from plant bioturbation, but instead from deposits of light hydrocarbons (Appendix C).

The sheet-like massive sandstone in the Bee Rock Member (Profile J) is 5 to 10 m thick, and laterally extensive for more than 200 m along the exposure. The outcrop face is approximately parallel to paleoflow as indicated by cross bedding located above and below the massive sandbody. Perpendicular to the outcrop face, the tabular bed can be traced for more than 50 m and may be equivalent to a tabular, massive sheet-like sandstone across a stream valley a few hundred meters distant. Sub-horizontal, diffuse laminations are apparent throughout the bed. Laminations at the base of the massive sheet-like sandstone are sub-parallel to the margins of convex-up scours. In many instances, quartz granules and pebbles are concentrated along the laminations. Also, sweeping, concave-up laminations several meters across are evident, apparently truncating one another (Fig. 15b). Weathered pockets 10 to 50 cm wide and present high-up within this facies, may represent casts of mudclasts.

Channel forms (S_{mc}). Steep-sided, slightly asymmetric to nearly symmetric channel-shaped massive sandstones are found in the Bee Rock, Rockcastle and Middlesboro Members (Fig. 16; Profiles J, LRD-1, and PM151E). The channel forms are less than 10 m wide and 2.5 to 5 m deep. The near-symmetric shape of the margins indicate that the channel trends at a high angle to paleoflow (as determined from surrounding cross beds). The length of the channels could not be determined. As with the sheet-like, massive sandstones, sediments are similar in composition and texture to the surrounding cross-bedded facies, including basal concentrations of quartz granules and pebbles.

Diffuse and discontinuous internal laminations are both sub-horizontal and concentric to the margins of the channel forms. In two of the channel forms (Figs. 16a and b), internal surfaces sub-parallel to the channel margin can be traced laterally out of the channel into

Figure 16. Channel-form massive sandstones (facies S_{mc}). a) Nearly 5 m-thick unit with lateral-accretion beds in Bee Rock Member (near Profile J). b) Channel form about 3 m thick in Rockcastle Member (Profile LRD-1). Bar scales = 1 m.


a
b

cross-bed sets. In one case (Fig. 16a), the bedding surfaces are continuous for 1 to 2 m into the channel form, but in the other example the surfaces are continuous to the base of the channel form. In addition, both of these examples have much wider tops giving the channel form "wings".

Fine-grained Facies

Cross-laminated sandstone (S_r). Sandstones containing cm-scale ripple laminations are not common in the Lee Formation except at the upper gradational contacts with shales of the Breathitt Formation. Sandstones are fine grained and are rich in mica grains and disseminated plant fragments. Bioturbation is rare. Generally, cross-laminated sets are less than 1 cm thick, less than 30 cm long and 4 to 6 cm wide. Trough-shaped and rare straight foresets are exposed on bedding planes. Only current-ripple laminations are found. Cosets of cross-laminated sandstone are up to 2 m thick and several meters in length and width. Climbing ripples are sometimes found in the lower parts of the cosets.

Mudstone (M). Discontinuous thin beds of thinly laminated, grey mudstone occur infrequently in the Lee Formation. Generally, these lenticular bodies are on the order of 30 cm thick by 1 to 2 m wide. Mudstone lenses are truncated by major erosional surfaces, but they are also found overlying erosional surfaces (Profile C-1 and LRD-2). One to several meter-thick beds of grey mudstone are found at the top of sandstone members. The lateral extent of these beds is unknown, but is probably regional. Grey mudstones are non-fossiliferous except for abundant disseminated plant fragments. Black shale beds gradationally overlie the thick mudstone beds.

LATERAL PROFILE ANALYSIS

Profile analysis involves examination of major bounding surfaces and the 2- and 3-dimensional arrangement of facies. Large-scale depositional elements are reconstructed from the information obtained in the analysis of the profiles. The Corbin (Figures 2 and 5, Profiles C-1, C-2, C-3 and C-4) and Cumberland Falls (Fig. 2, Profile CF) outcrops are particularly amenable to lateral profiling analysis because they are: 1) laterally extensive; 2) oriented nearly parallel to paleoflow and, in part, nearly perpendicular to flow; 3) tectonically undisturbed; and 4) inferred to be typical of Lee Formation sediments.

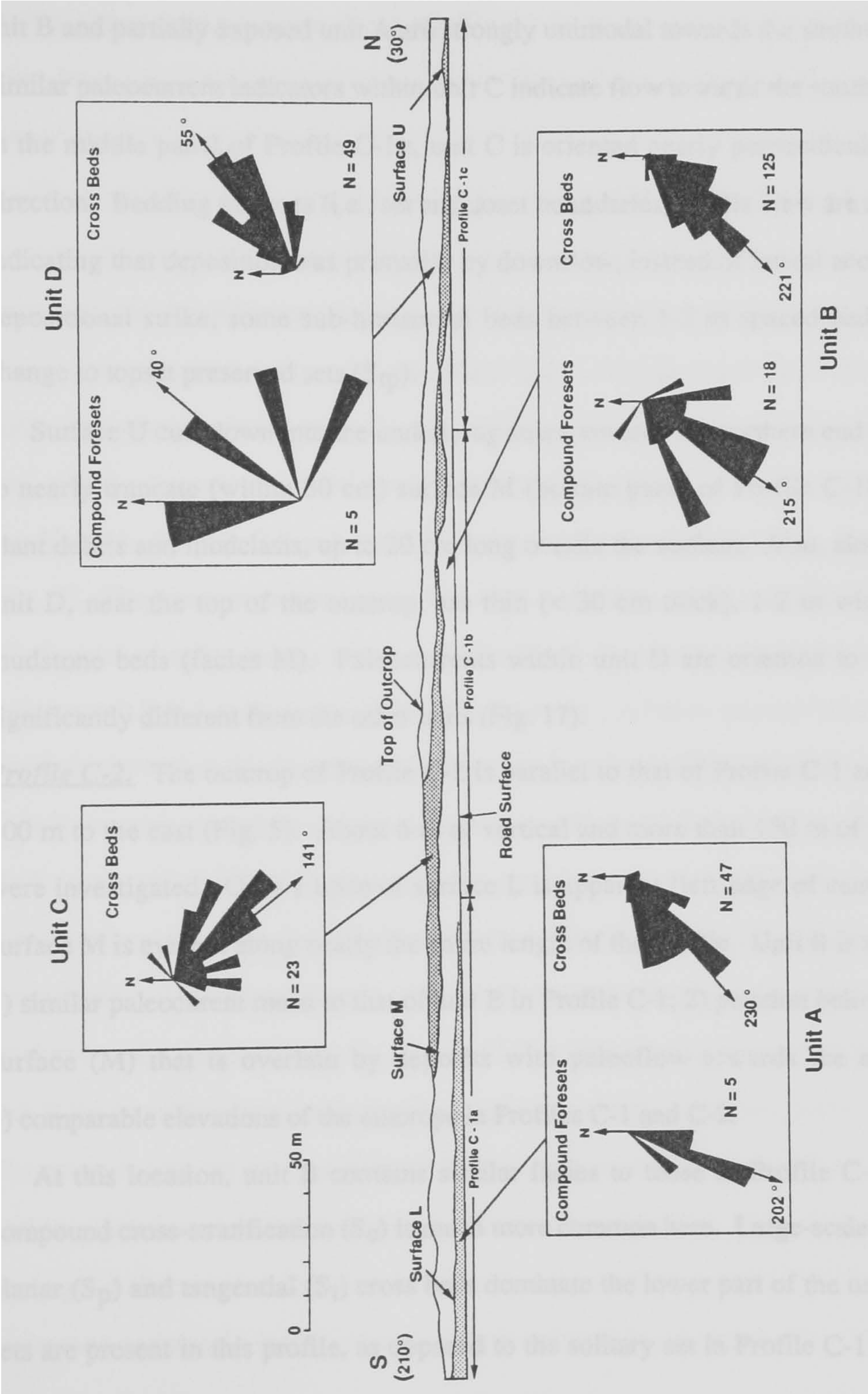
Profile Descriptions

Because the profiles are the source for the facies descriptions, only brief summaries of small- and intermediate-scale sedimentary features are provided in this section with emphasis placed rather on lateral and vertical relationships. The primary focus of this section thus is the recognition and description of large-scale sedimentary features, including their internal and external bounding surfaces.

Corbin Member. Outcrops of the Corbin Member examined in this study are located within the nearly horizontal strata of the western escarpment of the Cumberland Plateau (Fig. 2). Profiles C-1, C-2, C -3 and C-4 are situated in close proximity to each other (Fig. 5). Together, these profiles represent the upper 25 m of the Corbin Member, which is 30 to 60 m thick in this area (Rice, 1984).

Profile C-1. Profile C-1 embodies nearly 0.5 km of outcrop and is separated into three sections: C-1a, C-1b and C-1c (from north to south). Approximately 12 m of vertical section contains three sub-horizontal to low-angle erosional surfaces (from bottom to top: L, M and U), that separate four distinct units (A-D, Fig. 17). All three surfaces have local relief of a few meters. The surfaces are recognized by an abrupt facies transition from coarse-grained, often conglomeratic sandstone (facies S_{cg}) to cross-bedded medium-grained sandstone. No significant fine-grained facies are present within units. Internally, the units consist of tangential (S_t), planar-tabular (S_p) and rarely, trough (S_{tr}) cross bedding. A large planar set, about a meter above the base of unit B, is overlain by compound cross beds (S_c).

Figure 17. True-scale reduction of Profile C-1 with major erosional surfaces (U, M and L) and paleocurrents for intervening units (A-D).


S
(210°)
0
50 m
Surface L
Profile C - 1a
Surface M
Unit C
Cross Beds
141°
N = 23
Road Surface
Top of Outcrop
Profile C -1b
Profile C -1c
Unit D
Compound Foresets
40°
N = 5
Cross Beds
55°
N = 49
Surface U
N
(30°)
Unit A
Compound Foresets
202°
N = 5
Cross Beds
230°
N = 47
Unit B
Compound Foresets
215°
N = 18
Cross Beds
221°
N = 125

Paleocurrents from compound and simple (including compound intrasets) cross beds in unit B and partially exposed unit A are strongly unimodal towards the southwest (Fig. 17). Similar paleocurrent indicators within unit C indicate flow towards the southeast (Fig. 17.) In the middle panel of Profile C-1b, unit C is oriented nearly perpendicular to the flow direction. Bedding surfaces (i.e., set and coset boundaries) in this view are sub-horizontal indicating that deposition was primarily by downflow, instead of lateral accretion. Along depositional strike, some sub-horizontal beds between 1-2 m spaced bedding surfaces change to topset preserved sets (S_{tp}).

Surface U cuts down into the underlying strata towards the northern end of the outcrop to nearly truncate (within 30 cm) surface M (bottom panel of Profile C-1b). Abundant plant debris and mudclasts, up to 20 cm long overlie the surface. Also, along the base of unit D, near the top of the outcrop, are thin (< 30 cm thick), 1-2 m wide, dark-grey, mudstone beds (facies M). Paleocurrents within unit D are oriented to the northeast, significantly different from the other units (Fig. 17).

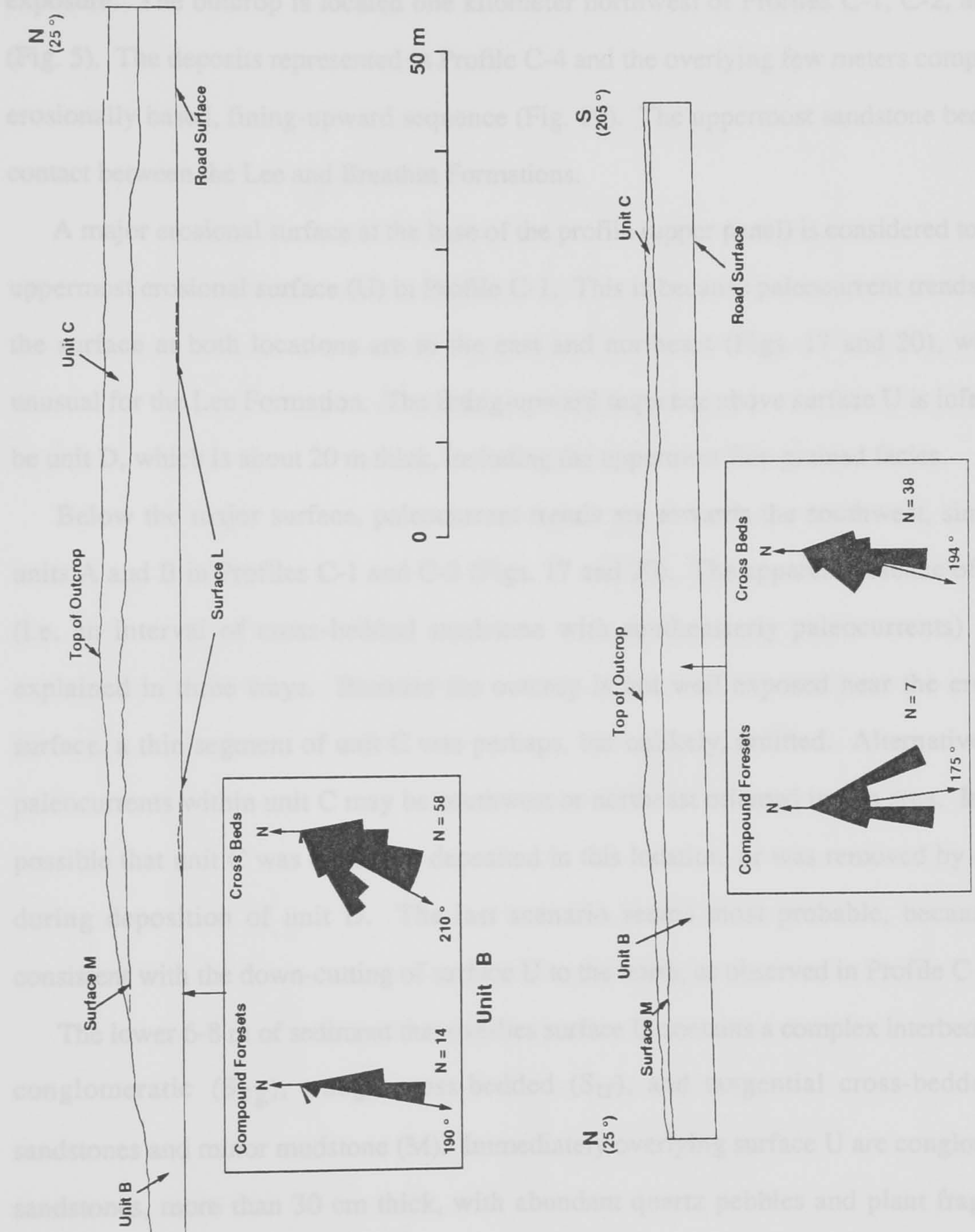
Profile C-2. The outcrop of Profile C-2 is parallel to that of Profile C-1 and is less than 100 m to the east (Fig. 5). About 6 m of vertical and more than 150 m of lateral section were investigated. Only a trace of surface L is apparent (left edge of center panel), but surface M is evident along nearly the entire length of the profile. Unit B is recognized by: 1) similar paleocurent mean to that of unit B in Profile C-1; 2) position below an erosional surface (M) that is overlain by deposits with paleoflow towards the southeast; and 3) comparable elevations of the outcrops in Profiles C-1 and C-2.

At this location, unit B contains similar facies to those in Profile C-1, except that compound cross-stratification (S_c) is much more common here. Large-scale tabular sets of planar (S_p) and tangential (S_t) cross beds dominate the lower part of the unit. Two basal sets are present in this profile, as opposed to the solitary set in Profile C-1; each is about

1 m thick. The lower set pinches out in the downflow direction, near the center of the profile, and is replaced by the upper set. Unlike the large-scale basal set in Profile C-1, the cross beds have long tangential bottomsets (S_{lb}), although the downflow set consists of planar foresets (S_p) near the southern end of the profile. Compound cross beds (S_c) overlie the large-scale sets in the same facies relationship as found in Profile C-1. Tops of compound cross beds are truncated by other compound sets, or by surface M. The upper truncation and downflow termination on the top of the large-scale basal sets bestow wedge-shape geometries to compound cross-bed sets. Paleocurrents for individual sets of compound cross beds are strongly unimodal, and because there is little variation among sets, reveal a strongly unimodal trend for unit B (Fig. 18a).

Profile C-3. Profile C-3 is located directly east, across the highway, from Profile C-2 (Fig. 5). The outcrop section, about 7 m high and 100 m wide, consists mostly of unit B (distinguished using same criteria as in Profile C-2). Surface L is not recognized, but because unit B is 5 m thick, it is inferred to be about 1 m below ground level. Surface M is nearly horizontal at this location and is overlain by a lenticular cross-bed set (S_s; Fig. 9b). Unit B contains several low-angle surfaces that delineate 1-2 m-thick cross-bed packages. Both compound and large-scale simple (tangential and planar) cross beds are common. In the downflow direction (south) there is a transition from primarily compound to simple cross beds. This downcurrent change appears to occur by preferential downflow thickening of a single intraset ('a', Profile C-3) within a compound cross-bed set. Within unit B the transition from compound to large-scale simple cross beds takes place in a step-like manner, with the stratigraphically higher compound cross-bed set changing in a more-upflow direction than the lower sets. Within the transition zone, it is common for a package to alternate between the simple and compound cross bedding. Paleocurrents for compound and simple cross beds are strongly unimodal to the south (Fig. 18b).

Figure 18. a) True-scale reduction of Profile C-2 with erosional surfaces L and M and paleocurrents for intervening unit B. b) True-scale reduction of Profile C-3 with erosional surface M and paleocurrents for intervening unit B.


S (205°)
N (25°)
Top of Outcrop
Unit C
Road Surface
Surface L
Surface M
Unit B
Compound Foresets
N
190°
N = 14
Cross Beds
N
210°
N = 58
Unit B
0
50 m
S (205°)
N (25°)
Unit C
Road Surface
Top of Outcrop
Unit B
Surface M
Compound Foresets
N
175°
N = 7
Cross Beds
N
194°
N = 38
Unit B

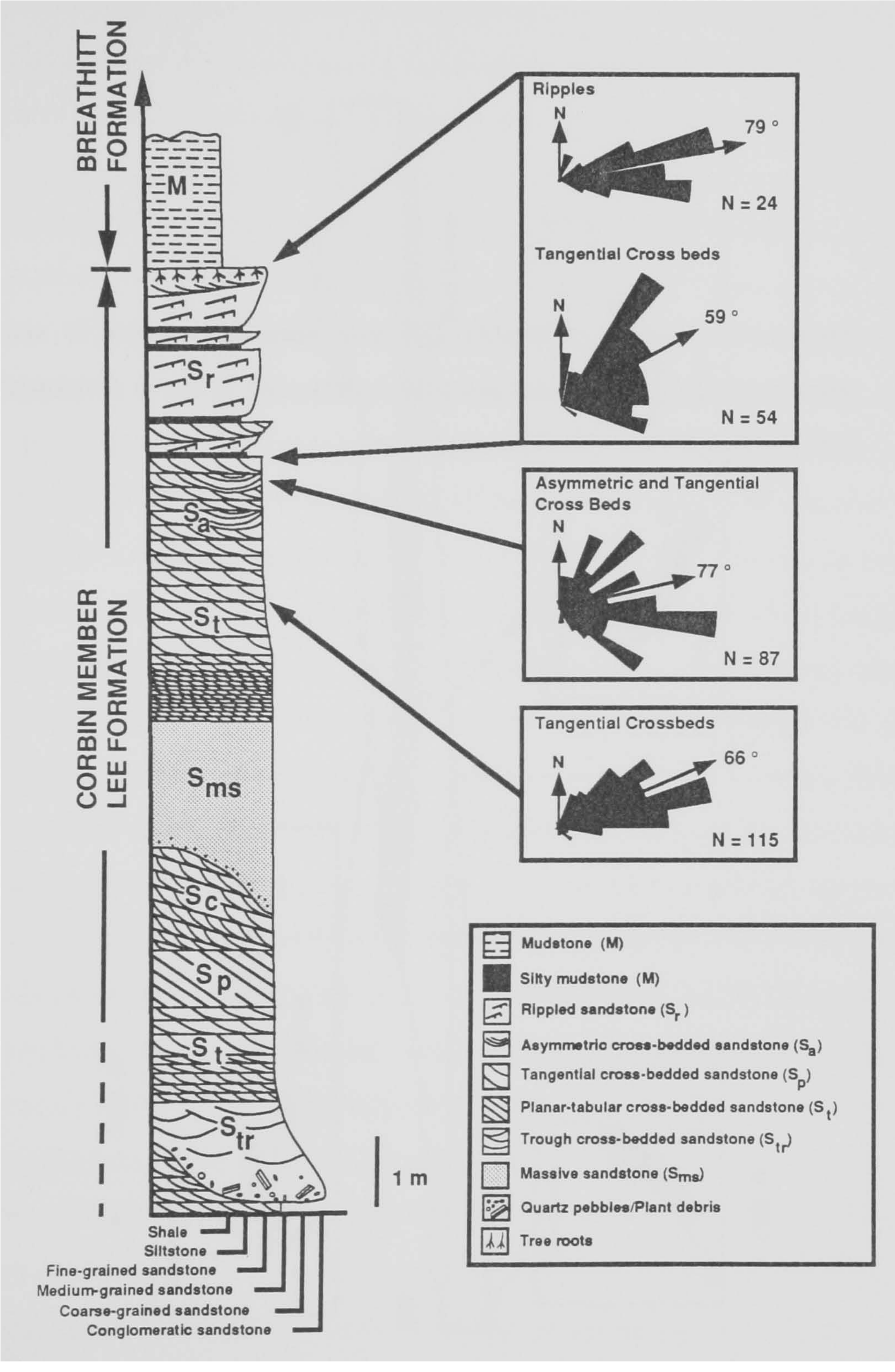
Profile C-4. Profile C-4 represents nearly 20 m of vertical section and 175 m of lateral exposure. The outcrop is located one kilometer northwest of Profiles C-1, C-2, and C-3 (Fig. 5). The deposits represented in Profile C-4 and the overlying few meters comprise an erosionally based, fining-upward sequence (Fig. 19). The uppermost sandstone bed is the contact between the Lee and Breathitt Formations.

A major erosional surface at the base of the profile (upper panel) is considered to be the uppermost erosional surface (U) in Profile C-1. This is because paleocurrent trends above the surface at both locations are to the east and northeast (Figs. 17 and 20), which is unusual for the Lee Formation. The fining-upward sequence above surface U is inferred to be unit D, which is about 20 m thick, including the uppermost fine-grained facies.

Below the major surface, paleocurrent trends are towards the southwest, similar to units A and B in Profiles C-1 and C-2 (Figs. 17 and 20). The apparent absence of unit C (i.e. an interval of cross-bedded sandstone with southeasterly paleocurrents) can be explained in three ways. Because the outcrop is not well exposed near the erosional surface, a thin segment of unit C was perhaps, but unlikely, omitted. Alternatively, the paleocurrents within unit C may be southwest or northeast oriented in this area. It is also possible that unit C was either not deposited in this location, or was removed by erosion during deposition of unit D. The last scenario seems most probable, because it is consistent with the down-cutting of surface U to the north, as observed in Profile C-1.

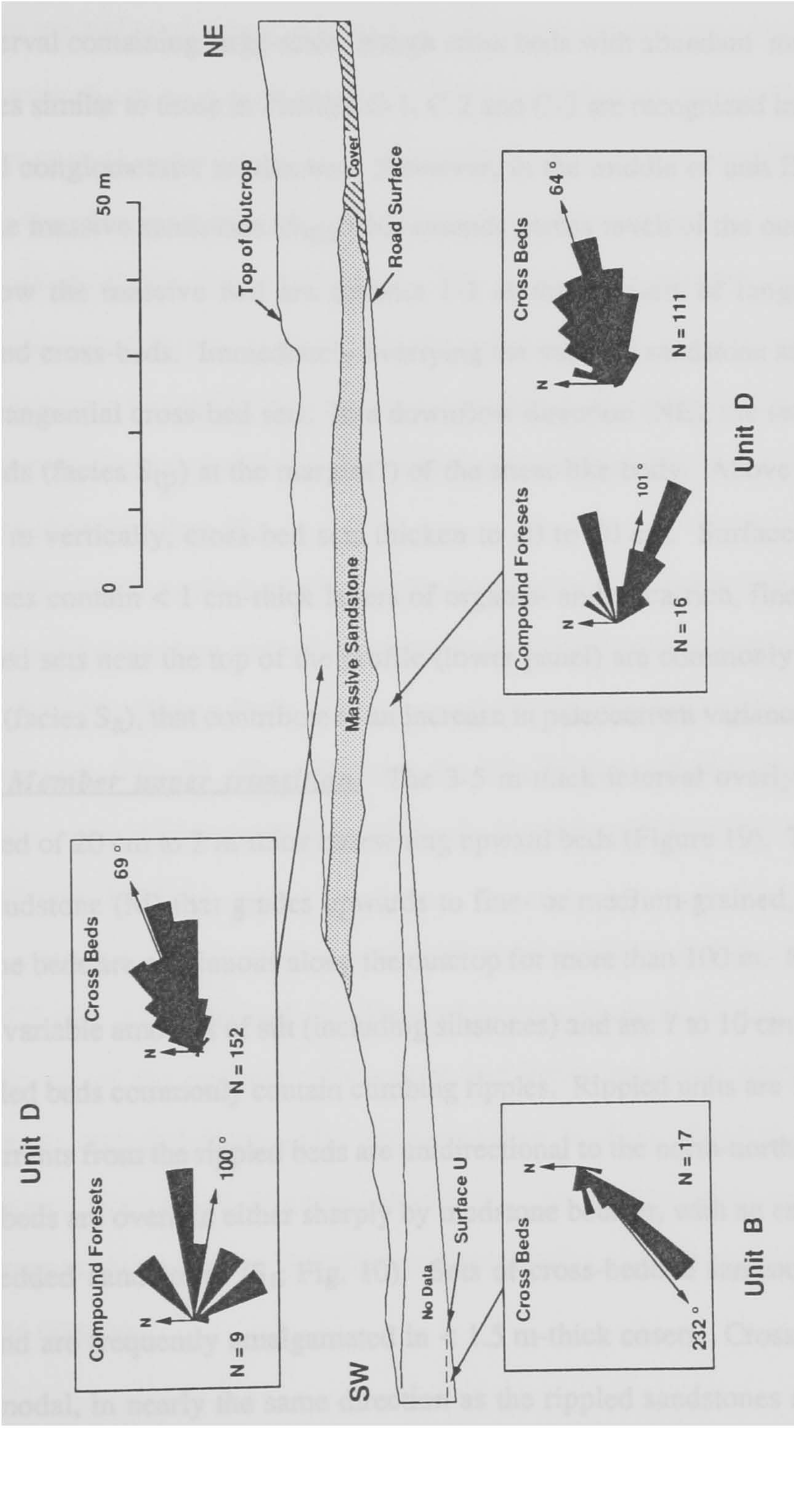
The lower 6-8 m of sediment that overlies surface U contains a complex interbedding of conglomeratic (S_{cg}), trough cross-bedded (S_{tr}), and tangential cross-bedded (S_t) sandstones and minor mudstone (M). Immediately overlying surface U are conglomeratic sandstones, more than 30 cm thick, with abundant quartz pebbles and plant fragments. Above this are large-scale, trough cross-bedded sandstones (S_t) with mudclast stringers. Tangential cross beds, with northeast paleocurrent trends, overlie the troughs. Near the top

Figure 19. Fining-upward sequence at the top of the Corbin Member (located along Rt. 770, Corbin, Ky.). Included are paleocurrents of cross-bedded sandstones (Profile C-4) and overlying fine-grained facies.



BREATHITT FORMATION
CORBIN MEMBER
LEE FORMATION
M
Sr
Sa
St
Sms
Sc
Sp
St
Str
1 m
Shale
Siltstone
Fine-grained sandstone
Medium-grained sandstone
Coarse-grained sandstone
Conglomeratic sandstone
Ripples
N
79 °
N = 24
Tangential Cross beds
N
59 °
N = 54
Asymmetric and Tangential Cross Beds
N
77 °
N = 87
Tangential Crossbeds
N
66 °
N = 115
Mudstone (M)
Silty mudstone (M)
Rippled sandstone (Sr)
Asymmetric cross-bedded sandstone (Sa)
Tangential cross-bedded sandstone (Sp)
Planar-tabular cross-bedded sandstone (St)
Trough cross-bedded sandstone (Str)
Massive sandstone (Sms)
Quartz pebbles/Plant debris
Tree roots

Figure 20. True-scale reduction of Profile C-4 with erosional surface U and paleocurrents of cross-bedded sandstones within unit D that is developed both above and below sheet-like massive unit.



SW
NE
0
50 m
Top of Outcrop
Road Surface
Cover
Massive Sandstone
No Data
Surface U
Unit D
Compound Foresets
N = 9
100 °
Cross Beds
69 °
N = 152
Unit B
Cross Beds
232 °
N = 17
Unit D
Compound Foresets
101 °
N = 16
Cross Beds
64 °
N = 111

of the trough cosets are at least two thin mudstone lens. The cross-bedded facies are truncated by a steeply-dipping, second erosional surface (U'). Above surface U' is a 2 m-thick interval containing large-scale, trough cross beds with abundant mudstone clasts.

Facies similar to those in Profiles C-1, C-2 and C-3 are recognized in Profile C-4 above the basal conglomeratic sandstones. However, in the middle of unit D is a < 5 m thick, sheet-like massive sandstone (S_{ms}) that extends across much of the outcrop. Both above and below the massive bed are distinct 1-2 m thick cosets of tangential, planar and compound cross-beds. Immediately overlying the massive sandstone are very-thin (about 10 cm) tangential cross-bed sets. In a downflow direction (NE), the sets form convex-up cross beds (facies S_{tp}) at the margin(?) of the sheet-like body. Above the thin sets, over about 5 m vertically, cross-bed sets thicken to 40 to 70 cm. Surfaces bounding cosets sometimes contain < 1 cm-thick layers of organic- and mica-rich, fine-grained material. Cross-bed sets near the top of the profile (lower panel) are commonly asymmetric-filled troughs (facies S_a), that contribute to an increase in paleocurrent variance (Fig. 19).

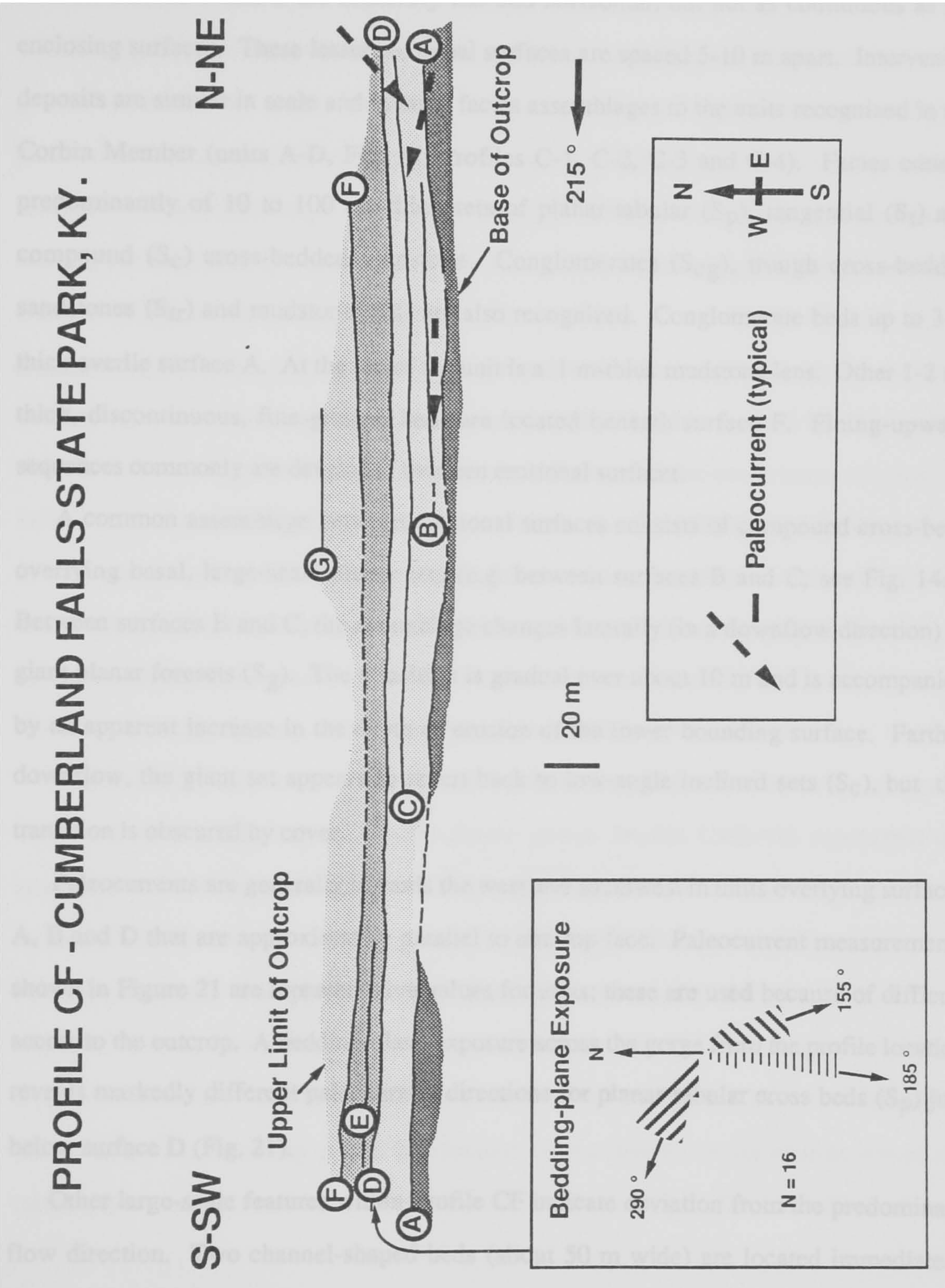
Corbin Member upper transition. The 3-5 m-thick interval overlying Profile C-4 is composed of 20 cm to 2 m-thick coarsening upward beds (Figure 19). The beds consist of basal mudstone (M) that grades upwards to fine- or medium-grained, rippled sandstone (S_r). The beds are continuous along the outcrop for more than 100 m. Mudstone beds (M) contain variable amounts of silt (including siltstones) and are 7 to 10 cm thick. The base of the rippled beds commonly contain climbing ripples. Rippled units are 10 cm to 2 m thick. Paleocurrents from the rippled beds are unidirectional to the north-northeast (Fig. 19). The rippled beds are overlain either sharply by mudstone beds or, with an erosional contact, by cross-bedded sandstones (S_t; Fig. 10). Sets of cross-bedded sandstone are 10-100 cm thick, and are frequently amalgamated in < 1.5 m-thick cosets. Cross-bed paleocurrents are unimodal, in nearly the same direction as the rippled sandstones and the underlying cross-bedded sandstone (Fig. 19). Tops of cross-bedded sandstone beds are flat to undulatory and are sharply overlain by mudstone. In one sequence an undulating sandstone top and overlying mudstone bed demarcate the preserved bedform (S_b; Fig. 10). The uppermost coarsening-upward unit (top of the Corbin Member) contains in-place, fossil tree roots of *Stigmaria ficoides*. Immediately overlying the mudstone/sandstone beds are more than 5 m of mudstone (M) gradational into black shales.

Rockcastle and Bee Rock Members. The Rockcastle Member was studied within the flat-lying stratigraphy of the Cumberland Plateau (Fig. 2). The study area of the Bee Rock Member is located in the central part of the synclinal Cumberland overthrust sheet, between steep-dipping rocks of the Cumberland and Pine Mountain outcrop belts (Fig. 2); structural dip is less than 5°.

Profile CF. A 40 m-thick section of the Rockcastle Member is exposed for more than 400 m along the western wall of the gorge immediately downstream of Cumberland Falls (Profile CF). The Barren Fork coal bed, overlying the member, is recognized near the top of the outcrop (Fig. 21, G). Although the base is not exposed, Profile CF represents nearly the entire thickness of the Rockcastle Member, which is up to 43 m thick in this region (Smith, 1963).

Six extensive, sub-parallel and sub-horizontal erosional surfaces are apparent within Profile CF (Fig. 21, surfaces A-F). Three of these surfaces (A, D, and F) can be traced the entire length of the outcrop. The upper and lower erosional surfaces (A and F) are concave-up and undulatory with approximately 5-10 m of relief. Surface D is the lower/upper Rockcastle Member boundary. An abrupt decrease in quartz pebble concentration above surface D facilitates identification, even though it is nearly horizontal along the outcrop.

Figure 21. True-scale reduction of Profile CF with major erosional surfaces (A-F), upper surface of the Rockcastle Member (G) and paleocurrent data. Stippled patterns separate extensive erosional surfaces (A, D, and F). Arrows indicate single measurement taken as representative of an individual unit (between erosional surfaces). Bedding-plane exposure, located across the gorge from the profile, reveals paleocurrent variation near the top of the unit above surface C. Each pattern in rose diagram represents an individual planar-tabular set (S_p).



PROFILE CF - CUMBERLAND FALLS STATE PARK, KY.
S-SW
N-NE
Upper Limit of Outcrop
Base of Outcrop
A
B
C
D
E
F
G
215 °
20 m
Paleocurrent (typical)
N
E
S
W
Bedding-plane Exposure
N
290 °
155 °
185 °
N = 16

Surfaces B, C and E are relatively flat and horizontal, but not as continuous as the enclosing surfaces. These lesser erosional surfaces are spaced 5-10 m apart. Intervening deposits are similar in scale and internal facies assemblages to the units recognized in the Corbin Member (units A-D, Fig. 17; Profiles C-1, C-2, C-3 and C-4). Facies consist predominantly of 10 to 100 cm-thick sets of planar-tabular (S_p), tangential (S_t) and compound (S_c) cross-bedded sandstone. Conglomerates (S_{cg}), trough cross-bedded sandstones (S_{tr}) and mudstones (M) are also recognized. Conglomerate beds up to 3 m thick overlie surface A. At the top of the unit is a 1 m-thick mudstone lens. Other 1-2 m-thick, discontinuous, fine-grained beds are located beneath surface F. Fining-upward sequences commonly are developed between erosional surfaces.

A common assemblage between erosional surfaces consists of compound cross-beds overlying basal, large-scale planar sets (e.g. between surfaces B and C; see Fig. 14a). Between surfaces B and C, the assemblage changes laterally (in a downflow direction) to giant planar foresets (S_g). The transition is gradual over about 10 m and is accompanied by an apparent increase in the depth of erosion of the lower bounding surface. Farther downflow, the giant set appears to revert back to low-angle inclined sets (S_c), but the transition is obscured by cover.

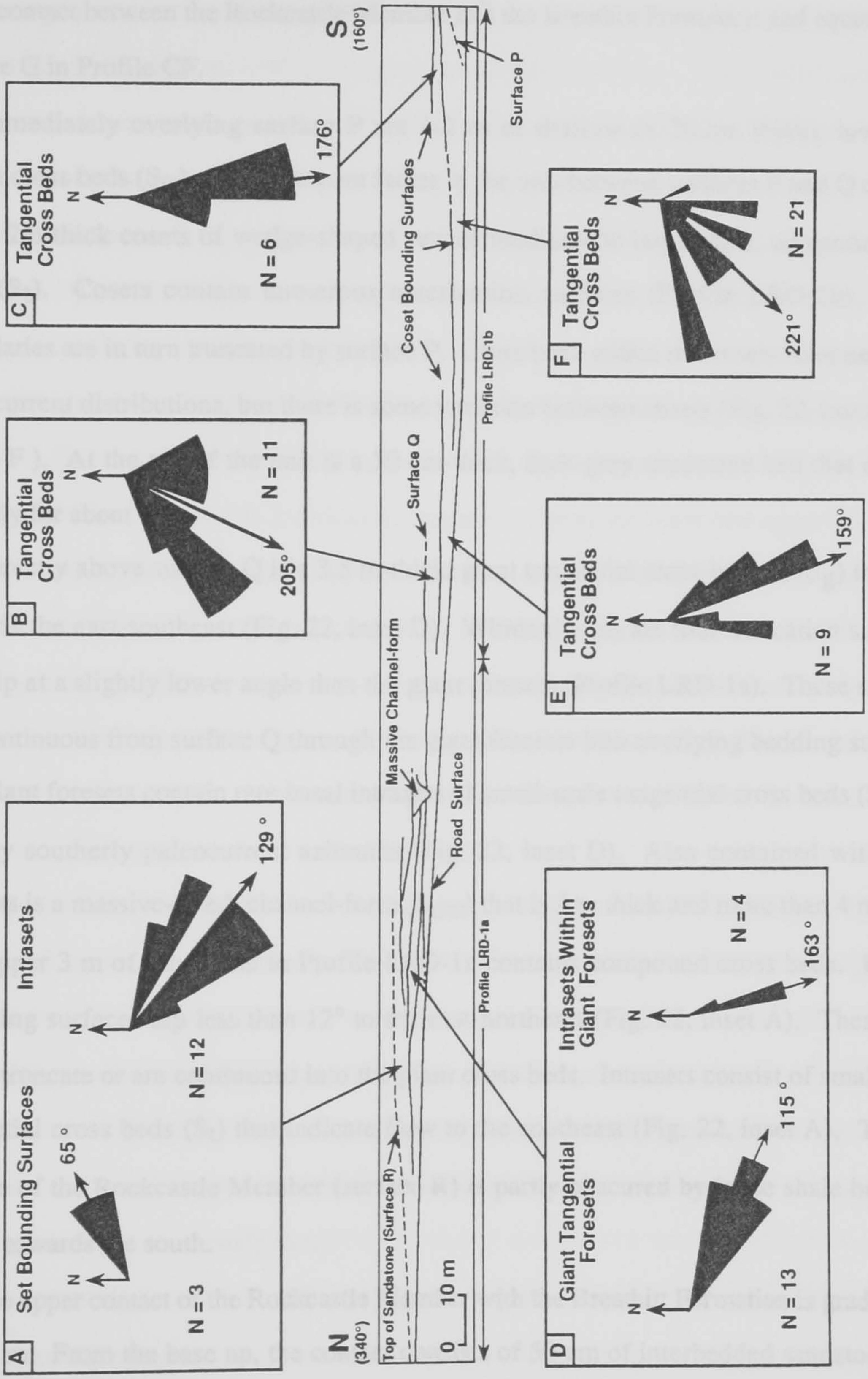
Paleocurrents are generally towards the west and southwest in units overlying surfaces A, B and D that are approximately parallel to outcrop face. Paleocurrent measurements shown in Figure 21 are representative values for units; these are used because of difficult access to the outcrop. A bedding-plane exposure across the gorge from the profile location reveals markedly different paleocurrent directions for planar-tabular cross beds (S_p) just below surface D (Fig. 21).

Other large-scale features within Profile CF indicate deviation from the predominant flow direction. Two channel-shaped beds (about 50 m wide) are located immediately beneath surface D at the southern end of the profile (Profile CF). These features, at the same stratigraphic level as the bedding-plane exposure, suggest flow highly oblique to the general paleocurrent trend. Deposits overlying surface F appear to have different paleocurrent azimuths to the underlying units, with flow estimated to be towards the east or southeast (no data available). On the southern end of the profile, these deposits contain bedding surfaces that dip about 10° towards the south(?) with intrasets of trough cross beds oriented oblique to the bedding surfaces.

Profiles LRD-1 and LRD-2. The upper Rockcastle Member outcrops on the road descending towards the base of Laurel River dam (Profile LRD-1, a and b). The profile extends vertically for more than 15 m and laterally for over 225 m but, because of cover, an individual bed can be traced for a maximum 100 m. Profile LRD-2, near the base of the dam, represents about 15 m of vertical and 100 m of lateral outcrop. The bottom third to half of this profile consists of the lower Rockcastle Member. The stratigraphic thickness of each of the lower and upper Rockcastle Members in this area is *ca.* 20 m (Donald Chesnut, Kentucky Geologic Survey, written communication, 1989).

Profile LRD-1 contains three significant surfaces (Fig. 22); two of these are erosional. Surface P at the base of the profile (lower panel, Profile LRD-1b) represents the upper/lower Rockcastle Member boundary. Although there is a textural change across surface P similar to that across surface D in Profile CF, these two surfaces cannot be directly correlated. Surface Q lies approximately in the middle of the upper Rockcastle Member. This surface is overlain by a thin (<10 cm thick) bed of conglomeratic sandstone (S_{cg}) with abundant mudclasts and quartz granules. This surface can be traced about 50 m before being concealed by cover, but because of the distinct facies change across the surface, it can be followed across the whole outcrop (including Profile LRD-2). Surface R

Figure 22. True-scale reduction of Profile LRD-1 with major erosional surfaces (P and Q), top of the Rockcastle Member (surface R), and paleocurrent data from giant foresets (S_g), set bounding surfaces, intrasets and tangential cross beds (S_t) within cosets.


A Set Bounding Surfaces
N
65°
N = 3
Intrasets
N
129 °
N = 12
B Tangential Cross Beds
N
205°
N = 11
C Tangential Cross Beds
N
176°
N = 6
N (340°)
S (160°)
Top of Sandstone (Surface R)
0 10 m
Massive Channel-form
Road Surface
Surface Q
Coset Bounding Surfaces
Surface P
Profile LRD-1a
Profile LRD-1b
D Giant Tangential Foresets
N
115 °
N = 13
Intrasets Within Giant Foresets
N
163 °
N = 4
E Tangential Cross Beds
N
159°
N = 9
F Tangential Cross Beds
N
221°
N = 21

is the contact between the Rockcastle Member and the Breathitt Formation and equates with surface G in Profile CF.

Immediately overlying surface P are 1-2 m of shallow (< 20 cm thick), low-angle trough cross beds (S_{tr}). The dominant facies in the unit between surfaces P and Q consists of *ca.* 2m-thick cosets of wedge-shaped sets of medium- to large-scale, tangential cross beds (S_t). Cosets contain numerous reactivation surfaces (Profile LRD-1b). Coset boundaries are in turn truncated by surface P. Cross beds within the cosets have unimodal paleocurrent distributions, but there is some variation between cosets (Fig. 22, insets B, C, E and F). At the top of the unit is a 50 cm-thick, dark-grey mudstone bed that extends laterally for about 6 m.

Directly above surface Q is a 3.5 m-thick, giant tangential cross-bed set (S_g) that dips towards the east-southeast (Fig. 22, inset D). Within the set are four truncation surfaces, that dip at a slightly lower angle than the giant foresets (Profile LRD-1a). These surfaces are continuous from surface Q through the giant foresets into overlying bedding surfaces. The giant foresets contain rare basal intrasets of small-scale tangential cross beds (S_t) that display southerly paleocurrent azimuths (Fig. 22, inset D). Also contained within the foresets is a massive-filled, channel-form (S_{mc}) that is 3 m thick and more than 4 m wide. The upper 3 m of sandstone in Profile LRD-1a contains compound cross beds. Intraset bounding surfaces dip less than 12° to the east-northeast (Fig. 22, inset A). These beds either truncate or are continuous into the giant cross beds. Intrasets consist of small-scale tangential cross beds (S_t) that indicate flow to the southeast (Fig. 22, inset A). The top surface of the Rockcastle Member (surface R) is partly obscured by loose shale but rises gently towards the south.

The upper contact of the Rockcastle Member with the Breathitt Formation is gradational over 2 m. From the base up, the contact consists of 50 cm of interbedded sandstone (S_r)

and dark gray mudstone, 50 cm of dark gray mudstone (M), and about 1 m of black shales that is sharply overlain by a 60 cm-thick coal bed (Barren Fork). Interbedded sandstones are less than 10 cm thick and contain abundant mica, disseminated plant fragments and cm-size clasts of gray mudstone. The dark gray mudstone fines upward, but there is a sharp contact with the overlying black shale. Both these units contain abundant impressions of plant material. The shale contains siderite nodules, 2-3 cm in diameter. There is no apparent bioturbation or body fossils at this locality, but in a section along the dam spillway creek (located about 1 km south of the outcrop), some bioturbation is present in rippled sandstone (S_r) at a similar stratigraphic level.

Surface P in Profile LRD-2 defines the contact between the lower and upper Rockcastle Members. Surface P is truncated by surface Q in the center of the top panel. Surface Q exhibits a broad, channel-shape geometry in the southeastern part of the profile. In the central profile area there is some uncertainty as to the exact position as to surface Q because of complex bedding relationships and poor exposure, but at least 7 m of erosional relief is evident along the surface. Thin lenses of grey mudstone, less than 30 cm thick and several meters long, are common above surface Q.

The upper Rockcastle Member contains several laterally extensive bedding surfaces that dip toward the eastern end of the profile and are truncated at the bottom by surfaces P and Q. These surfaces, spaced about 4 m apart, contain large-scale cross beds (S_g) between them. Despite the inaccessibility of the upper part of the outcrop, paleocurrent trends can be estimated for the cross beds. Low-angle stratification and trough-shapes immediately above the surface Q suggests flow to the southeast. Steeper cross beds near the top of the outcrop indicate more easterly flow.

The lower Rockcastle Member contains compound cross stratification (S_c). Three, sub-horizontal, compound-set (coset) bounding surfaces are spaced about 4 m apart and extend across the profile. The upper surface is truncated by surface P. Bounding surfaces to intrasets within the cosets are inclined at < 12°. Paleocurrents from the planar-tabular intrasets indicate flow towards the south (Fig. 23a), whereas set bounding surfaces dip parallel to slightly oblique to the dip directions of the intrasets (Fig. 23b).

Profile J. The Bee Rock Member in Profile J consists of a 10 m vertical section that is laterally continuous for more than 120 m. Because of significant relief in the roadcut face, the profile is distorted. The primary feature of interest in this outcrop is a massive sheet-like sandstone (S_{ms}), 5-10 m thick and extending for several hundred meters along the outcrop (Fig. 15a). A distortion-free profile would reveal the horizontal nature of the top of the massive bed.

An erosional surface is located at the base of the eastern end of the profile, but is largely obscured by cover. The surface is overlain by conglomeratic sandstone (S_{cg}) with abundant quartz pebbles, large plant fragments and mudclasts. Paleocurrent data from tangential (S_t) and planar-tabular (S_p) cross beds above the surface are to the west and southwest, but below the surface are to the southeast (Fig. 24). Above the massive bed are large-scale, planar-tabular cross beds (up to 2.5 m thick) that are inclined towards the southwest (Fig. 24). Also of interest in this outcrop is a nearly 10 m-thick interval of distorted tangential cross beds (S_t; Fig. 8) immediately below the massive sandstone bed on the western end of the profile.

Middlesboro Member. Several extensive exposures of the Middlesboro Member were examined along Interstate 75 on Pine Mountain, northern Tennessee. However, tectonic tilting and fracturing, in addition to oblique exposure angle, make many exposures unsuitable for lateral profiling. Nevertheless, three profiles were prepared in order to document important sedimentary features and to allow for comparison of the Middlesboro Member with the both the Corbin and Rockcastle Members.

Figure 23. Paleocurrents from planar-tabular cross beds (S_p) and enclosing set bounding surfaces within the lower Rockcastle Member at Laurel River dam (Profile LRD-2).

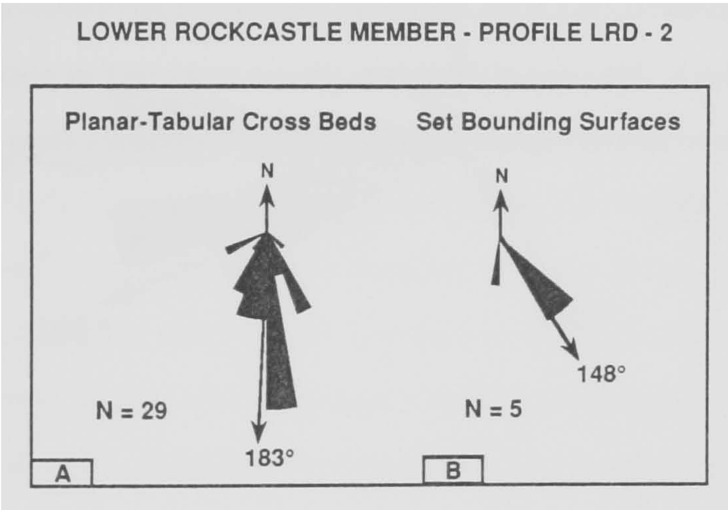
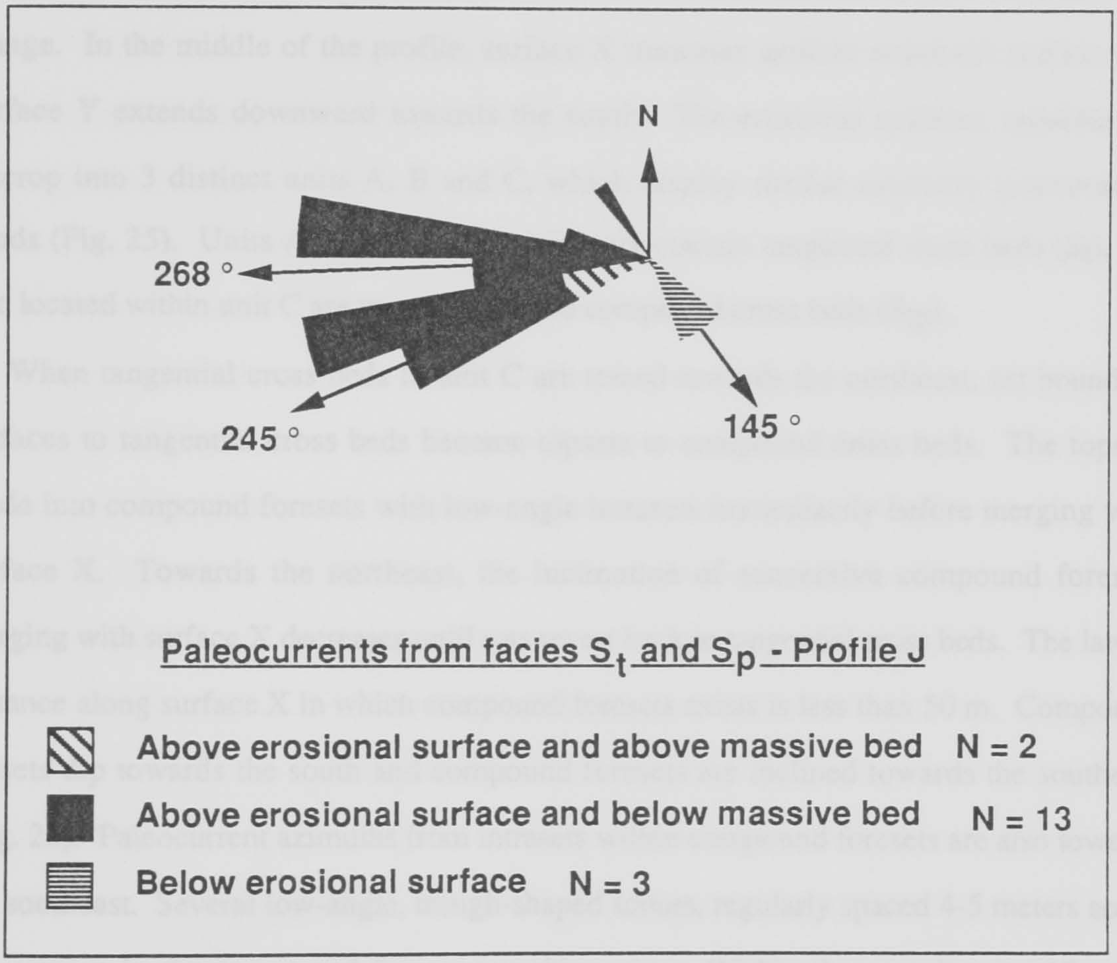

LOWER ROCKCASTLE MEMBER - PROFILE LRD - 2
Planar-Tabular Cross Beds
Set Bounding Surfaces
N
N
148°
N = 29
N = 5
183°
A
B

Figure 24. Paleocurrent data from tangential (S_t) and planar-tabular (S_p) cross beds in Profile J. Data are divided into groups according to position relative to a major erosional surface (near base of outcrop) and a sheet-like massive bed (Fig. 15a).

PROFILE J - PALEOCURRENTS



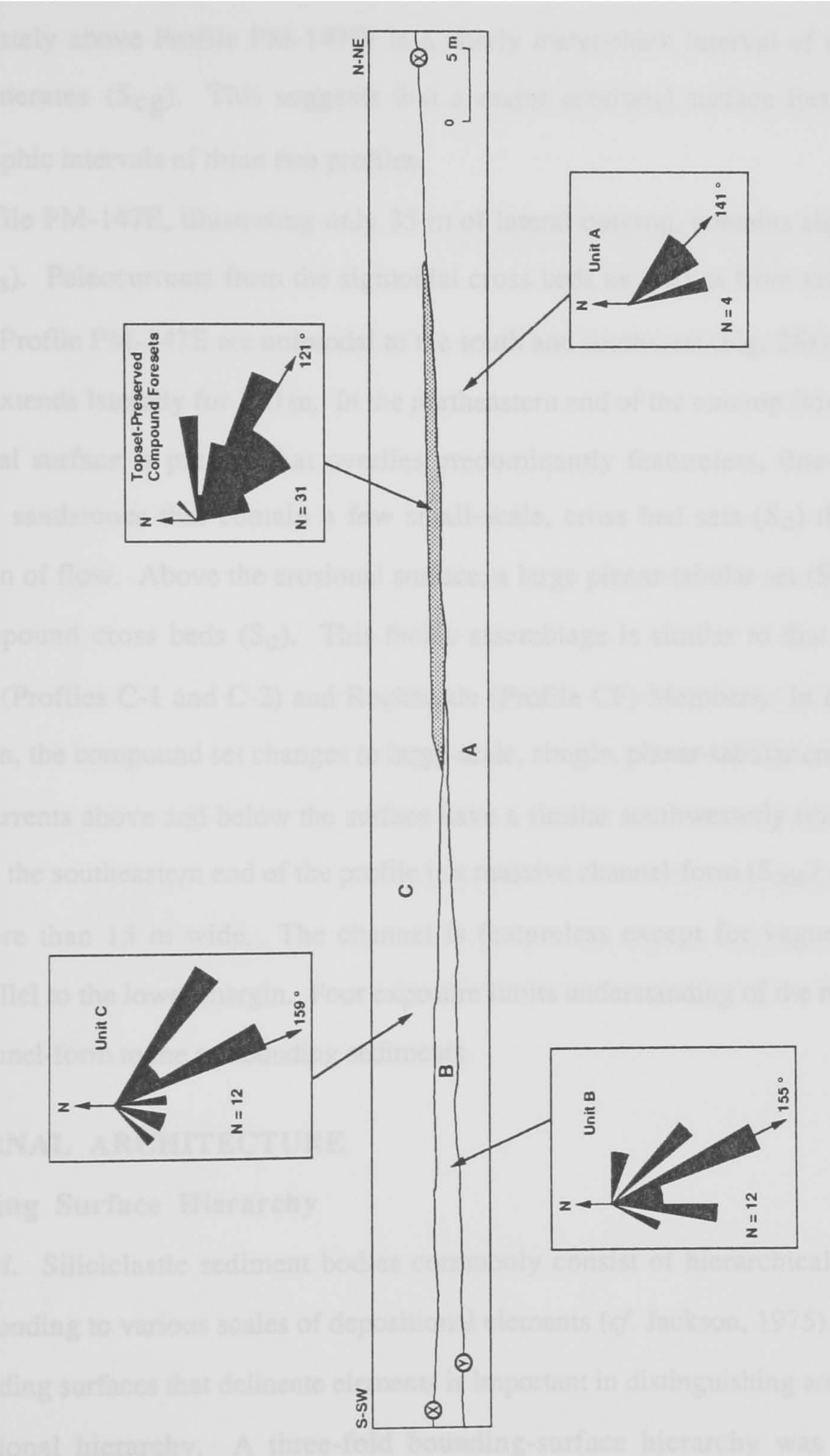
Profile PM-151. Profile PM-151 represents more than 5 m of vertical section in the lowermost part of the Middlesboro Member. The outcrop extends laterally for more than 100 meters. There is 20° of structural dip towards the southeast.

A low-angle, concave-up, erosional surface (X), midway up the profile, can be traced along the outcrop for more than 0.5 km. Across this surface there is no apparent grain-size change. In the middle of the profile, surface X truncates another erosional surface (Y). Surface Y extends downward towards the south. The erosional surfaces separate the outcrop into 3 distinct units A, B and C, which display similar southerly paleocurrent trends (Fig. 25). Units A, B and C predominantly contain tangential cross beds (S_t), but also located within unit C are topset-preserved compound cross beds (S_{tp}).

When tangential cross beds in unit C are traced towards the northeast, set bounding surfaces to tangential cross beds become topsets to compound cross beds. The topsets grade into compound foresets with low-angle intrasets immediately before merging with surface X. Towards the northeast, the inclination of successive compound foresets merging with surface X decreases until sets revert back to tangential cross beds. The lateral distance along surface X in which compound foresets exists is less than 50 m. Compound topsets dip towards the south and compound foresets are inclined towards the southeast (Fig. 25). Paleocurrent azimuths from intrasets within compound foresets are also towards the southeast. Several low-angle, trough-shaped scours, regularly spaced 4-5 meters apart, are present in the topset and foreset transition area. Within the troughs, stratification appears to dip parallel to the local paleocurrent direction (i.e. toward the southeast).

Profiles PM-147W and PM-147E. Profiles PM-147W and PM-147E, across the highway from one another, are located 6 km southwest of Profile PM 151. The profiles contains 5 m of section and are located below deposits comprising Profile PM-151. Structural dip is less than 15° to the southeast. Because of the dip, the westerly outcrop is

Figure 25. True-scale reduction of Profile PM-151 with major erosional surfaces (X and Y) and paleocurrent data from units A, B and C and from foresets of topset-preserved, compound cross beds (stippled area).


S-SW
N-NE
0
5 m
A
B
C
Topset-Preserved
Compound Foresets
N
N = 31
121 °
Unit A
N
N = 4
141 °
Unit C
N
N = 12
159 °
Unit B
N
N = 12
155 °

located stratigraphically just below the eastern exposure. In a poorly exposed outcrop immediately above Profile PM-147W is a nearly meter-thick interval of quartz pebble conglomerates (S_{cg}). This suggests that a major erosional surface lies between the stratigraphic intervals of these two profiles.

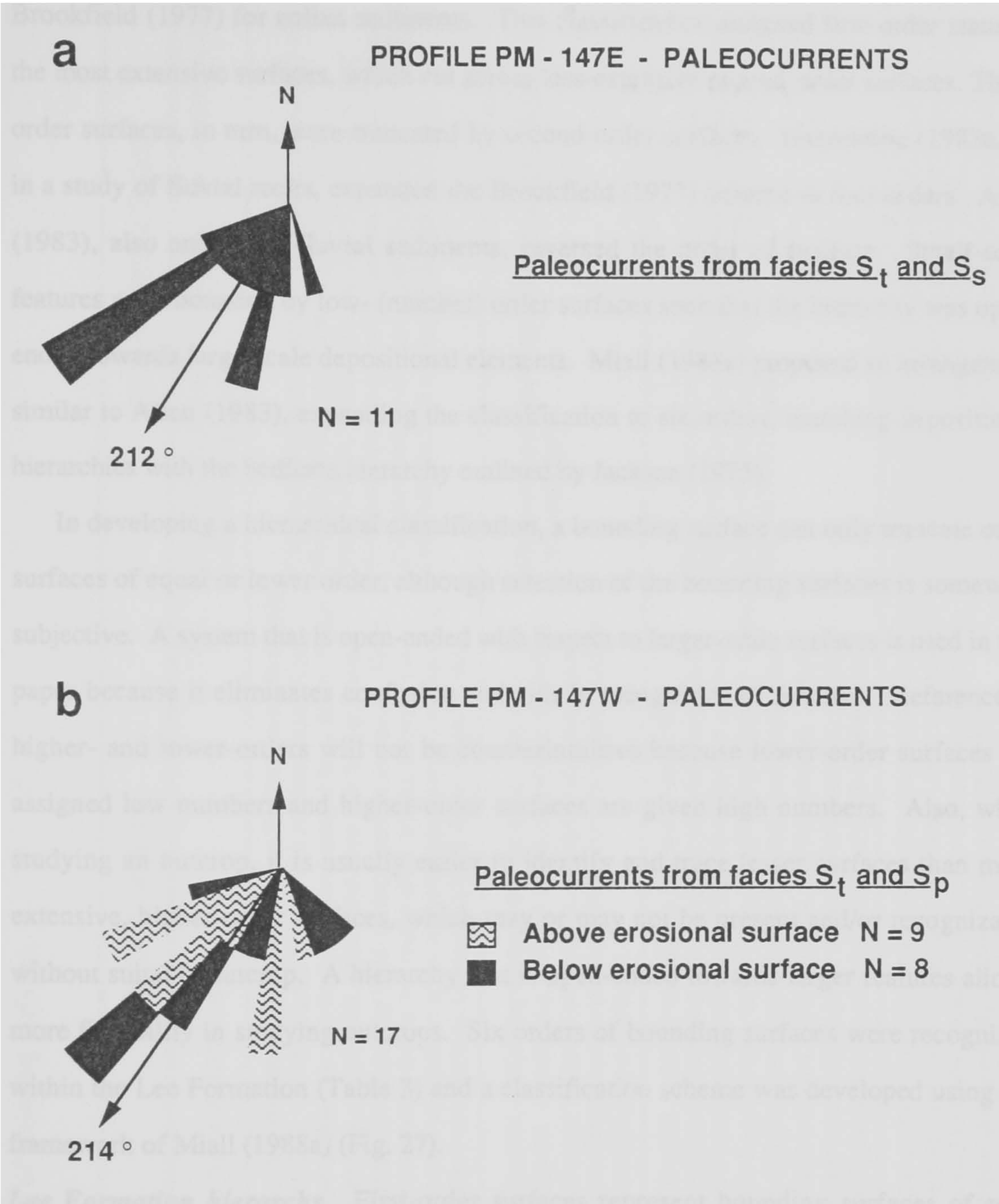
Profile PM-147E, illustrating only 35 m of lateral outcrop, contains sigmoidal cross beds (S_s). Paleocurrents from the sigmoidal cross beds as well as from tangential cross beds in Profile PM-147E are unimodal to the south and southwest (Fig. 26a). Profile PM-147W extends laterally for 130 m. In the northeastern end of the outcrop (lower panel), an erosional surface is present that overlies predominantly featureless, fine- to medium-grained sandstones that contain a few small-scale, cross bed sets (S_c) that dip in the direction of flow. Above the erosional surface, a large planar-tabular set (S_p) is overlain by compound cross beds (S_c). This facies assemblage is similar to that found in the Corbin (Profiles C-1 and C-2) and Rockcastle (Profile CF) Members. In the downflow direction, the compound set changes to large-scale, simple, planar-tabular cross beds (S_p). Paleocurrents above and below the surface have a similar southwesterly orientation (Fig. 26). At the southeastern end of the profile is a massive channel-form (S_{mc}) about 3 m tall and more than 15 m wide. The channel is featureless except for vague laminations subparallel to the lower margin. Poor exposure limits understanding of the relationship of the channel-form to the surrounding sediments.

INTERNAL ARCHITECTURE

Bounding Surface Hierarchy

General. Siliciclastic sediment bodies commonly consist of hierarchical subdivisions corresponding to various scales of depositional elements (*cf.* Jackson, 1975). Recognition of bounding surfaces that delineate elements is important in distinguishing and describing a depositional hierarchy. A three-fold bounding-surface hierarchy was proposed by

Figure 26. Paleocurrent data for Profiles PM-147E (a) and PM-147W (b).


a
PROFILE PM - 147E - PALEOCURRENTS
N
Paleocurrents from facies S_t and S_s
N = 11
212 °
b
PROFILE PM - 147W - PALEOCURRENTS
N
Paleocurrents from facies S_t and S_p
Above erosional surface N = 9
Below erosional surface N = 8
N = 17
214 °

Brookfield (1977) for eolian sediments. This classification assigned first-order status to the most extensive surfaces, which cut across less-extensive second-order surfaces. Third-order surfaces, in turn, were truncated by second-order surfaces. Haszeldine (1983a, b), in a study of fluvial rocks, expanded the Brookfield (1977) scheme to four orders. Allen (1983), also analyzing fluvial sediments, reversed the order of ranking. Small-scale features were bounded by low- (number) order surfaces such that the hierarchy was open-ended towards large-scale depositional elements. Miall (1988a) proposed an arrangement similar to Allen (1983), expanding the classification to six orders, matching depositional hierarchies with the bedform hierarchy outlined by Jackson (1975).

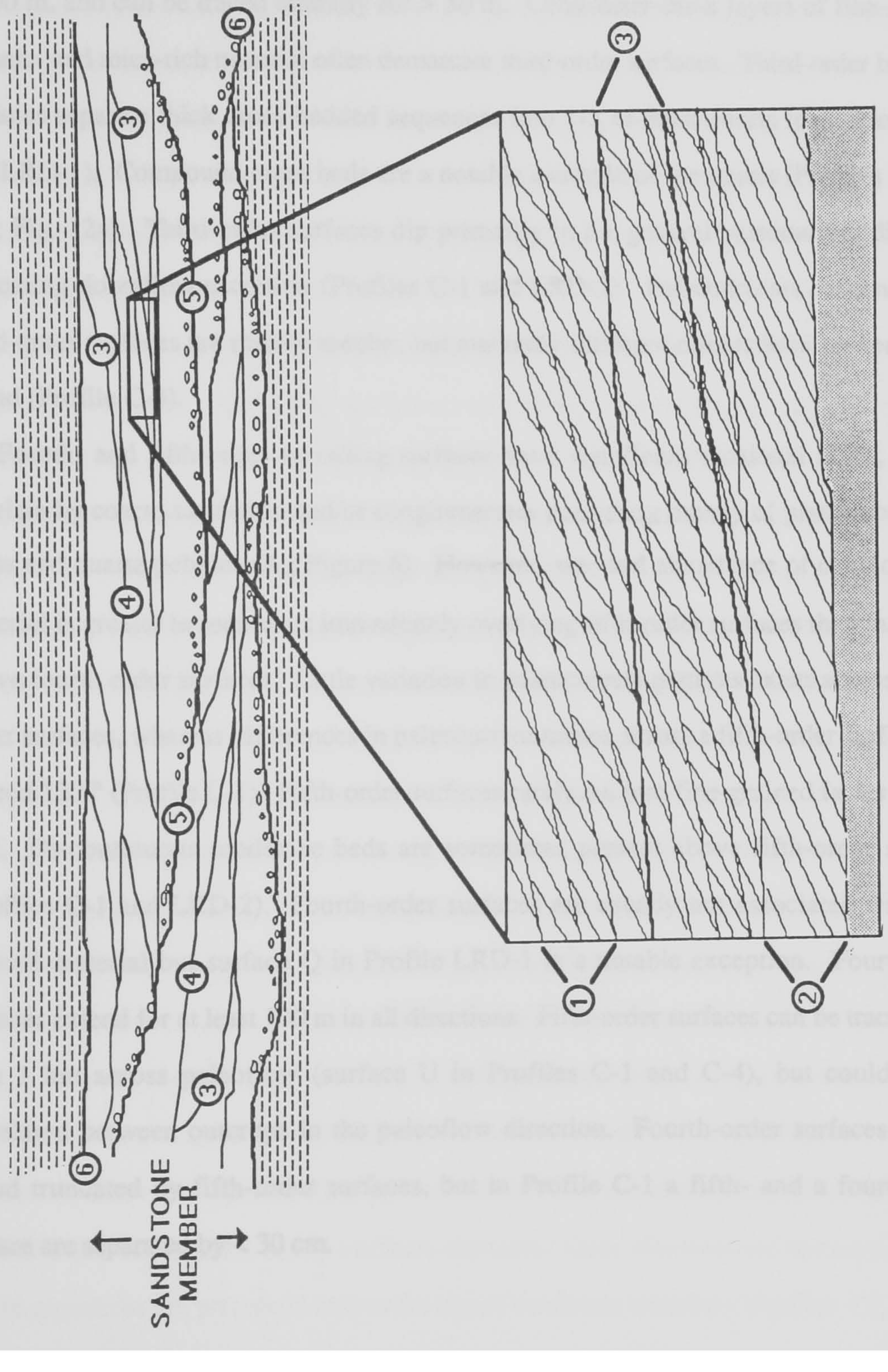
In developing a hierarchical classification, a bounding surface can only truncate other surfaces of equal or lower order, although selection of the bounding surfaces is somewhat subjective. A system that is open-ended with respect to larger-scale surfaces is used in this paper because it eliminates confusion and aids in recognition of features. Reference to higher- and lower-orders will not be counterintuitive because lower-order surfaces are assigned low numbers and higher-order surfaces are given high numbers. Also, when studying an outcrop, it is usually easier to identify and trace lesser surfaces than more extensive, higher-order surfaces, which may or may not be present and/or recognizable without suitable outcrop. A hierarchy that is open-ended towards larger features allows more flexibility in studying outcrops. Six orders of bounding surfaces were recognized within the Lee Formation (Table 3) and a classification scheme was developed using the framework of Miall (1988a) (Fig. 27).

Lee Formation hierarchy. First-order surfaces represent bounding surfaces of sets. Second-order surfaces separate cosets of different facies. Both first- and second-order surfaces are described in Table 2. Third-order bounding surfaces are low-angle (< 15°), and erosional; they truncate either downflow-dipping cross-bed sets (e.g., Profile C-1) or

Table 3. Summary of bounding surfaces.

ORDER	DESCRIPTION	INTERPRETATION
First and Second	See facies descriptions.	Represents boundaries between sets and cosets of mesoform deposits.
Third	Dip < 15° downflow (to slightly oblique?). Convex-up to slightly undulatory. Can be traced 10 to > 100 m along flow. Transverse to flow are sub-horizontal and extend for > 50 m. Often marked by ≤ 1 cm thick, plant- and mica-rich fine-grained material. Truncated by third- or fourth-order surfaces. Third-order surfaces define 1-2 meter thick cosets of cross beds. Paleocurrents across surfaces generally similar, but markedly different orientations between cosets sometimes found.	Surfaces separate deposits of dunes that moved down the face of a larger-scale bed configuration (macroform). Paleocurrents and downflow dips of surfaces indicate primarily down-flow accretion. Fine-grained material implies dunes sometimes flowed into relatively inactive areas of macroform surface.
Fourth	< 5 m relief, undulatory surface that extends > 100 m in all directions. Overlain by ≤ 50 cm of conglomeratic sandstone lag. Except for basal sandstone, no significant grain size, facies change across surface. Paleocurrent patterns similar above and below surfaces.	Surface separates large-scale depositional elements (macroforms).
Fifth	> 5 m relief, concave-up surface that extends >1 km transverse to flow direction. Probably extends much more than 1 km along the direction of flow. Overlain by < 3 m of facies S_{cg} with rare, large-scale, trough cross bedding. Facies S_{cg} contains abundant plant debris and mud clasts. Rare mudstone lenses, ≤30 cm thick, overlie surfaces. Paleocurrent means often show marked change across surface. Significant grain size change across surface in one case.	Major erosional surface, separates major-channel deposits. Frequently truncated by subsequent similar surfaces. Represents major avulsive event within the depositional system.
Sixth	Upper surface is gradational to dark gray mudstone and shale. Coal bed near top of Rockcastle Member. Lower surface (not included in study) similar to fifth-order surface, but overlying deposits are generally coarser than rest of member.	Individual sandstone members are wide channel-belts of fluvial system. Overall fining-up sequences indicate waning of sedimentation period. Pulses are probably tectonically controlled. Possible paleovalley-fill.

Figure 27. Bounding surface hierarchy developed for the Lee Formation.


SANDSTONE MEMBER

giant foresets (Profile LRD-1). In the flow direction, third-order surfaces extend 10 to > 100 m, and can be traced laterally for > 50 m. Centimeter-thick layers of fine-grained, organic- and mica-rich material often demarcate third-order surfaces. Third-order bounding surfaces separate thick cross-bedded sequences into 1-3 m-thick cosets (e.g., Profile C-4 and LRD-1). Compound cross beds are a notable example of the cosets (Profiles C-2 and C-3; Fig. 12a). Third-order surfaces dip primarily in the general paleocurrent direction, indicating downflow accretion (Profiles C-1 and LRD-2). Paleocurrents patterns across third-order surfaces are mainly similar, but markedly different orientations are sometimes found (Profile C-4).

Fourth- and fifth-order bounding surfaces have significant erosional relief, and are overlain by coarse sandstone and/or conglomerates consisting mostly of plant debris, mud chips and quartz pebbles (S_C; Figure 6). However, size and abundance of conglomeratic material is greater in sediments immediately overlying fifth-order surfaces than that found above fourth order surfaces. Little variation in paleocurrent patterns exists across fourth-order surfaces, whereas differences in paleocurrent means across a fifth-order surface may exceed 120 ° (Profile C-1). Fifth-order surfaces rarely cut into fine-grained facies (Profile CF). Discontinuous mudstone beds are sometimes present above fifth-order surfaces (Profiles C-1 and LRD-2). Fourth-order surfaces are usually not associated with fine-grained material but surface Q in Profile LRD-1 is a notable exception. Fourth-order surfaces extend for at least 100 m in all directions. Fifth-order surfaces can be traced more than 1 km across paleoflow (surface U in Profiles C-1 and C-4), but could not be correlated between outcrops in the paleoflow direction. Fourth-order surfaces are not found truncated by fifth-order surfaces, but in Profile C-1 a fifth- and a fourth-order surface are separated by < 30 cm.

Sixth-order surfaces define sandstone member boundaries. The bases of the members, not included in this study, are erosional into underlying strata (see Rice, 1984). Sediments overlying the base of the Rockcastle Member at Cumberland Falls are coarser than those of the rest of the member. A similar relationship was reported by Hester and Taylor (1977, 1981) for the Corbin Member. Members are gradational into the overlying and interfingering Breathitt Formation (Fig. 19; Profiles C-4 and LRD-1).

Architectural Elements

Architectural elements (Allen, 1983; Miall, 1985) in the Lee Formation are reconstructed using bounding surfaces, lithofacies assemblages and paleocurrent information. Depositional processes are interpreted at the element level of investigation.

Major channels are the primary component of the SW-NE trending sandstone members of the Lee Formation. Channel axes are generally aligned subparallel to sandbody orientations. Most other elements recognized in this study comprise deposits within major channel elements. Overbank elements are a relatively minor component of the sandbodies.

Major channel (CH) elements. Rock units between fifth-order bounding surfaces are interpreted as the deposits of system-scale, or major, channels. Although complete, major channel-forms were not found, relatively steep erosional surfaces (Profiles C-1, LRD-2 and CF) attest to channelization. Unidirectional paleocurrent data from strata overlying fifth-order surfaces supports a channel interpretation (Figs. 17 and 19). Channel elements are best seen in Profile CF, where the Rockcastle Member contains deposits of at least four major channels (Fig. 21, stippled areas). A fining-upward trend is apparent within channel fills. Basal conglomerates (S_{cg}) grade upwards into coarse- to fine-grained, cross-bedded sandstones that are capped by fine-grained facies. The sandstones consist of downflow-accretion and subordinate sandy bedform elements; these elements are discussed below. Entire sequences are preserved only at the top of sandstone members (Profiles CF, LRD-1, and C-4); otherwise fine-grained facies are completely or partially eroded by subsequent channels. In Profile CF, lenses of fine-grained facies are located at the top of channel-fill sequences below surfaces D and F.

Depositional processes. Fifth-order surfaces represent scour surfaces generated during formation of major channels. Channels were formed either by erosion due to lowering of base level or by avulsive events. The 10-20 m thicknesses of deposits between fifth-order surfaces in Profile CF represents the minimum amount of channel aggradation during channel-fill cycles. Fining-upward sequences indicate diminishing flow strength during channel aggradation. Basal conglomerates, up to 3 m thick (overlying surface A, Profile CF), indicate the most competent flow. Overlying sandstones reflect decreased competency during channel infill. Paleocurrents are strongly unimodal in the majority of a channel fill sequences, indicating steady, relatively high, flow velocities (*cf.* Banks and Collinson, 1974; Jones, 1977; Bridge, 1985). In the upper few meters of more completely preserved sequences, paleocurrent data vary by as much as 90° from the main flow direction (Figs. 19 and 21). High paleocurrent variance supports reduced flow competency and capacity (*cf.* Smith, 1972, Jones, 1977).

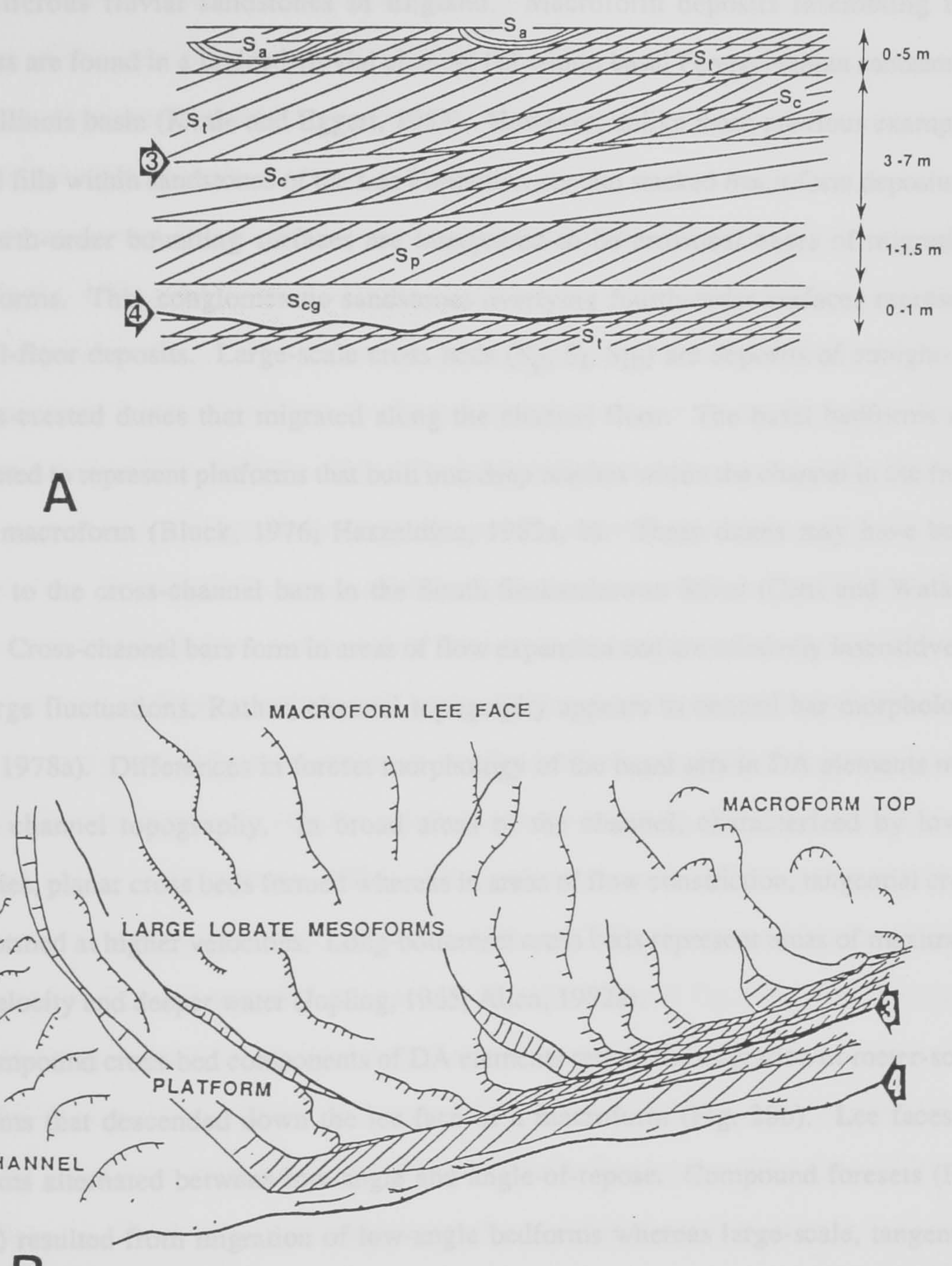
Downflow-accretion (DA) elements. Fourth-order surfaces separate major channel deposits into 5 to 10 m-thick units interpreted as downflow-accretion elements (e.g. Profile C-1 surfaces L and M; Profile CF, surfaces B, C, and E). Major channel elements in the Corbin and Rockcastle Members commonly contain two or more DA elements (Figs. 17 and 21; Profiles C-1 and CF). The widespread presence of DA elements in Profile CF and their presence in all other members indicates that these features dominated the depositional system of the Lee Formation. Two types of DA elements are recognized within channel-fill sandstones.

Type 1 DA elements. Characteristic facies assemblages of downflow-accretion elements consist of, from bottom to top: 1) thin conglomeratic sandstone (S_{cg}); 2) large-scale planar (S_p) or tangential (S_t and S_{lb}) cross beds; and 3) compound cross beds (S_c) that commonly grade in the direction of flow into large-scale tangential cross beds (S_t) (Profiles C-1, C-2, CF and PM-147W; Figs. 14a and 28a). Compound cross-bed components of DA elements within the Corbin Member contain downflow-dipping, third-order bounding surfaces (Profiles C-1, C-2 and C-3; Figs. 12a, b, and 28a). Paleocurrent azimuths from planar-tabular, tangential and compound intraset cross beds as well as from compound foresets and third-order surfaces are all sub-parallel (Figs. 17 and 18). Subordinate lateral accretion is recorded by topset-preserved compound cross beds (S_{tp}); paleocurrent azimuths from intrasets range from parallel to oblique with respect to compound foresets (Fig. 12d). Small- and medium-scale cross beds (S_t and S_a) often cap elements located at the top of channel-fill sequences (Profile C-4). In most elements, an upward decrease in set-thickness is evident. Above the basal conglomerate, no grain-size trends are developed in the downflow accretion (DA) elements.

In the across-flow direction, the extent of DA elements is unknown, but they are probably similar in scale to encompassing major channels. In the direction of flow, DA elements can be traced for more than a few hundred meters without significant changes in paleocurrent direction. In one instance, a DA element is laterally transitional into giant cross beds (S_g; Profile CF).

Depositional processes. Downflow-accretion elements are interpreted as macroform deposits within the major channels (Figure 28b). Recognition of macroforms is based on evidence of long-term lateral, oblique or downstream accretion of smaller bedforms on inclined bedding surfaces (Friend, 1983; Miall, 1988a, b). DA elements are very similar in internal geometry to the mid-channel macroforms described by Haszeldine (1983a, b) in

Figure 28. a) Characteristic sequence of facies above fourth- and fifth-order surfaces in channel-fills of the Lee Formation. b) Reconstructed macroform from sequence in (a). Vertical slice through macroform represents current-parallel outcrop view. Note hierarchy of superimposed bedforms.


Sa
Sa
St
0 -5 m
Sc
St
3
Sc
3 -7 m
Sp
1-1.5 m
Scg
4
0 -1 m
St
A
MACROFORM LEE FACE
MACROFORM TOP
LARGE LOBATE MESOFORMS
3
PLATFORM
4
CHANNEL
B

Carboniferous fluvial sandstones in England. Macroform deposits resembling DA elements are found in a braided-fluvial channel fill within basal Pennsylvanian sandstones of the Illinois basin (Kvale and Eggert, 1988). However, unlike these previous examples channel fills within sandstones of the Lee Formation contain stacked macroform deposits.

Fourth-order bounding surfaces are interpreted to be erosional bases of migrating macroforms. Thin conglomeratic sandstones overlying fourth-order surfaces represent channel-floor deposits. Large-scale cross beds (S_p, S_t, S_{lb}) are deposits of straight- to sinuous-crested dunes that migrated along the channel floor. The basal bedforms are interpreted to represent platforms that built into deep reaches within the channel in the front of the macroform (Bluck, 1976; Haszeldine, 1983a, b). These dunes may have been similar to the cross-channel bars in the South Saskatchewan River (Cant and Walker, 1978). Cross-channel bars form in areas of flow expansion and are relatively insensitive to discharge fluctuations. Rather, channel topography appears to control bar morphology (Cant, 1978a). Differences in foreset morphology of the basal sets in DA elements may reflect channel topography. In broad areas of the channel, characterized by lower velocities, planar cross beds formed whereas in areas of flow constriction, tangential cross beds formed at higher velocities. Long-bottomset cross beds represent areas of maximum flow velocity and deeper water (Jopling, 1965; Allen, 1982a).

Compound cross-bed components of DA elements reflect the migration of meter-scale bedforms that descended down the lee face of a macroform (Fig. 28b). Lee faces of bedforms alternated between low-angle and angle-of-repose. Compound foresets (Fig. 12b, d) resulted from migration of low-angle bedforms whereas large-scale, tangential cross beds developed from migration of angle-of-repose bedforms. Third-order surfaces separate deposits of individual bedforms and represent reactivation surfaces (*sensu* McCabe and Jones, 1977) on the macroform. Intrasets within compound cross-bed sets reflect migration of small-scale 2-D dunes on the meter-scale low-angle bedform. Bedding plane exposure of intrasets indicate that the small-scale dunes were commonly sinuous. Local flow fluctuations are induced by such bedforms (Allen, 1973) and may have produced the numerous reactivation surfaces within the intrasets.

Compound cross strata form when superposed bedforms migrate faster than the host form (Banks, 1973). Bedform superposition may be caused by unsteady flow conditions (Allen and Collinson, 1974), or develop in equilibrium states under which the larger bedforms produce boundary layers in which the small bedforms are locally stable (Rubin and McCulloch, 1980; Bridge, 1982). In the Lee Formation, constant grain size and similar paleocurrent orientations along the transition between simple and compound foresets (e.g. Profile C-3) argue against rapid changes in flow conditions. The downcurrent transition from compound to large-scale tangential cross beds may reflect a steady increase in flow velocities (*cf.* Jones, 1979). Lower velocities and decreased sediment supply would favor migration of smaller bedforms to produce compound cross bedding. Gradual increase in velocity results in smaller bedforms evolving into large-scale bedforms (Allen and Collinson, 1974). Flow velocities were sufficient to cause local overturning of intraset foresets (Profiles C-1, C-2 and C-3). Compound foresets that are oblique to the flow, as recorded by intrasets (Profile C-1 and Fig. 17), may indicate a lobate or strongly three-dimensional shaped crests for some of the meter-scale dunes (*cf.* Jones, 1979; Haszeldine, 1983a). If the area adjacent to the bedform flank was deep, then the entire set of dune cross strata was preserved, including topsets (Fig. 12b; *cf.* Bluck, 1979, Plate 1c). Thicknesses of topset-preserved sets yields an estimate of large scale dune height at 1.5 to 2 m.

Small- and medium-scale cross beds that cap DA elements are interpreted as macroform top deposits. This component is rarely preserved because subsequent macroforms or major channels erosively truncate macroform deposits unless the sequence is located at the top of a sandstone member. High paleocurrent variance and asymmetric cross beds (S_a) reflect low-stage deposits at the top of macroform deposits (*cf.* Haszeldine, 1983b; Kvale and Eggert, 1988)

In the same manner that meter-scale dunes alternated between low-angle and steep lee faces, the lateral transition of DA elements into giant cross beds (Profile CF) suggests that macroform lee faces fluctuated episodically from low-angle (DA elements) to angle-of-repose (giant cross beds). This change from DA elements to giant foresets may have been caused by significant flow fluctuations in the major channel. In the Hawkesbury Sandstone, Australia, a similar downflow transition has been interpreted to represent a falling (Conaghan and Jones, 1975), or rising (Allen, 1982a, p. 499) flood-stage sequence.

The absence of extensive mud drapes and mud-clast breccias above first-, second- or third-order surfaces suggests that flow was continuous and that all components of the macroform moved down the channel concurrently. Rare drapes of organic-and mica-rich, fine-grained material (commonly demarcating third-order surfaces) represent short periods of low discharge. Fine-grained sediment accumulated locally in depressions (sloughs?) on the macroform surface that were subsequently filled by downflow-migrating, large-scale dunes.

Type 2 DA elements. Giant cross beds located above surface Q in the Laurel River Dam outcrop (Profiles LRD-1 and LRD-2) exhibit a style of sedimentation distinct from that of type 1 DA elements. The bounding surface hierarchy developed for type 1 DA elements is applicable to type 2 DA elements, but internal facies associations are different.

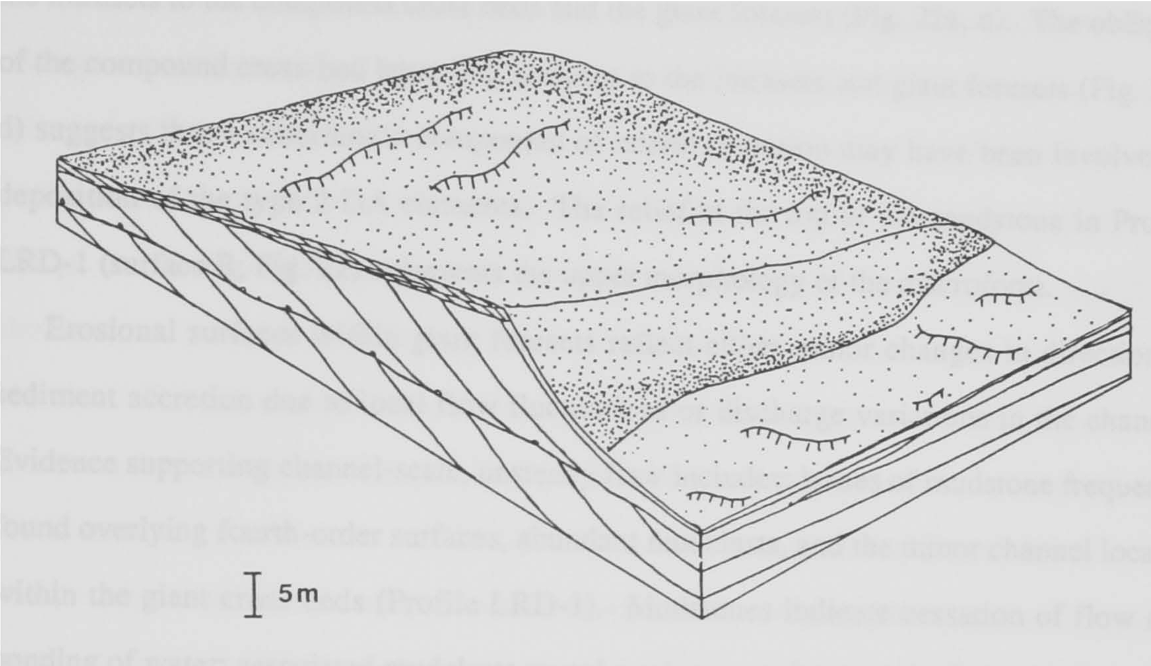
The channel-fill deposits overlying surface Q, a fifth-order bounding surface, consist entirely of type 2 DA elements. Individual DA elements in Profile LRD-2 are bounded by erosional fourth-order bounding surfaces. Erosional surfaces within the giant foresets, nearly parallel to the foresets (best seen in Profile LRD-1), are third-order surfaces. The relief of surface Q is apparent if relative thicknesses of strata containing type 2 DA elements in Profiles LRD-1 and LRD-2 are compared. The upper one third to two thirds of Profile LRD-2 consists of several DA elements whereas a single DA element in Profile LRD-1 overlies an erosional high.

Internally, type 2 DA elements consist primarily of simple giant cross beds, but include rare basal intrasets. The tangential shape of giant foresets and the wedge-shaped sets suggests trough-shaped deposits. The width of these features are estimated to be at least 20 meters, and depositional (non-truncated) elements are estimated to have been 7-8 m high and more than 25 m long. The type 2 DA element located between surfaces Q and R in Profile LRD-1 contains compound cross beds that overlie and are continuous into giant foresets.

Depositional processes. Type 2 DA elements are interpreted to be deposits of 3-D downflow-accreting macroforms. These macroforms migrated along the channel defined by surface Q. Deposits of several macroforms, each consisting of a giant cross-bed set, are found in the deeper part of the channel (Profile LRD-2). Rare intrasets within giant foresets reflect oblique migration of superposed small-scale dunes across the lee face of the macroform probably in response to turbulent eddies resulting from flow separation (*cf.* Allen, 1982a; Fig 29).

The single macroform deposit overlying the erosional high along surface Q in Profile LRD-1 is interpreted as a complete element, preserved because of rapid channel abandonment. Compound cross beds represent stoss deposits of the macroform. Intrasets of compound beds represent small-scale dunes that migrated along the stoss face in the direction of foreset progradation and supplied sediment to the lee face of the macroform

Figure 29. Reconstructed type 2 DA macroform from Profiles LRD-1 and LRD-2. Flow-parallel section represents view in Profile LRD-1.


5 m

(Fig. 29). This interpretation is supported by the parallelism of paleocurrent azimuths for the intrasets to the compound cross beds and the giant foresets (Fig. 22a, d). The obliquity of the compound cross-bed bounding surfaces to the intrasets and giant foresets (Fig. 22a, d) suggests that a subordinate component of lateral accretion may have been involved in deposition of the type 2 DA elements. The relief at the top of the sandstone in Profile LRD-1 (surface R; Fig. 22) represents the upper morphology of the macroform.

Erosional surfaces within giant foresets reflect either minor changes in direction of sediment accretion due to local flow fluctuations or discharge variations in the channel. Evidence supporting channel-scale, unsteady flow includes: lenses of mudstone frequently found overlying fourth-order surfaces, abundant mudclasts, and the minor channel located within the giant cross beds (Profile LRD-1). Mudstones indicate cessation of flow and ponding of water; associated mudclasts reveal a subsequent increase in flow velocity. The channel is interpreted as having been cut into the leeward slope of the macroform during a considerable drop in water level in the major-channel.

Type 2 DA elements, albeit smaller scale, are comparable to giant tangential cross beds described by McCabe (1977) and Casey (1980). These authors have interpreted giant cross beds to be downstream slipface deposits of alternate bank-attached bars (macroforms) in low-sinuosity channels. An alternative interpretation for the giant trough cross beds is that they represent the deposits of dunes. Large-scale 3-D dunes, up to 8 m tall, have been reported in the Yumana (Singh and Kumar, 1974) and Brahmaputra (Coleman, 1969) Rivers, as well as in other deep-water (> 4 m) environments (giant ripples, Reineck and Singh, 1980, p. 44-47). A macroform interpretation is preferred because evidence suggests that the depositional features persisted through significant discharge variations (*cf.* Miall, 1988b).

Lateral-accretion (LA) elements. Profile CF contains features that indicate lateral accretion. Lateral accretion is recognized by low-angle bedding surfaces that dip oblique to orthogonal to the transport direction, as determined from intrasets between the surfaces (above surrface F, Profile CF). The inclined beds are interpreted to be lateral-accretion surfaces on which bedforms migrated in the downflow direction (*cf.* Beutner et al., 1967; Bridge and Diemer, 1983, Fig. 9). Lateral accretion (LA) elements, like type 2 DA elements, are located above the uppermost major erosional surface within the Rockcastle Member. The presence of LA elements at a similar stratigraphic level to the type 2 DA elements, further supports the possibility that the latter developed, in part, in response to lateral accretion.

Minor elements within major channel elements. Other architectural elements found within the major channel elements consist of single facies, each of which has already been described. Inferred depositional processes for these elements are discussed below.

Gravity flow, channel bottom and minor channel elements. Massive sandstone bodies represent rapidly emplaced sand that was not sorted by hydraulic processes during deposition. Recent studies have proposed that massive sandstone bodies may be a product of upper flow-regime (Collinson, 1966, 1970b; Conaghan and Jones, 1975; Conaghan, 1980), subaqueous gravity-flow (Jones and Rust, 1983; Rust and Jones, 1987; Turner and Monro, 1987) and braid-channel (Hodgson, 1978) processes.

Differences in scale, internal features, and external geometries suggest that more than one mechanism was responsible for deposition of massive sandstone bodies in the Lee Formation. The sheet-like, massive element in the Profile C-4 was likely deposited as a **gravity flow** generated by channel-bank failure (*cf.* Rust and Jones, 1987; Turner and Munro, 1987). The transverse orientation of the basal scours with respect to cross-bed azimuths above and below the massive bed (Fig. 20), the faint marginal laminations and the flat top to the massive element support this interpretation. Furthermore, the occurrence of thin cross-bed sets above the gravity-flow deposit suggest that the massive bed aggraded locally in the channel to fill in most available space. Adjacent levee deposits (see below) as well as thinning of the massive body away from the inferred channel bank further support a gravity-flow interpretation. Undulose laminations in the thinner, more distal part of the flow are interpreted as deposits of large-scale bedforms that developed as the mass flow became diluted and approached more normal hydraulic conditions (*cf.* Turner and Munro, 1987). Climbing, large-scale stratification (undulatory laminations) resulted from very high sedimentation rates.

The considerable basal relief (up to 5 meters), thickness (up to 10 m) and lateral extent of the massive, sheet-like sandstone unit in Profile J are not compatible with a gravity-flow origin. Instead, they are similar to units described by Collinson (1970b) and Conaghan and Jones (1975) and are attributed to high-stage, **channel-bottom** deposition. The vertical sequence of facies in this outcrop may reflect a complete flood cycle (*cf.* Collinson, 1966, 1970b; Conaghan and Jones, 1975). Deformed cross beds develop in response to shearing of water saturated sediments by strong currents during rising stage flow when strong channel currents sheared water-saturated sediments (*cf.* Coleman, 1969; Doe and Dott, 1980; McKee, 1989). Abundant deformed cross beds below the massive sandstone thus are considered to record rising stage. As flow velocity increased to peak stage, the channel partly eroded the rising-stage sediments. Subsequent reduction of flow led to rapid deposition from the sediment-laden waters to produce the massive unit. Large-scale internal laminations, possibly representing giant cross beds, may be large-scale bedform deposits (*cf.* Collinson, 1966; Conaghan and Jones, 1975; Conaghan, 1980; Tyler and Ethridge, 1983). Collinson (1966) and Conaghan and Jones (1975) interpreted stratification associated with massive beds to be antidune deposits, but Jones and Rust

(1983, p.1256-1257) concluded that conditions necessary for antidune flow cannot be sustained in deep channels. Therefore, the cross beds within the massive facies are likely a product of low flow-regime bedforms. Large-scale, planar-tabular cross beds (S_p, up to 2.5 m thick) above the massive element represent waning stage deposits. The apparent transverse orientation of the basal scours with respect to the underlying cross beds (Fig. 15b) is perplexing. One possible explanation is that the high-stage channel elements occupy a sinuous thalweg (Bridge, 1985), that may have been oriented oblique to the direction of bedform migration. Conglomerates (S_{cg}) at the base of Profile C-4 are another example of a channel-floor sandstone body oriented at an apparent high angle relative to the migration direction of bedforms that produced overlying cross-beds.

Channel-forms occupied by massive sandstone in the Lee Formation were probably produced by incision into macroform deposits during significant reduction of water level in major channels. Cutting and infilling of this type has been attributed to gravity flows related to bank slumping (Jones and Rust, 1983; Rust and Jones, 1987; Turner and Munro, 1987) or 'normal' hydraulic flow (Hodgson, 1978). Cross-bed sets descending into the channels (Fig. 16) indicate that the secondary or **minor channels** were at least partially filled under conditions of 'normal' hydraulic flow. Because gravity flow scours are eroded and filled penecontemporaneously (see Monro and Turner, 1987), it is likely that 'normal' flow conditions also eroded the channel. The massive nature of the sandstones is ascribed to rapid deposition, without significant sorting. Hodgson (1978) proposed that similar massive channel fills were formed on transverse fluvial bars during falling stage and filled as the (secondary) channel flow was rapidly cut-off. However, flow within a channel rarely contains enough sediment to fill completely by rapid loss of carrying capacity. Instead, sediment probably was supplied from the upstream direction during increasing-flow stage, including slumping of the upstream channel edge, and filled the transverse secondary channel.

The channel forms below surface D in Profile CF are also considered to represent minor channels within a major channel. These features contain recognizable cross beds and thus were filled by migrating bedforms. However, in contrast to the above channel deposits, these are interpreted as low-stage, braid-channel phenomenon.

Sandy-bedform (SB) elements. Many cross-bedded facies cannot be related to deposits of macroforms. These include isolated sets or cosets of planar-tabular (S_p), tangential (S_t), or trough (S_{tr}) cross beds that are attributed to trains of straight-crested 2-D, sinuous-crested 2-D and sinuous crested 3-D dunes, respectively. Planar and tangential cross beds in sandy bedforms, as well as DA elements, are enigmatic. In general, dunes were straight to moderately sinuous and appear to have lacked significant lee-side erosional troughs. Flume experiments suggest that 2-D, lesser erosive, bedforms are relatively low velocity phenomena that occupy only small fields on bedform stability diagrams (e.g., Costello and Southard, 1981 and Harms et al., 1982). Unfortunately, many controls on bedform morphology are not well understood (Harms et al., 1982). In the Lee Formation deposits of 2-D dunes occur at all levels of channel fills and in both coarse- and fine-grained sandstones. Also, in Profile C-4, deposits with more erosive trough-like cross beds are found in the upper parts of channel fills. Similar relationships are found within channel fills in the Seaton Sluice Sandstone, England (Haszeldine, 1983a, b), the Massak Sandstone, Libya (Lorenz, 1987) and the Hawkesbury Sandstone (Rust and Jones, 1987). Depth may be an important control. Coleman (1969, p. 190) noted that irregular crestlines assume a high degree of continuity as the water depth becomes greater in channels of the modern Brahmaputra River.

Sigmoidal cross beds (S_S; Fig. 9) and cross beds with overturned foresets (Fig. 8) are also unrelated to macroform elements. Structures similar to the former have been attributed to high flow-stages (Simons et al., 1965, Kelling, 1969; Turner, 1977; Kohsiek and Terwindt, 1981; Saunderson and Lockett, 1983; Røe, 1987). Formation of overturned cross beds may be related to high-stage flow (Coleman, 1969; McKee, 1989).

Overbank (O) elements. Coarsening-upward units at the top of Profile C-4 (Fig. 19) are interpreted as **levee** deposits. The upward increase in grain size and the vertical and lateral sequence of sedimentary structures (Fig. 10) reflects an increase in bed shear stress related to individual flood events. Preserved bedforms on the top of one coarsening-upward unit implies a sudden decrease in flow. The presence of stacked units with non-erosive bases implies that the levee deposits are aggradational.

Similar deposits are preserved in the proximal floodplain of the modern Brahmaputra River in Bangladesh (Bristow, 1989; Bristow, personal communication, 1989). Flood deposits up to 2 m thick on natural levees occur as sheet-splays. Because the majority of overbank flow is a result of bank overtopping rather than crevassing, the bases of the sheet-splays are non-erosive and sediment transport is parallel to the channel axis (*cf.* Fig. 19). Also, elevated natural levees drain quickly during falling flood stages and deposition rarely takes place during that time. Rust and Jones (1987) have also reported fine-grained facies in the braided-fluvial Hawkesbury Sandstone that are similar to those in the Corbin Member, including preserved bedforms.

In-place, tree-root fossils at the top of the uppermost coarsening-upward unit suggest stabilization of the channel banks. However, because only the top bed of the levee elements are rooted, it is more likely that vegetation became established after channel abandonment. Overlying mudstone and shale were deposited by **vertical accretion** in ponded flood waters of either abandoned channel depressions or extensive overbank areas.

Poor exposure of vertical accretion elements does not allow differentiation of the two environments. The absence of any body fossils and abundance of plant fragments in the overbank elements strongly suggests non-marine deposition. Extensive intervals of fine-grained facies between sand bodies of the Lee Formation, considered tongues of the Breathitt Formation (Chesnut, 1988), are interpreted as deposits of floodplains, distant from the channel system. Although these intervals are largely non-marine, thin marine bands frequently are present between the members of the Lee Formation (Chesnut, 1981).

Channel depth. Depths of major channels can be estimated from the thickness of DA elements and of the complete fining-upward sequence in the Corbin Member (Fig. 19). In the Corbin and Rockcastle Members, the maximum thickness between fourth-order surfaces is consistently between 5 and 10 m. This translates to minimum estimated channel depths of between 5 and 20 m, assuming that the heights of the macroforms were between one-half and the total channel depth at bankfull stage (Bristow, 1987). The thickness of the sequence in the Corbin Member is about 20 m. However, because of the aggradational character of the channel and levee deposits, this estimate must be considered a maximum channel depth. In the Rockcastle Member at Laurel River dam (Profile LRD-1), the vertical distance between the uppermost, fifth-order, erosional surface and the top of the sandstone body is about 8 m. The thickness does not represent a good estimate of channel depth because of the relief exhibited by this surface. The channel deposit above surface F in Profile CF is about 15 m thick. This is likely an accurate estimate for bank-full depth because this channel-fill contains lateral accretion surfaces and is overlain by a gradational contact. Major channel (bank-full) depth is estimated to have been between 10 and 20 m and perhaps closer to 10 m during deposition of the Corbin Member and nearly 20 m during deposition of the Rockcastle Member.

FLUVIAL INTERPRETATION

Evidence from the bedform, macroform, channel and basin scale indicate a fluvial environment of deposition for the sandstone members of the Lee Formation. These criteria include: 1) Closely spaced erosional surfaces in multilateral and multistoried sandbodies that demarcate channels; 2) fining-upward, channel-fill sequences; 3) paleocurrent data demonstrating unidirectional paleocurrents for individual bedforms, macroforms and channel-fill sequences; 4) sedimentary structures indicative of channel scouring and subsequent high- to waning-stage deposition; 5) overbank deposits (natural levees) with paleoflow parallel to that of associated channel deposits; 6) in-place fossil tree roots; and 7) lack of body fossils and paucity of trace fossils in both channel sandstones and fine-grained facies. A fluvial interpretation is further supported by: numerous incised secondary channels of low stage origin; massive sandstone of bank-slump origin; abundant plant material; common overturned cross beds; frequent occurrence of thick conglomeratic units (clasts frequently < 5 cm, but reported up to 30 cm long, Hester and Taylor, 1977) and basin-wide unidirectional paleocurrent patterns (Potter and Siever, 1956). A complete absence of sedimentary features indicative of shallow marine or estuarine environment also favors a fluvial interpretation. Such absent features include:

1) wave-produced laminations;
2) bi-directional cross bedding;
3) flaser bedding or mud-draped foresets;
4) abundant reactivation surfaces;
5) cyclic changes in foreset thickness;
6) small-scale vertical sequences;
7) abundant planar and low-angle laminations;
8) autochthonous fauna;

9) trace fossils; and/or

10) glauconite (Reineck and Singh, 1980; Elliot, 1985; Terwindt, 1988). Reactivation structures are present in the Lee Formation but these can be produced by a range of processes in predominantly unidirectional or predominantly bidirectional flow systems (Collinson, 1970a; Allen, 1973; McCabe and Jones, 1977; Jones and McCabe, 1980; de Mowbray and Visser, 1984). The absence of mud drapes is significant because these are common features in channelized tidal environments, where slackwater periods are well developed due to strongly linear (non-rotational) tidal ellipses (Allen, 1982b; Yang and Nio, 1989).

Internal architecture of the Lee Formation also points to a fluvial origin. Downflow-accreting macroform elements (Fig. 28), the chief component of major channel fills, indicate continuous sedimentation, primarily during high to waning flow stages. Continuous channelized sedimentation at this scale is not compatible with an estuarine or nearshore marine setting (*cf.* Yang and Nio, 1989). In addition, major channels contain two or more macroform deposits with similar paleoflow directions. This supports an interpretation of strongly unimodal flow within the major channels. In nearshore marine or estuarine systems, flow is frequently segregated into separate pathways, each with strongly asymmetrical flows representing ebb and flood currents. In such cases, it is extremely unlikely that one particular paleocurrent orientation will predominate (Nio et al., 1980; Yang and Nio, 1989). Major channels are predominantly oriented sub-parallel to the northeast- to southwest-trending, sandstone belts of the Lee Formation (Chesnut, 1988). This arrangement of major channels, in addition to a unidirectional sediment transport direction (at all scales) along the trend of the sandbodies, further supports a fluvial origin for the Lee Formation.

Tidal Influence on Fluvial Deposition

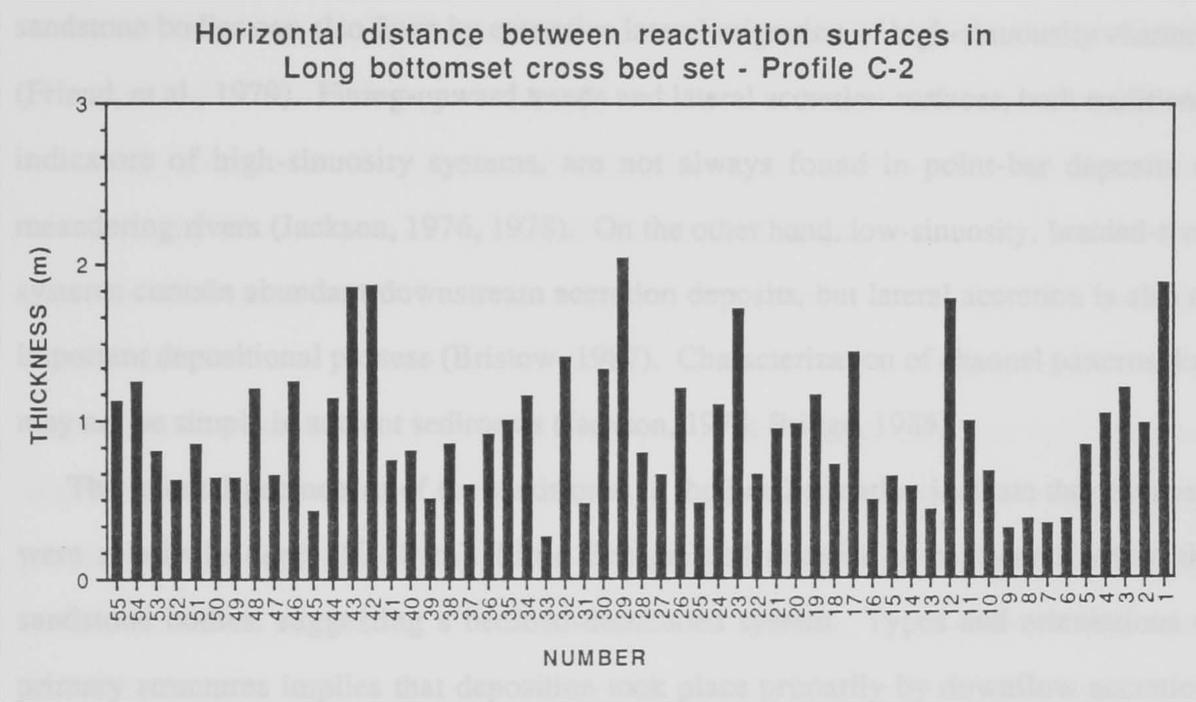
There are several *a priori* reasons that suggest a possible tidal influence on the fluvial system of the Lee Formation. First, the influence of tides on a river is not necessarily restricted to estuarine systems. Tidal currents can extend more than 100 km upstream (Allen and Truilhé, 1989). Another reason is that tidal-estuarine deposits have been documented in the Lower Pennsylvanian strata in the Illinois Basin (Kvale and Eggert, 1988; Kvale & Archer, in press; E. Kvale, personal communication, 1990). In addition, within the central Appalachian Basin, numerous thin marine zones are present throughout lower- and middle-Pennsylvanian, predominantly fluvio-deltaic rocks (Chestnut, 1981, Chesnut and Cobb, 1989). In light of the above, it is not unreasonable to expect some tidal *influence* on Lee Formation fluvial sedimentation, especially below the marine zones between the sandstone members.

Examination of Lee Formation sandstones within eastern Kentucky has not revealed unequivocal evidence of tidal influence. Thicknesses of bundles of foresets between reactivation surfaces in individual sets (facies S_t, S_{lb} and S_{tp}) does not reveal any apparent pattern. In one particular set of long-bottomset cross beds, an irregular pattern intimates periodicity (Fig. 30), but abundant background noise makes a positive interpretation tenuous. Although the pattern in Figure 30 does not reveal strong periodicity, tidal influence is not ruled out. Tidal periodicities in tropical regions can be quite complex (Archer, et al., 1990) and wind, floods, and local hydraulic influences may all act to diminish or mask a tidal influence.

Fluvial Styles

Numerous models have been developed for modern and ancient river systems and suggest a continuum of fluvial styles (Collinson, 1978; Miall, 1985). Channel pattern, specifically sinuosity, is frequently used to characterize end-members in the spectrum of

Figure 30. Thickness between organic- and mica-rich laminations in long bottomset cross-bed set, measured in the flow direction (Corbin Member, Profile 2).


Horizontal distance between reactivation surfaces in
Long bottomset cross bed set - Profile C-2
THICKNESS (m)
NUMBER
0
1
2
3
55 54 53 52 51 50 49 48 47 46 45 44 43 42 41 40 39 38 37 36 35 34 33 32 31 30 29 28 27 26 25 24 23 22 21 20 19 18 17 16 15 14 13 12 11 10 9 8 7 6 5 4 3 2 1

styles. High-sinuosity, or meandering streams often contain abundant fine-grained sediment. An exception to this generalization is the coarse-grained, high-sinuousity system (McGowen and Garner, 1970). Deposits of low-sinuosity, or braided rivers often consist of laterally extensive sand bodies with only minor fine-grained facies. However, sheet-like sandstone bodies can also form by extensive lateral migration of high-sinuousity channels (Friend, et al., 1979). Fining-upward trends and lateral accretion surfaces, both traditional indicators of high-sinuosity systems, are not always found in point-bar deposits of meandering rivers (Jackson, 1976, 1978). On the other hand, low-sinuosity, braided-river systems contain abundant downstream accretion deposits, but lateral accretion is also an important depositional process (Bristow, 1987). Characterization of channel patterns thus may not be simple in ancient sediments (Jackson, 1978; Bridge, 1985).

The internal geometries of the sandstones in the Lee Formation indicate that channels were relatively deep (10-20 m). Little fine-grained material is contained within the sandstone bodies, suggesting a bedload-dominated system. Types and orientations of primary structures implies that deposition took place primarily by downflow accretion. Type 1 DA macroforms were the primary depositional element within major channel fills demonstrating that downstream accretion was superior to lateral accretion in channel-scale, aggradational processes.

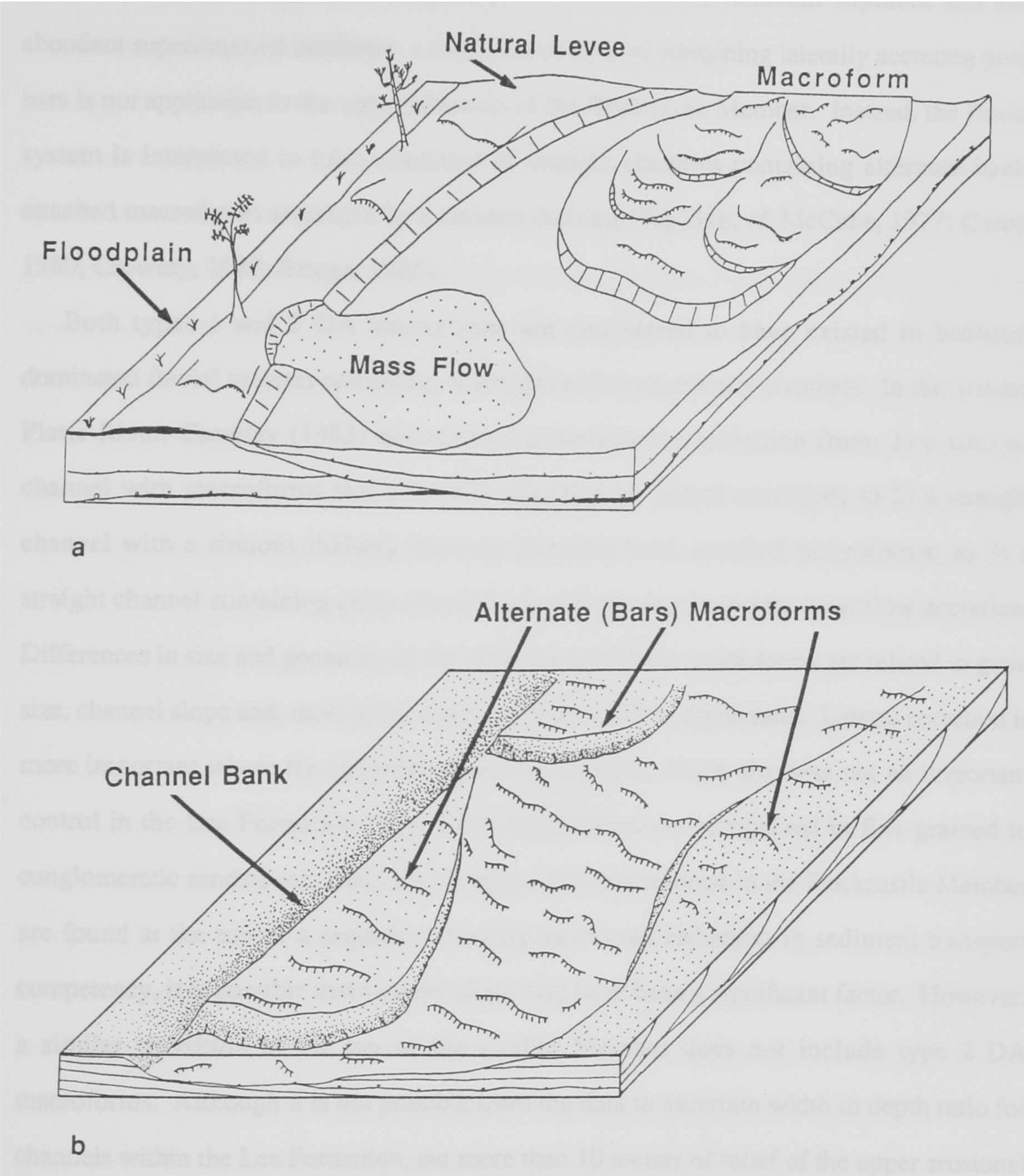
Paleocurrent data for channel fills within members in the Lee Formation indicate a spread in channel orientations roughly from west to east (e.g. Fig. 17). This suggests that: 1) the channels were highly sinuous (*cf.* Plint, 1983; Lorenz, 1987); 2) the macroforms were alternate, bank-attached features within low-sinuosity channels (*cf.* McCabe, 1977; Casey, 1980); or 3) the channels had a low sinuousity and changed orientation as a consequence of avulsion (*cf.* Bluck, 1980; Rust and Jones, 1987). A coarse-grained meandering model is not appropriate because downflow accretion and the characteristics of the facies are incompatible with such a system (e.g. McGowen and Garner, 1970; Jackson, 1976; Nijman and Puigdefabregas, 1978; Campbell and Hendrey, 1987). Alternate bars are also not a likely possibility because macroforms within channel deposits have subparallel paleocurrent trends (e.g. Fig. 17). A systematic variation between bar deposits within individual channel fills would be expected in a system containing alternate bars. Changes in channel orientation during aggradation within an alluvial valley appears the most likely explanation for major (regional) paleocurrent trends. Individual channels do not show significant variation in paleocurrent azimuths over distances of up to 1 km (e.g. unit D, Figs. 17 and 19), suggesting that the majority of the apparent spread in channel orientation is between and not within channels. Also, the uppermost channels within both the Corbin and Rockcastle Members trend more easterly than other channels within the members. When these channels are not considered, the channel orientations are predominantly to the south and west. Overall, evidence points to low-sinuousity channels for the fluvial system of the Lee Formation. Furthermore, reconstructed type 1 DA macroforms (Fig. 29b) are most similar to mid-channel features, as indicated by downflow-dipping cross-bed sets and basal platform components (*cf.* Bluck, 1976; Haszeldine, 1983a, b).

Sedimentation within channel fills containing type 1 DA elements took place during steady, relatively high-stage flow, but very high-stage flow (flood stage?) is suggested by channel bottom elements, overturned cross beds and sigmoidal cross beds. There is little evidence for low-stage reworking (secondary channels), suggesting either a low braiding-index, reworking of low-flow deposits by subsequent high-stage flows, or both. Stable banks are indicated by natural levee deposits, but evidence for channel bank slumping is also found.

Fluvial styles recognized within the Lee Formation do not fit either end-member classification, although a braided origin is more appropriate than a meandering origin. A model that incorporates deep, low-sinuousity channels that were filled by downflow-accreting hierarchical macroforms and not extensively reworked by low-stage flow seems to be appropriate for the majority of the Lee Formation. A summary diagram of the inferred fluvial depositional system for the Lee Formation containing type 1 DA macroforms is shown in Figure 31a. Comparable ancient deposits of deep, low-sinuosity systems include: the Devonian of Spitzbergen (Moody-Stuart, 1966) and eastern Canada (Cant and Walker, 1976); the Carboniferous Rhondda Beds (Kelling, 1969), Seaton Sluice Sandstone (Haszeldine, 1983a, b), 'fluvial' sandstone (Kirk, 1983), Turkey Run channel (Kvale and Eggert, 1988), and South Bar Formation (Rust and Gibling, 1990); the Triassic Hawkesbury Sandstone (Conaghan and Jones, 1975; Rust and Jones, 1987); and the Jurassic Kayenta Formation (Miall, 1988b) and Westwater Canyon (Campbell, 1976; Miall and Turner-Peterson, 1989) and Salt Wash (Tyler and Ethridge, 1983) Members of the Morrison Formation.

The deposits of the uppermost channels in the Rockcastle Member clearly reflect a change in fluvial style in comparison to the rest of the Lee Formation, although limited exposures restrict interpretations. Macroforms (type 2 DA) migrated primarily downstream in a bedload-dominated system, possibly with a component of lateral accretion. In contrast to channel fills containing type 1 DA elements, deposits of superimposed bedforms are not an important part of the uppermost channels. Significant variations in discharge in major-channels are indicated by relatively abundant mudstone and mudclasts and by minor channels of low-stage origin. LA elements in the upper channel in Profile CF suggest some sinuosity to the later channels. Hester and Taylor (1977, 1981) proposed a meandering stream origin for channels within the uppermost Corbin Member of northeast

Figure 31. Diagrammatic models of Lee Formation depositional systems. a) Majority of Lee Formation. b) Upper channels in Rockcastle Member as inferred from Profiles LRD-1 and LRD-2 (after Casey, 1980).


Natural Levee
Macroform
Floodplain
Mass Flow
a
Alternate (Bars) Macroforms
Channel Bank
b

Kentucky. Because type 2 DA macroforms have steep downstream slipfaces and lack abundant superimposed bedforms, a depositional system containing laterally accreting point bars is not applicable to the upper channels of the Rockcastle Member. Instead, the fluvial system is interpreted to have consisted of straight channels containing alternate bank-attached macroforms separated by a sinuous thalweg (Fig. 31b; *cf.* McCabe, 1977; Casey, 1980; Crowley; 1983; Bridge, 1985).

Both types 1 and 2 DA macroforms are considered to have existed in bedload-dominated fluvial systems consisting of straight or low-sinuousity channels. In the braided Platte River, Crowley (1983) documented a downstream transition from: 1) a sinuous channel with macroforms that primarily migrated by lateral accretion; to 2) a straight channel with a sinuous thalweg between alternate bank-attached macroforms; to 3) a straight channel containing cross-channel macroforms dominated by downflow accretion. Differences in size and geometry of the three types of Platte macroforms are related to grain size, channel slope and, most importantly, channel width to depth ratio. Lateral accretion is more important where the channel is more constricted. Grain size was not an important control in the Lee Formation: type 1 DA macroforms are recognized in fine-grained to conglomeratic sandstones. Because the type 2 DA macroforms in the Rockcastle Member are found at the top of a sequence showing an overall reduction in sediment transport competency, it is possible that channel slope may have been a significant factor. However, a similar transition at the top of the Corbin Member does not include type 2 DA macroforms. Although it is not possible from the data to ascertain width to depth ratio for channels within the Lee Formation, the more than 10 meters of relief of the upper erosional surfaces in the Rockcastle Member suggest that downcutting, and hence deep channels, may have been integral for type 2 macroform development. The alternate, bank-attached bars described by McCabe (1977) are also contained within channels that have been cut deeply into underlying sediments.

Analogous Modern Fluvial Systems

The braided South Saskatchewan River in Canada (Cant and Walker, 1978) is a fluvial system comparable, in some respects, to that proposed for the Lee Formation. The river is generally straight with few large bends, and is incised into a kilometer-wide valley with minor floodplain areas. Within the river are two or more channels up to 5 m deep that are separated by shallow to emergent sandflats, commonly dissected by minor channels. Sandflats develop by vertical and downstream accretion of compound dunes on shallow areas of cross-channel, straight-crested 'bars.' Most aggradation takes place during high to flood stage. Lateral migration of the channels is common. Preserved deposits of this type of river contain: channel elements, consisting of trough cross beds; channel-bar elements of large-scale planar-tabular cross beds; sandflat deposits, consisting of primarily complex arrangements of planar-tabular and trough cross beds; and minor floodplain deposits of mudstones and rippled sandstones (Cant, 1978a; Cant and Walker, 1978).

The river planiform, relatively deep channels and downstream accreting macroforms with hierarchical components, are attributes of the South Sakatchewan River that are similar to the Lee Formation fluvial system. A hypothetical stacking of the facies sequences (large-scale architecture), as envisioned by Cant (1978b, Fig. 8), contains a hierarchy of principal erosional surfaces similar to that found in the Lee Formation. However, facies architecture differs in many respects for the two systems. In the South Saskatchewan River, azimuths from bedforms that constitute sandflats are widely divergent, in contrast to the unidirectional nature of the components of type 1 DA elements. Other features dissimilar to the sandstones of the Lee Formation include abundant trough cross beds attributed to channel processes, and deposits resulting from low-stage reworking.

The Brahmaputra River in Bangladesh also is comparable to the fluvial system proposed for the Lee Formation. Bristow (1987) described a three-fold channel hierarchy in a braided reach of the river. The first-order channel comprises the entire river, that encompasses numerous second-order channels that are up to 40 m deep and separate large emergent areas. First- and second-order channels contain lateral, diagonal, medial and tributary bars (macroforms). Third-order channels dissect macroforms at low stage and can also contain bars. Deposition is controlled by the migration of second-order channels (up to 1 km lateral movement per year) and takes place, primarily during high discharges by complex processes of addition to bars, lateral accretion to banks, creation of mid-channel bars and channel abandonment. Extensive silt drapes are deposited during falling stage on the downstream margins of bars.

Large sandwaves within channels have migration rates of up to 500 m per day and develop at high and falling flow-stage (Coleman, 1969). Sandwaves are up to 15 m high and often have superposed dunes. Scour pools in the sandwave trough areas are present only during high stage. Coleman (1969) argues that only during falling stage will there be net aggradation and preservation of internal strata of sandwaves. As pointed out by Leeder (1982), the lee-side angle to the sandwaves of the Brahmaputra River is about 4°, suggesting that internal strata will be dominated by downstream-dipping cosets of cross stratification that formed as dunes migrated down the sandwave lee face. Coleman (1969) recorded downstream accretion rates of large sandwaves of more than 0.5 km per day.

Depositional processes in the Brahmaputra may lead to internal facies assemblages and geometries similar to those found in the Lee Formation. Current directions, obtained from cross beds in trenches at low stage, indicate unidirectional, channel-parallel flow on both medial and lateral bars (Coleman, 1969). Downstream accretion of sandwaves during waning stage flow probably produces deposits similar to type 1 DA elements. Little evidence for lateral accretion to macroforms, similar to that noted by Bristow (1987) in the Brahmaputra River, was found in the Lee Formation. However, it is not clear whether the deposits of the Brahmaputra River reflect channel processes as described by Coleman (1969), bar accretion processes as outlined by Bristow (1987), or both. Bristow (1987) predicts that recent sedimentation in the Brahmaputra River will produce a sheet-like sandbody 40 m thick and 20 km wide. Dimensions are controlled by lateral migration of the river within a channel belt (20 km) and by eventual avulsion of the channel belt to a new location within the floodplain. In contrast, channel-fill deposits between avulsive events within the Lee Formation are probably not more than 20 m thick. Furthermore, macroforms in the Lee Formation do not contain falling-stage drape deposits.

The Orinoco River in South America contains many features that make it an appropriate modern analog for the fluvial system of the Lee Formation. Nordin and Perez-Hernandez (1989) and McKee (1989) outline the depositional elements in the Orinoco River. In the study area the river consists of a sinuous channel that is about 30 m deep at bankfull stage. The river load is mostly well-sorted, sand-sized bedload, with very little suspended sediment. Medial and poorly-developed, bank-attached macroforms are common, but little is known about their migration histories. Bed features are largely preserved during low-stage flow (braiding is not well-developed), when 35% of the river bed is exposed. Continuous slip faces are found on the downflow end of macroforms during low-stage flow. Dunes, 2 to 3 m high, are superposed upon the macroforms. Straight to slightly-sinuous-crested dunes can be traced continuously for more than 1000 meters and only rare, localized, scour pools are found. Sounding records indicate that during high stage, dunes remain laterally continuous (2-D). Dunes adjacent to river banks have smaller, 30-40 cm high, dunes superposed on them.

Deposits of macroforms in the Orinoco River should contain a hierarchy similar to that of type 1 DA elements. Both the scale and morphology of the dunes in the Orinoco River are similar to those interpreted for the Lee Formation. Paleocurrent azimuths for all components of the macroforms within the Orinoco River are inferred to be unimodal. However, because little is known of the dynamics of the macroforms, internal geometries cannot be fully compared. It is likely that downflow accretion is the dominant mode of deposition of the macroforms in the Orinoco River, because channel meandering is negligible (Nordin and Perez-Hernandez, 1989). Minimal reworking of macroforms during low stage flow conforms with the proposed depositional model of the Lee Formation. The thickness of channel deposits, probably less than 20 m, is also similar to those in the Lee Formation.

DISCUSSION: DEPOSITIONAL CONTROLS

The internal architecture of the Lee Formation consists of a hierarchy of stacked facies assemblages and fining-upward sequences ('nested' cycles of Miall, 1980), that are discriminated by hierarchical bounding surfaces. Fourth-, fifth- and sixth-order surfaces are overlain by fining-upward sequences. The trends are more apparent with increased scale of the deposit. Fining-upward sequences of individual members (bound by sixth-order surfaces) are most pronounced and typically culminate in thin bands of marine strata (Chesnut, 1981). A concurrent upward change in fluvial style is apparent within the Rockcastle Member. An additional level of cyclicity is manifested by the coarsening upward beds in the levee deposits (Fig. 19). The repetitiveness or 'cyclicity' exhibited by the deposits of the Lee Formation represent responses of the fluvial system to various mechanisms that controlled sedimentation.

Internal architecture of fluvial deposits is a consequence of autocyclic (internal) and allocyclic (external) controls (Beerbower, 1964). Autocyclic mechanisms include crevassing, meander migration and avulsion whereas allocyclic controls include base level change, tectonics and climate. In deposits of mixed-load, high-sinuousity river system the architecture and controls on deposition are relatively well-understood (e.g. Allen, 1974, 1978, 1979; Bridge and Leeder, 1979; Friend et al., 1979; Allen and Williams, 1982; Behrensmeyer and Tauxe, 1982; Bridge and Diemer, 1983). Relationships between depositional architecture and controls on deposition of bedload-dominated river deposits remains poorly understood. A principal reason for this is the difficulty in recognizing erosional surfaces in relatively homogeneous deposits. The common occurrence of multiple channels (erosional surfaces) in these river systems complicates interpretation. Compounding the problem are effects of external controls on fluvial systems, especially in tectonically active regions. The problem of preservation potential limits the utilization of knowledge gained from modern studies to the analysis of ancient deposits. Interpretation of depositional controls is further hindered because responses of a fluvial system to a change of conditions is often complex (Parker, 1975; Schumm, 1977) and because similar sequences may be produced in different ways (Miall, 1980).

Autocyclic Controls: Fourth- and Fifth-Order Bounding Surfaces

In this study, both fourth- and fifth-order surfaces (beneath macroform and major channel elements, respectively) are interpreted to have been produced by autocyclic processes. The internal architecture within a channel-fill sequence, generally consisting of more than one macroform deposit separated by fourth-order surfaces, can be produced in two ways. The first model involves multiple channels within a larger braided river similar to the Brahmaputra and South Saskatchewan Rivers. The channels migrate laterally within the river belt, and combined with vertical aggradation produce a hierarchical architecture. A second model, after the Orinoco River, contains a single channel with relatively stable banks and mid-channel macroforms. With vertical aggradation the deposits of several macroforms will overlie one another. In both models, avulsion of the river channel to a different location in the alluvial valley is responsible for the formation of fifth-order surfaces.

In the first model, erosive bases to secondary channels are equated with fourth-order surfaces whereas the base of the entire river system is equivalent to a fifth-order surface. Lateral accretion of mid-channel and bank-attached macroforms should dominate the deposits because secondary channels migrate laterally within the river channel (Allen, 1970; Bristow, 1987). Previous architectural models of braided-stream deposits have incorporated laterally migrating secondary channels to produce sheet-like sandstones, elongated in the flow direction (*cf.* Campbell, 1976; Cant, 1978). Several studies have demonstrated that lateral accretion is an important component in the deposits of some braided-stream systems (e.g. Allen, 1983; Miall, 1988b; Miall and Turner-Peterson, 1989). Similarly, Bristow (1987) predicted that recent deposits of the Brahmaputra River will be dominated by second-order channels and their associated bars. An absence of lateral accretion features in this study deters application of this model to the sandstone members of the Lee Formation. However, in channels with high width/depth ratios, lateral-accretion surfaces will be very low angle (Moody-Stuart, 1966; Bristow, 1987). Therefore, lateral accretion components of macroform deposits that result from migrating secondary channels may be difficult to recognize. In addition, paleocurrents in multichannel systems should vary significantly within river belt deposits because secondary channels migrate within channel belts (e.g. 20 km wide for the Brahmaputra River). In the Lee Formation, paleocurrent azimuths of deposits between fifth-order surfaces (equivalent to river belt deposits in the Brahmaputra-Saskatchewan River model) do not vary significantly. Thus, the first model may not be appropriate for the Lee Formation.

The alternative model requires repetitive or semi-repetitive, downflow-migrating macroforms that move primarily during high and waning stages of flow. Rapid reduction in flow would not cause significant modification of the macroforms (Jones, 1977; Nordin and Perez-Hernandez, 1989). Vertical aggradation within the channel would preserve deposits that consist of stacked macroform deposits. Previous workers have described solitary downflow-accreting macroforms within individual channel-fill sequences (Kirk, 1983; Haszeldine, 1983a,b; Kvale and Eggert, 1988). There is no apparent reason that individual, low-sinuosity, single-channel deposits should not contain more than one macroform element. A possible drawback to applying this model to the Lee Formation is the extensive tabular shapes of the sandstone members. In order to construct sandbodies up to 100 km wide and 200 m thick, a single low-sinuosity channel must avulse at a sufficiently rapid rate (with respect to subsidence) to create a deposit with high interconnectedness of channels (*cf.* Allen, 1978, 1979; Bridge and Leeder, 1979). A rough approximation indicates that a 1 km wide channel that aggrades 20 m must avulse 1000 times to create sandbodies similar to those in the Lee Formation with no intercalated flood plain deposits. Chestnut and Cobb (1989) estimated that the period of time available for deposition of sandstone members is about 2.5 million years but, depending upon the actual length of the Pennsylvanian Period, may be as little as 1.4 million years (Klein and Willard, 1989a). Accordingly, the avulsion rate would be once per 1400 to 2500 years. These values lie within the range of channel avulsion rates for all river types, which are on the order of hundreds to thousands of years (Bridge, 1985). Therefore, at least theoretically, this model is applicable.

Interpretation is hindered by the lack of understanding of macroform processes and preservation in large, bedload-dominated fluvial systems and also by scale. The relationship between processes in these systems and the internal geometry of macroform deposits is not understood well enough to interpret the Lee Formation at the scale of outcrop in this investigation. A definitive distinction between a multi-channel or single-channel low-sinuousity fluvial system may require exposures at the scale of the channels (> 1 km) and oriented transverse to paleoflow. Nevertheless, data in this study, viz. paleocurrents and paucity of evidence documenting either lateral accretion or low-stage reworking, support a single-channel interpretation.

An autocyclic rather than allocyclic interpretation for fifth-order surfaces is favored by consideration of fluvial processes in modern systems in conjunction with the scale and internal facies characteristics of sequences above fifth-order surfaces. Although major channels in braided streams are often assumed to migrate laterally at rapid rates, spreading out to form sheet-like bodies, there is little evidence for this except in alluvial fans (Collinson, 1978, Bridge, 1985). Instead, the entire river channel (major channels) moves to another location on the alluvial plain by avulsion (Coleman, 1969; Bridge, 1985; Bristow, 1987; Wells and Dorr, 1987). Aggradation of the river with formation of an alluvial ridge eventually leads to destabilization and subsequent avulsion (Bridge, 1985; Wells and Dorr, 1987). Within a low-sinuosity river channel, sedimentation records a progressive loss of stream competency with a possible widening and shallowing of the channel before abandonment by avulsion (Moody-Stuart, 1966; Miall, 1977).

In the Lee Formation fluvial system, numerous overbank-flood events took place while the channel belt was in one position producing the gradational coarsening-upward sequences within the levee deposits. The overbank deposits are analogous to third-order cycles described by McLean and Jerzykiewicz (1978) and flood-generated sedimentation sequences of Bridge (1984) and indicate that major channels were relatively stable and vertically aggrading between avulsive events. Major channel-fill deposits in the Lee Formation reflect a progressive loss of stream competency. Apparently, abandonment by avulsion was relatively rapid with accumulation of only minor thicknesses of vertical-accretion deposits at the top of channel fills (e.g. Profile CF). Relief along fifth-order surfaces in the Lee Formation, up to 10 meters, is not necessarily indicative of an allocyclic control on the fluvial system; during avulsive events rivers may downcut more than 5 meters into the alluvial plain (Miall, 1977; Tyler and Ethridge, 1983; Hopkins, 1985). Allocyclic processes may have indirectly contributed to the formation of fifth-order surfaces. For example, avulsions may be triggered by external controls such as tectonism (Alexander and Leeder, 1987).

Fifth-order surfaces are produced by avulsions and represent significant time breaks. The fifth-order surface at the lower/upper Rockcastle Member boundary (surface D, Profile CF; surface P, Profiles LRD-1 and LRD-2) is an example of such a time break. The abrupt decrease in grain size across this surface can be explained by the river system shifting by avulsing across a wide alluvial valley before returning to the original site of deposition at a time that the paleoslope was less steep than before. Tabular sandbodies, numerous erosional surfaces attributed to avulsions and rare, complete fining-upward sequences, characteristics of the Lee Formation, also indicate a wide alluvial valley. Rivers within wide valleys avulse more frequently and thus their channel-fill deposits contain fewer complete fining-upward sequences than rivers within narrow valleys (Rust and Gibling, 1990). Other deep, low-sinuousity channel systems with low-braiding indexes (single channel) that migrated by avulsion and were located within relatively wide alluvial valleys have been described by Moody-Stuart (1966), Tyler and Ethridge (1983) and Rust and Gibling (1990).

Allocyclic Controls: Sixth-order Bounding Surfaces

Member-scale depositional cycles in the Lee Formation are interpreted to be a response to allocyclic mechanisms. It is unlikely that internal controls could have produced the distribution and scale of individual members. Gross fining-upward trends and changes in fluvial style such as that within the Rockcastle Member and documented for the Corbin Member in northeastern Kentucky (Hester and Taylor, 1977, 1981) are characteristics of paleovalley fills (Fisk, 1944, 1947; Brown et al., 1973; Sedimentation Seminar, 1978; Blakey and Gubitosa, 1984). Early Pennsylvanian paleovalleys have been identified at several localities in eastern Kentucky (Rice, 1984; Rice and Schwietering, 1988). If individual sandstone members of the Lee Formation represent paleovalley fills, then valleys would have been more than 100 km wide. Fluvial downcutting and creation of paleovalleys followed by aggradational infilling can be caused by climate change (Hall, 1990), sea-level change (Fisk, 1944, 1947; Aubrey, 1989), and/or tectonism (Blakey and Gubitosa, 1984). Each of these controls is now discussed then evaluated with respect to the Lee Formation.

Regional climatic change to an arid environment results in a reduction in plant cover, leading to increased runoff and sediment supply to a fluvial system (Baker, 1978; Hall, 1990). Channel dissection and high sediment (bedload) yield characterize fluvial systems during arid phases. Return to more humid conditions results in decreased runoff with concomitant aggradation of channels.

Downcutting and creation of paleovalleys may also result from a base-level drop related to a regional tectonic uplift or a drop in sea level (*cf.* Schumm, 1977). Infilling of the valley takes place as sea-level rises or the basin subsides. Because the relative effects of tectonism and eustatic change are the same, they may be impossible to discriminate.

Large eustatic sea-level changes may be caused by global-scale tectonic processes, which can change volume of the ocean basin (see Pittman, 1978), or by transfer of water between oceans and land-based ice sheets, commonly thought to be contolled by variations in the Earth's orbit (e.g. Hays et al., 1976). Eustatic sea-level fluctuations are well documented in the geologic past, including the Late Paleozoic (Ross and Ross, 1987).

Tectonic subsidence in foreland basins is primarily related to downward buckling (deflection) of a lithospheric plate by loading of a fold-thrust belt and can be approximated by a thin-plate isostatic flexural model (Beaumont, 1978, 1981; Jordan, 1981). In modeling, the rheology of the plate may be considered as elastic (Jordan, 1981) or as viscoelastic (Beaumont, 1981). In either case flexure of the plate creates a topographical high, or forebulge, at the distal margin. The primary difference between the two models is the response of the plate to loading. An elastic plate remains stable with time until load conditions are changed (e.g. erosion or further loading), whereas a viscoelastic plate relaxes under the applied load to effectively deepen and narrow the basin (Beaumont, 1978). Flemings and Jordan (1989, 1990) quantitatively modeled stratigraphic responses to episodic thrust loading in an elastic flexural model and compared the results to those predicted for a viscoelastic model. In the elastic model, sedimentary cycles are separated by unconformities caused by basinward migration of the forebulge as the basin narrows in response to thrusting. Later subsidence is produced by sediment loading and widening of the basin. In the viscoelastic model, unconformity-bound cycles result from inward migration of the forebulge during relaxation and subsidence related to subsequent thrust loading. Detailed knowledge of stratigraphy, especially with respect to timing of thrusting events, is necessary to distinguish which model is appropriate for foreland-basin evolution.

Substantial debate exists regarding the relative effects of paleoclimate, sea-level changes and tectonics on deposition in the Appalachian basin during the Pennsylvanian. Cecil (1990) proposed cyclic paleoclimatic variations, driven by orbital forcing and associated with sea-level change, as the primary control on siliciclastic sedimentation. Cecil (1990) suggested that the periods of these cycles were on the order of 10^4 - 10^5 years. Busch and

Rollins (1984) and Chesnut and Cobb (1990) proposed that cyclic sedimentation in the Appalachian basin resulted from eustatic sea-level changes of similar period to that proposed by Cecil (1990). In addition, cyclic sedimentation in the Pennsylvanian of the central Appalachian basin has been interpreted as a response to episodic thrust loading of a viscoelastic lithosphere (Beaumont, 1981) during collisional tectonics of the Alleghanian orogeny (Tankard, 1986; Klein and Willard, 1989b).

It is unlikely that climatic changes *alone* could have produced changes of the magnitude observed between sixth-order surfaces in the Lee Formation. Channel trenching due to a change to arid climate conditions as described by Hall (1990) will result in only a few meters of erosion. Instead, a significant relative drop and subsequent rise in base level are necessary to produce paleovalley sequences similar to those in the Lee Formation. Climatic changes coinciding with significant base-level changes may have influenced the fluvial system. For example, arid conditions may accompany eustatic sea-level falls (see Baker, 1978). Alternatively, tectonic uplift may cause both regional uplift (relative drop in base level) and arid conditions due to a rain-shadow effect.

Tectonic processes appear to be the major control on the sandstone members of the Lee Formation. Because the Lee Formation was deposited near the edge of the Appalachian foreland basin (Fig. 2), the tectonic behavior of a forebulge may have been instrumental in the determining stratigraphic sequence. The westward shift of the fluvial system of the Lee Formation through time (Rice, 1984; Chesnut, 1988) suggests that thrust loading caused increased lithospheric flexure and progressive migration of the foreland basin to the west. A similar migration of fluvial systems in the Himalayan foreland basin has been attributed to episodic tectonism (Burbank and Reynolds, 1984). Migration of depositional facies away from the orogen with sequential thrust loading is predicted in both elastic and viscoelastic models (Beaumont, 1981; Jordan, 1981; Quinlan and Beaumont, 1984;

Flemings and Jordan, 1990). Furthermore, the eastward rotation of major channels as indicated by paleocurrent azimuths at the top of the Corbin and Rockcastle Members may reflect a change in paleoslope caused by the onset of basinward forebulge migration (i.e., thrust loading in the elastic lithosphere model or relaxation in the viscoelastic model) in the central Appalachian area.

It is difficult to rule out the influence of eustatic sea-level changes on sedimentation. Sea-level oscillations are well documented in the Upper Pennsylvanian of the midcontinent United States (Heckel, 1986). However, sea-level changes during the Lower Pennsylvanian were less frequent and of lesser magnitude than in the Upper Pennsylvanian (Ross and Ross, 1987). Nevertheless, effects of sea-level change have been documented as an important control on the depositional architecture and processes of fluvial systems up to several hundred kilometers from the sea (Fisk, 1944, 1947; Aubrey, 1989). In turn, it is plausible that some erosional surfaces within sandstone members of the Lee Formation may have been related to eustatic sea-level fluctuations, but available evidence is not sufficient to evaluate this possibility.

CONCLUSIONS

Detailed sedimentological analysis of sandstone members of the Lee Formation utilizing lateral profiling and bounding surface analysis demonstrate that these lower Pennsylvanian quartzose sandstones were deposited in a low-sinuosity, bedload-dominated river system. Regional stratigraphic and paleocurrent azimuth trends support this interpretation. No evidence was recognized for deposition in a nearshore marine environment.

Sandstone members of the Lee Formation consist primarily of amalgamated channel-fill deposits less than 20 m thick. Vertically and horizontally stacked channel units contain a complex hierarchy of erosional surfaces and bedforms that are interpreted as deposits of downstream-accreting macroforms. Channel elements were dominated by macroform deposits. Type 1 downstream-accreting macroform elements are deposits of large-scale bedforms that consisted of a three-fold bedform hierarchy. Small-scale sinuous-crested dunes were frequently superimposed on large-scale, slightly sinuous-crested dunes, which in turn migrated on the surface of the macroforms. Deposition was primarily by downflow accretion. Channels at the top of the Rockcastle Member contain type 2 downstream-accreting macroform elements that indicate deposition primarily by downflow accretion, but with a component of lateral accretion. Sedimentation in these channels is attributed to low-sinuousity fluvial channels with bank-attached, alternate macroforms. The fluvial system underwent relatively large fluctuations in flow stage.

Other depositional elements recognized within channel elements are minor-channel, sandy-bedform, gravity-flow (attributed to bank slumping), and channel-bottom elements. The latter element is contained within a facies sequence that suggests rising- to flood- to waning-stage deposition. In general, deposition was probably during relatively high stage: little evidence of low-stage flow was recognized. Subordinate fine-grained facies found within the sandstones are interpreted as levee and overbank deposits.

Modern, deep, low-sinuosity, bedload-dominated, fluvial systems with either low or high braiding indexes are suitable analogues for the depositional system of the Lee Formation. Large-scale fluvial architectural analysis could not resolve whether deposition was in a multi-channel or a single-channel system. However, lack of evidence for lateral accretion, minor deposits indicating low-stage reworking and paleocurrent data favor the latter. This conclusion suggests that deposits of several macroforms within individual channel fills developed largely by downstream accretion and vertical aggradation. The Orinoco River is a good analogue for sedimentological features found in the Lee Formation, but the long-term behavior of the Orinoco River system is not known.

The position of channels in the alluvial plain was largely controlled by avulsion of the river from fully aggraded channel belts to other areas of the plain. Allocyclic controls (i.e. climate, tectonism and eustatic sea-level changes) on avulsion processes, although likely to have occurred, cannot be ascertained. Allocyclic mechanisms acting on the Lee Formation fluvial system controlled the distribution and scale of individual sandstone members. A general upward decrease of grain-size and change in fluvial style within the Rockcastle and Corbin Members are similar to paleovalley sequences. It is likely that tectonism was the predominant allocyclic mechanism responsible for the sequences, but climactic and eustatic controls may also have been important.

CHAPTER 2: ORIGIN OF FLUVIAL QUARTZOSE SANDSTONES IN A FORELAND BASIN: EVIDENCE FROM THE PENNSYLVANIAN LEE FORMATION, CENTRAL APPALACHIANS

ABSTRACT

The origin of the quartzose sandstones of the Pennsylvanian Lee Formation in the central Appalachian foreland basin was determined through the analysis of stratigraphic, sedimentologic and petrographic data. These data indicate that the sandstones were deposited in a bedload-dominated fluvial system, with source area and climate functioning as primary controls on composition. Source areas were composed primarily of quartz-rich sedimentary rocks and were located chiefly to the northeast/north. A east/southeast source area supplied subordinate and low-grade metamorphic rock fragments. Intense weathering, associated with humid tropical climates, acted upon the detritus throughout the sedimentation cycle. Less important controls on composition were tectonics and transport/depositional processes that extended exposure of the sediments to the severe climatic conditions. Quartzose sandstones of the Lee Formation reflect lower rates of tectonic subsidence and greater recycling of sand-sized grains during transportation and temporary deposition on the alluvial plain, relative to lithic time equivalents to the east.

INTRODUCTION

Until recently, quartz-rich sandstones were considered by most authors to be polycyclic, deposited by eolian and/or shallow-marine processes in stable tectonic settings (for a thorough discussion see Chandler, 1988). However, numerous studies have demonstrated that not all quartz-rich sandstones are polycyclic; first-cycle quartzose sands are recognized in Holocene river systems (Potter, 1978; Franzinelli and Potter, 1983; Johnsson *et al.*, 1988). Sandstone composition is determined not only by tectonic setting (e.g., Graham *et al.*, 1976; Dickinson and Suczek, 1979) and depositional environment (e.g., Davies and Ethridge, 1975; Mack, 1978), but can also be strongly influenced by climate (e.g., Mann and Cavaroc, 1973; Basu, 1976; Johnsson *et al.*, 1988), source-area composition (e.g., Blatt, 1967; Mack, 1981; Dickinson and Suczek, 1979) and diagenesis (e.g., McBride, 1985, 1987).

Quartzose sandstones of the lower to middle Pennsylvanian Lee Formation in the central Appalachian foreland basin have been previously interpreted as barrier-island deposits based primarily on lithology as well as on stratigraphic relationships (Ferm *et al.*, 1971; Miller, 1974; Horne *et al.*, 1978; Ferm and Horne, 1979). However, several studies have challenged the stratigraphic framework utilized in the barrier model (Ettensohn, 1980; Rice, 1984; Chesnut, 1988). Detailed sedimentological analysis of the Lee Formation in southeastern Kentucky and northern Tennessee demonstrated deposition in a fluvial environment (chapter 1). Because quartzose sandstones are not traditionally developed in fluvial environments or foreland basins, the Lee Formation sandstones reflect unusual conditions.

In this article, the origin of the quartzose sandstones of the Lee Formation is evaluated by analyzing petrographic, sedimentologic, and paleogeographic information. From these data, the various controls on sandstone composition are assessed with respect to the Lee Formation. Finally, the depositional/tectonic setting of the Lee Formation is compared to possible modern and ancient analogs and a model for the genesis of the quartzose sandstones is proposed.

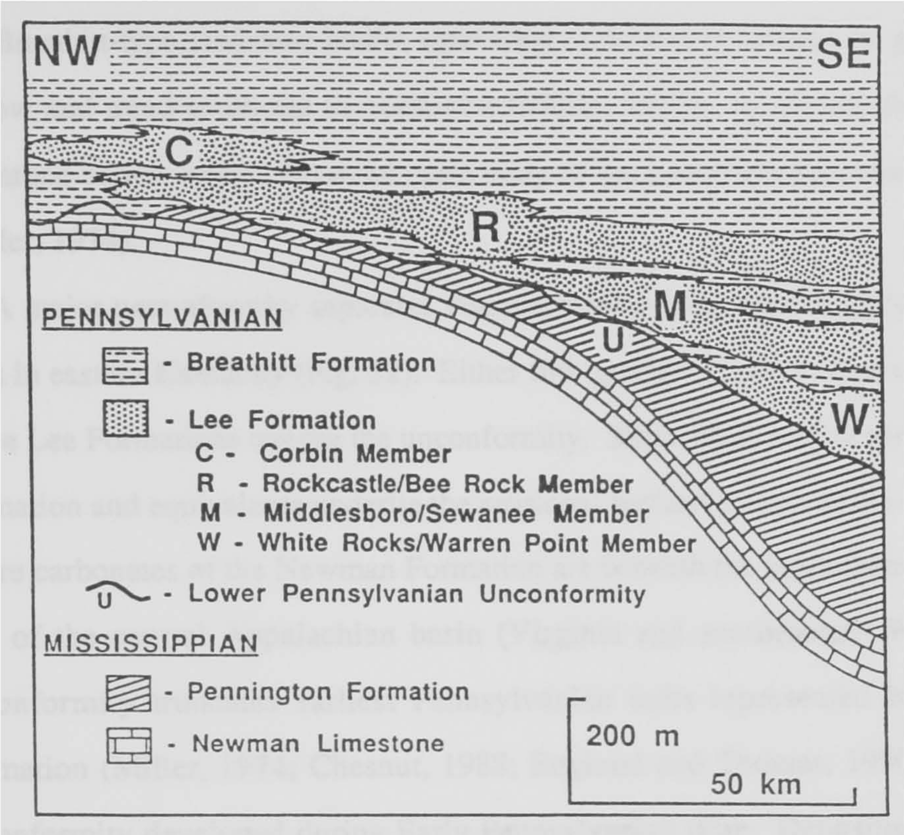
LEE FORMATION IN THE CENTRAL APPALACHIAN BASIN

Stratigraphy

Basal Pennsylvanian stratigraphic units in the central Appalachian basin are composed primarily of quartzose, commonly conglomeratic, sandstones and intervening mudstones and shales. In eastern Kentucky the sandstones are known as the Lee Formation and consist of several extensive bodies that are stratigraphically located between older, shallow-marine strata (Newman Limestone, Pennington Formation and equivalents) and younger, coal-bearing, fluvio-deltaic sediments (Breathitt Formation and equivalents) (Fig. 32; Rice, 1984; Chesnut, 1988). Individual sandstone bodies (members) of the Lee Formation trend parallel to the basin axis as elongate belts (up to 100 km wide and 200 m thick and several hundred km long) arranged in a stair-step manner, with younger members to the west (Fig. 32). Sediments of the Breathitt Formation that overlie older members of the Lee Formation in the eastern part of the basin are laterally equivalent to younger sandstone members to the west (Fig. 32; Rice, 1984, 1986; Chesnut, 1988). The percentage of sandstone in Early Pennsylvanian sequences varies within the central Appalachian basin with the highest percentages of sandstone (more than 80%) corresponding to areas where the interval is thin (Wanless, 1975). In general, Pennsylvanian rocks in the central Appalachian basin thicken towards the southeast, where they are truncated by the present day erosional surface (Englund, 1979; Rice 1986).

The Lee Formation is a lithostratigraphic unit and is distinguished from other units of similar age by its comparatively *clean* composition. Lateral equivalents of the Lee Formation to the east consist of fine-grained sediments, primarily shale and siltstone with lesser amounts of sandstone and coal (Miller, 1974; Chesnut, 1988). The sandstones are *dirtier* than those of the Lee Formation, generally containing more feldspar, lithic fragments, mica and clay matrix (Miller, 1974; Englund, 1979). Because of the lithologic similarity of the eastern lateral equivalents of the Lee Formation to the overlying Breathitt Formation and the stratigraphic relationships mentioned above, these units are commonly grouped together as 'Breathitt Formation' (Fig. 32; Rice, 1984, 1986; Chesnut, 1988). Lee-type and Breathitt-type sandstones contain *ca.* 90-95% and 70-80% quartz,

Figure 32. Schematic diagram of stratigraphy of the Lee Formation. See Figure 2-4 for location of cross section. After Rice (1984) and Chesnut (1988).


NW
SE
C
R
M
U
W
PENNSYLVANIAN
- Breathitt Formation
- Lee Formation
C - Corbin Member
R - Rockcastle/Bee Rock Member
M - Middlesboro/Sewanee Member
W - White Rocks/Warren Point Member
U - Lower Pennsylvanian Unconformity
MISSISSIPPIAN
- Pennington Formation
- Newman Limestone
200 m
50 km

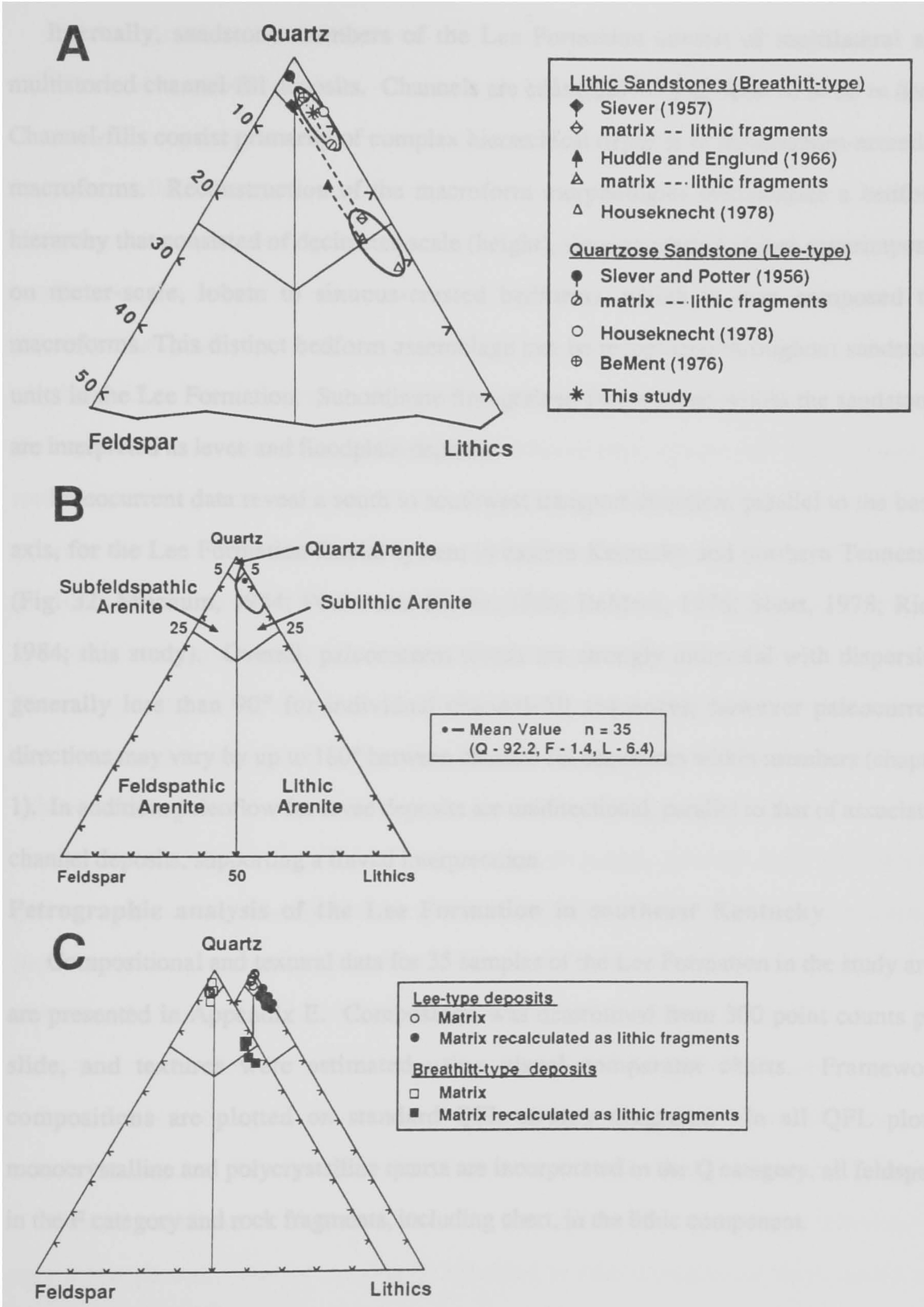
respectively (Fig. 33a). Subsurface lithofacies mapping in southwestern Virginia indicates that Breathitt-type sandstone bodies equivalent to the Lee Formation are generally relatively narrow and trend southeast to northwest (Miller, 1974). In several locations, quartzose sandstone bodies extend from the main body of the Lee Formation towards the southeast (Miller, 1974).

A major unconformity separates Pennsylvanian strata from underlying Mississippian units in eastern Kentucky (Fig. 32). Either fine-grained Breathitt-type units or sandstones of the Lee Formations overlie the unconformity. Siliciclastic sediments of the Pennington Formation and equivalents underlie the erosional surface, except in the extreme northwest where carbonates of the Newman Formation are beneath the unconformity. In the eastern part of the central Appalachian basin (Virginia and southeastern West Virginia) the unconformity truncates earliest Pennsylvanian units represented by the Pocahontas Formation (Miller, 1974; Chesnut, 1988; Englund and Thomas, 1990). Therefore, the unconformity developed during Early Pennsylvanian time. Deposition appears to have been continuous across the Mississippian-Pennsylvanian systemic boundary in the eastern part of the basin (Englund *et al.*, 1986).

Fluvial interpretation for the Lee Formation

Because of the stratigraphic position and mature composition of the Lee Formation and the lack of marine fossils, there has been significant controversy regarding the origin of the quartzose sandstones. Detailed sedimentological facies analyses of the Lee Formation in southeastern Kentucky and Pine Mountain area in Tennessee support deposition of the lower Pennsylvanian quartzose sandstones in a bedload-dominated river system (chapter 1). This interpretation is supported by the presence of abundant plant fossils and the lack of sedimentary features indicative of deposition in a nearshore marine environment.

Figure 33. Ternary composition (QFL) diagrams (classification after Pettijohn et al., 1972). a) Mean values of both Lee-type and Breathitt-type sandstones in the central Appalachian basin. Initial examination of data reveals significant variation in the composition of the sandstones, attributed to the relative amounts of matrix and lithic fragments (pseudomatrix; Fig. 2-4c) recognized in each study. Matrix-rich samples of Siever and Potter (1956), Siever (1957), and Huddle and Englund (1966) was recalculated, as matrix was originally lithic fragments. The recalculated data are comparable in composition to the matrix-poor samples of Houseknecht (1978) and this study. Dashed lines connect end members of matrix and lithic fragment spectrum. Outlined areas represent 'typical' Lee-type and Breathitt-type sandstones. b) Lee Formation as determined in this study (n = 35). c) Average composition of Lower Pennsylvanian sandstones for Appalachian, Michigan and Illinois basins. Data from Siever (1957), plotted as matrix-rich (right) and as pseudomatrix (left).


A
Quartz
10
20
30
40
50
Feldspar
Lithics
Lithic Sandstones (Breathitt-type)
Siever (1957)
matrix -- lithic fragments
Huddle and Englund (1966)
matrix ---lithic fragments
Houseknecht (1978)
Quartzose Sandstone (Lee-type)
Siever and Potter (1956)
matrix ---lithic fragments
Houseknecht (1978)
BeMent (1976)
This study
B
Quartz
Quartz Arenite
5
5
Subfeldspathic Arenite
Sublithic Arenite
25
25
Mean Value n = 35
(Q - 92.2, F - 1.4, L - 6.4)
Feldspathic Arenite
Lithic Arenite
Feldspar
50
Lithics
C
Quartz
Lee-type deposits
Matrix
Matrix recalculated as lithic fragments
Breathitt-type deposits
Matrix
Matrix recalculated as lithic fragments
Feldspar
Lithics

Internally, sandstone members of the Lee Formation consist of multilateral and multistoried channel-fill deposits. Channels are estimated to have been 10 to 20 m deep. Channel-fills consist primarily of complex hierarchical deposits of downstream-accreting macroforms. Reconstruction of the macroform morphologies demonstrate a bedform hierarchy that consisted of decimeter-scale (height), sinuous-crested, dunes superimposed on meter-scale, lobate to sinuous-crested bedforms, which in turn composed the macroforms. This distinct bedform assemblage can be recognized throughout sandstone units in the Lee Formation. Subordinate fine-grained facies found within the sandstones are interpreted as levee and floodplain deposits.

Paleocurrent data reveal a south to southwest transport direction, parallel to the basin axis, for the Lee Formation fluvial system in eastern Kentucky and northern Tennessee (Fig. 32; Mitchum, 1954; Potter and Siever, 1956; BeMent, 1976; Short, 1978; Rice, 1984; this study). Overall, paleocurrent trends are strongly unimodal with dispersion generally less than 90° for individual channel-fill sequences; however paleocurrent directions may vary by up to 180° between channel-fill sequences within members (chapter 1). In addition, paleoflow for levee deposits are unidirectional, parallel to that of associated channel deposits, supporting a fluvial interpretation.

Petrographic analysis of the Lee Formation in southeast Kentucky

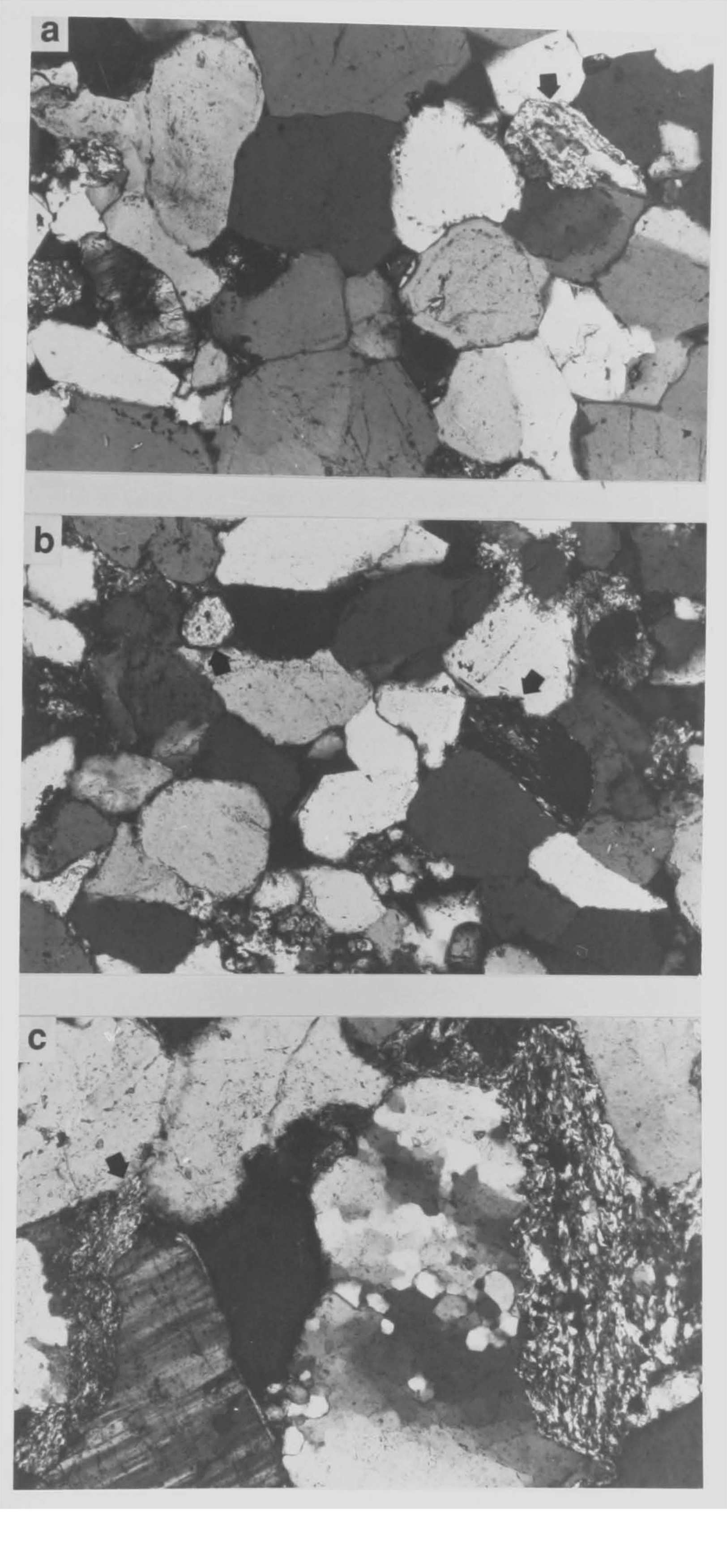
Compositional and textural data for 35 samples of the Lee Formation in the study area are presented in Appendix E. Composition was determined from 300 point counts per slide, and textures were estimated using visual comparator charts. Framework compositions are plotted on standard QFL ternary diagrams. In all QFL plots, monocrystalline and polycrystalline quartz are incorporated in the Q category, all feldspars in the F category and rock fragments, including chert, in the lithic component.

Sandstones of the Lee Formation, utilizing a modified version of the Pettijohn *et al.* (1972) classification, are sublitharenites to quartz arenites (Fig. 33b). The dominant framework component is monocrystalline quartz with subordinate polycrystalline quartz (monocrystalline to polycrystalline ratio is about 10-20 to 1), rock (lithic) fragments, feldspars, micas and heavy minerals. Polycrystalline quartz commonly consists of sub-parallel elongate subgrains, often with sutured sub-grain boundaries. Lithic fragments are primarily low-grade metamorphics including (muscovite- or chlorite-quartz) schists and phyllites (Fig. 34a, b). Many fine-grained micaceous rock fragments could not be distinguished as either shale or low-grade metamorphic grains and are counted as metamorphic fragments. Most of the fine-grained fragments are foliated. Minor amounts of sedimentary rock fragments include predominantly chert, and rarely, very fine-grained sandstone, mudstone and carbonate clasts. Feldspars are chiefly orthoclase and microcline with only minor plagioclase. Feldspar grains are both unaltered and heavily seritcized. Potassium feldspars are albitized. Muscovite is the dominant mica; only traces of biotite and chlorite are found. Heavy minerals consist almost exclusively of zircon and subordinate tourmaline.

Grain sizes of the samples varies from fine to coarse grained, most are medium grained. Generally the sandstones are moderate- to well-sorted, but some samples (commonly conglomeratic) are very poorly sorted. Overall roundness was difficult to determine because of quartz overgrowths and pressure solution suturing of grain edges. Rare dust rims define well-rounded grains (Fig. 34a). In general, the sandstones are estimated to be sub-angular to rounded. Both rock fragments and feldspars are commonly rounded or angular. Zircon and tourmaline grains are frequently well rounded (Fig. 34b).

Detrital matrix material is a minor part of the sandstones, comprising a maximum of only a few percent. Matrix consists of silt-sized to very fine-grained mica, quartz and

Figure 34. Photomicrographs of Lee Formation thin sections (all with cross-nicols and field of view about 1.5 mm). a) Sublithic arenite from Corbin Member. Note metamorphic rock fragment (arrow, quartz and muscovite schist) and authigenic overgrowths on quartz grains. b) Sub-lithic arenite from Corbin Member. Note rounded zircon (left arrow) and hydrocarbon residue in metamorphic rock fragment (right arrow). c) Quartz arenite from Rockcastle Member. Grains include polycrystalline (center) and monocrystalline (upper left) quartz, potassium feldspar (lower left) and metamorphic/sedimentary rock fragments. Note apparent embayments in quartz grains, secondary porosity (left center) and plastically deformed rock fragment (arrow).


a
b
c

feldspar. Extensive compaction of micaceous lithic fragments to produce pseudomatrix is common in many samples (Fig. 34c). Lithic grains were distinguished from pseudomatrix according to methods outlined by Dickinson (1970), the most important of which is similarity to non-compacted lithic grains.

Cements consist primarily of quartz overgrowths on quartz grains, commonly forming an interlocking texture. Minor authigenic clays include illite and kaolinite (rarely replacing feldspars), rarely chlorite and (possibly) smectite. Calcite cement is rare, but many samples contain secondary porosity chiefly as dissolution of cements, probably calcite. In all samples there is minor, partial leaching of feldspars and lithic fragments. Several samples contain petroleum hydrocarbon residues (Appendix B), principally in secondary porosity within rock fragments (Fig. 34b).

White well-rounded pebbles, commonly referred to as *vein* quartz, are common in the sandstones of the Lee Formation. Pebbles are generally 0.5 to 2 cm in diameter. In eastern Kentucky, pebbles consist of nearly equal amounts of monocrystalline and polycrystalline quartz and 5 to 10 % sedimentary rock fragments, that are predominantly sandstone (BeMent, 1976).

Provenance

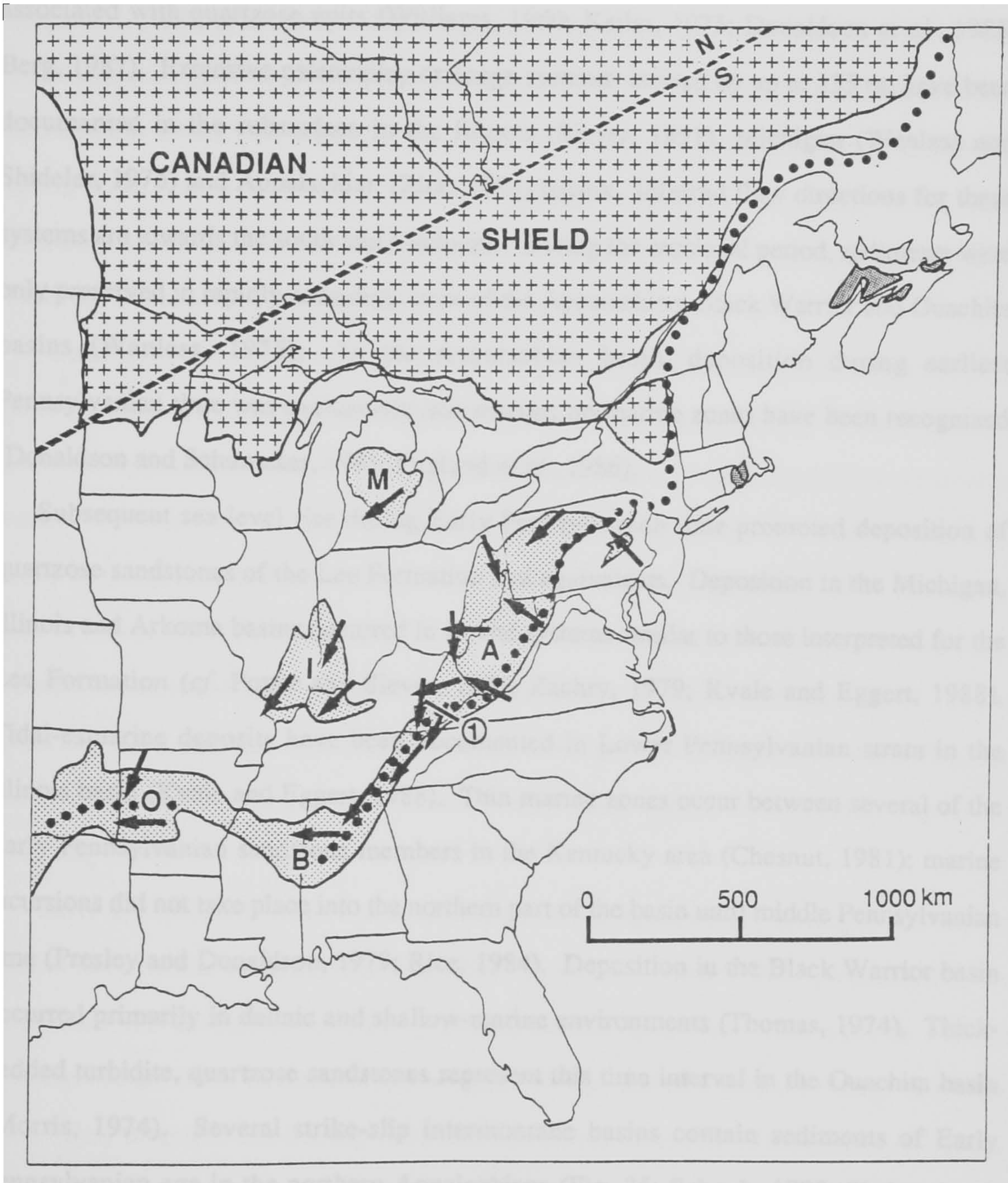
Inferred source areas for the Lee Formation are to the southeast and east (e.g., Ferm *et al.*, 1971; Miller, 1974; Ferm and Horne, 1979; Thomas, 1979; Hatcher et al, 1989), to the north (Mitchum, 1954; Rice, 1984; Rice and Schwietering, 1988), or a combination of both (Siever and Potter, 1956; Donaldson *et al.*, 1985; Chesnut, 1988).

In this study, the sediment transport direction for the Lee Formation in southeast Kentucky was from northeast to southwest, as determined by paleocurrents in the context of a fluvial depositional environment. Sandstones contain both metamorphic and sedimentary fragments. The sedimentary rock source is inferred primarily from rock fragments; Seiver and Potter (1956) also recognized 'second-cycle quartz grains' in the Lee Formation. The source rocks for the metamorphic fragments were probably schists, phyllites and slates. Because some sandstones contain up to 15% labile lithic fragments, which do not survive extensive transport (*cf.* Cameron and Blatt, 1971; Abbott and Peterson, 1978; Mack, 1981), a local (southeast/east?) source is indicated. Local westerly oriented paleocurrent directions in the Lee Formation (e.g., Rice, 1984) and the northwest to southeast trend of quartzose and Breathitt-type sandstone bodies that are eastern equivalents of the Lee Formation (Miller, 1974), further support a nearby Appalachian source. However, the easterly equivalent sandstones in Virginia rarely contain quartz pebbles (Mitchum, 1954; Miller, 1974; Rice, 1984). Thus, it is unlikely that the Lee Formation was derived entirely from the southeast/east. A decrease in the size and number of quartz pebbles from north to south within the sandstone members of the Lee Formation (Englund, 1964; BeMent, 1976; Rice, 1984) supports a northern source area. Therefore, it is likely that the northeast to southwest-trending sandstone bodies of the Lee Formation in the study area were derived primarily from a northern source area with minor influx of sediment from the southeast/east.

SOURCE AND DISPERSAL OF EARLY PENNSYLVANIAN QUARTZOSE SANDSTONES IN THE EASTERN UNITED STATES

Quartzose sandstones are common constituents of Early Pennsylvanian sediments throughout the eastern United States (Fig. 35; Siever and Potter, 1956; Meckel, 1967; Morris, 1974; Mack, 1982; this study). Like the Lee Formation, these units commonly overlie the extensive Early Pennsylvanian unconformity, which is attributed to a major drop in relative sea-level (Pryor and Sable, 1974; Donaldson and Shumaker, 1981). Tectonic tilting and uplift of eastern North America and creation of the unconformity was likely related to major thrusting in the southern Appalachians and the Ouachitas (Beaumont *et al.*,

Figure 35. Map showing locations of Lower Pennsylvanian sediments, primarily quartzose sandstones, in eastern North America and cross-section in Figure 1 (1). Major structural basins are: **A** - Appalachian; **M** - Michigan; **I** - Illinois; **B** - Black Warrior; and **O** - Ouachita/Arkoma. Arrows indicate paleocurrent means in basins; data from Potter and Siever (1956), Schlee (1963), Meckel (1967), Morris (1974), BeMent (1976), Short (1978), Presley and Donaldson (1979), Zachry (1979), Houseknecht (1980), Rice (1984) and this study (chapter 1). Dashed line denotes position of equator during early Pennsylvanian period (Scotese and McKerrow, 1990). Also included are strike-slip intermontane basins (dark stipple pattern), western (northern) margin of Appalachian (Ouachita) deformation (dotted lines), outcrop of Precambrian Canadian shield (pattern).


CANADIAN
SHIELD
N
S
M
I
A
O
B
1
0
500
1000 km

1987). Widespread erosion, and intense weathering of the exposed surface is manifested by karst topography on carbonate bedrock, and by soils and high-aluminum clays associated with quartzose units (Williams, 1960; Keller, 1975; Donaldson *et al.*, 1985; Berg, 1987). Extensive paleovalley drainage systems, incised by up to 135 m, have been documented in the subsurface in the Illinois (Siever, 1951), Michigan (Wanless and Shideler, 1975) and Appalachian (Rice, 1984) basins. Inferred flow directions for these systems are towards the south and southwest. During the erosional period, sediments were only preserved in rapidly subsiding areas of the Appalachian, Black Warrior and Ouachita basins (Wanless, 1975). In the Appalachian basin, deposition during earliest Pennsylvanian time was exclusively non-marine; no marine zones have been recognized (Donaldson and Schumaker, 1981; Englund *et al.*, 1986).

Subsequent sea-level rise during Early Pennsylvanian time promoted deposition of quartzose sandstones of the Lee Formation and equivalents. Deposition in the Michigan, Illinois and Arkoma basins occurred in fluvial systems similar to those interpreted for the Lee Formation (*cf.* Potter and Siever, 1956; Zachry, 1979; Kvale and Eggert, 1988). Tidal-estuarine deposits have been documented in Lower Pennsylvanian strata in the Illinois basin (Kvale and Eggert, 1988). Thin marine zones occur between several of the Early Pennsylvanian sandstone members in the Kentucky area (Chesnut, 1981); marine incursions did not take place into the northern part of the basin until middle Pennsylvanian time (Presley and Donaldson, 1979; Rice, 1984). Deposition in the Black Warrior basin occurred primarily in deltaic and shallow-marine environments (Thomas, 1974). Thick-bedded turbidite, quartzose sandstones represent this time interval in the Ouachita basin (Morris, 1974). Several strike-slip intermontane basins contain sediments of Early Pennsylvanian age in the northern Appalachians (Fig. 35; Schenk, 1978; Shekan et al, 1986), indicating active tectonism in that area.

A well-defined southerly paleoslope for most of the eastern United States during the Early Pennsylvanian is indicated by regional paleocurrent data for the quartzose sandstones (Fig. 35). The paleoslope was inherited from a similar drainage system of Late Mississippian age or older (Potter and Pryor, 1961; Pryor and Sable, 1974; Donaldson and Schumaker, 1981). Regional paleocurrent trends indicate a longitudinal, axis-parallel, filling of the central Appalachian basin as suggested by Graham *et al.* (1975). In the Black Warrior and Ouachita basins a similar axial drainage pattern was developed, although the basin fill is more complicated (Thomas, 1979).

Continued rise in sea-level and increased siliciclastic sediment supply from the Ouachita and Appalachian highlands characterize the Middle Pennsylvanian epoch. In the central and southern Appalachian, Illinois, Michigan and Black Warrior basins, deposition occurred in extensive fluvio-deltaic systems, transverse to orogenic highlands, with marine influences more apparent than in Early Pennsylvanian sediments (Pryor and Sable, 1974; Donaldson and Schumaker, 1981; Hatcher *et al.*, 1989). These marine and terrestrial sediments formed consistent, repetitive deposits, or cyclothems (see Klein and Willard, 1989). In the Ouachita basin, turbidite deposition continued, and high rates of deposition nearly filled the basin; southward-building deltaic systems characterized Middle Pennsylvanian deposition (Morris, 1974).

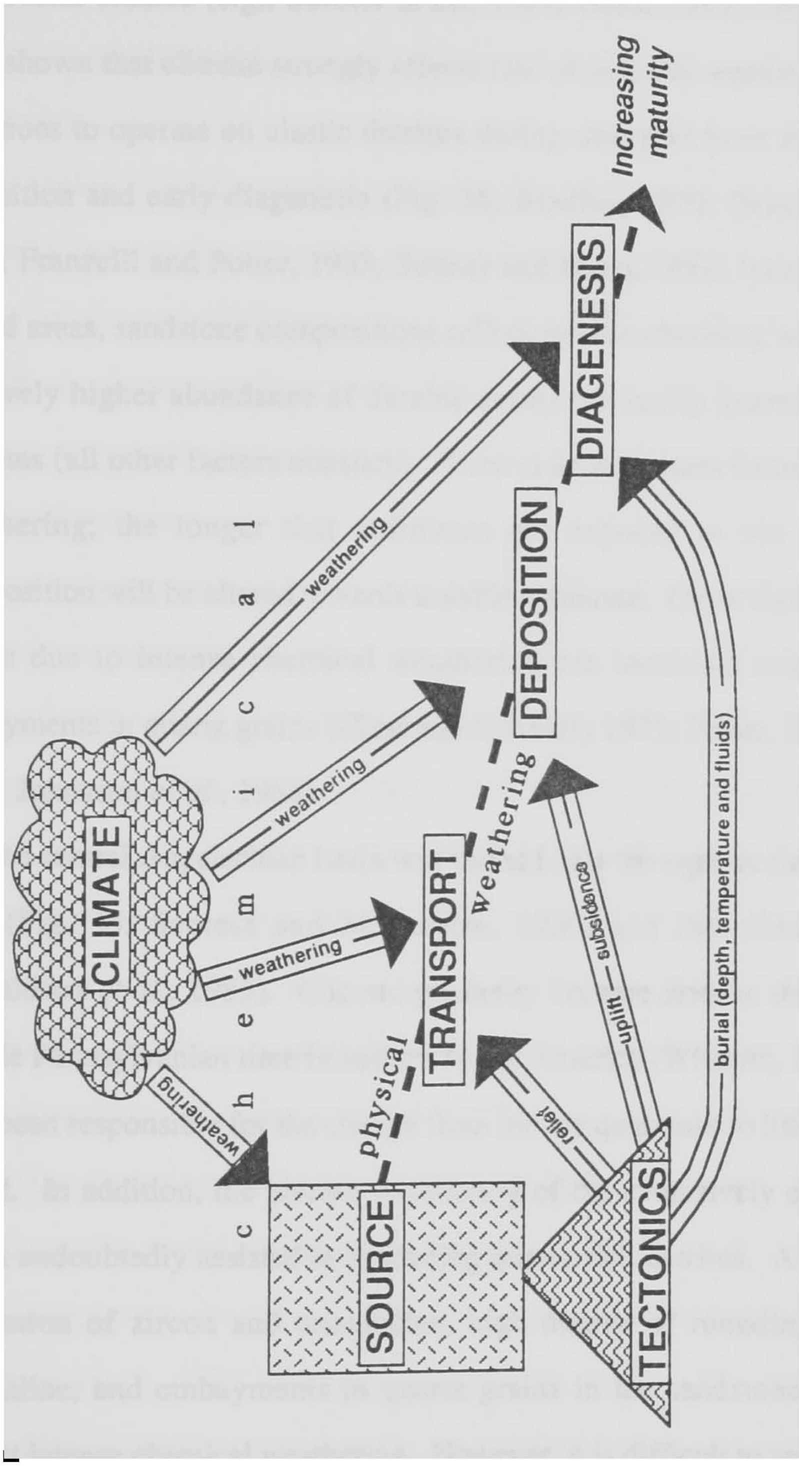
Early and Middle Pennsylvanian sandstones of the Appalachian, Michigan and Illinois basins have similar compositions (Fig. 33c). Siever and Potter (1956) proposed that the compositional evolution from Early Pennsylvanian, quartzose (Lee type) to Middle Pennsylvanian, lithic (Breathitt-type) sandstones was due to a gradual change in source area. They proposed that initially there was a dominant Canadian shield (Lee-type) source, which was gradually replaced by a northern Appalachian source (Breathitt-type). Petrographic and paleocurrent data (Fig. 35) substantiate a dual source for the northern part of the Appalachian basin (Fuller, 1955; Meckel, 1967; Heimlich *et al.*, 1975; Presley and Donaldson, 1979). The cratonic source area included sedimentary rocks as indicated by sedimentary rock fragments, abraded quartz overgrowths and Devonian fossils (Fuller, 1955; Siever and Potter, 1956; Meckel, 1967). In the Black Warrior and Ouachita basins there is a similar compositional evolution (*cf.* Morris, 1974; Mack, 1982). However, in the Black Warrior basin, a southern, partially volcanic source, especially apparent in Middle Pennsylvanian deposits, complicates the trend (Mack, 1982).

CONTROLS ON COMPOSITION OF SANDSTONES OF THE LEE FORMATION

Source

Source rocks contain indelible characteristics that are maintained, to varying degrees, during subsequent physical and chemical weathering (Fig. 36). The composition of a sandstone depends intrinsically on the relative amount of durable grains in the source rock. Because the Lee Formation was partially derived from sedimentary rocks, the percentage of durable grains (e.g., quartz, zircon and tourmaline) in the source was relatively high. On the other hand, the metamorphic source area supplied rock fragments (schists and phyllites) that are susceptible to physical (transport) and chemical (weathering and diagenesis) destruction. Cameron and Blatt (1971) have demonstrated that most schist fragments are mechanically destroyed in less than 25 km of fluvial transport. Bradley (1970), Cleary and Connolly (1971), Mann and Cavaroc (1973), Abbott and Peterson (1978) and Mack (1981) have documented the chemical instability and rapid destruction of low-grade metamorphic lithic fragments, particularly in humid climates. Therefore, the nature of both inferred source areas was integral to the composition of the Lee Formation.

Figure 36. Schematic diagram summarizing primary controls (capitol letters) on framework-grain composition throughout a sedimentary cycle. Text in arrows indicate important processes (lower case letters) that assist maturity. Outlined words represent processes that act to inhibit maturity. See text for details.


CLIMATE
SOURCE
TRANSPORT
DEPOSITION
DIAGENESIS
TECTONICS
c h e m i c a l
weathering
weathering
weathering
weathering
physical
weathering
relief
uplift
subsidence
burial (depth, temperature and fluids)
Increasing maturity

Climate

Several studies (e.g., Suttner *et al.*, 1981; Basu, 1985; Grantham and Vebel, 1988) have shown that climate strongly affects rate of surficial weathering. In addition, climate continues to operate on clastic detritus during transport from the source area and during deposition and early diagenesis (Fig. 36; Bradley, 1970; Baker and Penteado-Orellana, 1978; Franzelli and Potter, 1983; Suttner and Dutta, 1986; Johnsson *et al.*, 1988). In hot humid areas, sandstone compositions reflect intense chemical weathering by containing a relatively higher abundance of durable grains (primarily quartz) than sandstones in arid climates (all other factors constant). Time is an important factor with regard to chemical weathering; the longer that sediments are exposed to *the elements*, the more the composition will be altered towards a stable residuum. Other significant changes in detrital grains due to intense chemical weathering are increased rounding and formation of embayments in quartz grains (Clear and Connolly 1971; Potter, 1978; Franzelli and Potter, 1983; Johnsson *et al.*, 1988).

The central Appalachian basin was located near the equator during Early Pennsylvanian time (Fig. 35; Scotese and McKerrow, 1990) and the climate was hot and humid (Donaldson *et al.*, 1985). Climate gradually became drier in the basin during Early and Middle Pennsylvanian time in eastern North America (Winston, 1990), and this trend may have been responsible for the change from mostly quartzose to lithic sandstones during this period. In addition, the intense weathering of the extensively exposed North American craton undoubtedly assisted in producing quartz-rich detritus. A restricted heavy mineral population of zircon and tourmaline, high degree of rounding of quartz, zircon and tourmaline, and embayments in quartz grains in the sandstones of the Lee Formation support intense chemical weathering. However, it is difficult to separate the relative effects of climate from other possible influences. For instance, a sedimentary, or recycled, source will yield a similar (chemically and physically) mature composition. Also, possible dissolution by calcite cement (see Suttner and Dutta, 1986) may have produced embayment features on quartz grains.

Tectonics

Tectonism, like climate, has a considerable effect on sandstone composition throughout the sedimentary cycle (Fig. 36). Most important, tectonic setting largely determines the type of source rock (Dickinson and Suczek, 1979). For the Lee Formation, uplift of the cratonic source areas to the north and of mountains to the northeast and east/southeast (Fig. 35) was responsible for supplying sedimentary and metamorphic detritus to the central Appalachian basin.

Tectonic processes also control subsidence (or local uplift) in the basin and affects depositional processes, which together determine basin-fill geometry. The asymmetric shape of the central Appalachian foreland basin (Fig. 32) and eastward change from erosion and the major unconformity, to continuous sedimentation (Englund *et al.*, 1986) suggest differential subsidence, which can be explained by a flexural lithospheric model (*cf.* Beaumont *et al.*, 1987). Because deposition of the Lee Formation was confined to the western part of the basin where subsidence was less, the residence time of the sediment in the zone of weathering was relatively longer than for sediments in the eastern part of the basin. Reworking of sediment in the alluvial plain of the fluvial systems (discussed below) would then be greater in the western part (*cf.* Bridge and Leeder, 1979). Reincorporation of temporarily stored, intensely weathered, sediment in the alluvial plain by channel migration increases the maturity of bed material of modern rivers in tropical South America (Johnsson *et al.*, 1988; Johnsson and Meade, 1990). In a similar manner, quartzose sandbodies aligned transverse to the southwest trending Lee Formation sandstone bodies may reflect reworking, or cannibalization, of proximal foreland basin sediments during uplift (Houseknecht, 1980). In this case the alluvial plain also acts as a sediment storage area, but sediments are reintroduced to the system by uplift instead of fluvial processes.

In addition, subsidence probably controlled the distribution of quartz pebbles in the basin. Quartz pebbles, as mentioned earlier, came from the northern part of the basin. BeMent (1976) suggested that Precambrian and Lower Paleozoic sandstones were the source, and determined that the maximum thickness of possible source units is in New York and Pennsylvania. Conglomeratic units with abundant quartz pebbles in central and eastern Pennsylvania (Meckel, 1967) may have had a similar source. If the source of the pebbles was that far north, then significant transport (> 500 km) must have occurred. Slower subsidence, as inferred for the western part of the basin, facilitates the transport of coarse sediment in alluvial basins (Paola, 1988).

Transportation/deposition

Several authors have documented the effects of transportation on the composition of clastic detritus (e.g., Cameron and Blatt, 1971; Davies and Ethridge, 1975; Ethridge, 1977; Mack, 1978). Most studies have concentrated on the changes produced by physical weathering, which appear to be negligible in rivers and significant in nearshore marine systems. As alluded to earlier, climate, in particular chemical weathering, is the foremost control on sandstone composition. Because climatic processes continue to operate throughout sedimentary cycle the longer the time between erosion of the parent rock and deep burial, the more time available for chemical alteration. In fluvial systems the period of chemical alteration is lengthened by temporary storage of sediments in alluvial terraces, floodplains or bars (Bradley, 1970; Baker and Penteado-Orellana, 1978; Johnsson and Meade, 1990). In the world's large rivers that are located within tropical climates, this storage effect contributes to a characteristic downstream maturing of the sediments (Franzelli and Potter, 1983; Johnsson *et al.*, 1988).

The eastern equivalent sediments of the Lee Formation are fluvial deposits that generally contain high percentages of fine-grained facies and are considered to be meandering-stream deposits (e.g., Miller, 1974; Chesnut, 1988). The inferred fluvial system is unlike the bedload-dominated fluvial system of the Lee Formation (chapter 1), which did not contain significant fine-grained material. This difference in depositional systems may account for dissimilar framework compositions. Intraformational, fine-grained, sedimentary lithic fragments are more likely produced in a meandering-river system than in a bedload-dominated system, because of reworking of fine-grained floodplain material during channel migration. It is possible that some of the matrix material that was recalculated as lithic fragments for the Breathitt-type sandstones may be intraformational lithic fragments, however, because it is difficult to petrographically discriminate the origin of fine-grained lithic fragments this proposal could not be fully evaluated. In addition, the relatively sand-rich material stored in the floodplain of the Lee system was likely incorporated into the channel during avulsion. Analyses of erosional surfaces in the multilateral and multistoried sandstones of the Lee Formation indicate that significant reworking of sand-sized material took place during avulsive events (chapter 1). Reworking of intensely weathered floodplain sediments during channel migrations is an important process of increasing compositional maturity in modern tropical rivers (Johnsson and Meade, 1990). The interpretation of the quartzose sandstones of Lee Formation as deposits of a large fluvial system (Graham *et al.*, 1975; chapter 1) is consistent with the observations of big rivers within tropical climates in the present day.

Diagenesis

Composition of detrital sands can be significantly altered by diagenetic processes that operate from conditions immediately after burial in the zone of weathering to the deep subsurface (McBride, 1985). In addition, after uplift and exposure, outcrop weathering results from contact with meteoric waters. Diagenetic processes of greatest importance are the total dissolution of grains of feldspar, rock fragments and heavy minerals as well as the replacement of these grains by carbonates, clays and zeolites (McBride, 1985). The degree of modification of sandstones may be severe enough to cause the formation of diagenetic quartz arenites (McBride, 1985, 1987). Sandstones of the Lee Formation contain some petrographic evidence of diagenetic alteration of framework grains in the Lee Formation (i.e., minor partial replacement of potassium feldspar by kaolinite), but the degree of alteration is not appreciable. Comparison of sandstones of the Breathitt and Lee Formations to the sands in modern rivers that drain metamorphic and sedimentary source rocks in the humid southern Appalachians (e.g. Cleary and Connolly, 1971; Mack, 1981), reveals similar compositions and supports the contention that diagenetic alteration is not significant. Outcrop dissolution of labile grains and iron oxides staining (Liesegang rings) is apparent, but comparison of outcrop samples with cores (see Siever and Potter, 1956 and Houseknecht, 1978) demonstrates minor compositional alteration.

DISCUSSION

In many ways, the Andean foreland basin is an appropriate modern physiographic analogy for the Appalachian basin. The Orinoco River transports quartzose sands in a deep, low-sinuousity, bedload-dominated river system with a low braiding index comparable to that reconstructed for the Lee Formation (see chapter 1; Nordin and Perez-Hernandez, 1989). Intense tropical weathering has destroyed labile detrital components derived from eastern cratonic (Guyana Shield) and western orogenic (Andes Mountains) sources (Johnsson *et al.*, 1988). Tributaries from the Andean highlands have constructed alluvial fans, which force the Orinoco River against the exposed craton (Nordin and Perez-Hernandez, 1989). A similar situation is inferred for the central Appalachian basin, where the Lee Formation fluvial system was situated at the western extreme of the basin, adjacent to the exposed North American craton, and east of the alluvial system of the time-equivalent Breathitt Formation.

Aspects of the Triassic Sydney basin in Australia (Conaghan *et al.*, 1982) and the Devonian clastic wedge of Arctic Canada (Embry and Klovan, 1976) also resemble the central Appalachian basin. These foreland basins contain quartzose sandstones, deposited in longitudinal fluvial systems, that are associated with coal-bearing sequences. In addition, sediments in both basins were derived from cratonic sources thought to be the likely cause of the compositional maturity of the quartzose sandstones. Chemical weathering in a humid climate may also have been an important compositional control in the Sydney (Dutta and Wheat, 1988) and Canadian (Embry and Klovan, 1976) basins.

Petrography and paleocurrent data demonstrate that the principal source area for the Lee Formation was to the north. Source areas probably included the Lower Paleozoic cover rocks on the Canadian shield area and the northern and central Appalachians. In addition, some sediment was derived from a nearby southeast/east metamorphic source, but volumetrically was probably not significant. This conclusion differs from those of recent studies (e.g., Thomas, 1979; Hatcher *et al.*, 1989; Englund and Thomas, 1990) that invoked a dominant southeast/east source area for the quartzose sandstones of the Lee Formation, based on regional stratigraphy, lithofacies distributions and sediment dispersal patterns .

Several studies have proposed that climate was the governing control on sandstone composition in the central Appalachian basin (e.g., Cecil *et al.*, 1985; Donaldson et al, 1985). Although a climatic drying trend is recognized from Early to Middle Pennsylvanian time (Winston, 1990), it is also likely that a change in sediment dispersal (including source areas) and drainage patterns (*cf.* Donaldson and Schumaker, 1981) caused the change from lithic to quartzose sandstones in the central Appalachian basin.

Ferm and Horne (1979, p. 370) suggested that the solution to 'the orthoquartzite problem' of the Lee Formation is that 'the sands have been cleaned of non-quartz material due to reworking' in barrier environments. However, stratigraphic and sedimentologic data indicate that the barrier island model is not viable for the sandstones of the Lee Formation.

CONCLUSIONS

Sandstones of the Lee Formation are attributed to deposition in a bedload-dominated fluvial system, with source area and climate functioning as primary controls on composition. Source rocks were primarily composed of quartz-rich sediments and metamorphic rocks. Intense weathering, associated with humid tropical climates, acted upon the detritus throughout the sedimentation cycle. Less important controls on sandstone composition were tectonics and transport/depositional processes that promoted prolonged exposure of the detritus to the intensive chemical weathering environment. The quartzose sandstones of the Lee Formation reflect lower rates of tectonic subsidence and greater recycling of sand-sized grains during transportation and temporary deposition on the alluvial plain, relative to the lithic time equivalents to the east.

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APPENDIX A: PHOTOMOSAICS OF OUTCROPS: USEFUL PHOTOGRAPHIC TECHNIQUES

ABSTRACT

Photomosaics can be useful tools for understanding and communicating geologic features expressed on outcrop faces. In order to properly utilize photomosaics maximum resolution and minimum geometric distortion of the features is necessary. Maximum resolution is best obtained by using quality equipment and by attending to proper technique. In some cases increasing contrast will improve resolution; various methods can be utilized.

Sufficient overlap of photographs in the construction of photomosaics will remove distortion in most situations. A common problem is perspective distortion, the convergence of vertical lines. This occurs when the film plane and outcrop face are not parallel and results in curved or "smiling" mosaics. When it is not feasible to obtain parallelism, several methods can be used to help correct this problem. In many situations fitting a 35 mm camera with a perspective control lens is the simplest and most economical strategy for reducing or eliminating the problem.

INTRODUCTION

Recent studies of sedimentary rocks that incorporate two- and three-dimensional exposures (see Miall, 1988a for a partial list) have reemphasized the usefulness of photographs as an aid in description and interpretation. Maximum resolution of detail and minimum distortion of linear features are qualities necessary to properly convey the sedimentary relationships manifested in an outcrop. This paper describes photographic techniques that will ensure the best possible photographic reproduction of rock outcrops.

Representation of extensive rock outcrop by photographic means may require piecing together several photographs into a single picture (i.e., a photomosaic). Unfortunately, when joining several photographs together, offsets resulting from distortion will multiply and commonly result in unacceptable levels of error. Furthermore, a significant problem may come about when working with tall or inclined outcrops (Figure A-1). Photographs of such outcrops may show convergence of vertical lines (Figure A-2). When joined together, mosaic curving, or "smiling" results, which distorts the true geometries of sedimentary elements (Figure A-3a).

An understanding of several basic photographic principles and application of a few techniques will help improve the quality of a photomosaic. In this paper, improvement of image detail resolution and elimination of distortion are discussed. For simplicity only black and white print film will be discussed, however, most concepts are directly applicable to color film. The methods described herein are suggested for photomosaics of sedimentary rocks, but can be applied to any type of geologic study which incorporates two-dimensional representations of extensive rock outcrops.

RESOLUTION

Resolution is the ability to distinguish detail in a photographic image. Integral to the resolution of a feature is the sharpness (i.e., the distinctiveness of the boundaries of a feature against its background). Image sharpness depends on several factors, including: quality of the lens, correct focusing, sufficient depth of field, elimination of camera (or subject) motion, film sensitivity, and contrast. The first two factors are self explanatory, the other three are discussed below.

Depth of field refers to the nearest and farthest parts of the subject that can be rendered sharp at a given focus setting (Hedgecoe, 1980). Depth of field increases with decreased aperture and is also greater for a lens with shorter focal length. To obtain a sharp image of a subject that has considerable relief along the line of sight, minimum aperture opening is necessary. To maximize the depth of field for a particular situation, it is important to keep

Figure A-1. Schematic diagram showing camera positions relative to outcrop faces for cases where perspective distortion a) is non-existent; b) results from tilting the camera back to include all of outcrop into the picture when outcrop height is large relative to camera to outcrop distance (this effect will be lessened as camera is moved away from outcrop); and c) is caused by sloped outcrop face.

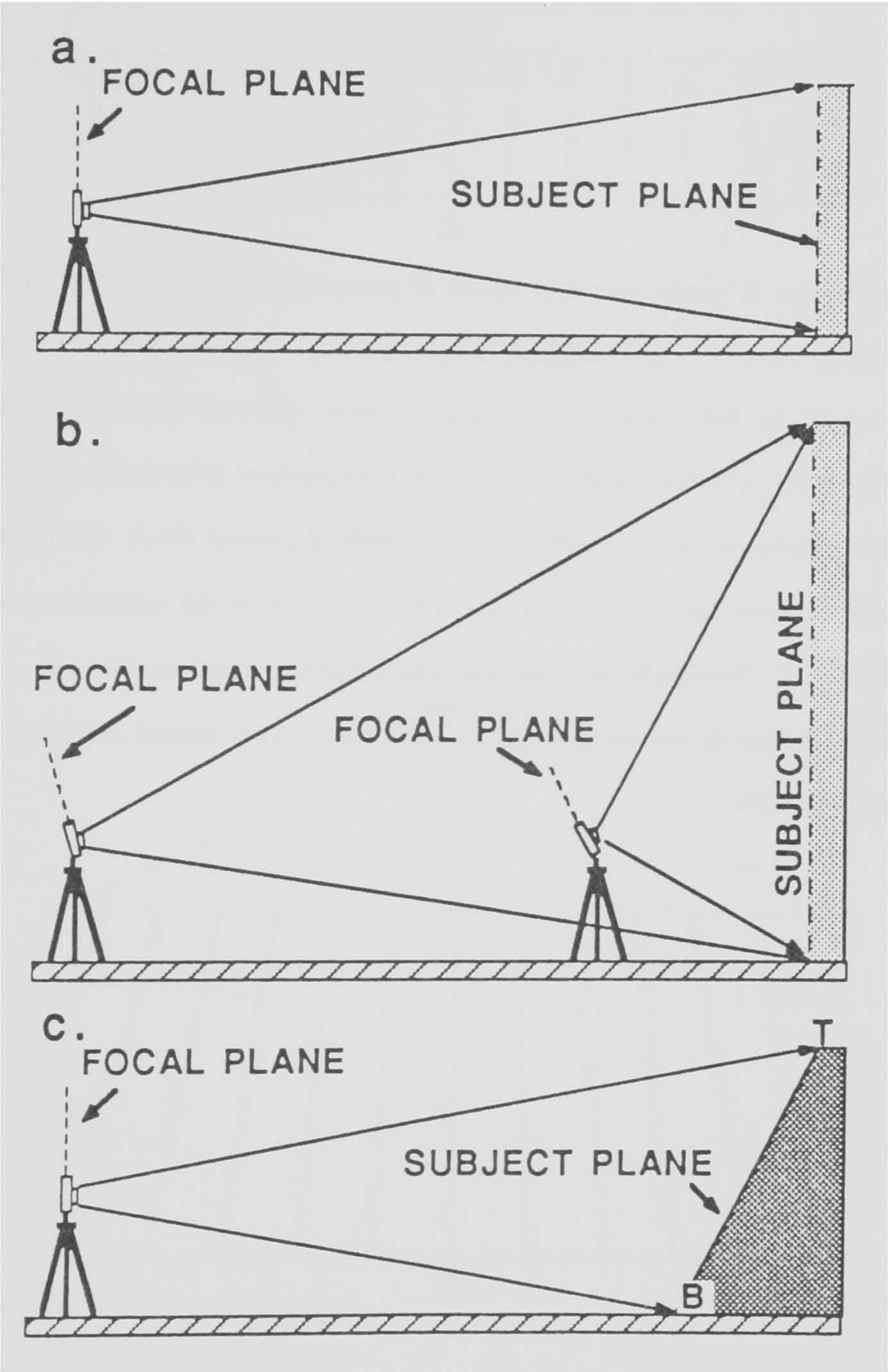
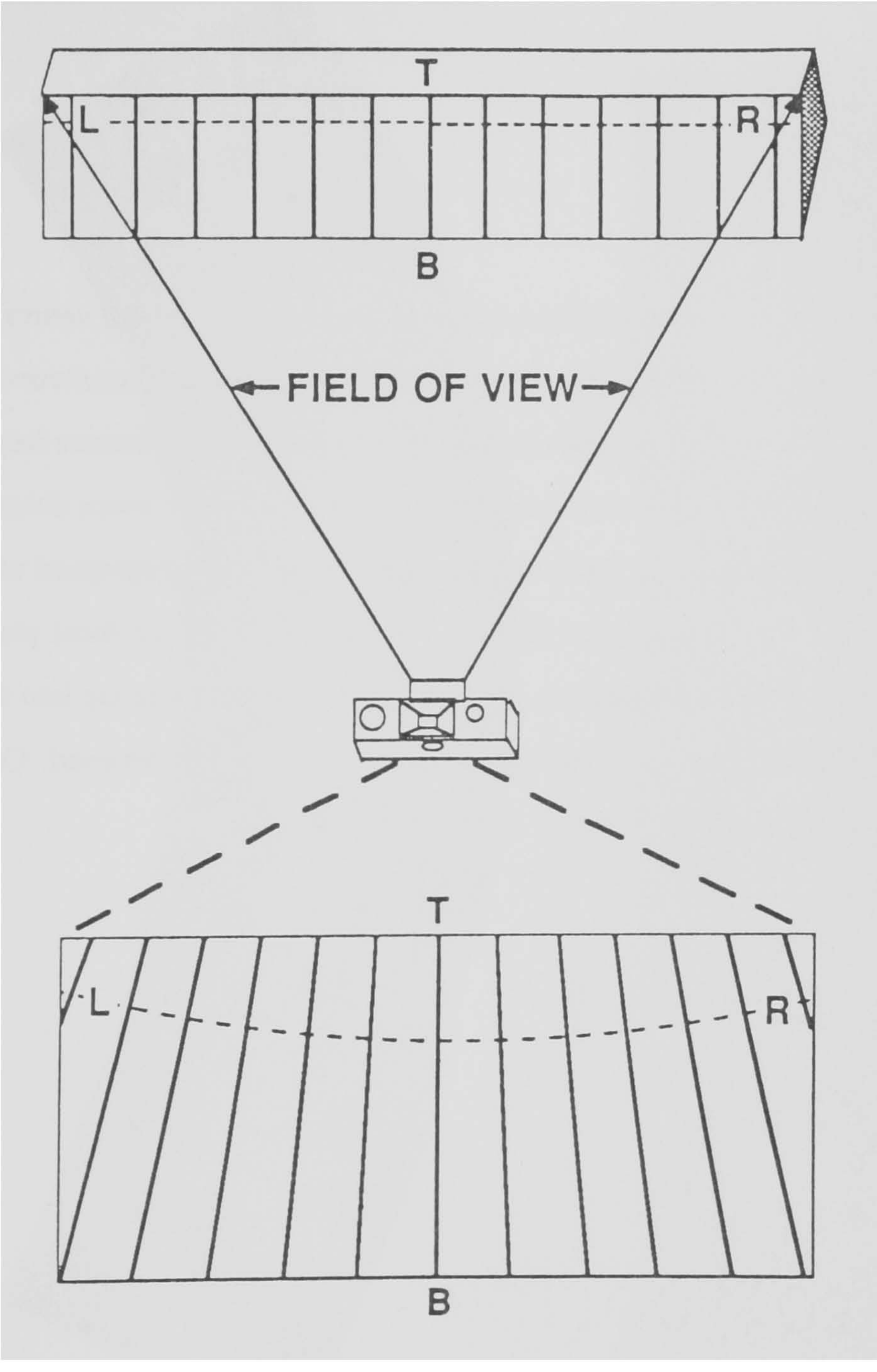

a.
FOCAL PLANE
SUBJECT PLANE
b.
FOCAL PLANE
FOCAL PLANE
SUBJECT PLANE
c.
FOCAL PLANE
SUBJECT PLANE
T
B

Figure A-2. Schematic diagram showing causes of perspective distortion for case when outcrop face is sloped. Closely spaced parallel lines represent vertical lines on outcrop face (e.g., drilled blast tubes), dashed line is straight and horizontal. a) View from above of inclined surface (Figure A-1c). Because the camera field of view increases with distance from the camera, it encompasses more outcrop in the upper portion of the outcrop than in the base. b) Resultant photograph for situation depicted in 1c and 2a. Convergence of vertical lines in top part distorts true vertical and lateral linear relationships.


T
L
R
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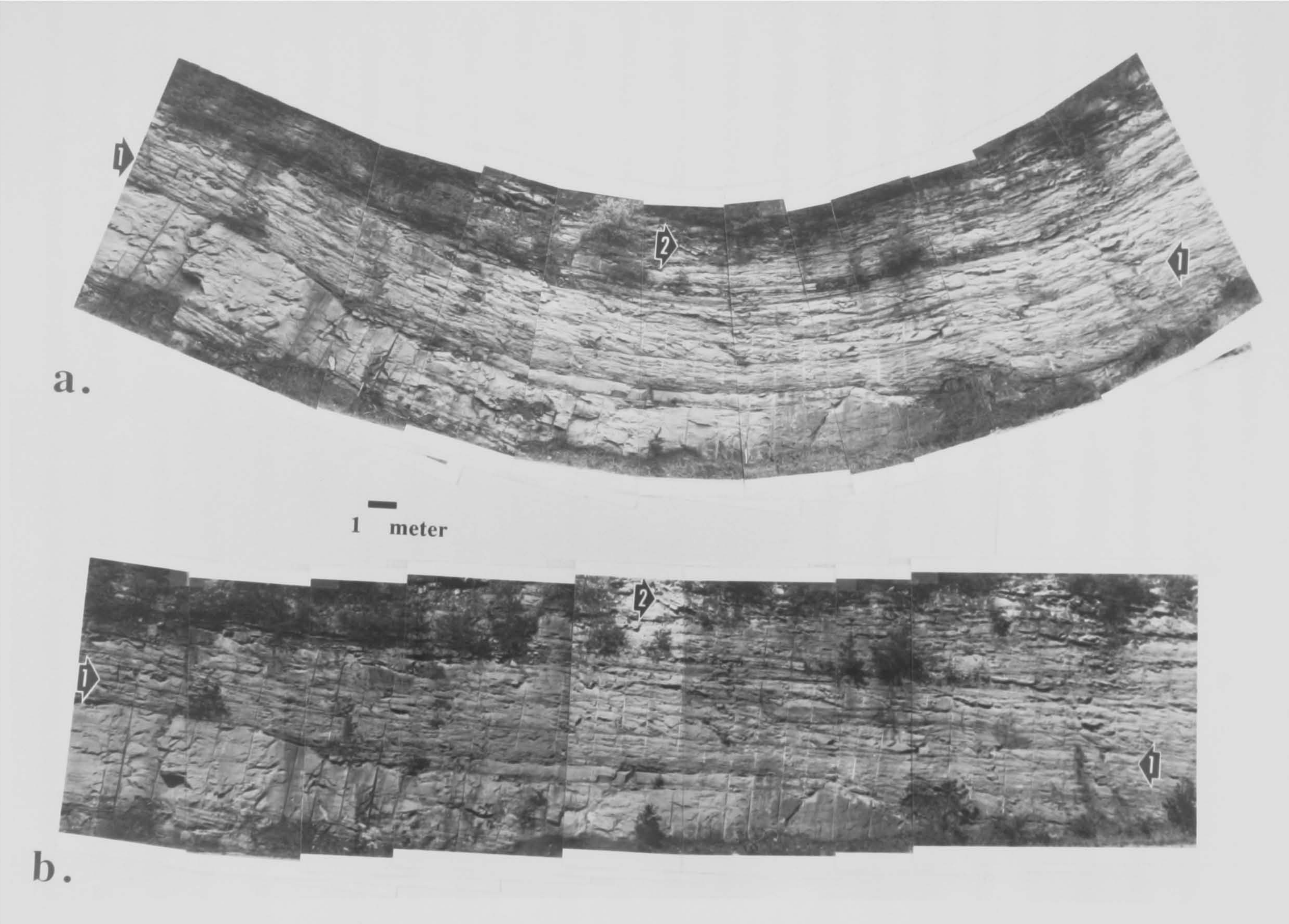
FIELD OF VIEW
T
L
R
B

Figure A-3. Pictures of outcrop with face inclined 30 to 35 degrees from vertical, taken from approximately 20 meters from outcrop. Significant lateral curvature in the outcrop requires large overlap. a) Photographed with camera focal plane vertical. Note excessive "smiling" of photomosaic and severe distortion of linear features (e.g., bedding planes (arrow 1)). b) Photographed using PC lens. Camera was tilted forward from vertical until the focal plane and outcrop face were parallel, PC lens adjusted to bring outcrop into the film format. Distortion of horizontal surface (arrow 1) is removed. Compare larger size of feature (arrow 2) with that in mosaic a.


1
2
1
a.
1 meter
1
2
1
b.

in mind that depth of field extends farther behind (about 3/5 of total depth of field) the distance for which a lens is focused, than in front.

For a given film sensitivity, shutter speed and aperture must be matched for correct exposures. When maximizing depth of field (decreased aperture), slower shutter speeds are required. Camera motion becomes a problem when using slow shutter speeds, less than 1/60 second for a standard 35 mm camera with a 50 mm lens. Furthermore, film sensitivity is directly related to the obtainable resolution. In general, the slower the film speed, the better the sharpness. Therefore, since slow speed film warrants longer exposure times (all else constant), slow shutter speeds are desired. Best results are obtained by using slow film (ASA/ISO 100 or less), a small aperture and a slow shutter speed, in conjunction with a tripod, which prevents camera motion.

Contrast refers to the range in tonal variation in an image. For black and white prints, a high contrast image contains mostly black and white, whereas a low contrast image contains many intermediate grey tones. Low contrast, or apparent blending of tones across the edges of an important feature can result in low resolution, whereas a feature that contrasts highly with its surroundings is easily distinguished.

Quite often it is desirable to increase the contrast in a photograph, to better show geologic (sedimentologic) features. To a limited extent, contrast of an image is increased with decreased aperture size. Significant contrast increase can be obtained by using different films, camera filters and printing techniques.

Most black and white films are panchromatic (i.e., sensitive to all colors of visible light). However, high contrast films have variable color sensitivities, specific colors are either accented (darkened) if the film type is sensitive to that color or muted (lightened) if it is not, producing a high contrast image. Since only a few color-sensitive types are manufactured, these special films are limited in application. Alternatively, lens filters can be utilized to produce the same effect on panchromatic black and white film for a wide variety of colors. By using an appropriate filter, the contrast can be increased between two different colors that would otherwise photograph as similar tones of gray without a filter (Kodak Professional Photoguide, 1986). In general, a filter of similar color to the object photographed will lighten it, whereas a filter of a complementary color will darken it, thus selectively increasing the contrast.

Contrast can also be adjusted during printing. Photographic paper can be obtained with different grades of contrast, from low to high. Printing on high contrast paper yields a high contrast image. However, most paper used is polycontrast, where the contrast can be adjusted by using contrast filters when printing.

PERSPECTIVE DISTORTION

To accurately portray the spatial relationships in a two-dimensional outcrop, it is imperative that the camera film, or focal plane is parallel to the outcrop face (Figure A-1a). This is simply because photographs, two-dimensional representations of a three-dimensional features, display closer objects larger than far away objects. If the focal plane is oblique to the outcrop face (Figures A-1b and A-1c), then part of the outcrop will appear relatively smaller (Figures A-2 and A-3). This type of distortion is called *perspective distortion*. For a single photograph, this may not be a problem, but when constructing a photomosaic, severe mismatch from photograph to photograph as a result of perspective distortion may cause difficulties.

Of course, not all outcrop faces easily lend themselves to being photographed such that the focal and outcrop plane are parallel. For instance, road cuts are often 3-dimensional, their faces following the curvature of the road. Lateral deviations from an ideal plane are normally easily corrected by positioning the camera equidistant from the face for each photograph. However, vertical deviations are sometimes more difficult to correct and may require special techniques.

Two examples of vertical perspective distortion are shown in Figure A-1. Rock outcrops can particularly cause a problem if they are tall, relative to the distance from which the photograph is taken (Figure A-1b). By tilting the camera body to include the entire outcrop within the viewfinder, the focal plane is no longer parallel to the image plane, causing the upper part of the outcrop to appear farther away than the lower part, resulting in distortion. Similar distortion occurs for outcrops that tilt away from the vertical plane (Figure A-1c). In each case, perspective distortion will produce convergence of vertical lines and substantial curvature of horizontal lines (Figure A-2). In many situations, perspective distortion can be severe and must be eliminated or significantly reduced in order to properly convey linear relationships in a photograph.

Correction

There are several ways to reduce vertical perspective distortion when photographing an outcrop. First, the camera may be positioned so that the relationship between the focal plane and the outcrop face is as close to parallel as possible *and* the subject is kept within the view finder. This may require simply raising the camera towards the (vertical) center of the exposure, for example by using a ladder. In this manner much of the distortion may be eliminated. Also, by increasing the subject to camera distance the effect is reduced (Figure A-1b). However, in many situations these simple solutions may not be practical or sufficient to correct the distortion.

Increasing the field of view by using a linear corrected, super-wide-angle lens (18 to 20 mm) will reduce the distortion, but the resultant image will only be a fraction of the film format and significantly off from the center of the negative (Kodak Professional

Photoguide, 1986). This may require increased enlargement of the image to get the desired size, possibly resulting in decreased resolution and loss of important details.

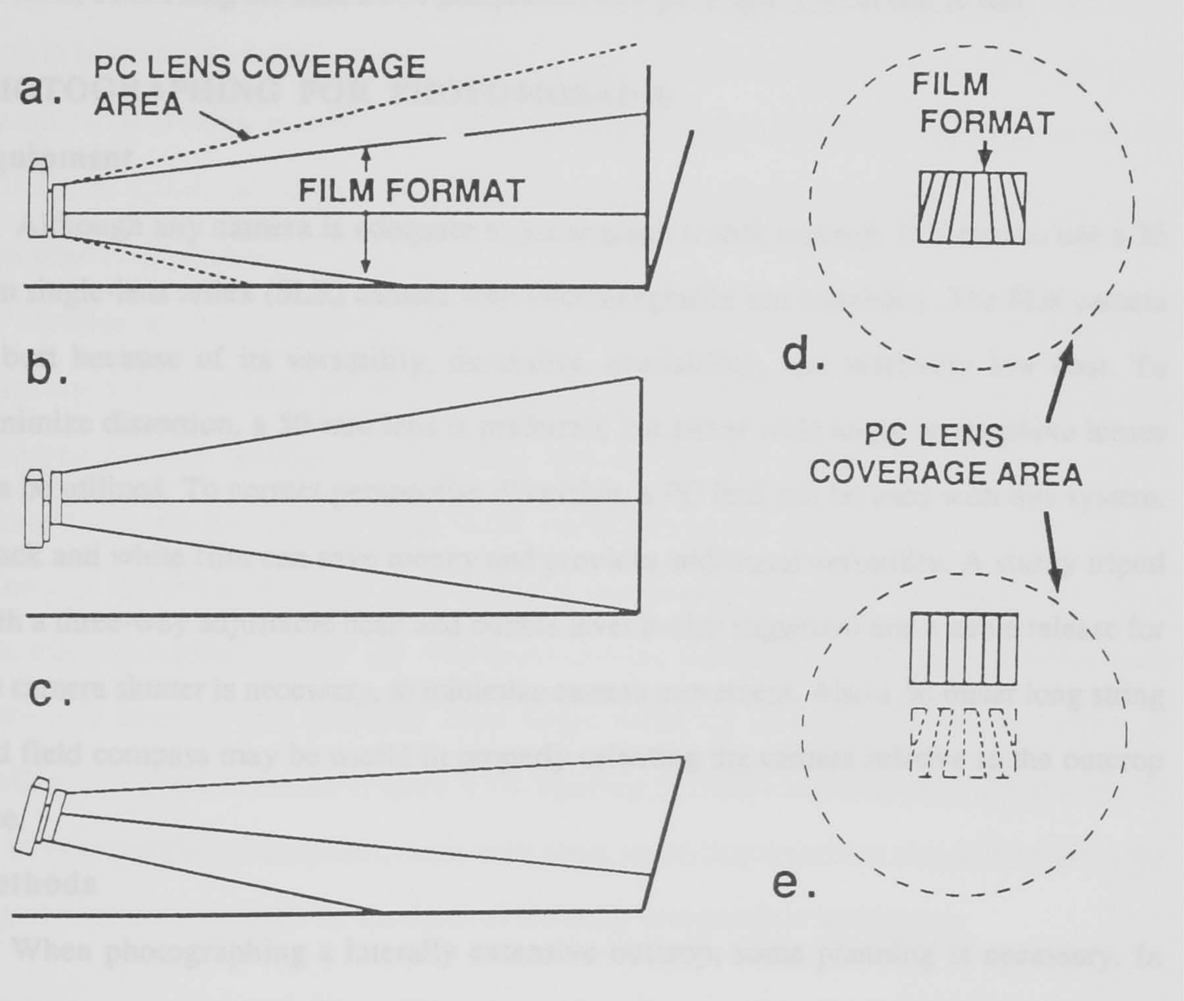
Another alternative for correcting perspective distortion are large-format view cameras, which allow independent control of the lens with respect to the camera body. The focal plane is placed parallel to the outcrop face and the lens elements are adjusted to bring the subject entirely within the viewing screen, not necessarily an easy operation. There will be no distortion in the photograph as long as the focal plane and outcrop face remain parallel.

Drawbacks to large-format view cameras include high price, excessive size and bulk of equipment, and relative complexity of operation. Also only one frame of film at a time can be loaded in the camera. However, because of the large negative size, a sharper and better quality image is produced with these cameras.

With 35 mm cameras one does not normally have the capability to separately manipulate the lens with respect to the focal plane. Perspective control (PC) lenses are a special type of lens where limited movement of the lens allows some control of the image perspective (Figure A-4). The main camera manufacturers offer slightly wide angle (28 and/or 35 mm focal length) PC lenses. They are predominantly utilized by architects for photographing tall buildings and others who wish to control perspective distortion in their photographs. The main advantages of PC lenses are relative ease of operation and compatibility with standard 35 mm film format. Unfortunately, although 35 mm photographic systems are popular, PC lenses are more costly than standard lenses.

A 35 mm camera fitted with a PC lens reduces or eliminates distortion in pictures taken from relatively close to the outcrop face, where the outcrop face is tall or inclined, without having to elevate the camera (Figure A-3). The principle behind PC lenses is similar to a large-format view camera lens, although movement is restricted to one direction at a time

Figure A-4. Schematic diagram illustrating function of PC lens. In (a) camera positioned with vertical focal plane, inadequate to photograph tall or inclined outcrops (bold lines). Dashed lines emanating from camera show field of view covered by PC lens, other lines are the film format (i.e., actual picture coverage). Corrected position of camera and PC lens for tall outcrop (b) and inclined outcrop (c). Relationship of field of view and film format for uncorrected (d) and corrected (e) positions.


a.
PC LENS COVERAGE AREA
FILM FORMAT
b.
c.
FILM FORMAT
d.
PC LENS COVERAGE AREA
e.

and the range of movement is limited. PC lenses have a very wide field of coverage area (Figure A-4). A knob controls movement of the film format (actual recorded image) within the field, controlling the amount of perspective change (Figures A-4d and A-4e).

PHOTOGRAPHING FOR PHOTOMOSAICS

Equipment

Although any camera is adequate to photograph a rock outcrop, it is best to use a 35 mm single-lens reflex (SLR) camera with interchangeable lens capability. The SLR camera is best because of its versatility, durability, availability, and relatively low cost. To minimize distortion, a 50 mm lens is preferred, but either wide angle or telephoto lenses can be utilized. To correct perspective distortion, a PC lens can be used with this system. Black and white film can save money and provides additional versatility. A sturdy tripod with a three-way adjustable head and bubble level is also suggested and a cable release for the camera shutter is necessary, to minimize camera movement. Also a 50 meter long string and field compass may be useful in properly orienting the camera relative to the outcrop face.

Methods

When photographing a laterally extensive outcrop, some planning is necessary. In order to keep the focal plane parallel to the outcrop face and include the full height, a proper distance away from the outcrop must be found. When there are limitations on the distance away from the outcrop which disallow this, then corrections for distortion must be made. Once the optimum camera-to-outcrop distance is determined, a line representing an equal distance from the face should be constructed. The closer to the outcrop that the camera is placed the more crucial the accuracy in constructing this line. Within 50 meters, it is a good idea to measure the distance with a string and repeatedly mark the distance off along the outcrop as required.

When setting up the camera, be sure the tripod is level, this acts as a datum surface for subsequent positions along the outcrop. Adjust the camera so that the focal plane, effectively the back of the camera, and subject plane are parallel (a compass with adjustable level is helpful), then lock the tripod head in place. At each photograph position the tripod should be re-leveled, assuring that the two planes remain parallel. In the case where the inclination of the outcrop face changes, the camera must be readjusted, however, some distortion is inevitable.

Overlap of photographs is necessary in construction of mosaics, because distortion, resulting from curvature of the lens, will occur to some degree in all photographs. For outcrops prone to perspective distortion, spacing between photographs is an important variable. As can be seen in Figure A-2b, vertical lines are straight in the center and the effects of distortion are more pronounced at the photograph edges. It is clear that when constructing a photomosaic any additional overlap between photographs will eliminate more and more of the distorted outer edges reducing the accumulated distortion or "smiling" of a photomosaic (Figure A-3). Spacing for fully corrected photographs should be at least one-half photograph, those with some remaining distortion should be closer, the degree necessary depends on the detail of the study and practical limitations.

APPENDIX B: LATERAL PROFILES AND ARCHITECTURAL ELEMENTS

Line drawings prepared from photomosaics of outcrops could not be reproduced in a continuous manner without loss of detail because of the lateral extent of most outcrops and the desire to keep true-scale (i.e., no vertical exaggeration). Profiles, located in the back pocket, are sectioned to ensure reduction of line drawings to dimensions appropriate to fit on 11 x 17 in-size paper with maximal detail and ease of reproduction. The right end of each profile section (panel) is continued on the left end of the section below (Figure B-1). Panels of individual profiles may be joined together for a continuous profile.

Facies described in the text are included on the profiles (Table B-1) as are higher-order (i.e., third-, fourth-, fifth- and sixth-order) bounding surfaces, which are annotated with numerals representative of their rank. Architectural elements (Table B-2) are identified on true-scale reductions of the profiles (Figures B-2 to B-11).

Table B-1. Facies symbols used in profiles.

Symbol	Facies
S_{cg}	Conglomerate and conglomeratic sandstone
S_t	Tangential cross beds
S_{lb}	Long-bottomset cross beds
S_s	Sigmoidal cross beds
S_a	Asymmetric troughs
S_p	Planar-tabular cross beds
S_c	Compound cross beds
S_{tp}	Topset-preserved cross strata
S_{tr}	Trough cross beds
S_g	Giant cross beds
S_{ms}	Massive sheet-like sandstone
S_{mc}	Massive channel-form sandstone
M	Mudstone

Table B-2. Symbols used in Figures B-2 to B-10.

Symbol	Architectural Element
CH	Major channel
DA_1	Type 1 downstream accretion
DA_2	Type 2 downstream accretion
LA	Lateral accretion
SB	Sandy bedform
GF	Gravity flow
CB	Channel bottom
ch	Minor channel

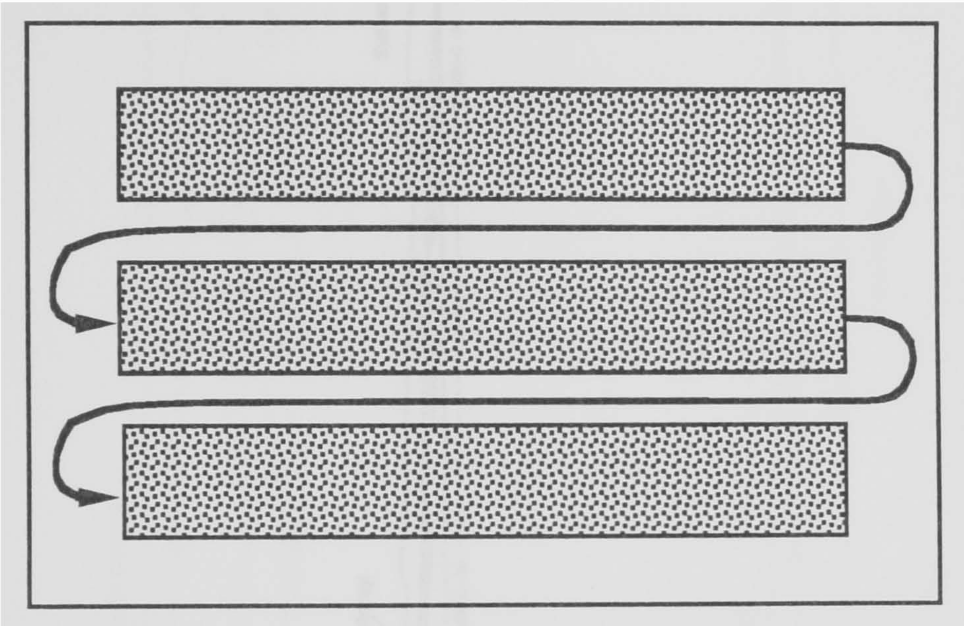


Figure B-1. Schematic diagram demonstrating spatial relationships of panels in profiles.

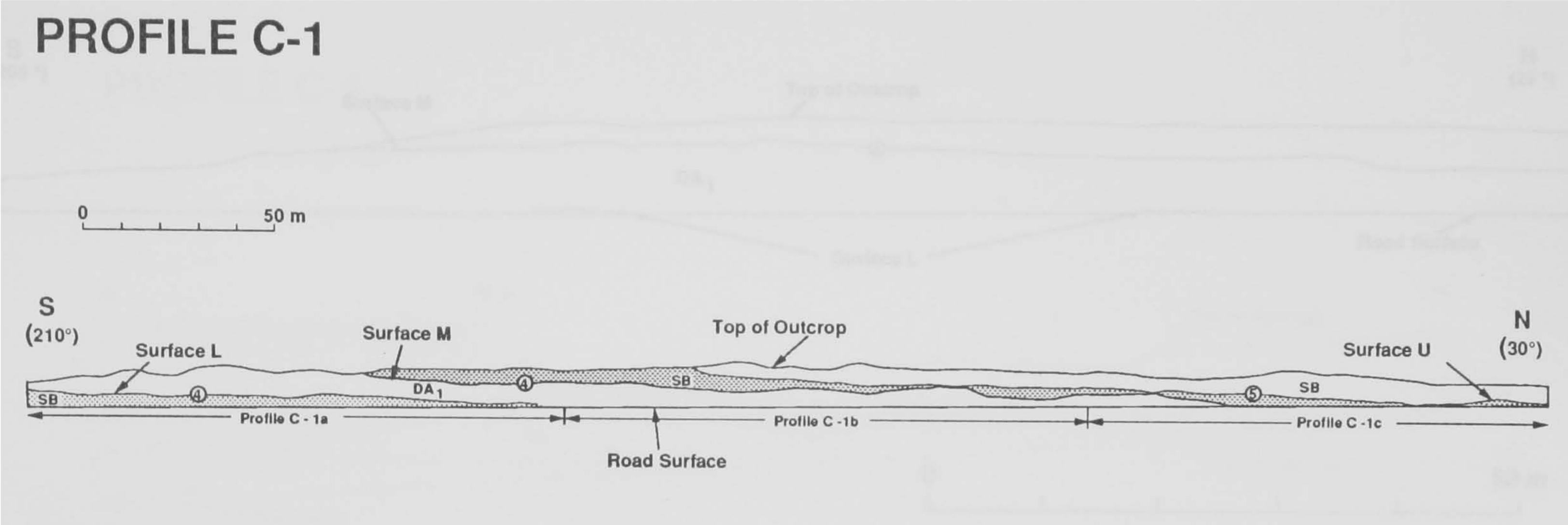


Figure B-2. True-scale reduction of Profile C-1 indicating architectural elements.and higher-order bounding surfaces.

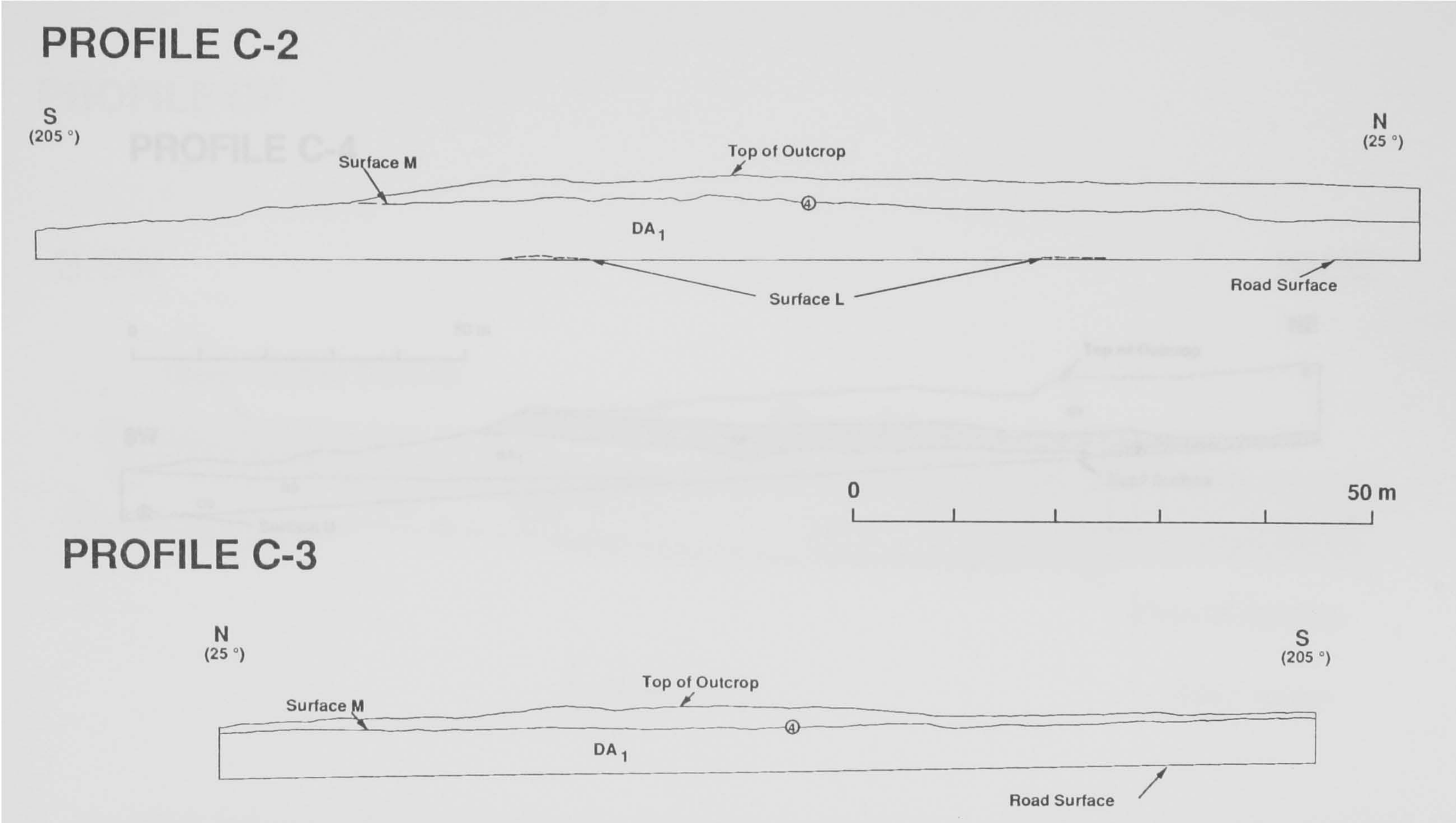


Figure B-3. True-scale reduction of Profiles C-2 and C-3 indicating architectural elements. and higher-order bounding surfaces.

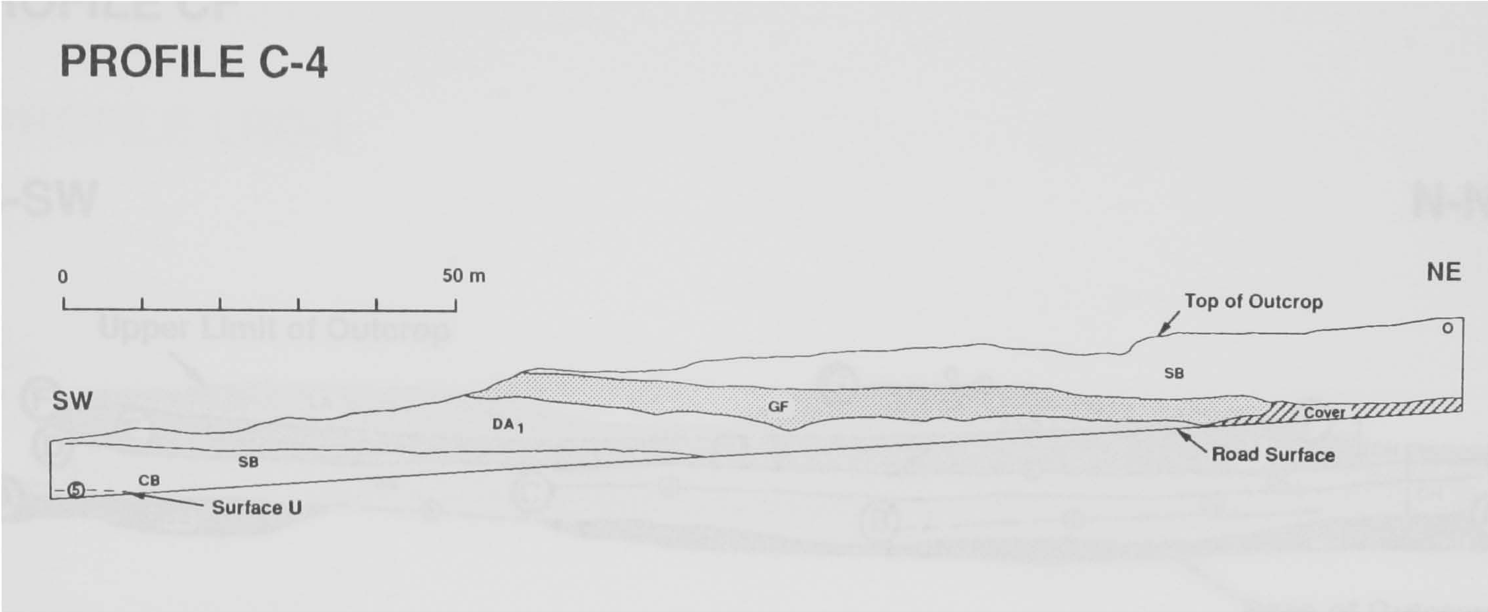


Figure B-4. True-scale reduction of Profile C-4 indicating architectural elements.and higher-order bounding surfaces.

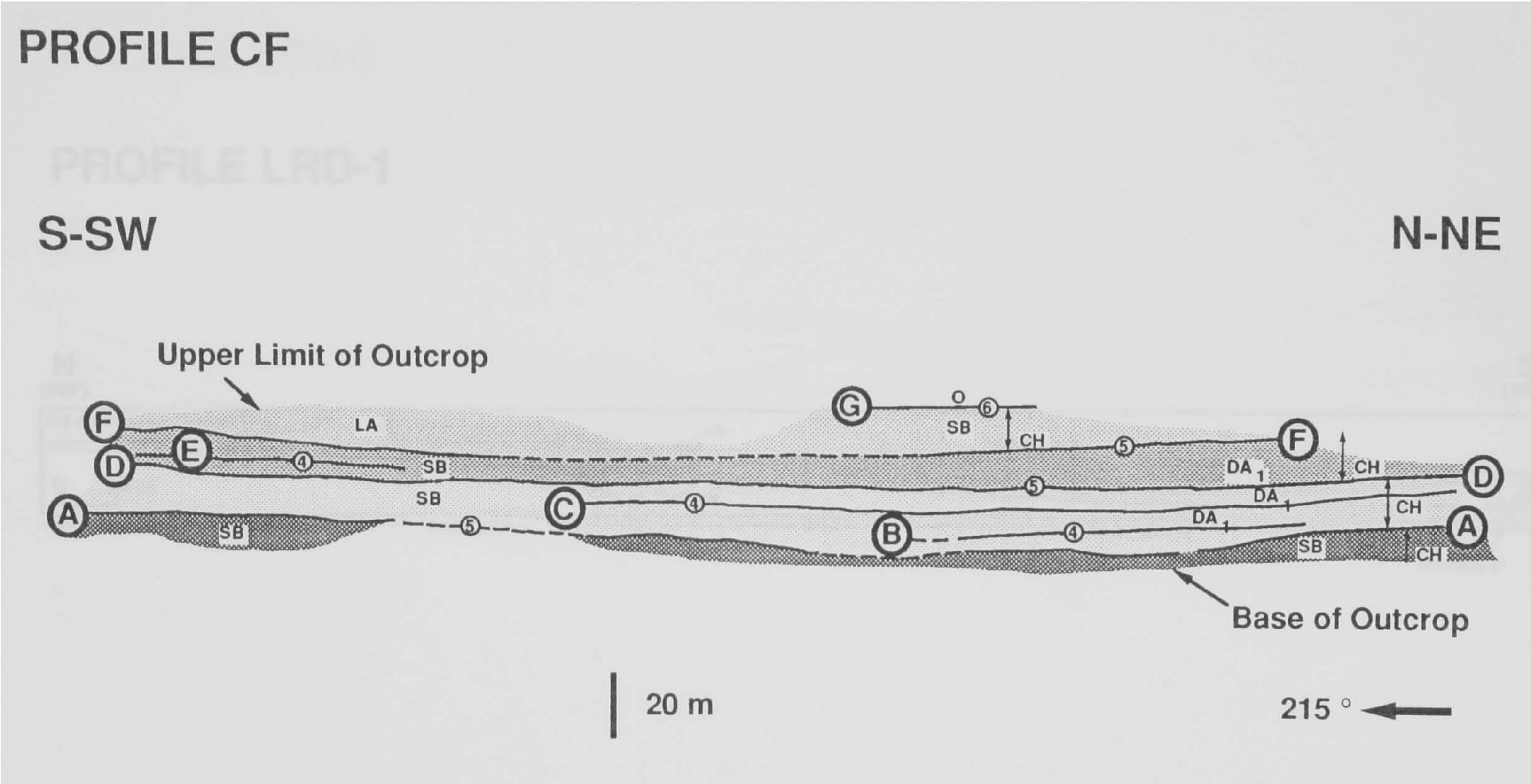


Figure B-5. True-scale reduction of Profile CF indicating architectural elements.and higher-order bounding surfaces.

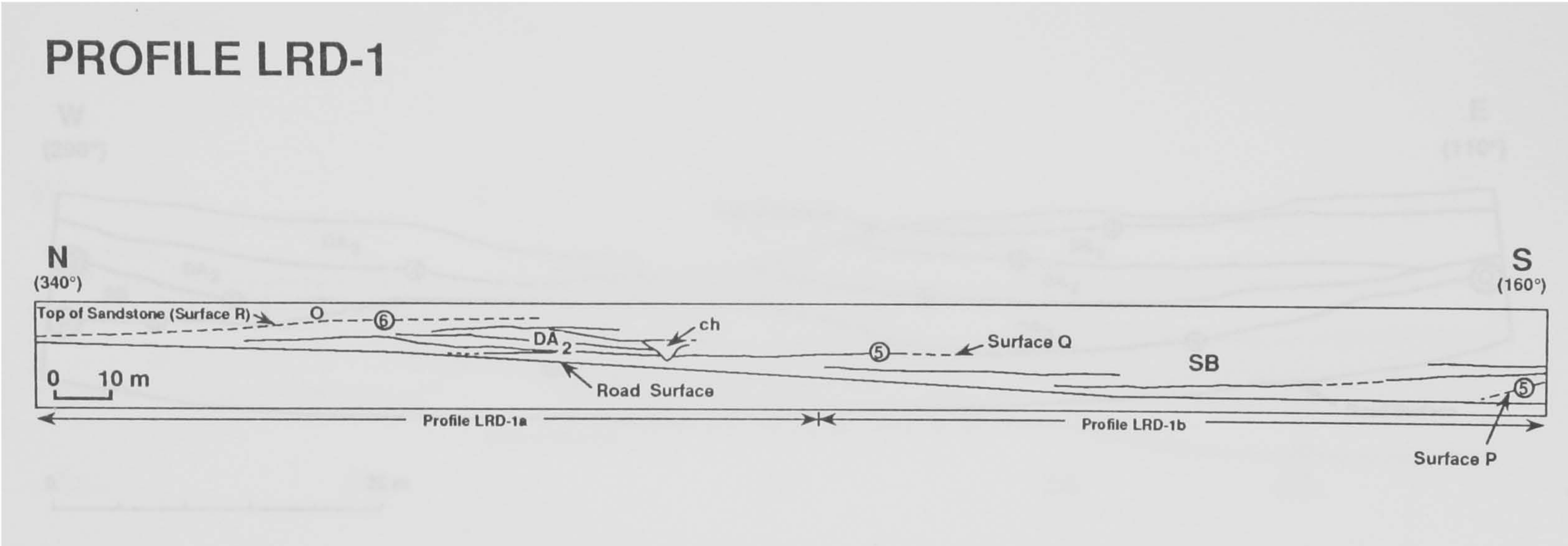


Figure B-6. True-scale reduction of Profile LRD-1 indicating architectural elements.and higher-order bounding surfaces.

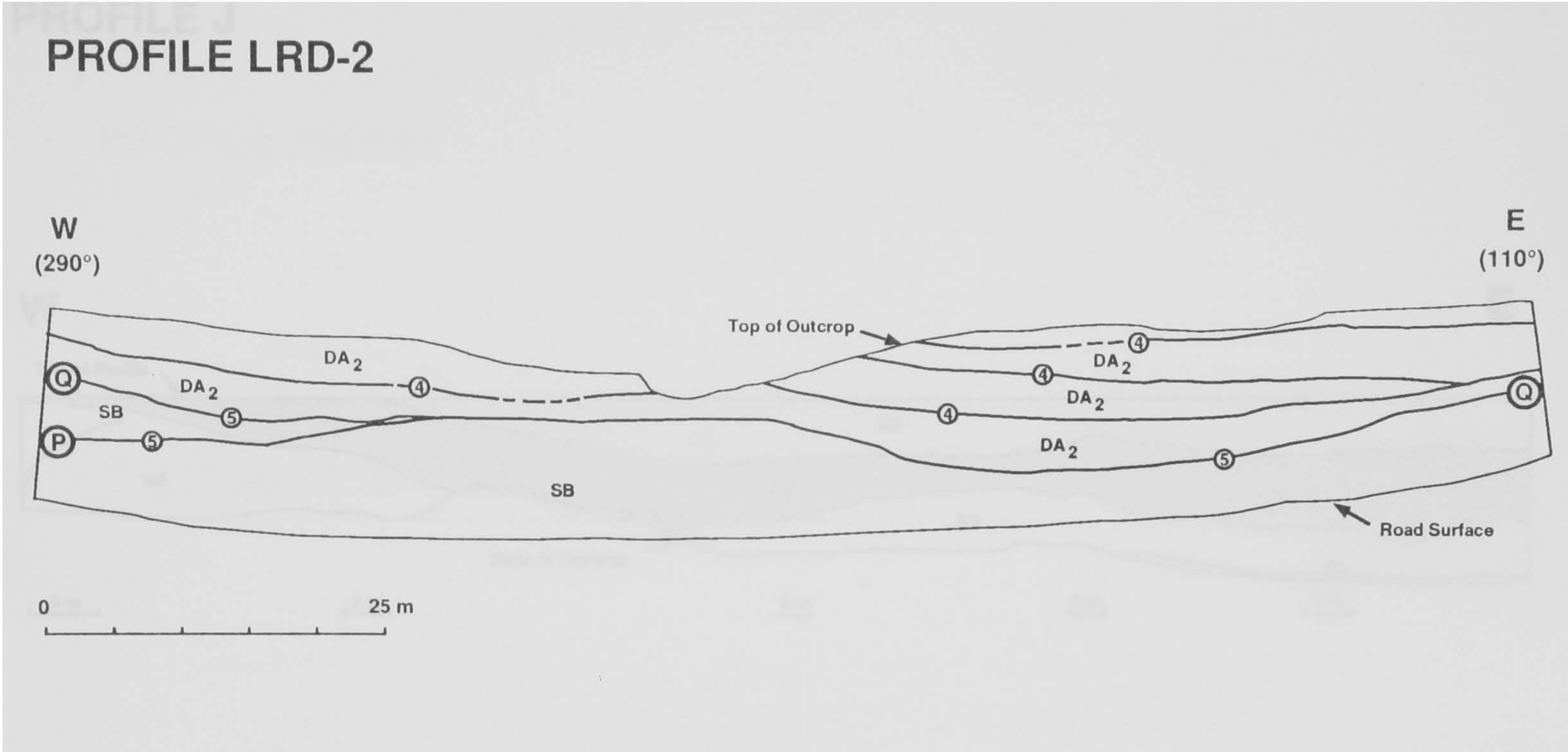


Figure B-7. True-scale reduction of Profile LRD-2 indicating architectural elements.and higher-order bounding surfaces.

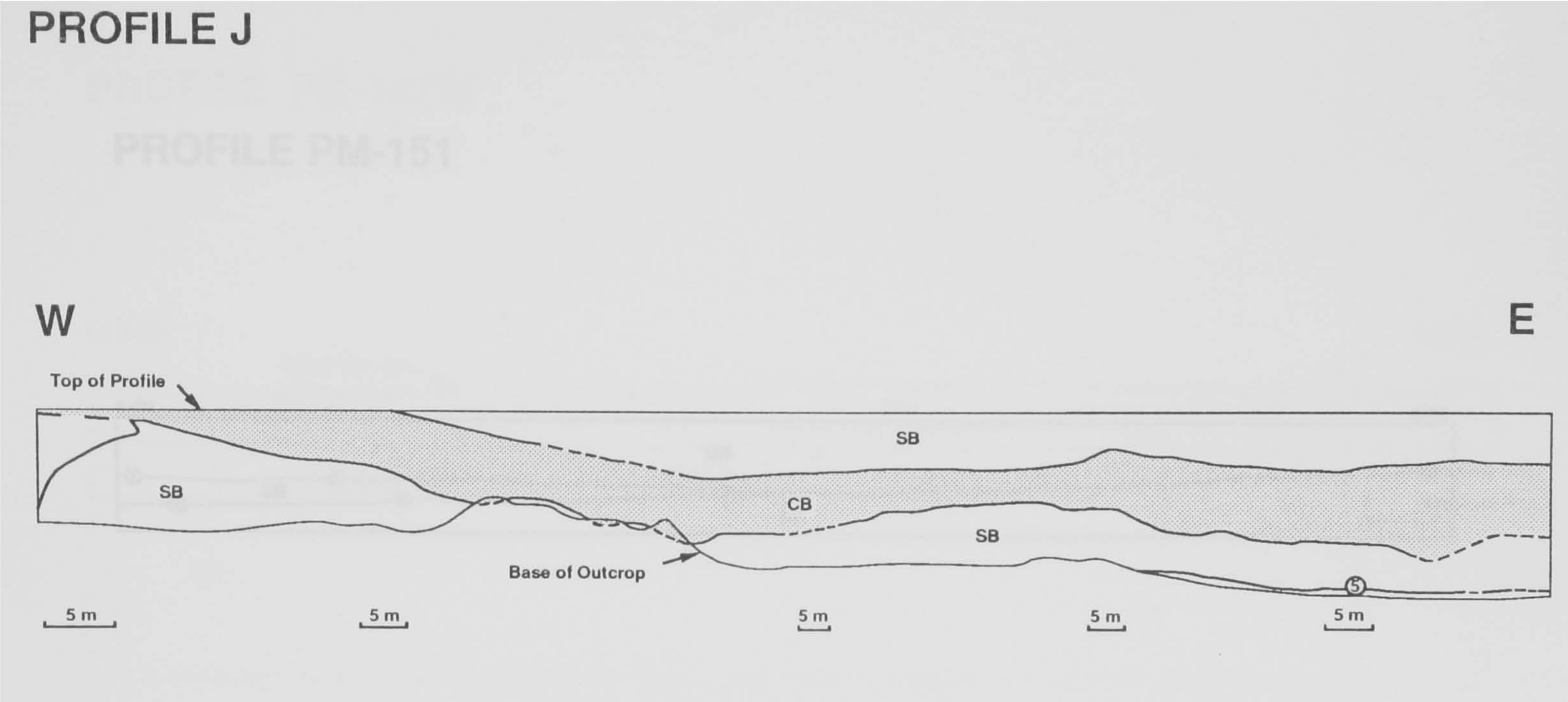


Figure B-8. True-scale reduction of Profile J indicating architectural elements.and higher-order bounding surfaces.

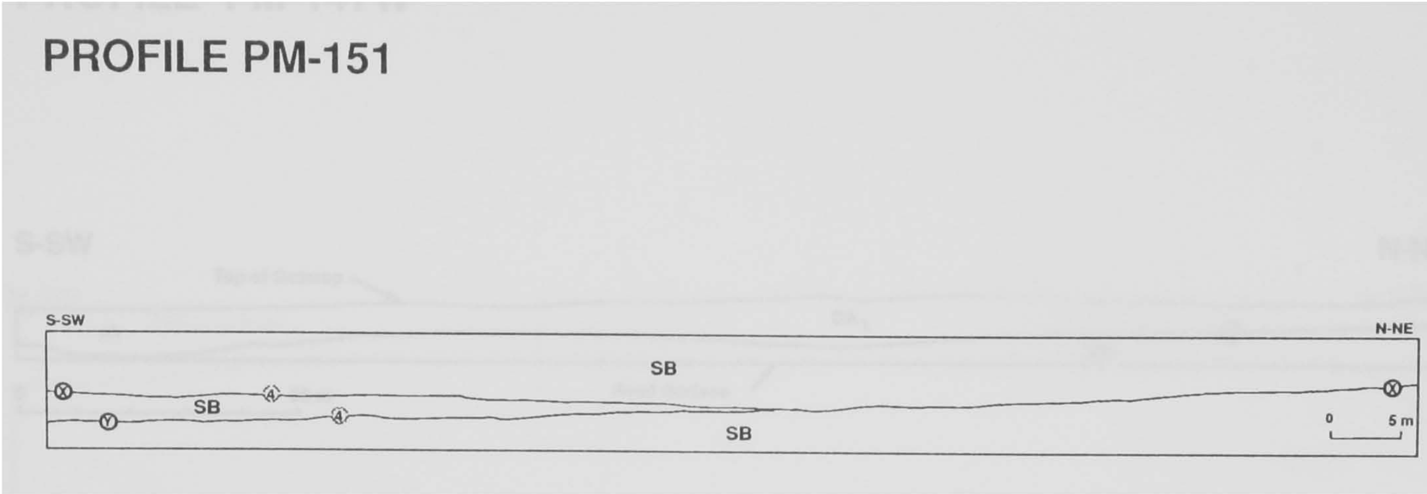


Figure B-9. True-scale reduction of Profile PM-151 indicating architectural elements.and higher-order bounding surfaces.

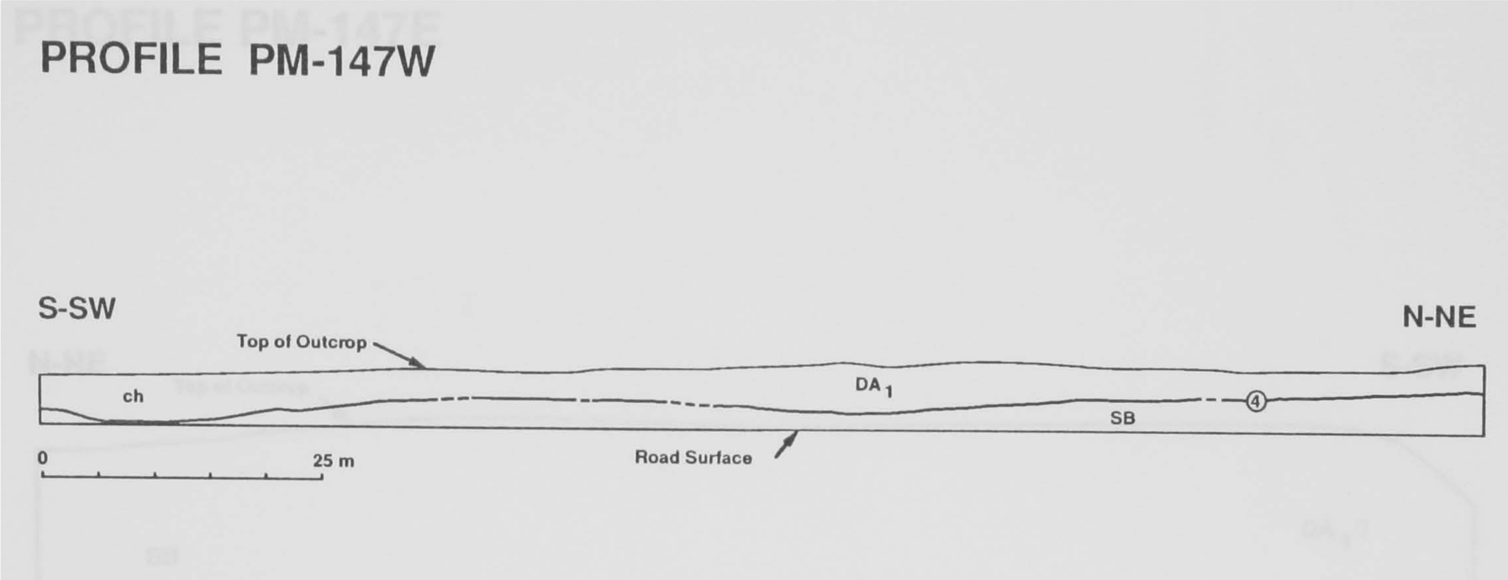


Figure B-10. True-scale reduction of Profile PM-147W indicating architectural elements.and higher-order bounding surfaces.

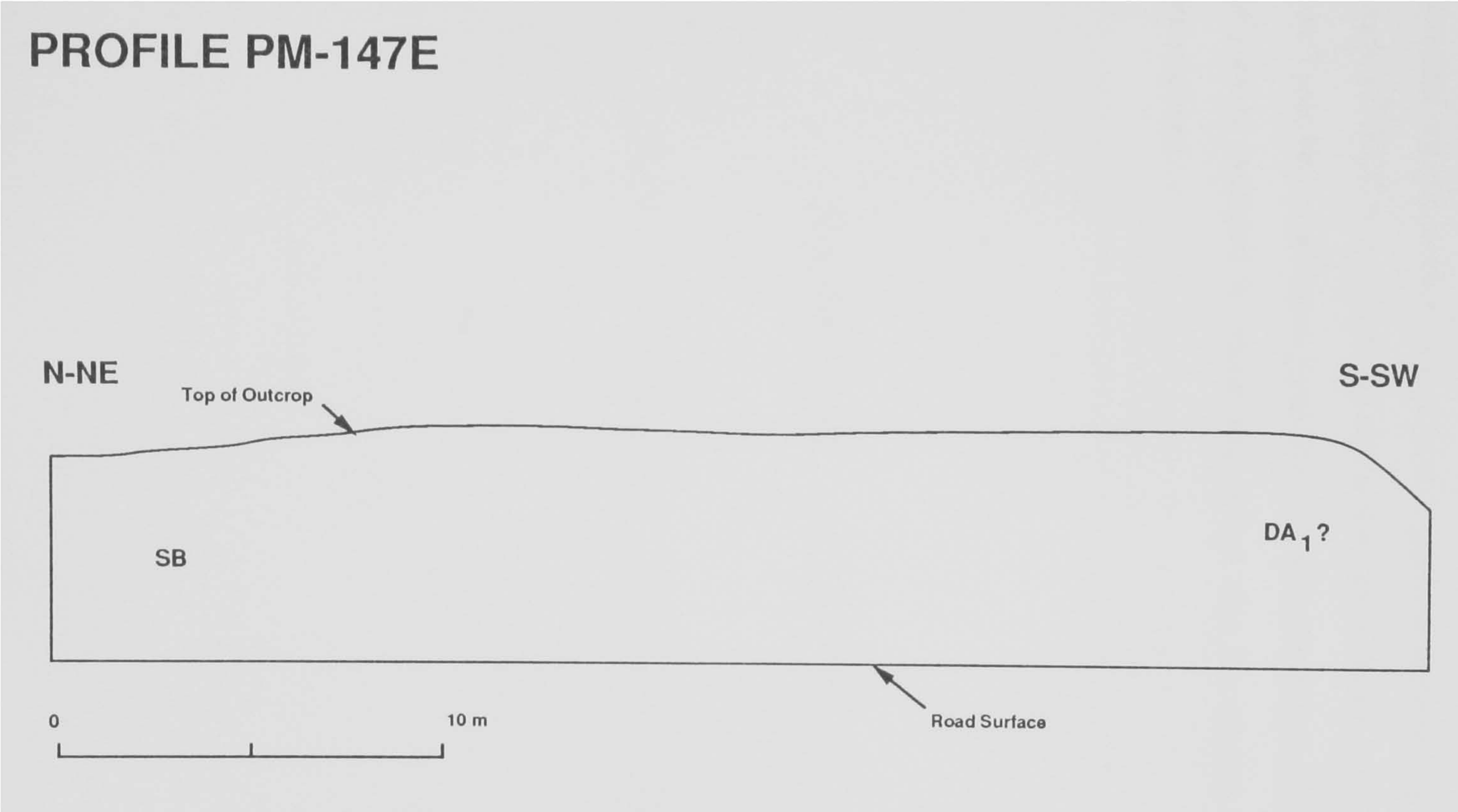


Figure B-11. True-scale reduction of Profile PM-147E indicating architectural elements.and higher-order bounding surfaces.

APPENDIX C: GEOCHEMICAL ANALYSES OF HYDROCARBONS WITHIN THE MASSIVE FACIES

Freshly broken surfaces of the massive facies reveal cm-scale, round, patchy, brown-grey spots. Thin section examination of these features indicates that they consist of an interstitial organic residue concentrated in grains of reworked low-grade metamorphics. A sample (25-ECK-3; Appendix E) was sent to Brian Turner at The University of Newcastle upon Tyne, Newcastle upon Tyne, Great Britain for geochemical tests. The report of the analyses is included in whole and reveal that the organic deposits are petroleum hydrocarbons.

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SAMPLE 25-ECK-3

ANALYTICAL RESULTS

NRG Newcastle Research Group
in Fossil Fuels and Environmental Geochemistry

SAMPLE 25-ECK-3

ANALYTICAL RESULTS

Transmitted Light Microscopy

Microscopical examination of a thin section in transmitted light showed it to be a fine grained sandstone; the grains of which were well rounded and moderately sorted. Pore spaces were partially filled with clay minerals which had then been coated with dark material with the appearance of bitumen. Observation of the dark material under rotated cross polars indicated that it was not siderite. The supposed bitumen accounted for only a small proportion of the rock pore spaces.

Reflected Light Microscopy

Reflected light microscopical examination of the sample mounted in a polished block revealed the characteristic presence of bitumen (hydrocarbon) staining in the sandstone through this bitumen was present in low amounts.

Organic Geochemical Analyses

Liquid chromatographic separation of the EOM on a column, packed with alumina and silica, using successive elutions with light petroleum (BP 40°-60°C) and 50% DCM in light petroleum provided aliphatic and aromatic hydrocarbon fractions respectively. The gravimetrically determined composition of the EOM was aliphatic hydrocarbons 37%, aromatic hydrocarbon 27% and NSO fraction (by difference) 36%. This EOM composition is not dissimilar to many crude oils.

Gas chromatographic analyses of the aliphatic and aromatic hydrocarbon fractions (see Figs 1 and 2, respectively) showed them to be typical of hydrocarbon fractions from a moderately to severely biodegraded crude oil. Both chromatograms are dominated by large unresolved complex mixtures (UCMs), with few resolved components. However, a number of resolved peaks are present in the high molecular weight (biomarker) region of the aliphatic hydrocarbon chromatogram. These peaks (marked B) are probably triterpanes and steranes, GC-MS analysis of which would provide more information on the extent of biodegradation and possibly on the source-rock type of these petroleum hydrocarbons.

Conclusion

Sample 25-ECK is a fine grained sandstone containing relatively low amounts of migrated petroleum hydrocarbons. These hydrocarbons have suffered moderate to severe biodegradation, though the timing of the biodegradation (i.e. before or after migration into the sample) was not ascertained.

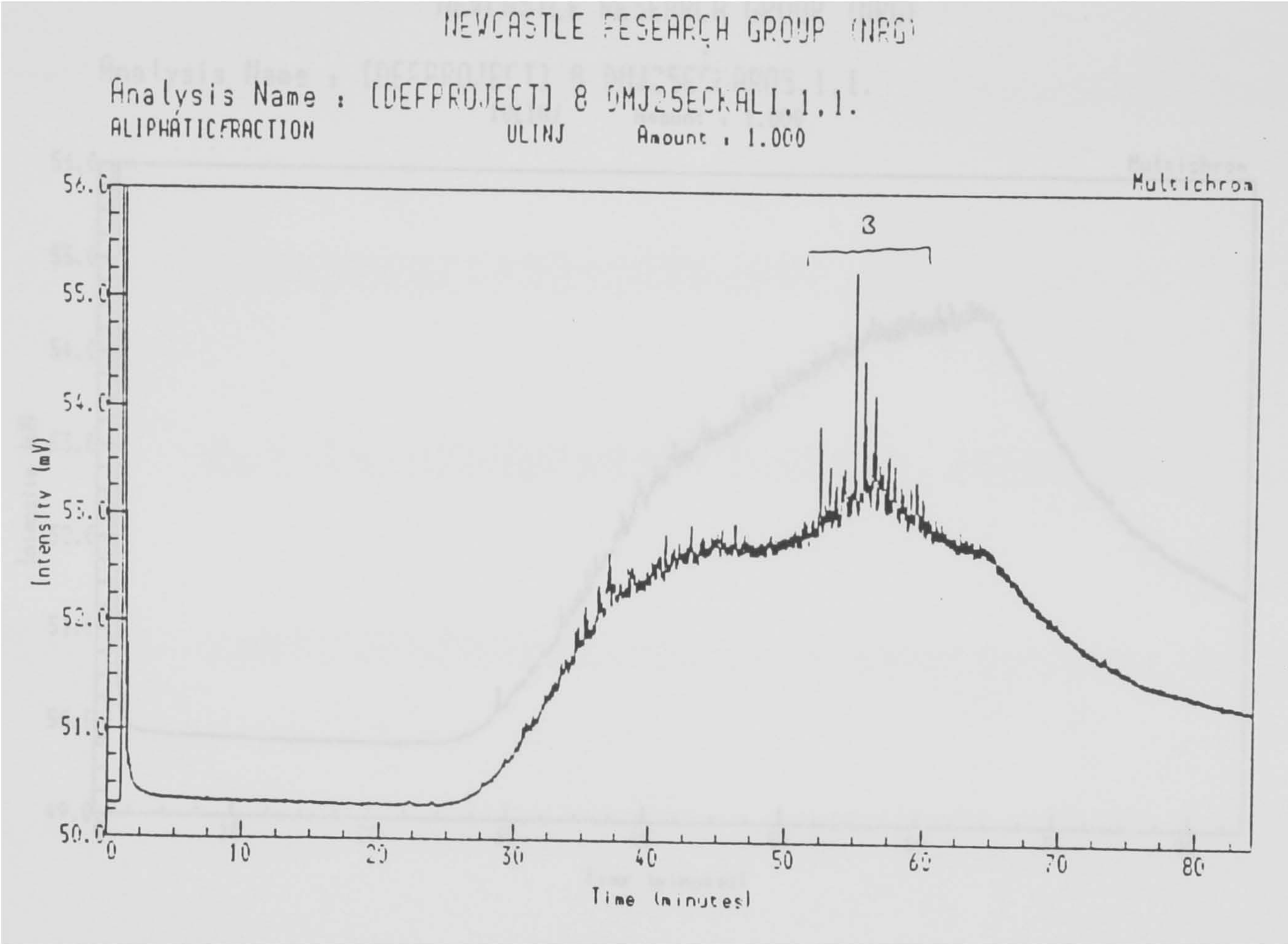

Figure C-1. Gas chromatogram of aliphatic hydrocarbon fraction from sample 25-ECK-3

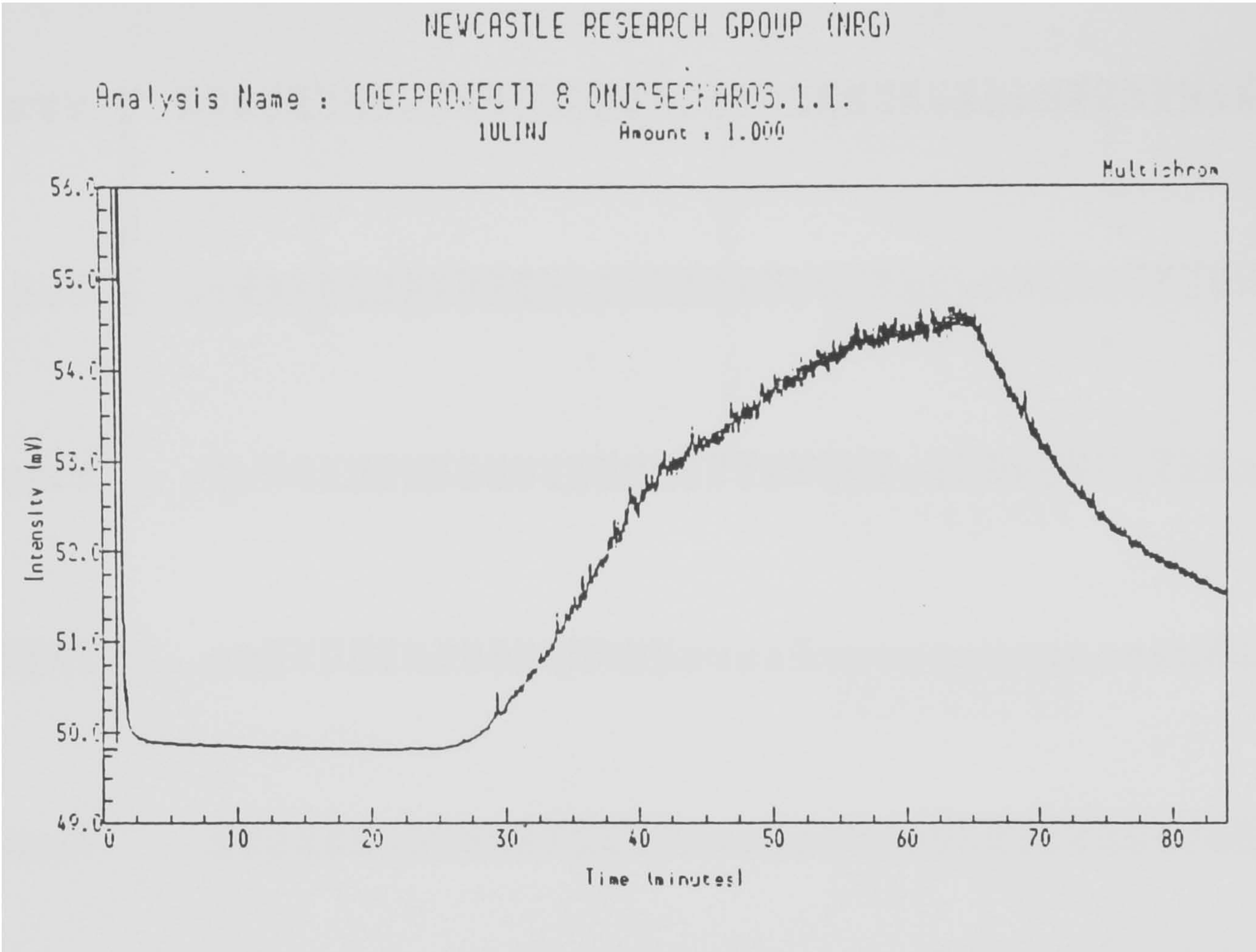

Figure C-2. Gas chromatogram of aromatic hydrocarbon fraction from sample 25-ECK-3

APPENDIX D: PALEOCURRENT DATA

Profile C-1

Planar to tangential cross beds

225	233	195	190	210	226
252	245	192	185	261	245
260	214	45	60	247	255
44	228	201	345	220	205
250	220	195	15	160	230
242	261	206	62	240	251
231	252	212	52	210	216
251	225	200	45	227	245
250	222	239	60	220	247
237	245	220	55	232	200
240	225	196	25	190	220
223	226	295	34	220	205
240	210	212	32	252	214
250	252	210	60	240	220
260	230	70	50	220	46
205	228	104	85	227	166
235	225	230	36	210	45
230	209	235	52	220	60
235	200	231	355	199	55
220	236	190	49	200	65
165	200	206	35	180	80
240	210	201	37	185	25
235	206	216	55	196	20
260	232	265	222	175	216
258	210	235	230	207	178
220	228	220	185	214	222
220	170	240	222	258	215
225	236	219	231	240	221
210	240	229	222	205	237
245	261	222	190	210	209
219	205	240	207	223	198
219	200	264	210	220	214
220	202	233	182	230	210
242	225	255	219	238	68
242	187	280	228	245	55
275	200	205	223	240	51
195	195	197	215	228	50
240	190	230	210	233	69
208	223	33	62	221	97
199	47	232	56	186	30
248	3				

Trough cross beds

30	160	31	130	22	110
30	110	26	117	31	155
16	155	26	125	32	145
18	145	27	170	32	135
22	135	31	130		

Set bounding surfaces

305	120	210	195	227	86
225	180	240	250	200	210
205	78	190	345	232	175
185	110	185	356	244	200
150	32	88	90	210	177
230	230	80	75	244	200
225	248	90	92		

Plant debris

220/40	160/340	275/95	90/270	180/0	55/235
230/50	70/250	220/40	60/240	20/200	5/185
235/55	40/220	265/75	160/340	270/90	25/205
330/150	10/195	120/300	30/210	245/65	7/187
250/70	65/245	72/252	50/230	180/0	130/310
275/95	57/237	266/86	50/230	300/120	75/225
265/85	265/85	10/190	75/255	290/110	90/270
307/127	255/75	80/260	65/245	50/230	100/280
192/12	335/155	55/235	80/260	350/170	

Profile C-2

Planar to tangential cross beds

190	225	230	196	190	178
215	227	195	235	210	200
205	215	225	207	202	178
200	220	245	180	230	174
172	224	236	202	240	180
224	228	230	195	194	200
182	178	275	194	190	205
190	217	225	208	220	185
250	235	200	216	197	212
180	230	185	240	190	195
192	235	195	105	215	215
215	180	260	180		

Compound foresets

185	200	181	200	185	184
180	240	174	248	170	175
212	184				

Profile C-3

Planar to tangential cross beds

195	162	230	180	230	170
190	188	221	185	215	195
210	180	202	190	205	192
192	160	210	184	208	165
175	210	178	193	178	174
189	195	215	207	182	174
166	210	204	187	184	162
220	204	193	188		

Compound foresets

185	156	155	185
182	165	175	190

Profile C-4

Planar to tangential cross beds

88	92	105	55	47	95
84	270	35	48	75	285
75	60	92	65	90	35
120	100	70	60	78	100
75	55	89	65	70	110
42	52	50	78	5	70
65	45	81	58	65	65
92	54	49	32	110	0
90	10	86	71	35	65
35	80	90	65	90	5
140	110	45	70	30	35
84	75	35	80	79	350
79	300	70	115	55	70
84	180	70	105	91	65
76	70	81	345	85	55
155	90	110	80	104	25
90	80	90	80	134	20
120	105	45	70	105	45
50	85	90	96	110	78
85	100	87	62	15	55
42	80	101	44	20	55
70	85	65	125	0	85
62	90	66	100	20	65
68	80	72	85	120	70
70	42	56	70	97	55
72	40	56	75	99	58
80	120	80	101	105	52
80	18	77	120	65	63
78	70	92	90	82	74
72	55	80	135	125	70
66	25	95	80	90	55
74	30	78	80	40	45
71	75	25	85	180	25
65	40	110	75	50	75
75	120	81	246	78	70
62	42	50	95	50	65
94	25	58	185	30	85
5	50	55	115	66	70
60	10	75	70	50	45
75	100	86	80	100	93
90	30	90	80	68	76
55	105	81	60	18	106
55	31	78	77	45	17
45	35	75	2	7	66
34	90	79	90	64	86

Profile C-4

Planar to tangential cross beds, continued

30	95	105	120	34	35
105	75	81	120	5	120
125	90	77	64	70	102
105	68	90	92	30	75
48	50	75	80	45	37
117	5	18	150	110	19
337	50	80	80	36	34
95	31	32	85	44	78
155	79	24	91	57	80
335	115	23	85	6	78
18	50	12	80	75	64
65	118	86	30	46	86
40	81	94	70	80	45
62	25	22	55	107	68
110	115	20	110	10	66
60	87	20	80	46	95
270	9	15	70	60	54
60	90	20	78	130	90
100	77	10	93	39	58
25	80	58	80	114	55
91	76	10	88	155	135
54	131	357	85	85	85
95	65	30	87	52	81
93	42	60	50	212	224
72	22	114	70	75	235
15	65	69	85	85	199
75	85	26	66	220	228
112	80	55	60	270	229
28	45	61	70	110	223
72	35	55	350	69	210
78	30	86	90	53	211
42	80	45	30	263	213
52	160	60	51	251	228
9	0	99	35	256	235

Asymmetric and tangential cross beds

106	100	105	89	15	0
17	130	60	82	20	20
86	140	45	125	40	120
90	65	65	90	140	97
44	120	350	40	42	99
15	70	35	180	41	105
5	20	160	85	125	65
95	90	25	55	85	120
95	30	63	96	87	35
115	75	170	95	120	89
70	112	72	115	110	130
345	42	65	120	40	60
80	52	40	90	25	50
125	9	28	95	95	95
32	20	130			

Compound foresets

80	125	135	104	120	50
40	115	122	160	112	75
110	102	146	150	112	70
145	127	78	130	77	117
155	31	102	130	115	117
89	7				

Fine-grained facies

Tangential cross beds

46	86	212	10	60*	45*
78	30	75	29	72*	42*
94	105	80	357	35*	76*
92	50	39	55	45*	75*
80	21	56	56	100*	38*
125	62	30	0	35*	60*
5	42	43	35	55*	105*
88	102*	38*	80*	65*	
76*	98*	37*	69*		

Current ripples

15	90	55	78	100	112
85	87	75	93	23	93
70	82	71	92	65	105*
62	62	85	78*	70	90*

* Preserved bedform facies

Cumberland Falls

Planar to tangential cross beds

263	165	275	280	290	285
240	185	305	275	150	145
280	195	310	280	290	230
285	180	145	270	150	230
295	185	155	245	160	

Profile LRD-1

Giant Foresets

115	115	120	105	118	165*
124	122	105	124	164*	155*
102	115	100	117	165*	

* Intraset

Planar to tangential cross beds

197	169	205	142	185	210
210	180	142	166	237	180
170	190	150	110	205	140
169	180	150	130	255	242
175	155	93	234	230	110
185	190	90	110	256	135
185	180	185	150	235	245
145	198	110	140	218	231
179	280	105	109	175	210
185	182	190	250	188	163
125	190	145	190	175	222
195	105	220	135	224	205
250	232	150	220	150	110
203	180	250	145	225	155
223	247	187	130	140	

Compound foresets

70	58
115	75

Massive channel edge

85	80
100	115

Profile LRD-2

Planar and tangential cross beds

160	177	240	145	202	165
180	202	195	240	190	135
184	180	160	153	186	155
150	185	155	178	150	170
170	165	160	194	173	170
175	207	180	155	195	180
170	175	184	180	185	165
240	175	192	186		

Compound foresets

100	122	185
120	115	

Profile PM-151

Topset preserved compound foresets

86	118	111	133	142	167
88	149	119	100	104	111
88	141	98	148	134	47
150	135	89	109	112	123
130	118	121	162	125	127
155					

Planar to tangential cross beds

131	125	159	200	150	159
141	200	187	96	125	144
121	193	171	105	187	134
167	138	154	134	128	176
150	227	152	160		

Profile J

Planar to tangential cross beds

140	270	252	320	130	255
151	260	249	275	245	280
240	250	254	274	274	260

Profile PM147W

Planar to tangential cross beds

245	255	201	225	244	165
215	224	218	144	180	185
225	155	198	138	232	

Profile PM147E

Planar to tangential cross beds

235	185	220	215	200	198
248	165	238	232	194	

APPENDIX E: SANDSTONE PETROGRAPHY

SAMPLE	25ECK-2	25ECK-3	25ECK-4	25ECK-5	25ECK-5A
PROFILE	C-4	C-4	C-4	C-4	C-4
FRAMEWORK GRAINS					
Quartz monocrystalline	253	226	254	233	253
Quartz polycrystalline	13	24	11	11	17
K-feldspar	6	14	10	10	4
Plag-feldspar	1	0	0	0	0
Metamorphic rock frag.	23	31	20	45	24
Sedimentary rock frag.	1	2	1	0	0
Chert	0	0	0	0	0
Muscovite	3	3	1	0	1
Biotite/chlorite	0	0	0	1	0
Opaques + heavy mins	0	0	3	0	1
Total Counts	300	300	300	300	300`
QFL					
Total Quartz	89.56	84.18	89.53	81.61	90.60
Total Feldspar	2.36	4.71	3.38	3.34	1.34
Total Lithics	8.08	11.11	7.09	15.05	8.05
GRAIN SIZE	lM - uF	uM - uF	lM - lF	uF - lF	uM - lF

SAMPLE	25ECK-6A	25ECK-6B	25ECK-7	25ECK-9	25ECK-10
PROFILE	C-4	C-4	C-4	C-4	C-4
FRAMEWORK GRAINS					
Quartz monocrystalline	232	242	260	257	247
Quartz polycrystalline	14	13	7	14	7
K-feldspar	12	11	13	9	6
Plag-feldspar	0	0	0	1	0
Metamorphic rock frag	38	33	17	19	40
Sedimentary rock frag	2	0	1	0	0
Chert	0	0	0	0	0
Muscovite	2	0	2	0	0
Biotite/chlorite	0	0	0	0	0
Opaques + heavy mins	0	1	0	0	0
Total Counts	300	300	300	300	300
QFL					
Total Quartz	82.55	85.28	89.60	90.33	84.67
Total Feldspar	4.03	3.68	4.36	3.33	2.00
Total Lithics	13.42	11.04	6.04	6.33	13.33

GRAIN SIZE	uM - lF	lC - lF	uM - lF	UM - lF	lM - uVF

SAMPLE	25ECK-11	25ECK-13	25ECK-14	25ECK-15	25ECK-16A
PROFILE	C-4	C-4	C-4	C-4	C-4
FRAMEWORK GRAINS					
Quartz monocrystalline	258	261	259	270	268
Quartz polycrystalline	8	6	13	6	8
K-feldspar	6	7	4	5	10
Plag-feldspar	0	0	0	0	0
Metamorphic rock frag	28	24	22	16	14
Sedimentary rock frag	0	0	0	0	0
Chert	0	0	0	0	0
Muscovite	0	2	2	1	0
Biotite/chlorite	0	0	0	0	0
Opaques + heavy mins	0	0	0	2	0
Total Counts	300	300	300	300	300
QFL					
Total Quartz	88.67	89.60	91.28	92.93	92.00
Total Feldspar	2.00	2.35	1.34	1.68	3.33
Total Lithics	9.33	8.05	7.38	5.39	4.67
GRAIN SIZE	lM - lF	lC - lF	lC - lF	uM - lF	lC - lF

SAMPLE	25ECK-16B	25ECK-17A	25ECK-17B	25ECK-18	75SS-1
PROFILE	C-4	C-4	C-4	C-4	C-1
FRAMEWORK GRAINS					
Quartz monocrystalline	269	253	247	260	274
Quartz polycrystalline	11	4	8	9	11
K-feldspar	3	10	12	7	2
Plag-feldspar	0	0	0	0	0
Metamorphic rock frag	17	31	29	23	12
Sedimentary rock frag	0	0	0	1	0
Chert	0	0	0	0	1
Muscovite	0	2	2	0	0
Biotite/chlorite	0	0	0	0	0
Opaques + heavy mins	0	0	2	0	0
Total Counts	300	300	300	300	300
QFL					
Total Quartz	93.33	86.24	86.15	89.67	95.00
Total Feldspar	1.00	3.36	4.05	2.33	0.67
Total Lithics	5.67	10.40	9.80	8.00	4.33
GRAIN SIZE	uM - lF	lM - uVF	lM - uVF	lC - lF	lC - lF

SAMPLE	75SS-2	75SS-3	75SS-4	75SS-5	PPR-1A
PROFILE	C-1	C-1	C-1	C-1	LRD-1
FRAMEWORK GRAINS					
Quartz monocrystalline	269	264	254	253	227
Quartz polycrystalline	5	8	15	10	38
K-feldspar	3	5	5	10	13
Plag-feldspar	0	0	0	0	0
Metamorphic rock frag	21	21	25	27	19
Sedimentary rock frag	0	1	0	0	2
Chert	0	0	0	0	0
Muscovite	2	1	1	0	0
Biotite/chlorite	0	0	0	0	0
Opaques + heavy mins	0	0	0	0	0
Total Counts	300	300	300	300	299
QFL					
Total Quartz	91.95	90.97	89.97	87.67	88.63
Total Feldspar	1.01	1.67	1.67	3.33	4.35
Total Lithics	7.05	7.36	8.36	9.00	7.02
GRAIN SIZE	lM - lF	lM - uF	uM - uVF	uM - lF	uVC - lF

SAMPLE	PPR-1B	PPR-2	PPR-3	PPR-4	RC
PROFILE	LRD-1	LRD-1	LRD-1	LRD-1	LRD-2
FRAMEWORK GRAINS					
Quartz monocrystalline	257	263	245	251	256
Quartz polycrystalline	19	19	18	16	18
K-feldspar	4	2	9	11	6
Plag-feldspar	1	0	0	0	0
Metamorphic rock frag	16	15	27	18	18
Sedimentary rock frag	1	0	0	2	0
Chert	0	1	1	1	1
Muscovite	2	0	0	1	1
Biotite/chlorite	0	0	0	0	0
Opaques + heavy mins	0	0	0	0	0
Total Counts	300	300	300	300	300
QFL					
Total Quartz	92.62	94.00	87.67	89.30	91.64
Total Feldspar	1.68	0.67	3.00	3.68	2.01
Total Lithics	5.70	5.33	9.33	7.02	6.35
GRAIN SIZE	lC - lF	uC - lF	lM - uVF	uM - lF	lC - uF

SAMPLE	J/L-1A	J/L-1B	PM-151A	PM-151B	PM-147E
PROFILE	J	J	PM-151	PM-151	PM-147
FRAMEWORK GRAINS					
Quartz monocrystalline	281	276	281	268	264
Quartz polycrystalline	6	7	3	8	20
K-feldspar	0	0	1	3	0
Plag-feldspar	0	0	0	0	0
Metamorphic rock frag	12	16	12	19	12
Sedimentary rock frag	0	1	1	0	1
Chert	1	0	2	1	1
Muscovite	0	0	0	0	1
Biotite/chlorite	0	0	0	0	0
Opaques + heavy mins	0	0	0	1	1
Total Counts	300	300	300	300	300
QFL					
Total Quartz	95.67	94.33	94.67	92.31	95.30
Total Feldspar	0.00	0.00	0.33	1.00	0.00
Total Lithics	4.33	5.67	5.00	6.69	4.70
GRAIN SIZE	uC - lF	uC - lF	lM - lF	lM - uVF	lVC - uVF

MEAN VALUES

	Corbin	Rockcastle	Bee Rock
Total Quartz	88.89	90.64	95.00
Total Feldspar	2.60	2.56	0.00
Total Lithics	8.51	6.80	5.00

	Middlesboro	Lee-Total	Lee-Total*
Total Quartz	94.09	89.98	92.20
Total Feldspar	0.45	2.26	1.40
Total Lithics	5.46	7.76	6.40

* Mean for the Lee Formation recalculated using weighted averages of individual members.

VITA

Michael Charles Wizevich was born in Meriden, Connecticut on May 12, 1956 and graduated from The Cheshire Academy in June of 1974. He earned a B.S. in Material Science and Engineering with a concentration in Metallurgy and Manufacturing at Cornell University in 1979. Following a few years of employment in the aerospace industry, Michael returned to school at the University of Illinois at Chicago, completing a M.S. degree in 1985. The title of his thesis is *Heavy Mineral Distributions in the Lower William River and Lake Athabasca Sediments*. During the summer of 1988, he was employed by Standard Alaska Production Company (now BP Alaska) in Anchorage, Alaska.

Michael married Karen Lee Jamison on December 24, 1990. Upon completion of the Ph.D., Michael will begin a two year post-doctoral fellowship at Victoria University at Wellington, New Zealand.