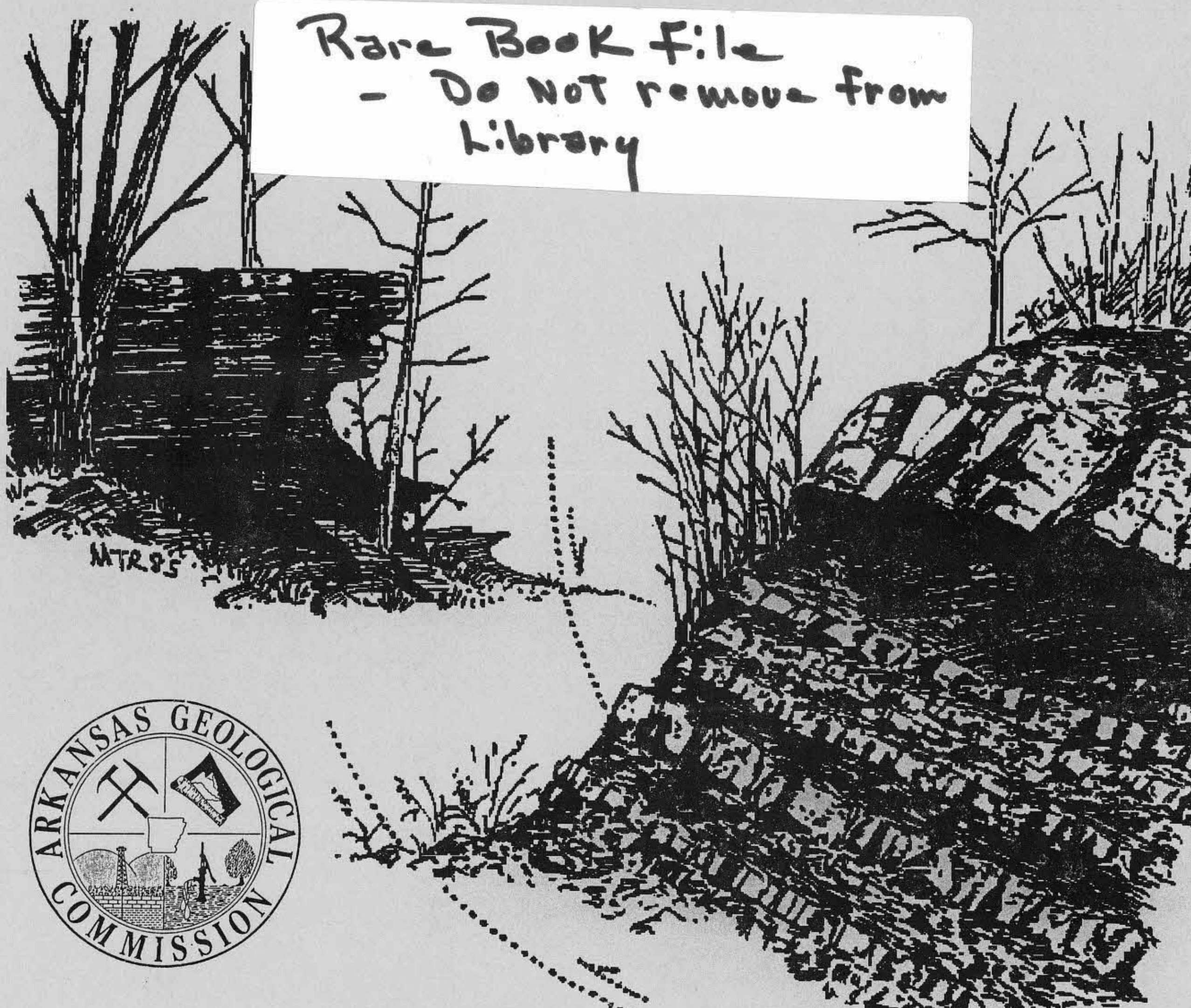
Guidebook to Carboniferous and Early Paleozoic Shelf and Basin Facies In Central Arkansas

By Charles G. Stone & John C. Meredith



ARKANSAS GEOLOGICAL COMMISSION

October 19-22, 1993

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GUIDEBOOK TO THE CARBONIFEROUS AND EARLY PALEOZOIC BASIN AND SOME SHELF FACIES IN CENTRAL ARKANSAS

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Charles G. Stone and John C. Meredith

INTRODUCTION

The field trip is scheduled for three days and consists of 28 stops (Fig. 1), several of which may be optional. The stops (Fig. 2) are designed to examine some seventeen Early to Middle Pennsylvanian submarine fan and other deep-water deposits; five Mississippian through Early Ordovician deep-water facies; three Middle Pennsylvanian deltaic progradations; two Ordovician and Devonian mostly pelagic "porcellanite" accumulations; and finally but not least - scenic thermal springs and tufa (Hot Springs) then onto the largest clear quartz crystal mine in North America.

The first day is a traverse north from the frontal Ouachita Mountains and "Maumelle chaotic belt" at Little Rock to the southeastern Arkoma basin at Conway and Morrilton, then return back to the south through the frontal Ouachitas to Perryville and finally proceeding to our destination in the central Ouachitas at Hot Springs. We will view respectively-deep (proximal) to shallow marine rocks in the morning and shallow to mostly proximal deep marine strata in the afternoon (Fig. 9). The second day is a transect to the Athens Plateau (Fig. 4) region of the southeastern Ouachita Mountains in the Lake DeGray area near Friendship, Arkadelphia, Bismarck, and Hollywood. The entire day is scheduled to examine the submarine fan sequence comprising the 6,000+ foot thick Jackfork Formation. To most recent investigators, these rocks have recognizable sheet-like clastic intervals of both upper and middle submarine fan environments of deposition. On day three, we will visit fine exposures of Lower Mississippian through Lower Ordovician (generally) pre-orogenic strata in the central Ouachita Mountains in the vicinity of Hot Springs, Mountain Pine, Blue Springs, and Crows. The classic exotic-bearing proximal fan-channel deposits in the Crystal Mountain Formation at the Ron Coleman quartz crystal mine will be a highlight.

We wish to express our sincere thanks to all the gracious individuals who have granted us permission to be on their property or who have been of assistance in other ways! Some of these people include: Hugh Durham, IV of International Paper Company, Kenny Hand of the Old Big Rock Quarry, Brian Westfall of the U.S. Corps of Engineers at Lake DeGray, and Randy Frazier at Pinnacle Mountain State Park.

Lastly, we wish to thank all the superb geologists who have helped to interpret these unique rock formations. Special thanks are due to the following: Boyd Haley, Rufus LeBlanc, Doug Jordan, Roger Slatt, Arnold Bouma, Martin Link, Fred Keller, Rod Tillman, Jim Coleman, Michael Roberts, Alan Thomson, Ernie Glick, Kaspar Arbenz, Lloyd Yeakel, Lewis Cline, Robert Morris, Tom Hendricks, and last but not least -- Hugh Miser!

ITINERARY FOR FIELD TRIP IN CENTRAL ARKANSAS

Charles G. Stone and John C. Meredith, October 19-22, 1993

<u>Tuesday, Oct. 19, 1993</u>	Arrive in Little Rock at the Holiday Inn West Holidome (501) 223-3000.
6:00 p.m.	Happy Hour and Orientation.
6:30 p.m.	Overview of Regional Geology and Depositional Facies Ouachita Mountains and Arkoma Basin (Stone).
<u>Wednesday, Oct 20, 1993</u>	Little Rock to Morrilton to Hot Springs, Arkansas.
Stop 1	Pinnacle Mountain State Park, Middle Jackfork, Canyon and Exotic Fill.
Stop 2	I-430 Roadcut, Little Rock, Upper Jackfork Downdip. Channel Fill and Levee.
Stop 3	Old Big Rock Quarry, North Little Rock, Upper Jackfork, Canyon and Channel Fill.
Stop 4	Old Park Hill Quarry, North Little Rock, Upper Jackfork, Channel Fill and Levee Deposits.
Stop 5	Roadcut near Levy, Middle Jackfork, Wildflysch-Slope Facies.
Stop 6	Jeffrey Quarry. Upper Jackfork, Channel Fill and Levee.
LUNCH	
Stop 7	Bayou Meto Anticline, Cabot, Lower Atoka, 1,000' Mid-fan Turbidites, Channels and Lobes.
Stop 8	Round Mountain Shale Pit, El Paso, Upper Atoka Coal Beds.
Stop 9	I-30 Roadcut, Conway, Middle Atoka Delta.
Stop 10	Morrilton Roadcuts, Four Upper Atoka Deltaic Exposures, Morrilton Anticline.
Stop 11	Roadcuts at Tom's Mountain (Perryville), Lower Atoka, Mid-fan Turbidites, Channels and Lobes.
Stop 12	Rock Shelter and Roadcut, Williams Junction, Middle Jackfork, Channel Fill.
Stop 13	Roadcut Highway #9, north of Paron, Lower Jackfork, Channel Fill and Slump mass.

HOLIDAY INN LAKE HAMILTON (501) 525-1891

Thursday, Oct. 21, 1993	Hot Springs to Lake DeGray, Hollywood, and Hot Springs.
Stop 14	Friendship Roadcut I-30, Upper Jackfork, Channels and Lobes.
Stop 15	Murray Quarry, Upper Jackfork, Channel Lenses to Sheet-like Anatomy.
Stop 16	Lake DeGray Spillway, Classic Exposures of Middle and Upper Jackfork, Fan Channels and Lobes.
Stop 17	Lake DeGray Intake, Downdip Correlations of Lake DeGray Spillway Section.
Stop 18	Lake DeGray Dam Powerhouse, Upper Jackfork, Channel Levee Complex.
Stop 19	DeRoche Ridge Roadcut, Initial Lowermost Jackfork Turbidite (Mid-fan) Deposits.
Stop 20	Hollywood Quarry, Upper Jackfork, Channels and Levees.

RETURN TO HOLIDAY INN LAKE HAMILTON IN HOT SPRINGS (501) 525-1891.

Friday, Oct. 22, 1993

Stop 21	Bypass around Hot Springs, Middle Stanley, Outer Fan Deposits.
Stop 22	Hot Springs National Park Tour, Thermal Springs, Bath Houses, Tufa, Zigzag Mountains, etc.
Stop 23	City Quarry, Hot Springs, Bigfork Chert, "Porcellanite" - Turbidite Facies.
Stop 24	Hot Springs Water Works, Hot Springs Sandstone Member, Channel Fill and Relationship to Novaculite ("Diatomite").
Stop 25	Blakely Mountain Dam, Blakely Sandstone, Channel and Levee Fill.
Stop 26	Ron Coleman's Quartz Crystal Mine, Crystal Mountain Sandstone, Exotic-bearing Canyon and Upper Fan Channel Fill.
Stop 27	Tripoli Mine, Malvern Minerals Company, Arkansas Novaculite Formation, Deep Marine Novaculite, Conglomerates and Carbonaceous Shales.
Stop 28	Roadcut near Crows, Arkansas, Mazarn Formation, Abyssal Plain Deposits.

PROCEED TO LITTLE ROCK AIRPORT

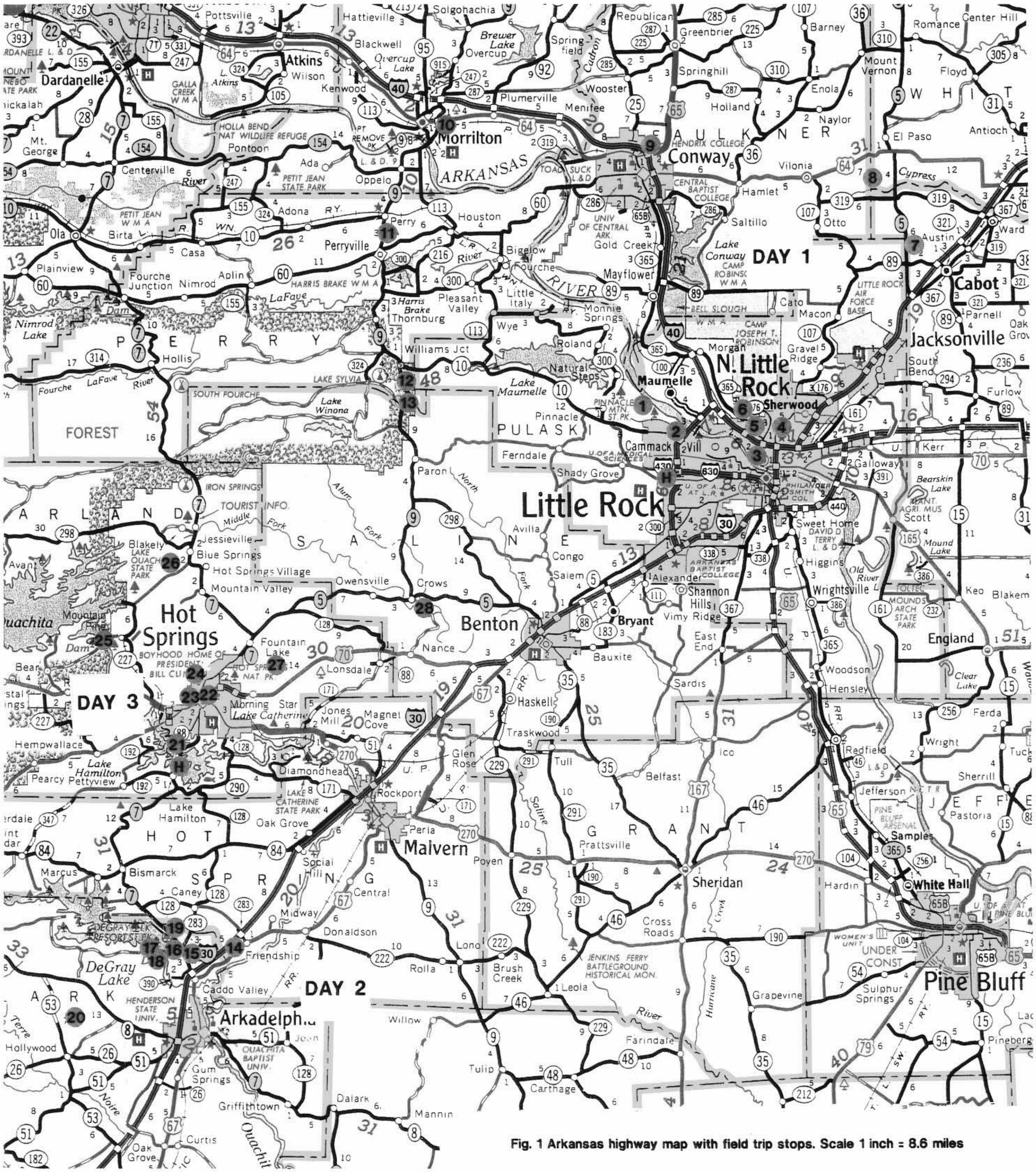


Fig. 1 Arkansas highway map with field trip stops. Scale 1 inch = 8.6 miles

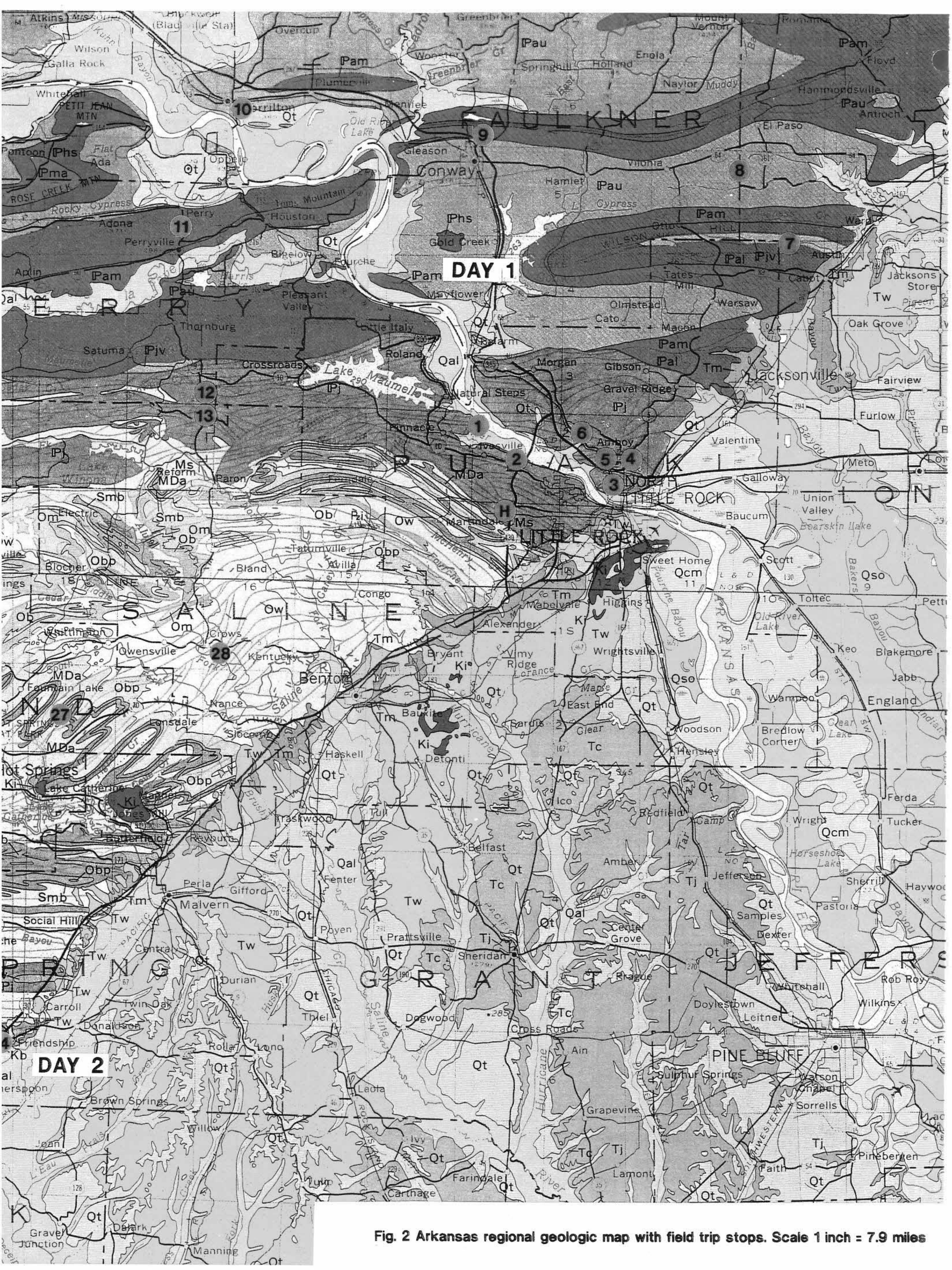
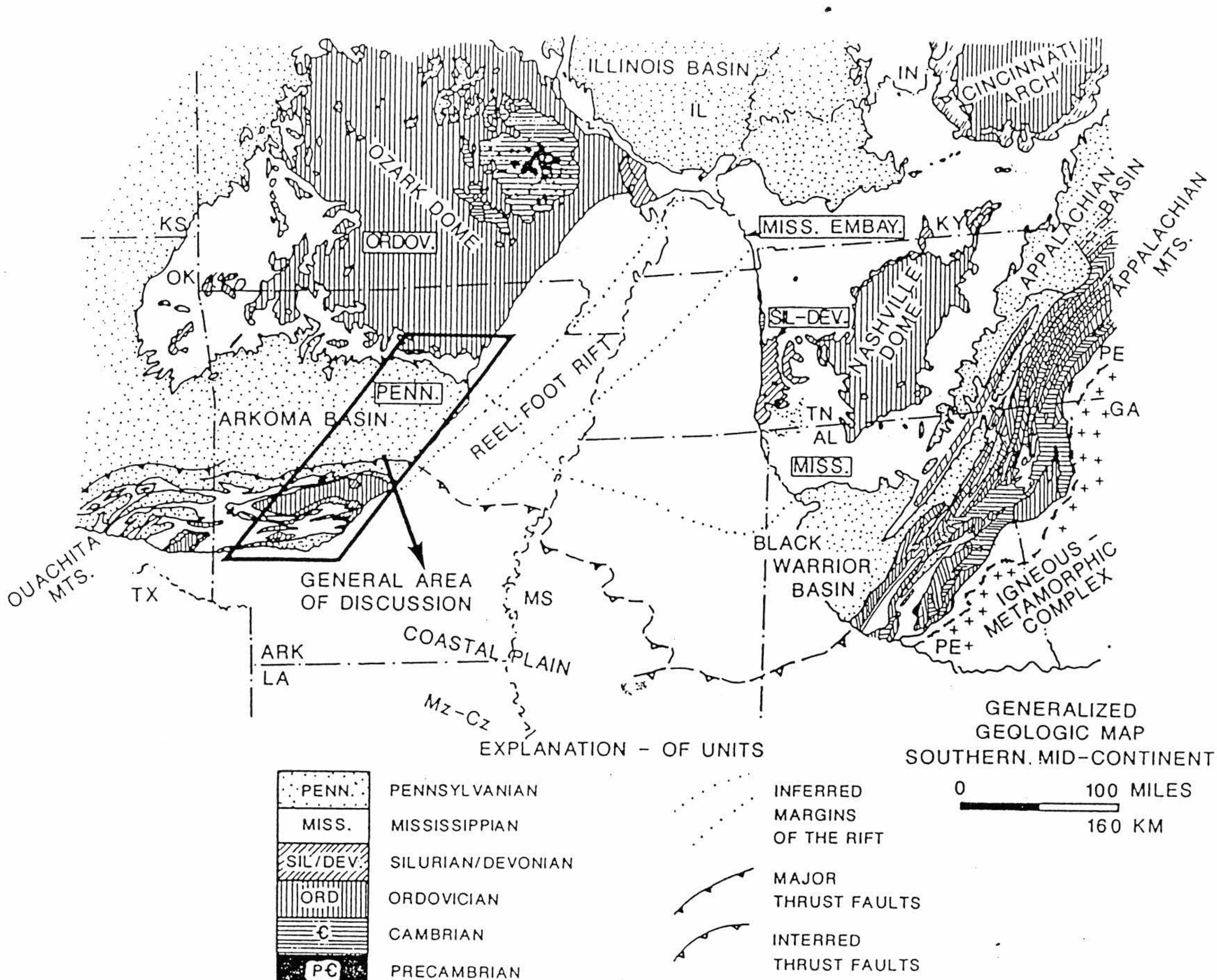


Fig. 2 Arkansas regional geologic map with field trip stops. Scale 1 inch = 7.9 miles



From Link and Roberts, 1986

Fig. 3

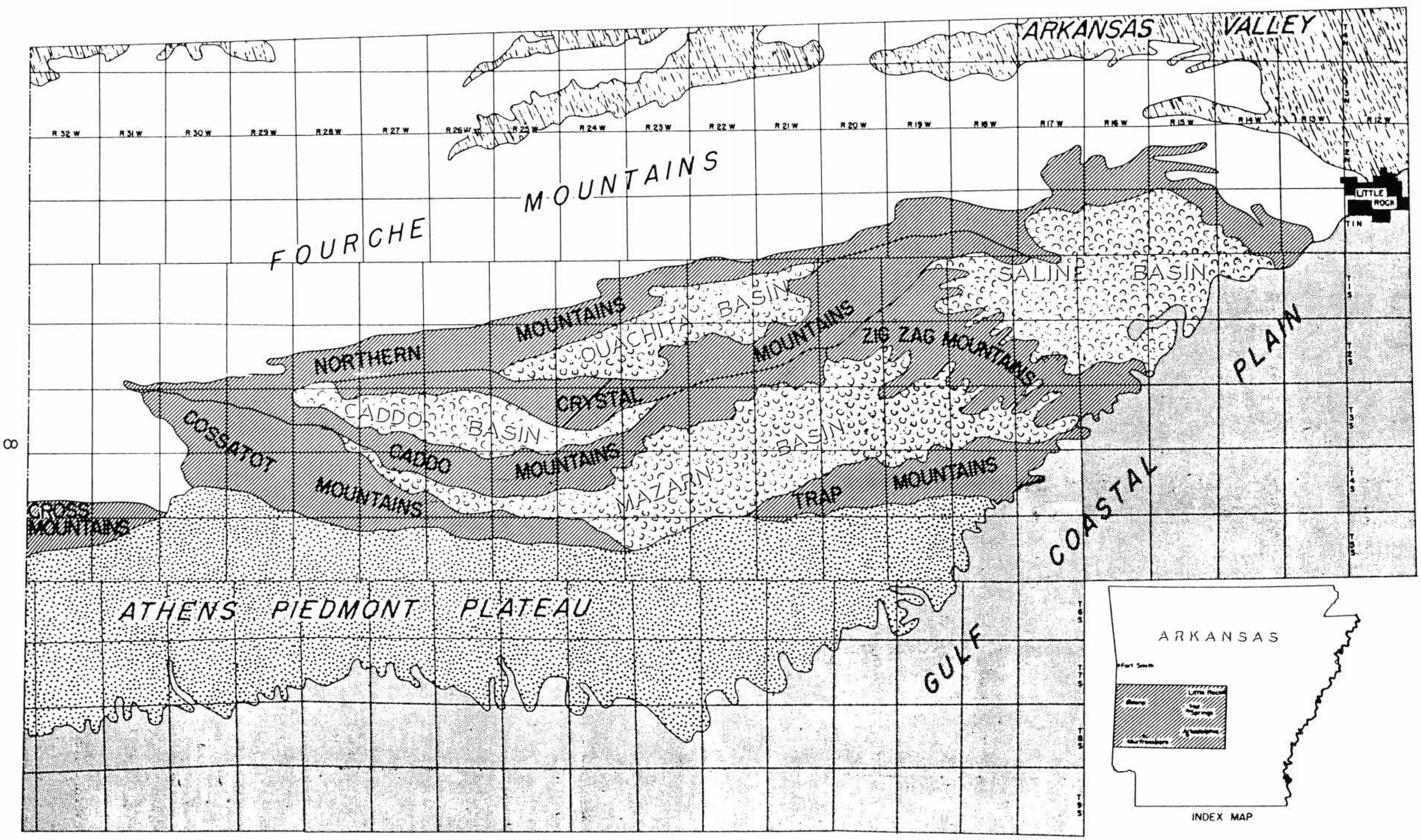


Figure 4. *Map showing major physiographic features in the Ouachita Mountain region of Arkansas.*

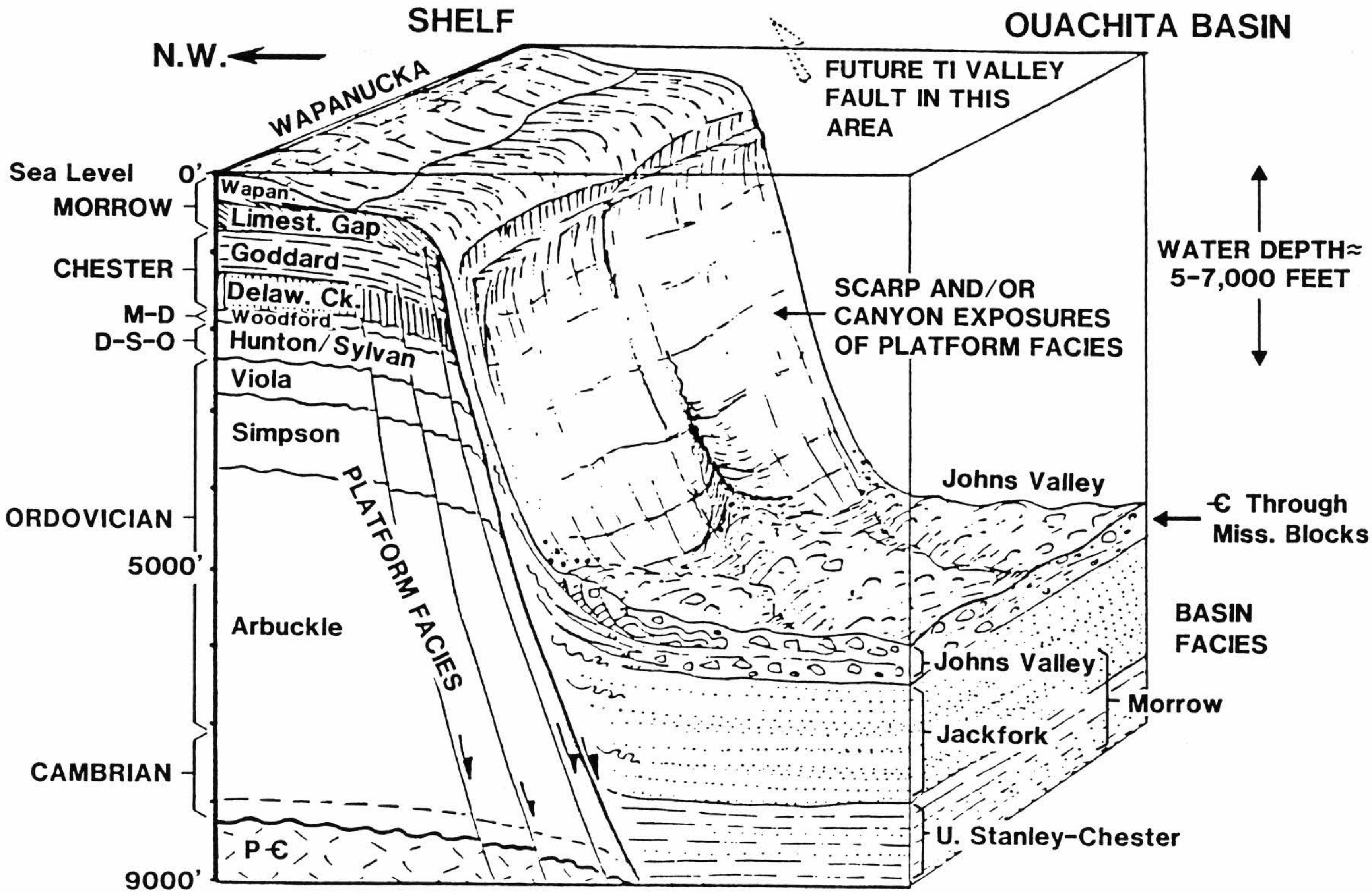


Figure **5**. North to south "correlation" of shallow and deep water formations of Missíssipian to Upper Morrow formations (M. Roberts, 1987).

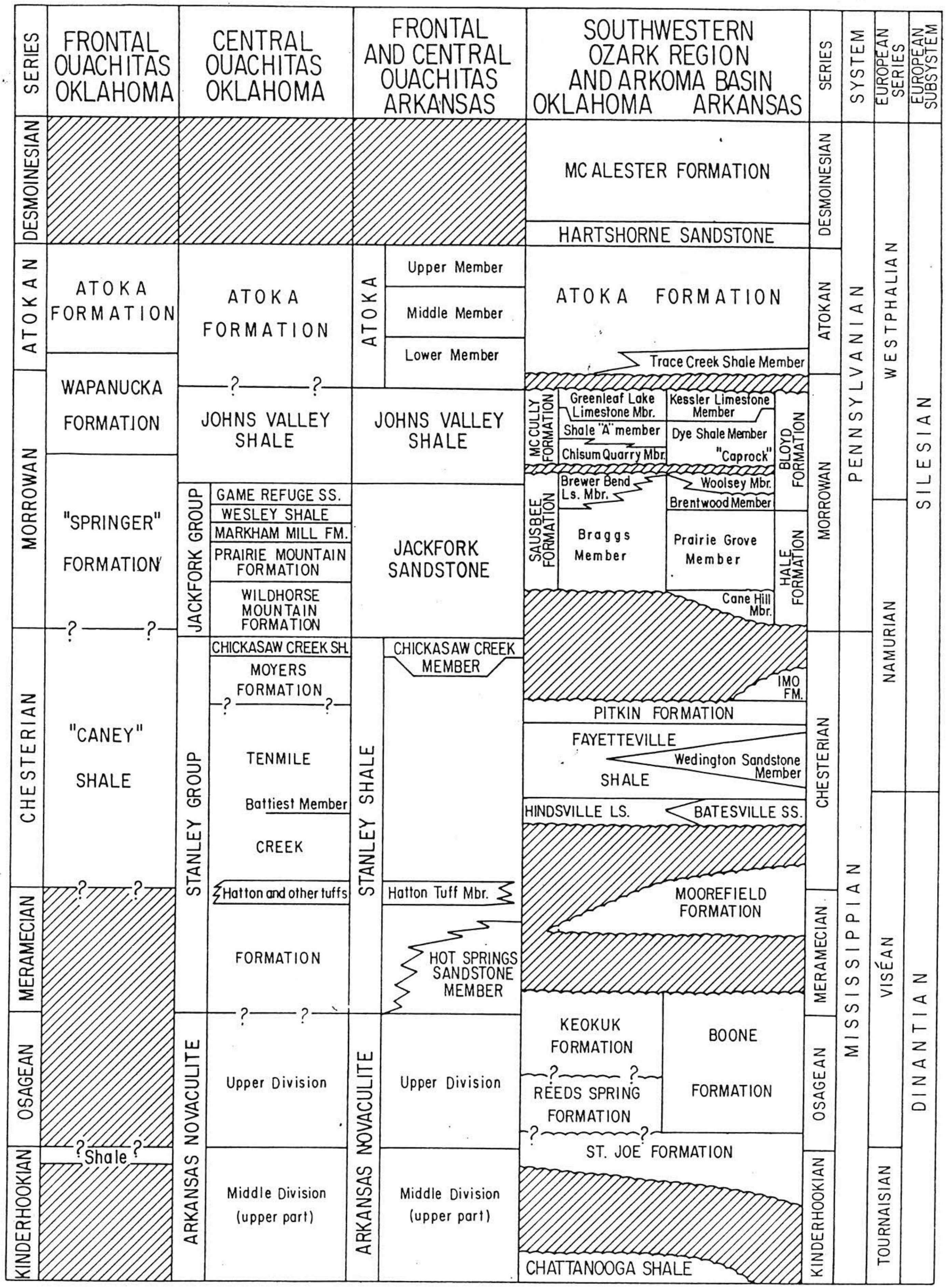
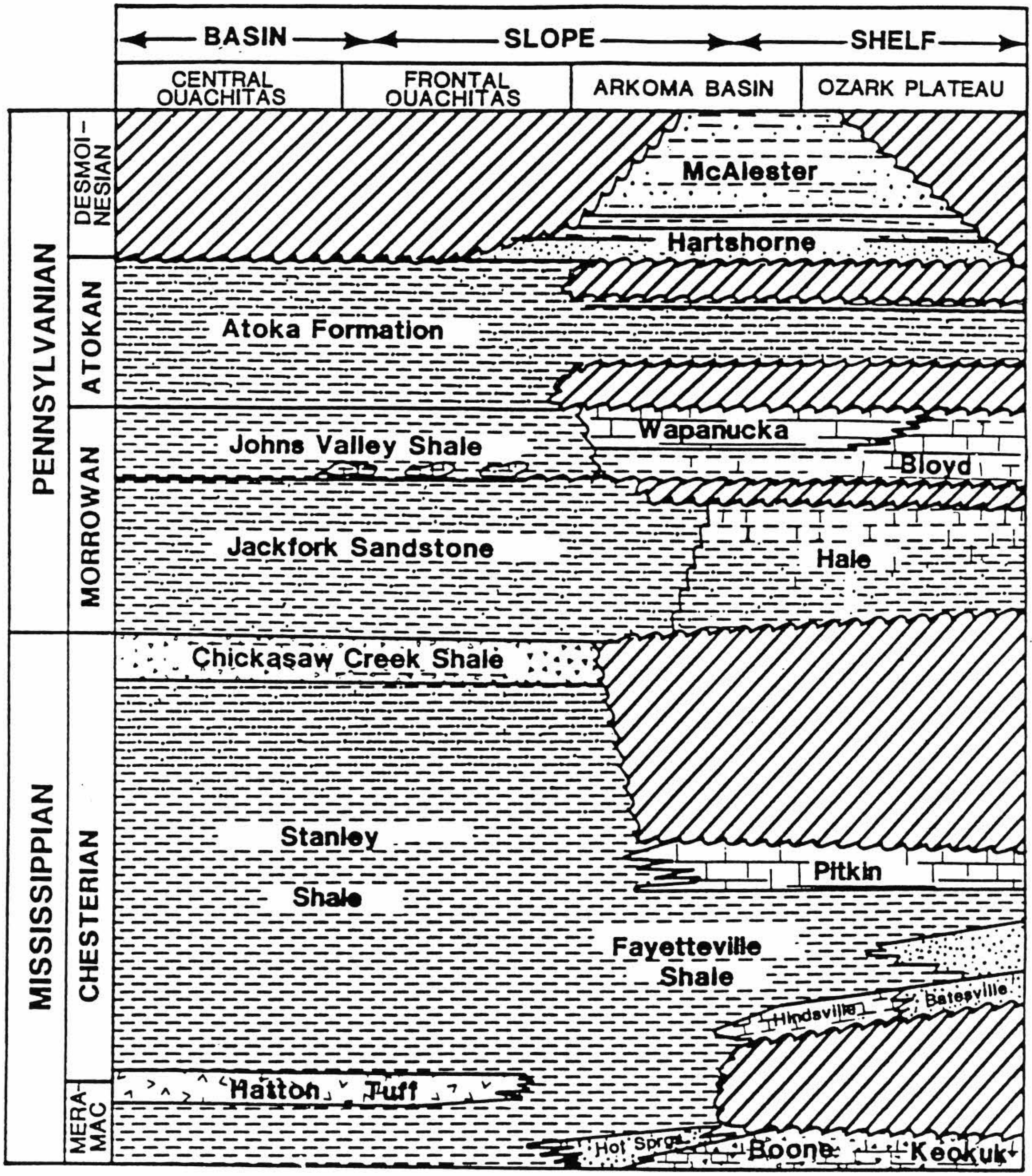

SERIES
FRONTAL OUACHITAS OKLAHOMA
CENTRAL OUACHITAS OKLAHOMA
FRONTAL AND CENTRAL OUACHITAS ARKANSAS
SOUTHWESTERN OZARK REGION AND ARKOMA BASIN
OKLAHOMA ARKANSAS
SERIES
SYSTEM
EUROPEAN SERIES
EUROPEAN SUBSYSTEM
DESMOINESIAN
ATOKAN
MORROWAN
CHESTERIAN
MERAMECIAN
OSAGEAN
KINDERHOOKIAN
MC ALESTER FORMATION
HARTSHORNE SANDSTONE
ATOKA FORMATION
ATOKA
Upper Member
Middle Member
Lower Member
ATOKA FORMATION
Trace Creek Shale Member
WAPANUCKA FORMATION
JOHNS VALLEY SHALE
JOHNS VALLEY SHALE
MC CULLY FORMATION
Greenleaf Lake Limestone Mbr.
Shale "A" member
Chisum Quarry Mbr.
Kessler Limestone Member
Dye Shale Member
"Caprock"
BLOYD FORMATION
"SPRINGER" FORMATION
JACKFORK GROUP
GAME REFUGE SS.
WESLEY SHALE
MARKHAM MILL FM.
PRAIRIE MOUNTAIN FORMATION
WILDHORSE MOUNTAIN FORMATION
JACKFORK SANDSTONE
SAUSBEE FORMATION
Brewer Bend Ls. Mbr.
Braggs Member
Woolsey Mbr.
Brentwood Member
Prairie Grove Member
Cane Hill Mbr.
HALE FORMATION
"CANEY" SHALE
STANLEY GROUP
CHICKASAW CREEK SH.
MOYERS FORMATION
TENMILE
Battiest Member
CREEK
Hatton and other tuffs
FORMATION
STANLEY SHALE
CHICKASAW CREEK MEMBER
Hatton Tuff Mbr.
HOT SPRINGS SANDSTONE MEMBER
IMO FM.
PITKIN FORMATION
FAYETTEVILLE
Wedington Sandstone Member
SHALE
HINDSVILLE LS.
BATESVILLE SS.
MOOREFIELD FORMATION
KEOKUK FORMATION
REEDS SPRING FORMATION
BOONE FORMATION
ST. JOE FORMATION
CHATTANOOGA SHALE
ARKANSAS NOVACULITE
Upper Division
Middle Division (upper part)
ARKANSAS NOVACULITE
Upper Division
Middle Division (upper part)
Shale
PENNSYLVANIAN
MISSISSIPPIAN
WESTPHALIAN
NAMURIAN
VISÉAN
TOURNAISIAN
SILESIAN
DINANTIAN

Fig. 6

OUACHITA CARBONIFEROUS STRATIGRAPHY



(adapted from Briggs 1974, and Sutherland and Manger, 1979)

Fig. 7

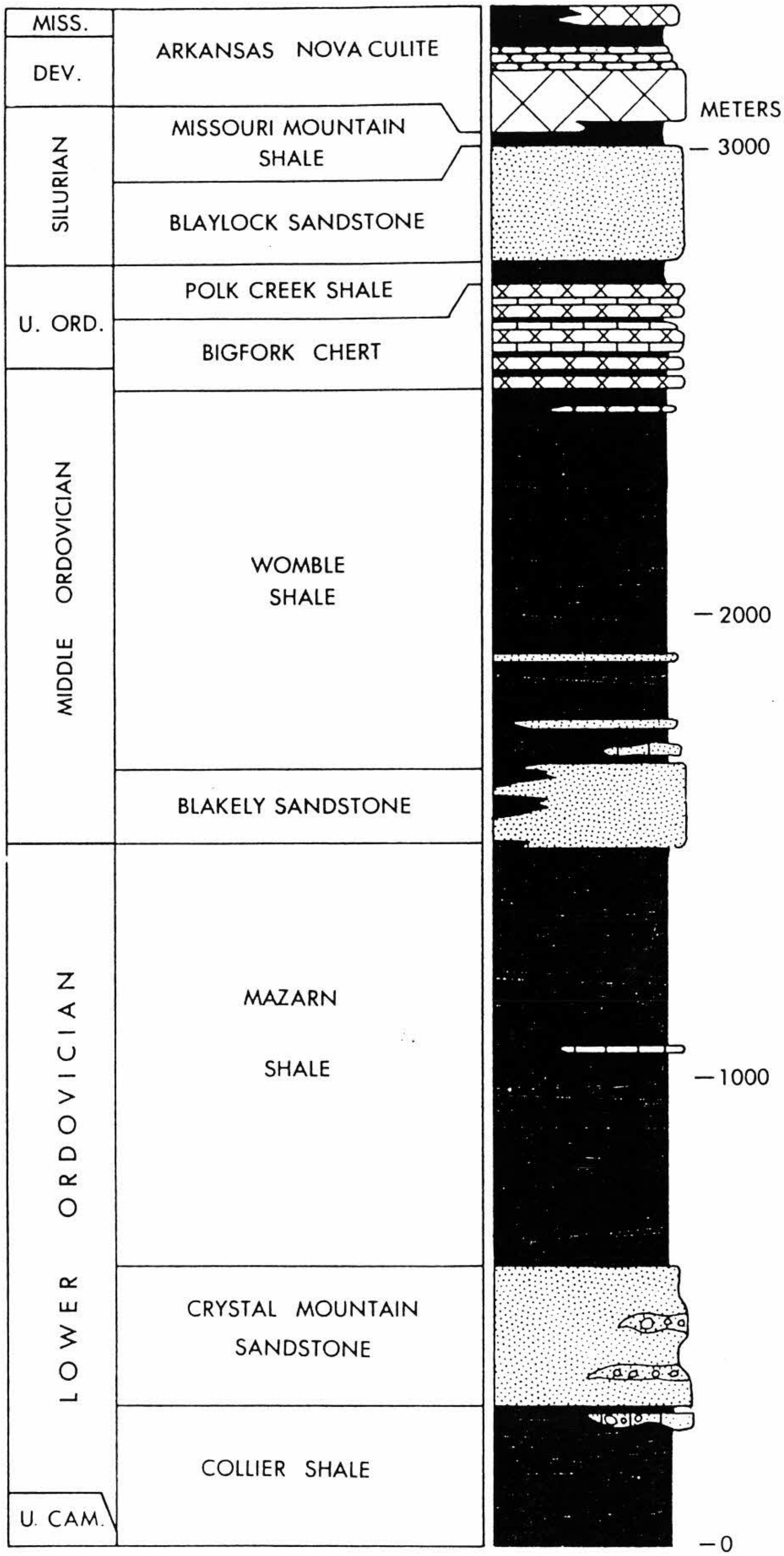


Figure 8. General stratigraphic column of pre-orogenic strata in the Ouachita Mountains. Lithologies and thicknesses represent units in the southern part of the Benton uplift, Arkansas. Revised from Lowe (1985).

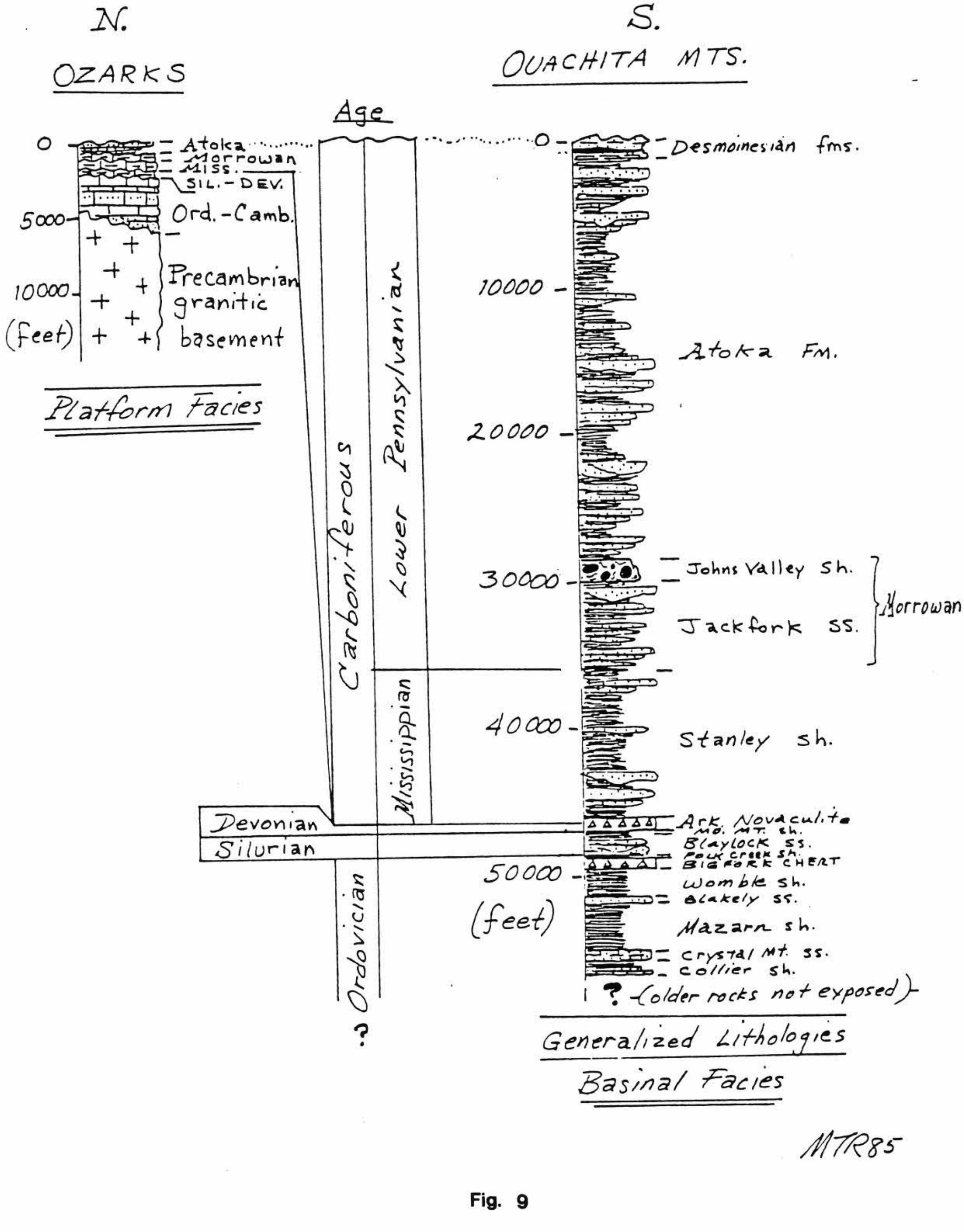

N.
OZARKS
S.
OUACHITA MTS.
Age
Atoka
Morrowan
MISS.
SIL.-DEV.
Ord.-Camb.
Precambrian granitic basement
0
5000
10000
(feet)
Platform Facies
Carboniferous
Lower Pennsylvanian
Mississippian
Devonian
Silurian
Ordovician
?
Desmoinesian fms.
Atoka Fm.
Johns Valley Sh.
Jackfork SS.
Morrowan
Stanley sh.
Ark. Novaculite
Mo. MT. sh.
Blaylock SS.
Polk Creek sh.
Bigfork CHERT
Womble sh.
Blakely SS.
Mazarn sh.
Crystal Mt. SS.
Collier sh.
? (older rocks not exposed)
0
10000
20000
30000
40000
50000
(feet)
Generalized Lithologies
Basinal Facies
MTR85

Fig. 9

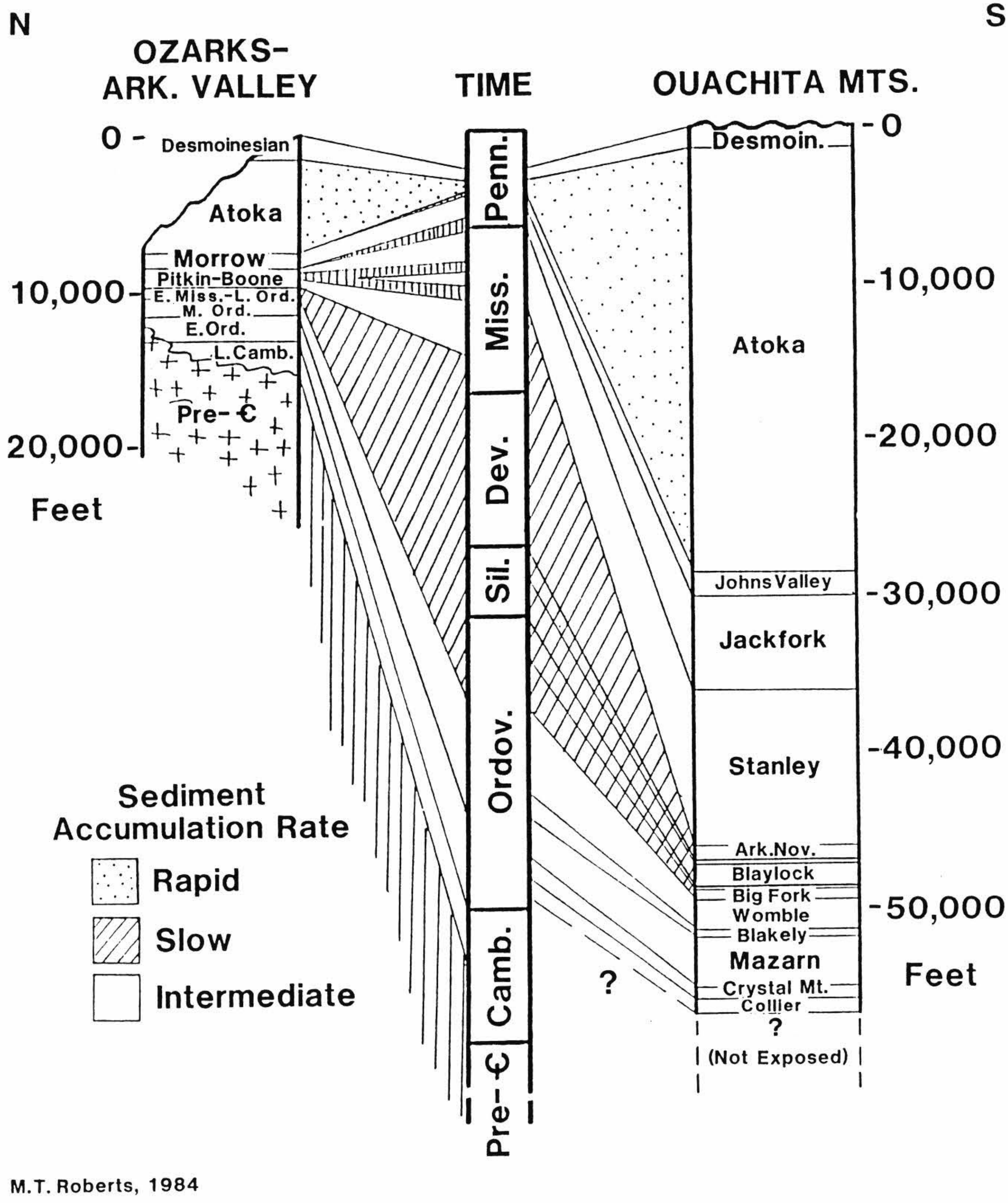

N
S
OZARKS-
ARK. VALLEY
TIME
OUACHITA MTS.
0 -
Desmoinesian
Atoka
Morrow
Pitkin-Boone
10,000-
E. Miss.-L. Ord.
M. Ord.
E.Ord.
L.Camb.
Pre-Є
20,000-
Feet
Penn.
Miss.
Dev.
Sil.
Ordov.
Camb.
Pre-Є
-0
Desmoin.
-10,000
Atoka
-20,000
Johns Valley
-30,000
Jackfork
-40,000
Stanley
Ark.Nov.
Blaylock
Big Fork
Womble
-50,000
Blakely
Mazarn
Crystal Mt.
Collier
?
(Not Exposed)
Feet
?
Sediment
Accumulation Rate
Rapid
Slow
Intermediate
M.T. Roberts, 1984

Fig. 10

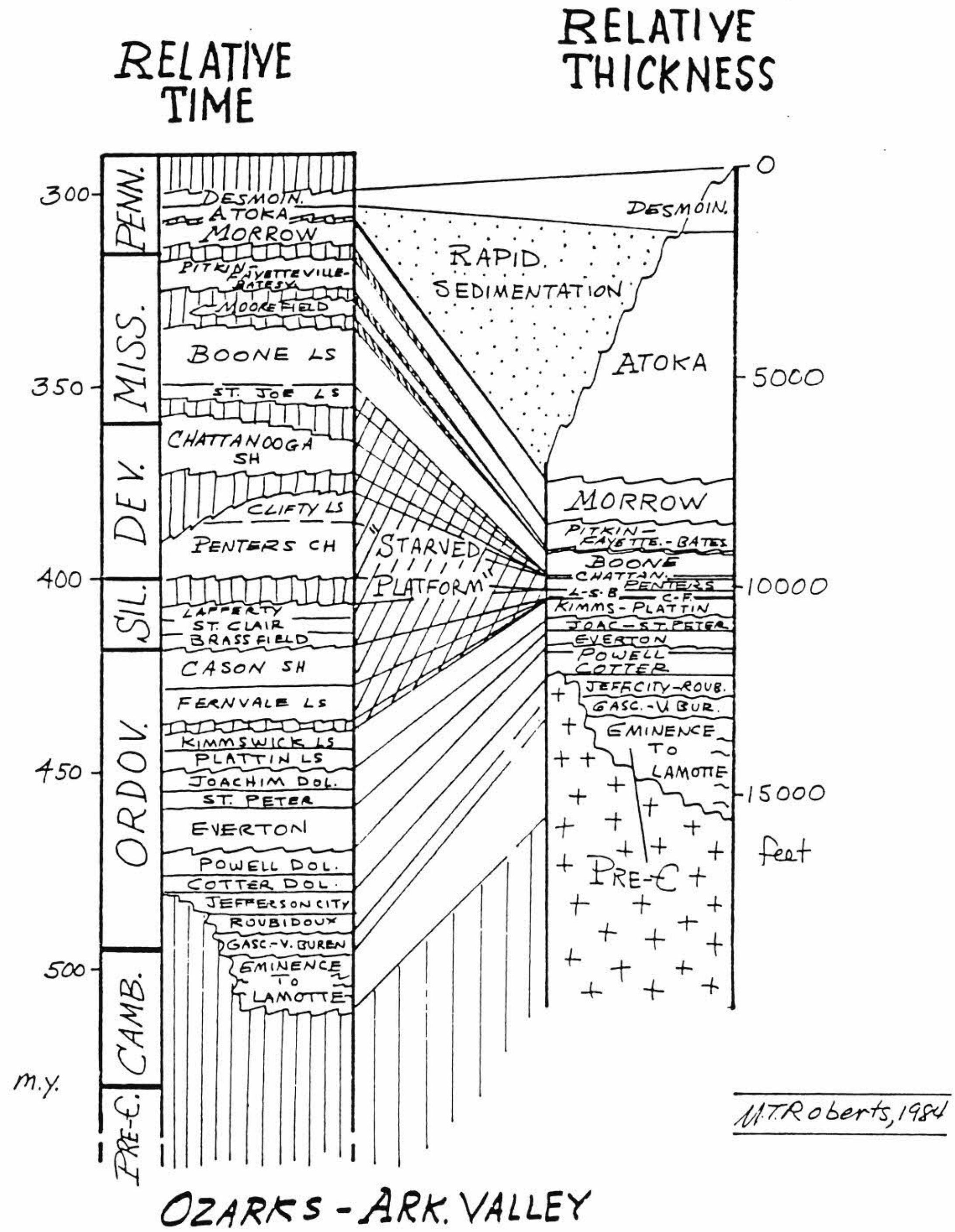

RELATIVE TIME
RELATIVE THICKNESS
300
350
400
450
500
m.y.
PENN.
MISS.
DEV.
SIL.
ORDOV.
CAMB.
PRE-Є.
DESMOIN.
ATOKA
MORROW
PITKIN-FAYETTEVILLE-BATESV.
MOOREFIELD
BOONE LS
ST. JOE LS
CHATTANOOGA SH
CLIFTY LS
PENTERS CH
LAFFERTY
ST. CLAIR
BRASSFIELD
CASON SH
FERNVALE LS
KIMMSWICK LS
PLATTIN LS
JOACHIM DOL.
ST. PETER
EVERTON
POWELL DOL.
COTTER DOL.
JEFFERSON CITY
ROUBIDOUX
GASC.-V. BUREN
EMINENCE TO LAMOTTE
RAPID SEDIMENTATION
STARVED PLATFORM
0
5000
10000
15000
feet
DESMOIN.
ATOKA
MORROW
PITKIN-FAYETTE.-BATES
BOONE
CHATTAN.
PENTERS
L-S-B
C-F
KIMMS-PLATTIN
JOAC-ST. PETER
EVERTON
POWELL
COTTER
JEFF.CITY-ROUB.
GASC.-V. BUR.
EMINENCE TO LAMOTTE
PRE-Є
M.T.Roberts, 1984
OZARKS - ARK. VALLEY

Fig. 11

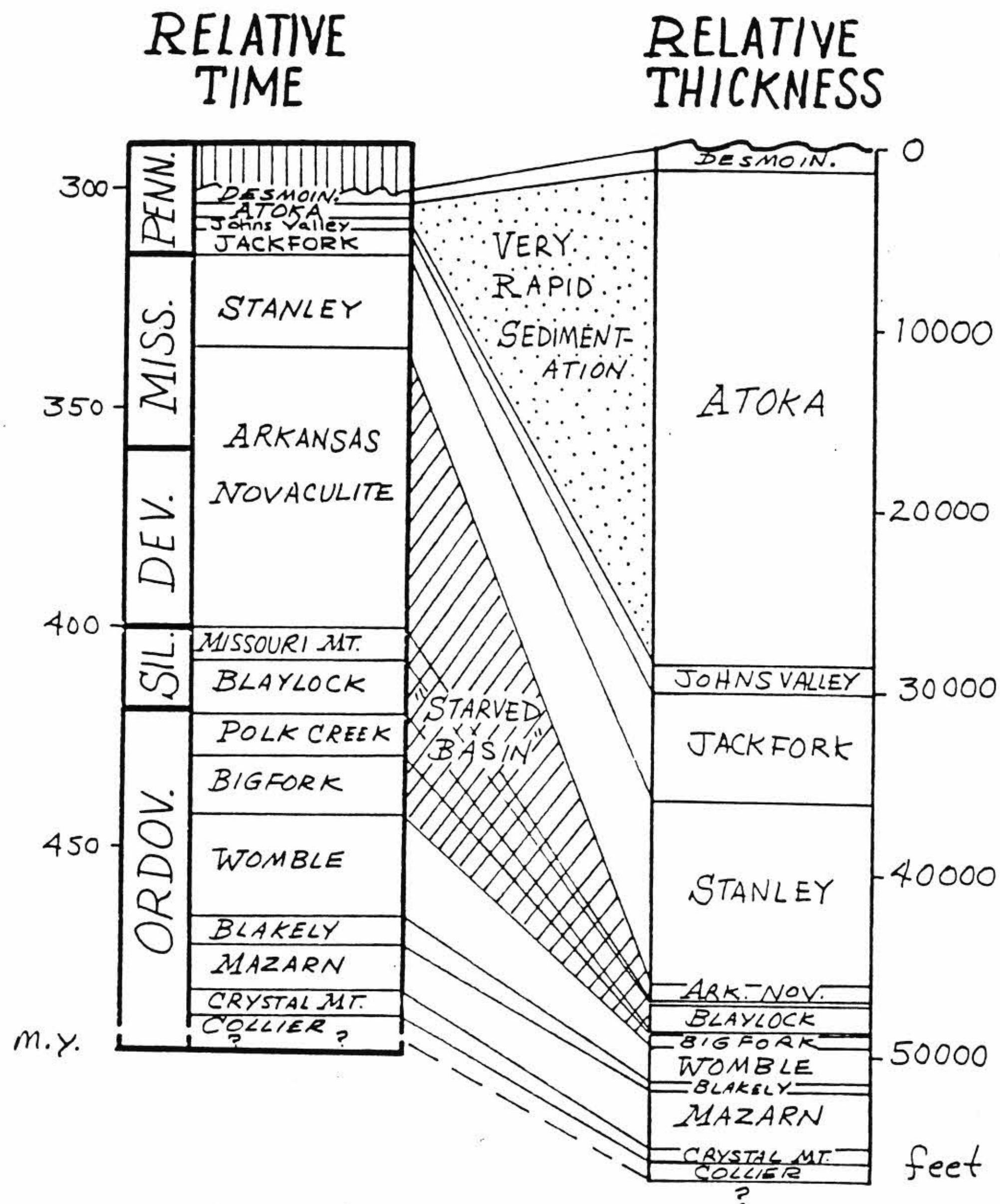
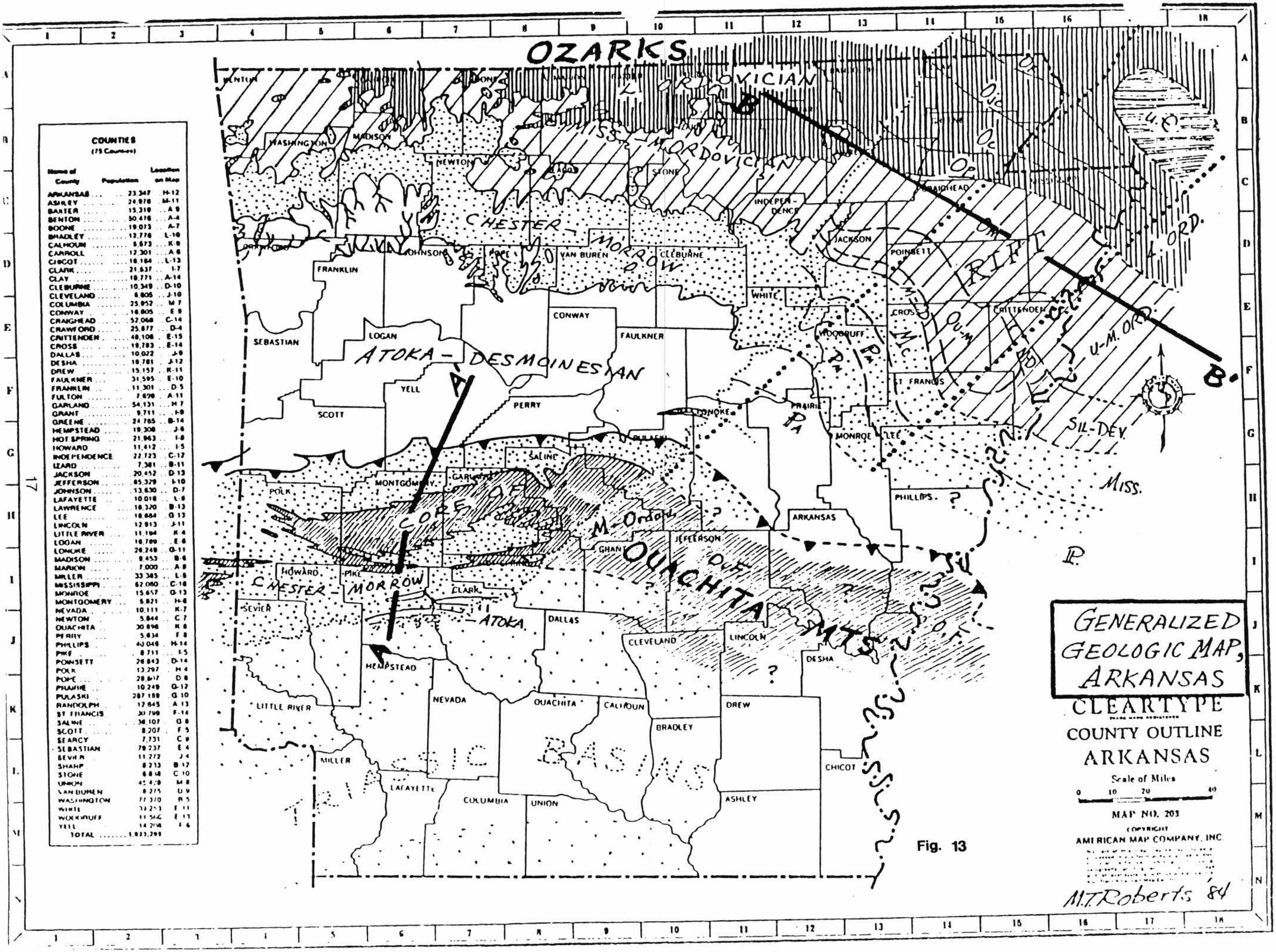


Fig. 12 OUACHITA MTS.

M.T. Roberts, 1984
(Revised from MTR 1980; Thickness data from Stone & Haley, 1984; time scaling reflects new numerical ages in Odin, 1984)



COUNTIES
(75 Counties)

Name of County	Population	Location on Map
ARKANSAS	23,347	H-12
ASHLEY	24,976	M-11
BAXTER	15,319	A-9
BENTON	50,476	A-4
BOONE	19,073	A-7
BRADLEY	12,778	L-10
CALHOUN	5,573	K-9
CARROLL	12,301	A-6
CHICOT	18,164	L-13
CLARK	21,537	I-7
CLAY	18,771	A-14
CLEBURNE	10,349	D-10
CLEVELAND	6,605	J-10
COLUMBIA	25,952	M-7
CONWAY	16,805	E-9
CRAIGHEAD	52,068	C-14
CRAWFORD	25,677	D-4
CRITTENDEN	48,106	E-15
CROSS	19,783	E-14
DALLAS	10,022	J-9
DESHA	18,761	J-12
DREW	15,157	K-11
FAULKNER	31,595	E-10
FRANKLIN	11,301	D-5
FULTON	7,699	A-11
GARLAND	54,131	H-7
GRANT	9,711	I-9
GREENE	24,765	B-14
HEMPSTEAD	19,308	J-6
HOT SPRING	21,963	I-8
HOWARD	11,412	I-5
INDEPENDENCE	22,723	C-12
IZARD	7,381	B-11
JACKSON	20,452	D-13
JEFFERSON	85,329	I-10
JOHNSON	13,630	D-7
LAFAYETTE	10,018	L-6
LAWRENCE	16,320	B-13
LEE	18,884	G-13
LINCOLN	12,913	J-11
LITTLE RIVER	11,194	K-4
LOGAN	16,789	E-6
LONOKE	26,249	G-11
MADISON	9,453	B-6
MARION	7,000	A-8
MILLER	33,385	L-5
MISSISSIPPI	62,060	C-16
MONROE	15,657	G-13
MONTGOMERY	5,821	H-6
NEVADA	10,111	K-7
NEWTON	5,844	C-7
OUACHITA	30,896	K-8
PERRY	5,634	F-8
PHILLIPS	40,046	H-14
PIKE	8,711	I-5
POINSETT	26,843	D-14
POLK	13,297	H-4
POPE	28,607	D-8
PRAIRIE	10,249	G-12
PULASKI	287,189	G-10
RANDOLPH	12,645	A-13
ST FRANCIS	30,799	F-14
SALINE	36,107	G-8
SCOTT	8,207	F-5
SEARCY	7,731	C-9
SEBASTIAN	79,237	E-4
SEVIER	11,272	J-4
SHARP	8,233	B-12
STONE	6,838	C-10
UNION	45,428	M-8
VAN BUREN	8,275	D-9
WASHINGTON	77,370	B-5
WHITE	39,253	E-11
WOODRUFF	11,566	E-13
YELL	14,208	F-6
TOTAL	1,923,295	

Fig. 13

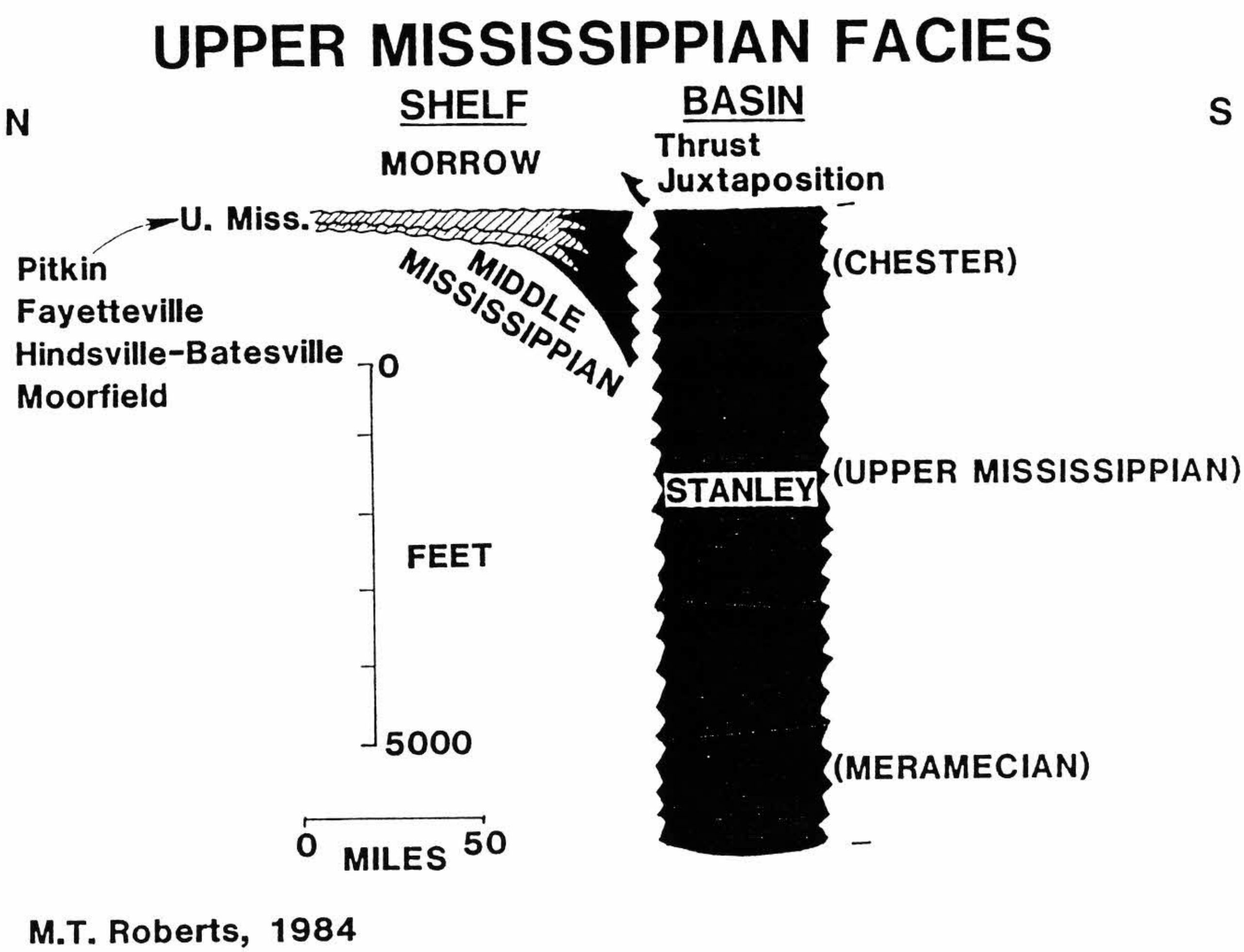

UPPER MISSISSIPPIAN FACIES
N
SHELF
BASIN
S
MORROW
Thrust
Juxtaposition
U. Miss.
Pitkin
Fayetteville
Hindsville-Batesville
Moorfield
MIDDLE
MISSISSIPPIAN
(CHESTER)
0
STANLEY
(UPPER MISSISSIPPIAN)
FEET
5000
(MERAMECIAN)
0
MILES
50
M.T. Roberts, 1984

Fig. 14

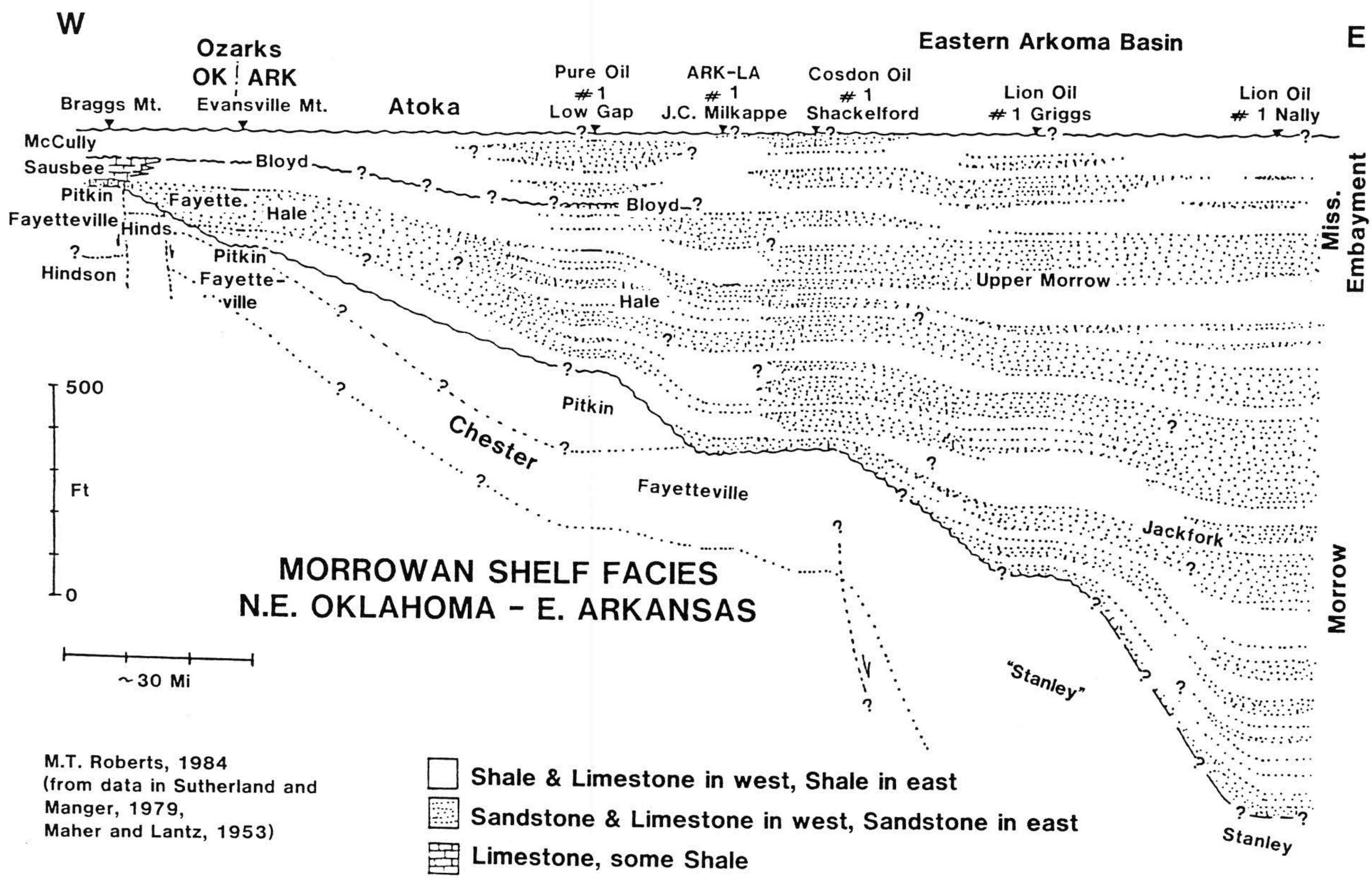
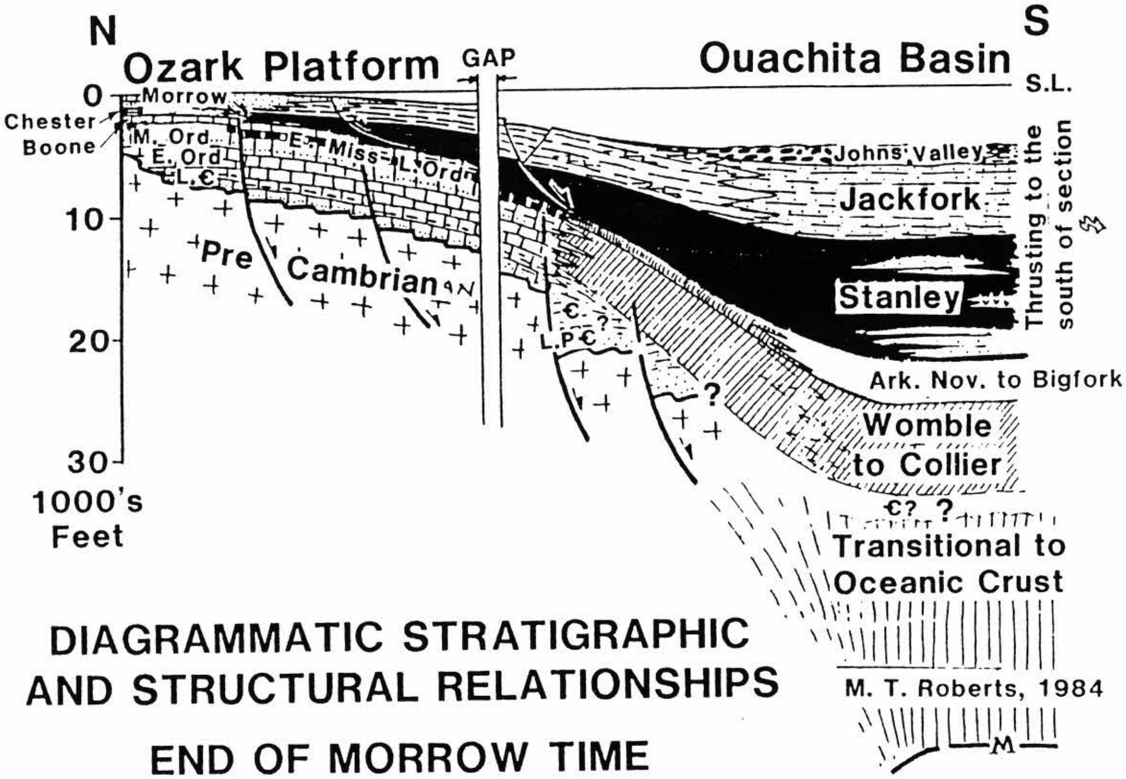

W
Ozarks
OK ¦ ARK
Eastern Arkoma Basin
E
Braggs Mt.
Evansville Mt.
Atoka
Pure Oil
#1
Low Gap
ARK-LA
1
J.C. Milkappe
Cosdon Oil
1
Shackelford
Lion Oil
#1 Griggs
Lion Oil
1 Nally
McCully
Sausbee
Pitkin
Fayetteville
Hindson
Bloyd
Fayette.
Hale
Hinds.
Pitkin
Fayette-ville
Bloyd
Hale
Upper Morrow
Miss. Embayment
500
Ft
0
~30 Mi
Chester
Pitkin
Fayetteville
Jackfork
Morrow
"Stanley"
Stanley
MORROWAN SHELF FACIES
N.E. OKLAHOMA - E. ARKANSAS
M.T. Roberts, 1984
(from data in Sutherland and Manger, 1979,
Maher and Lantz, 1953)
Shale & Limestone in west, Shale in east
Sandstone & Limestone in west, Sandstone in east
Limestone, some Shale

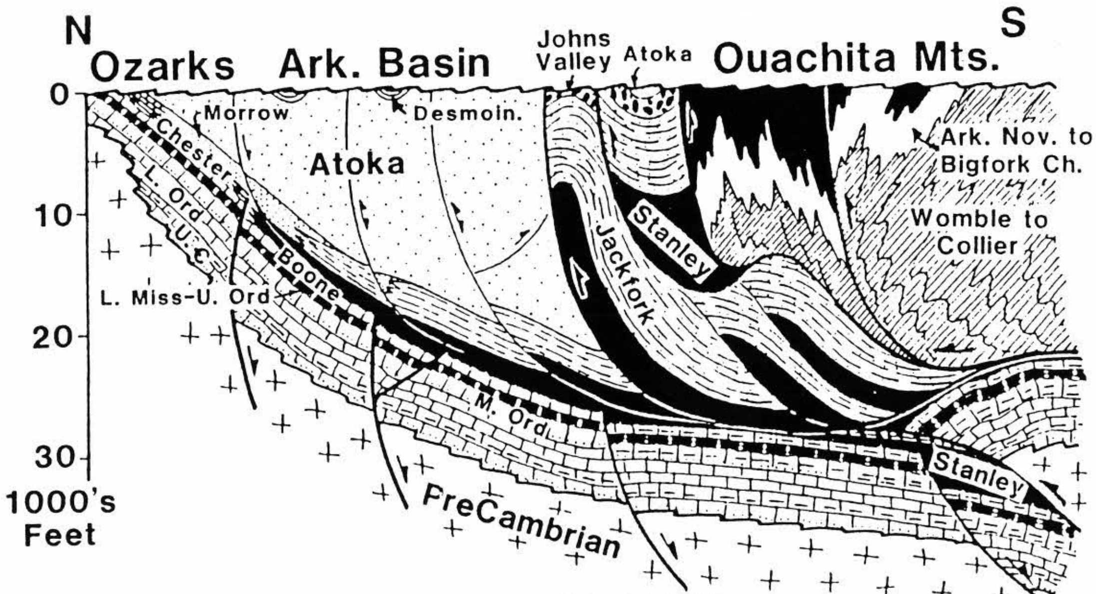
Fig. 15



DIAGRAMMATIC STRATIGRAPHIC AND STRUCTURAL RELATIONSHIPS

END OF MORROW TIME

M. T. Roberts, 1984

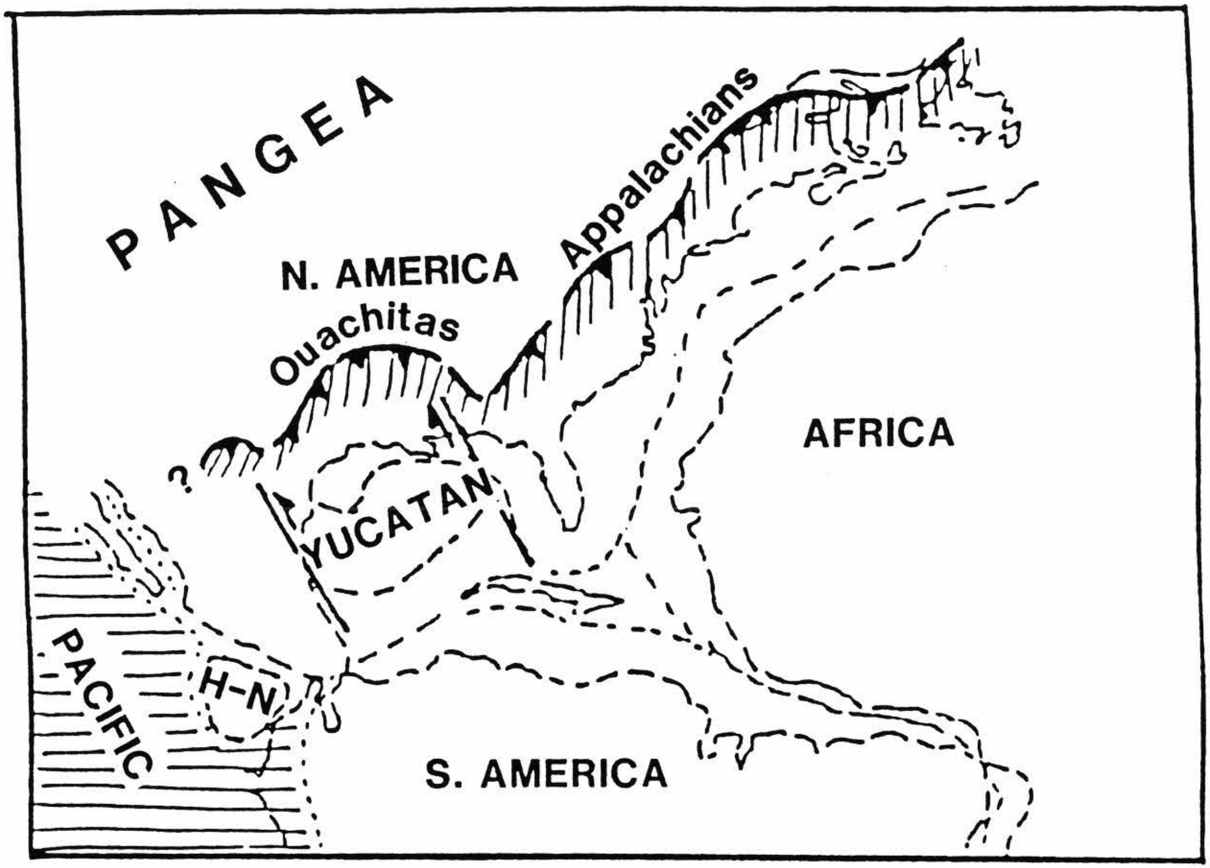


DIAGRAMMATIC STRATIGRAPHIC AND STRUCTURAL RELATIONSHIPS

CENTRAL ARKANSAS

M. T. Roberts, 1984

Fig. 16



A COLLISION MODEL FOR ORIGIN OF OUACHITA MOUNTAINS

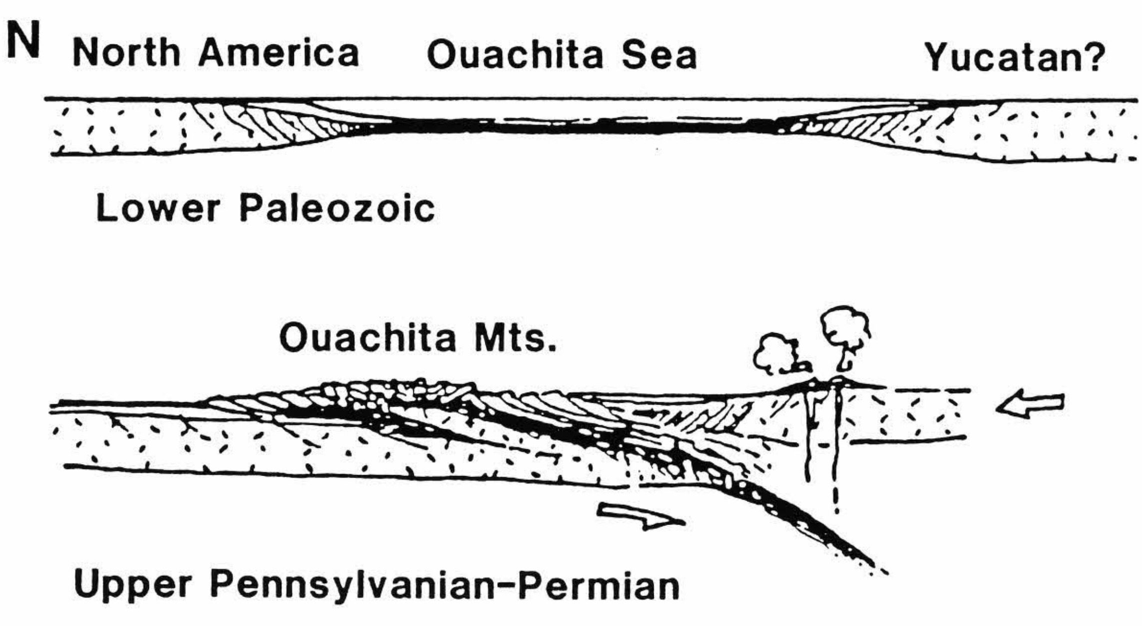


Figure 17. Collision model for origin of Ouachita Mountains (M. Roberts, 1987). See text for discussion of figure.

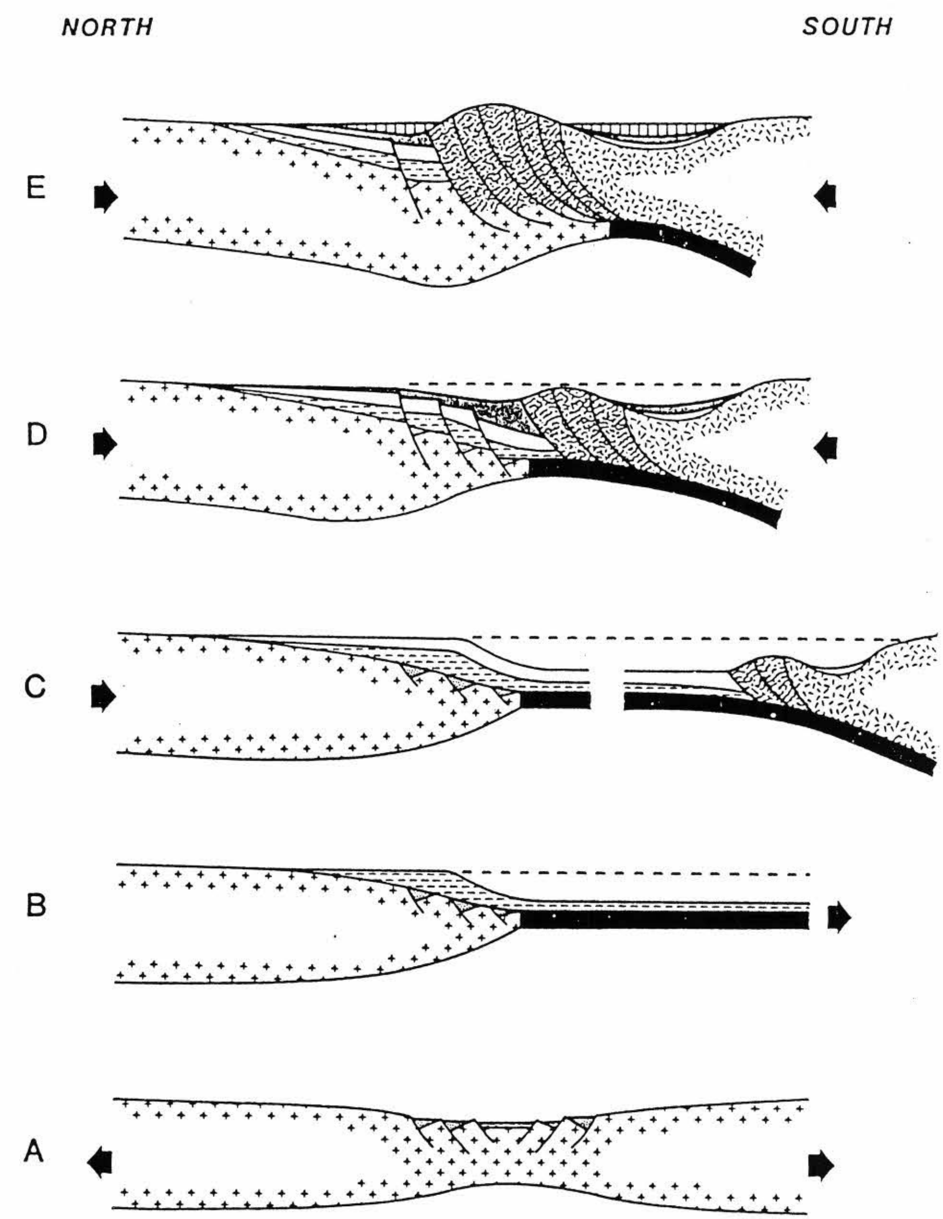


Figure **18.** Hypothetical cross-sections depicting the tectonic evolution of the southern margin of North America during:

A. late Precambrian--earliest Paleozoic

B. late Cambrian--earliest Mississippian

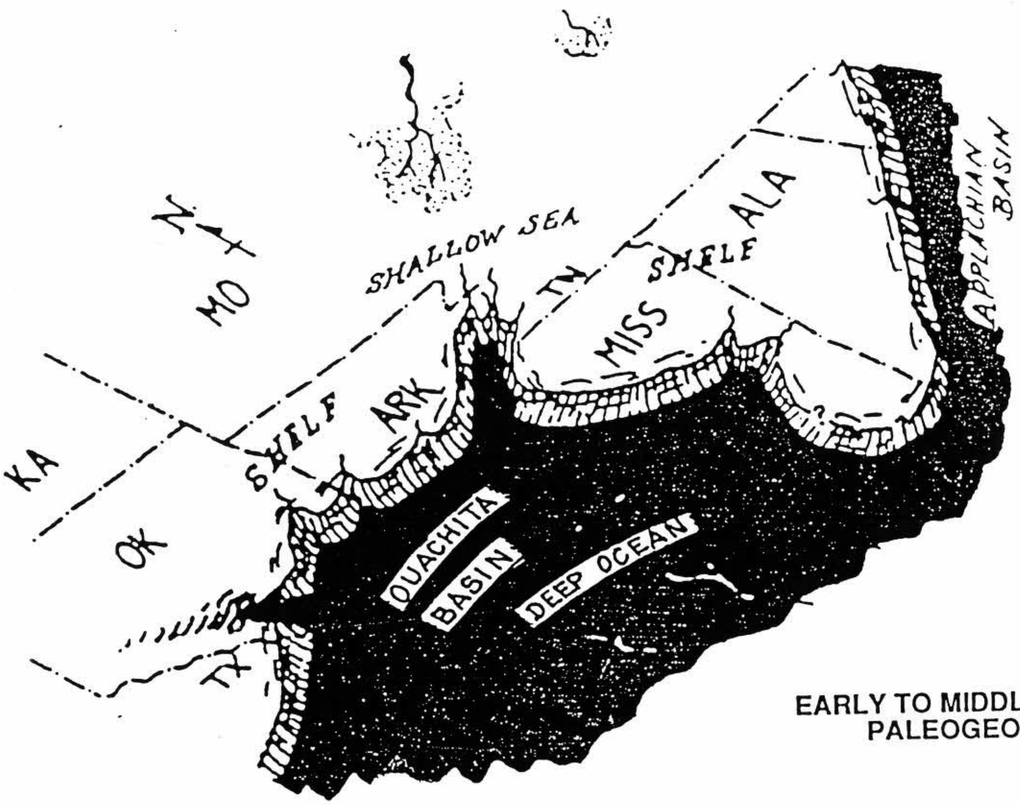
C. early Mississippian--earliest Atokan

D. early Atokan--middle Atokan

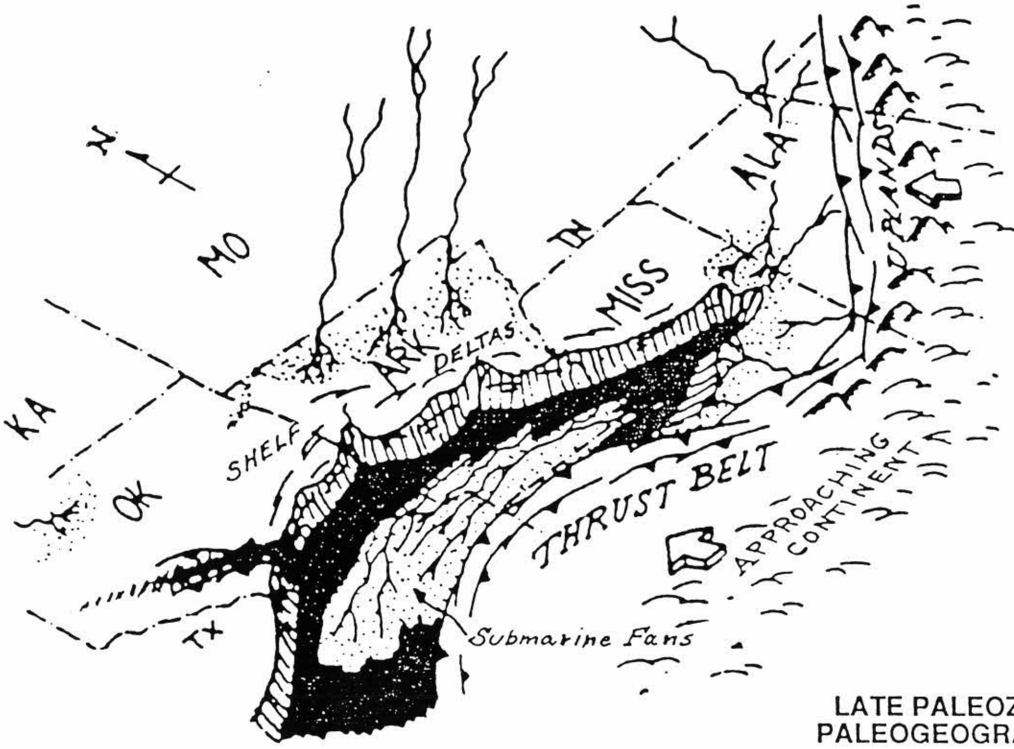
E. late Atokan--Desmoinesian

Key to patterns: crosses = continental crust (undifferentiated in A, North American in B-E); straw hatches = "Llanorian" crust; black = oceanic crust; sand = earliest Paleozoic strata; shale = late Cambrian-Devonian strata; white = Mississippian-earliest Atokan strata; shaded = early-middle Atokan strata; vertical lines = late Atokan-Desmoinesian strata; mottled = imbricated Paleozoic strata forming subduction complex; black triangle = magmatic arc volcanoes.

From Houseknecht and Kacena, 1983



EARLY TO MIDDLE PALEOZOIC
PALEOGEOGRAPHY



LATE PALEOZOIC
PALEOGEOGRAPHY

Fig. 19

From Link and Roberts, 1986

STANLEY (MISSISSIPPIAN) PALEOGRAPHY

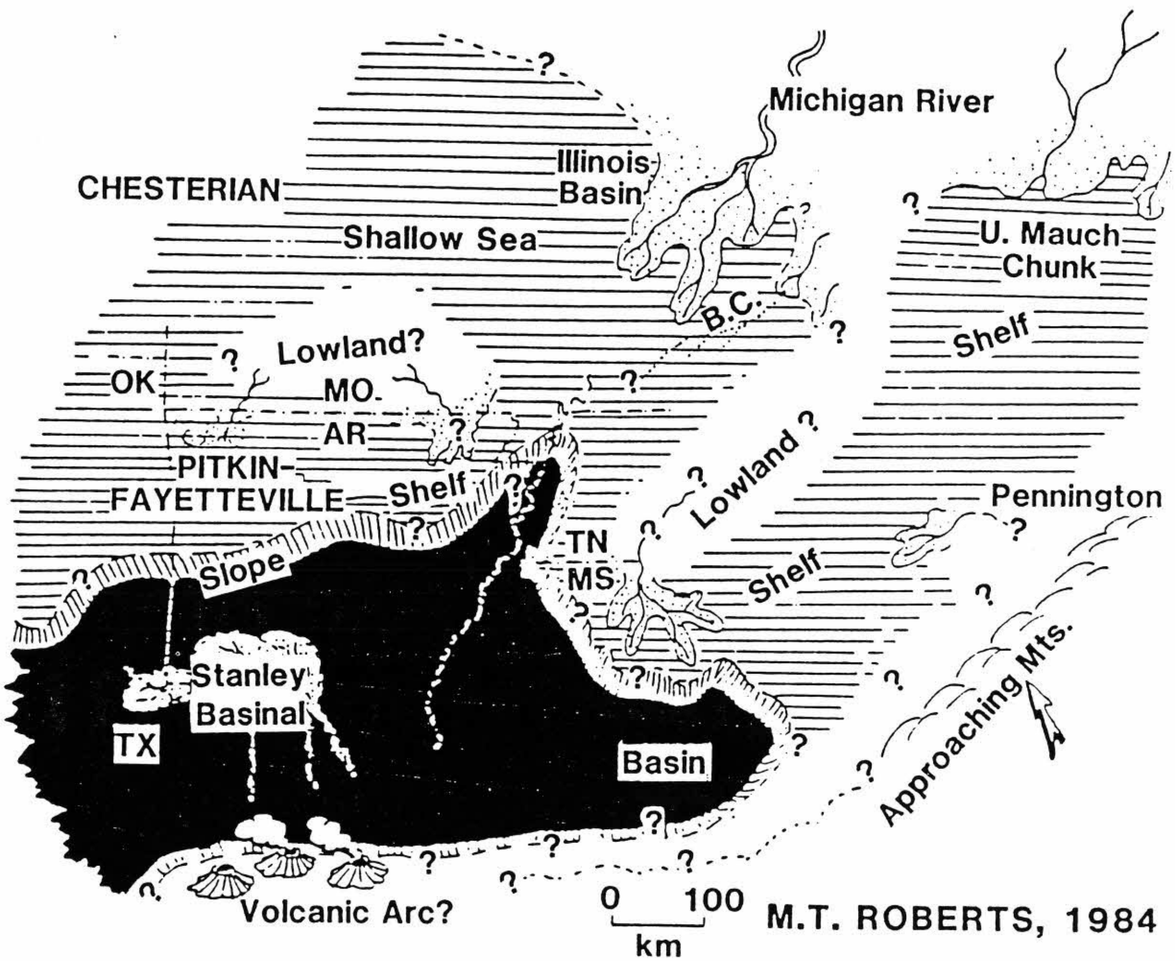


Fig. 20

EARLY MORROWAN SCHEMATIC PALEOGEOGRAPHIC MAP

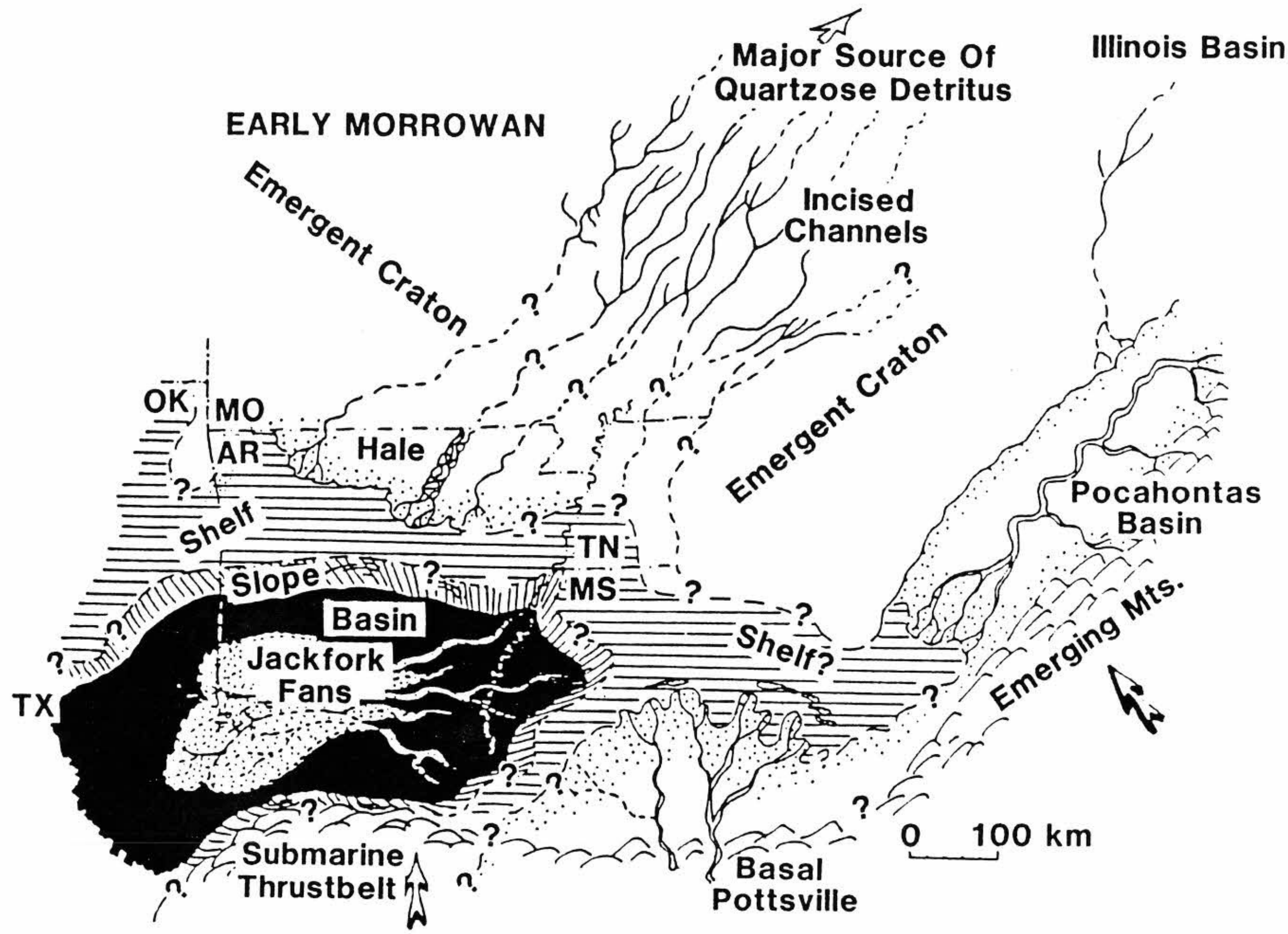


Fig. 21

SCHEMATIC JACKFORK PALEOGRAPHY

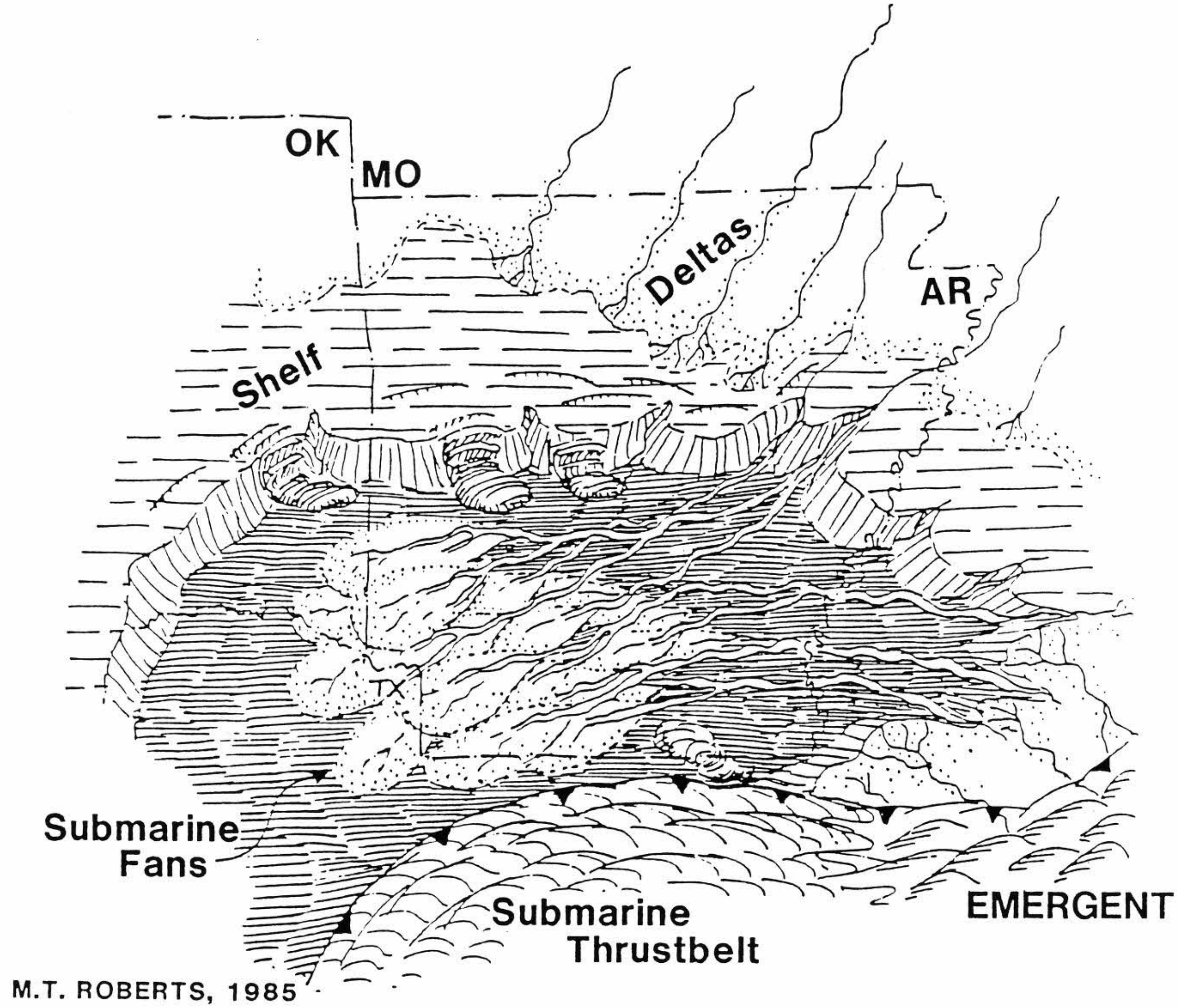
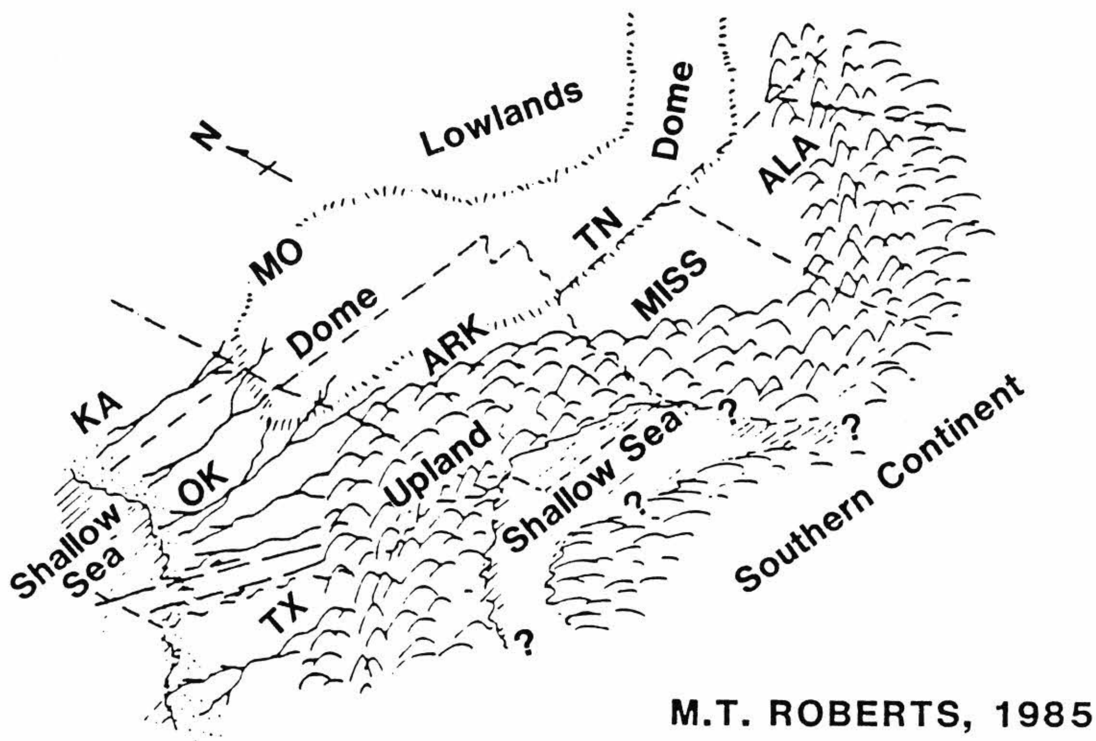


Fig. 22

END OF PALEOZOIC, PALEOGRAPHY



LATE PALEOZOIC PALEOGRAPHY

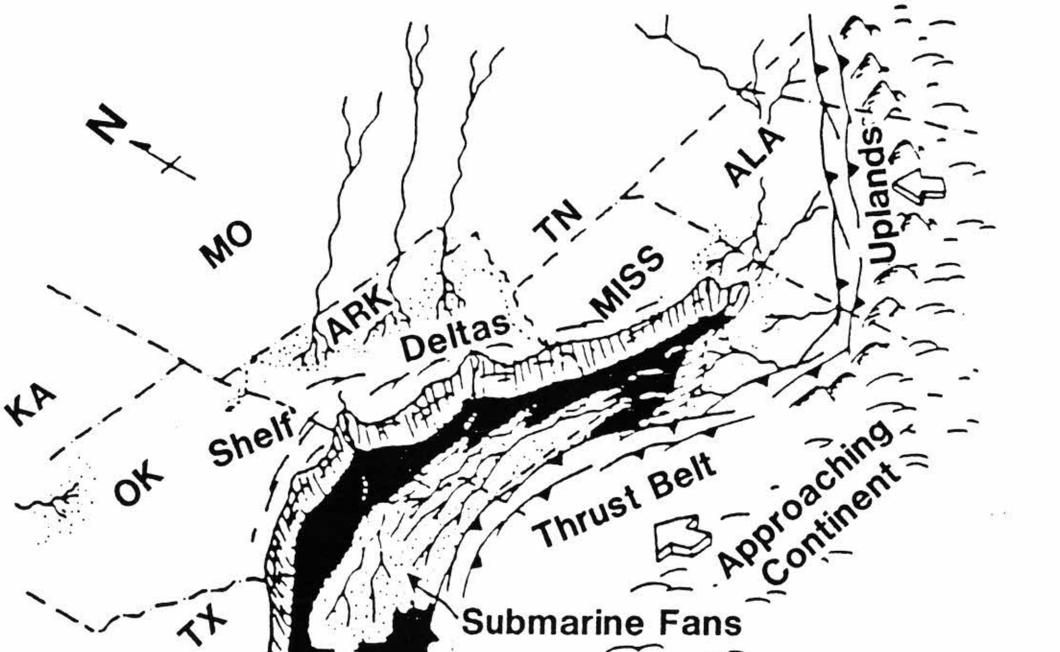


Fig. 23

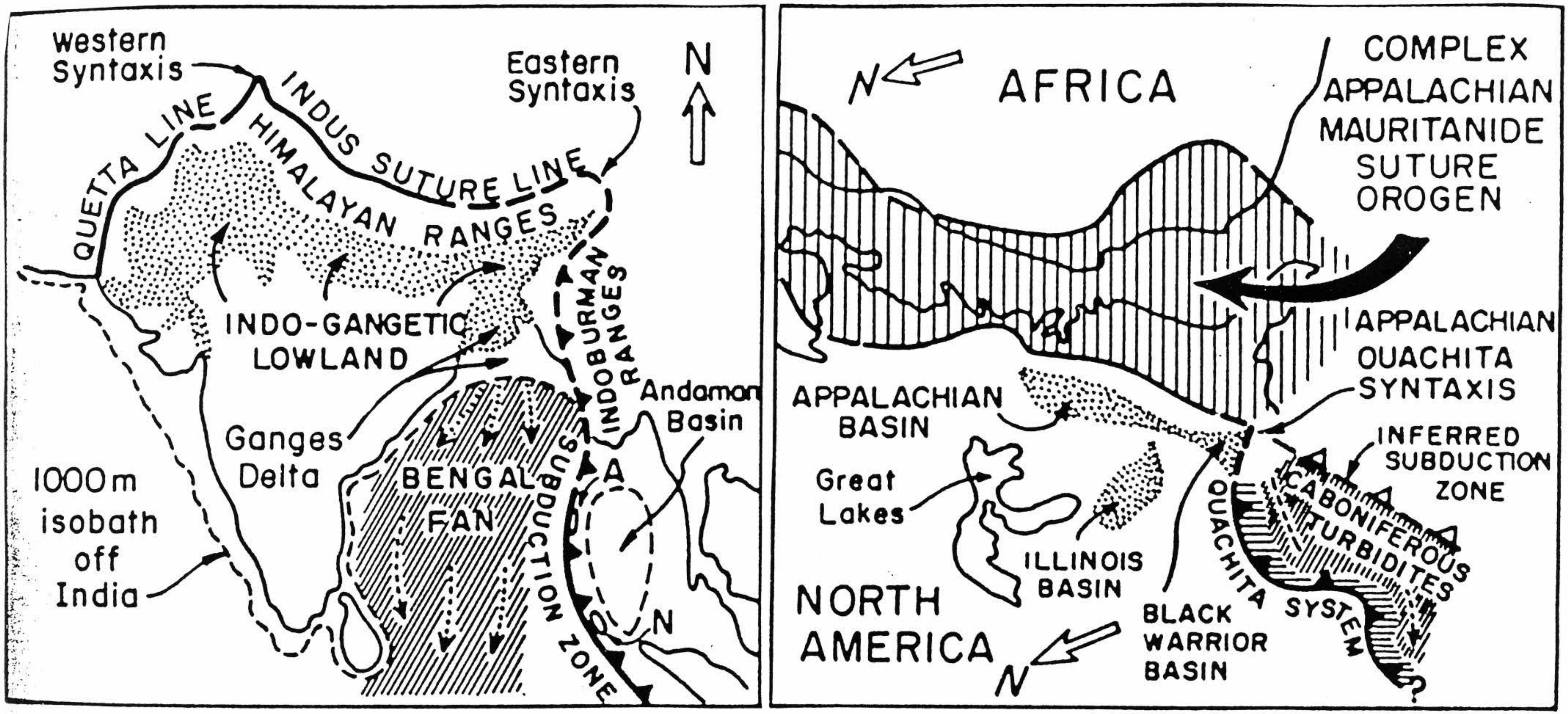
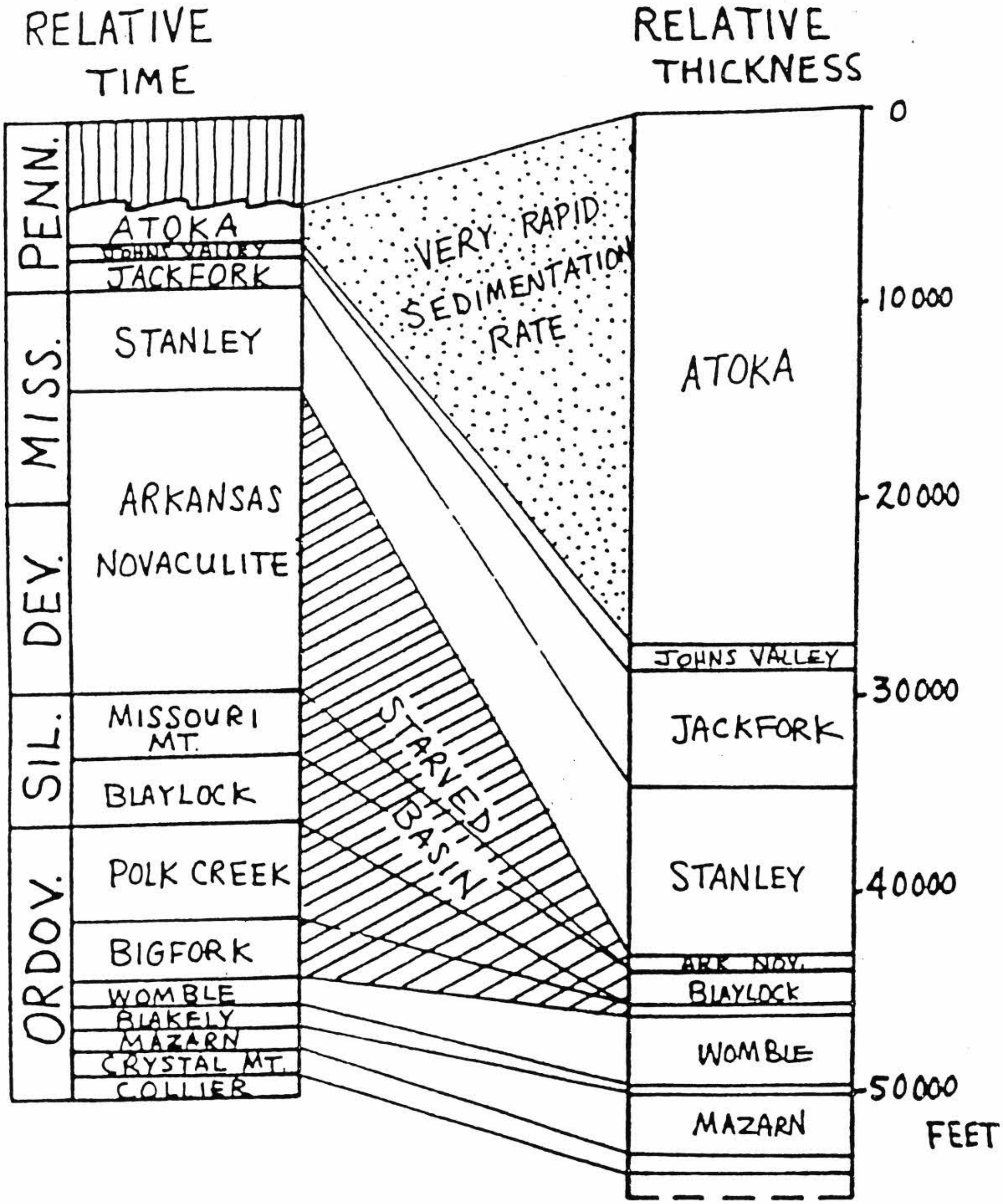


Fig. 24. Comparison, at same scales, of Cenozoic Himalayan-Bengal system (left) and Carboniferous Appalachian-Ouachita system (right). Stipples indicate major depocenters for nonmarine and shallow marine strata. Diagonal rules indicate deep-sea turbidite fans. On right, inferred subduction zone is analogous to Andaman (A) - Nicobar (N) subduction zone on left, and Carboniferous turbidite fan is show occupying a remnant ocea basin prior to a Carboniferous arc-continent collision that formed the Ouachita orogenic system (also shown known today by horizontal rules beside thrust front). From Graham, Ingersoll, and Dickinson, 1976.



(MTR 80 from data in Stone et al., 1973; Haley, 1976)

PALEOZOIC UNITS
OUACHITA MTS., ARKANSAS

Figure **25.** Paleozoic stratigraphic section showing the relative thickness of units in the basinal part of the Ouachita basin as related to time. Note the starved-basin (reduced sedimentation) phase for the Ordovician Bigfork and Polk Creek, Silurian Blaylock and Missouri Mountain, and Devonian-Mississippian Arkansas Novaculite formations. Note also the period of very rapid sedimentation in Atoka time.

From Link and Roberts, 1986

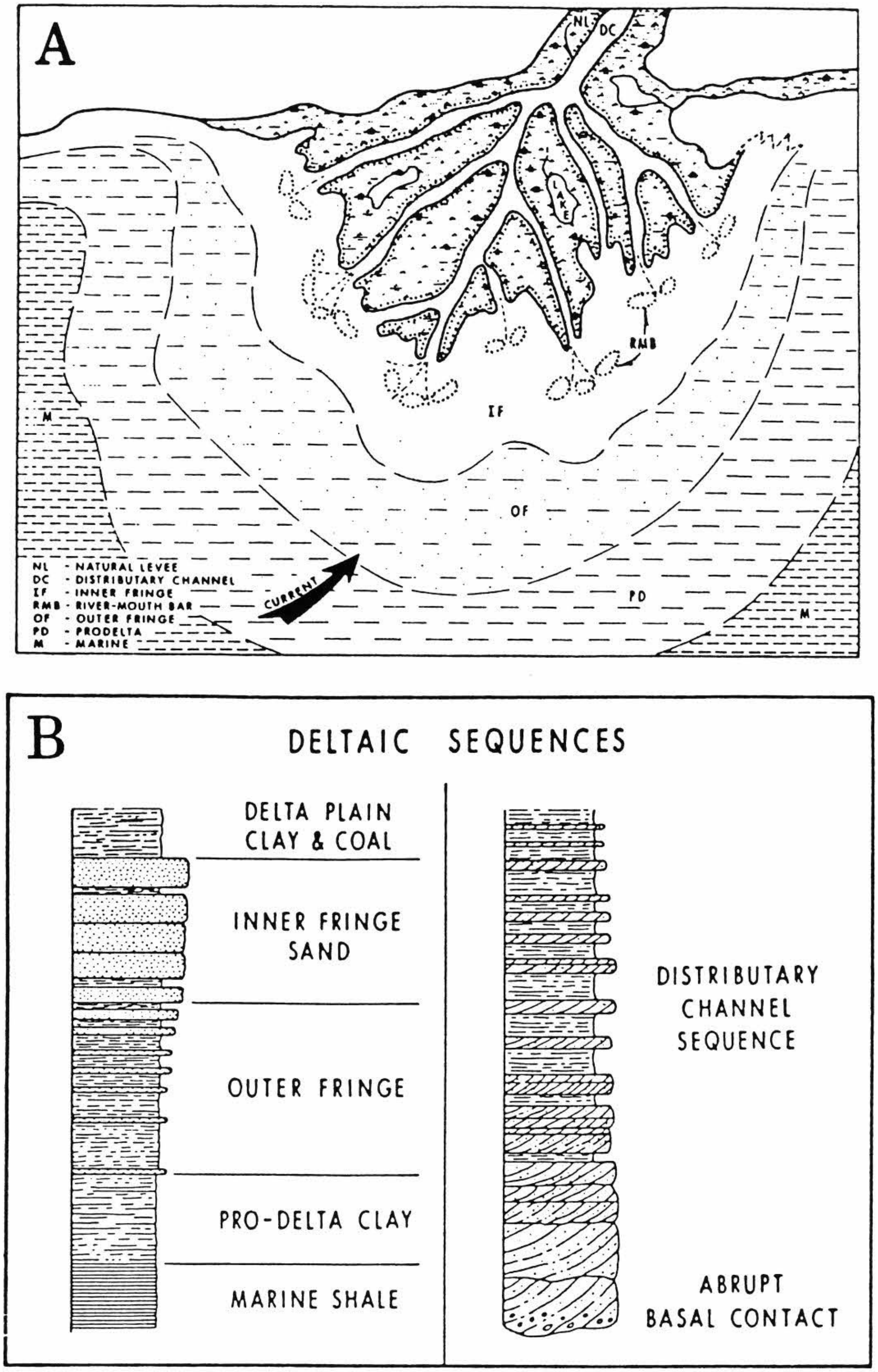


Figure **26.** *A*, depositional environments and sequences typical of modern deltas used as model for interpreting sequences seen during first part of today's trip (from LeBlanc, 1977). *B*, typical sequences of deltaic deposits.

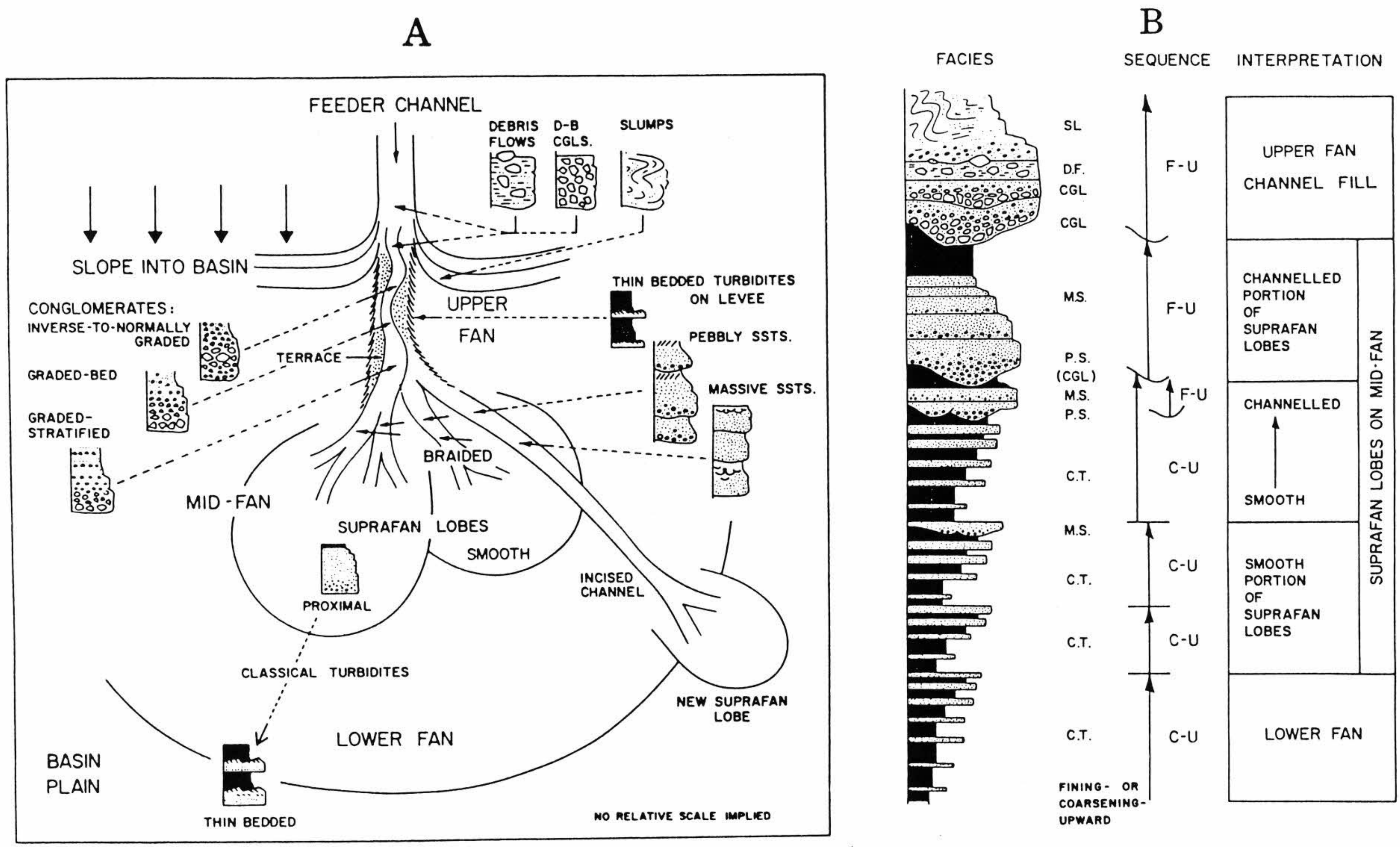


Figure **27.** *A*, Submarine-fan model and associated turbidite facies of Walker (1978). *B*, Hypothetical stratigraphic sequence that could be developed during fan progradation; *C-U* represents thickening- and coarsening-upward sequence; *F-U* represents thinning- and fining-upward sequence; *C.T.*, classic turbidites; *M.S.*, massive sandstones; *P.S.*, pebbly sandstones; *CGL*, conglomerate; *D.F.*, debris flows; *SL*, slumps; from Walker (1978).

BOUMA SEQUENCE

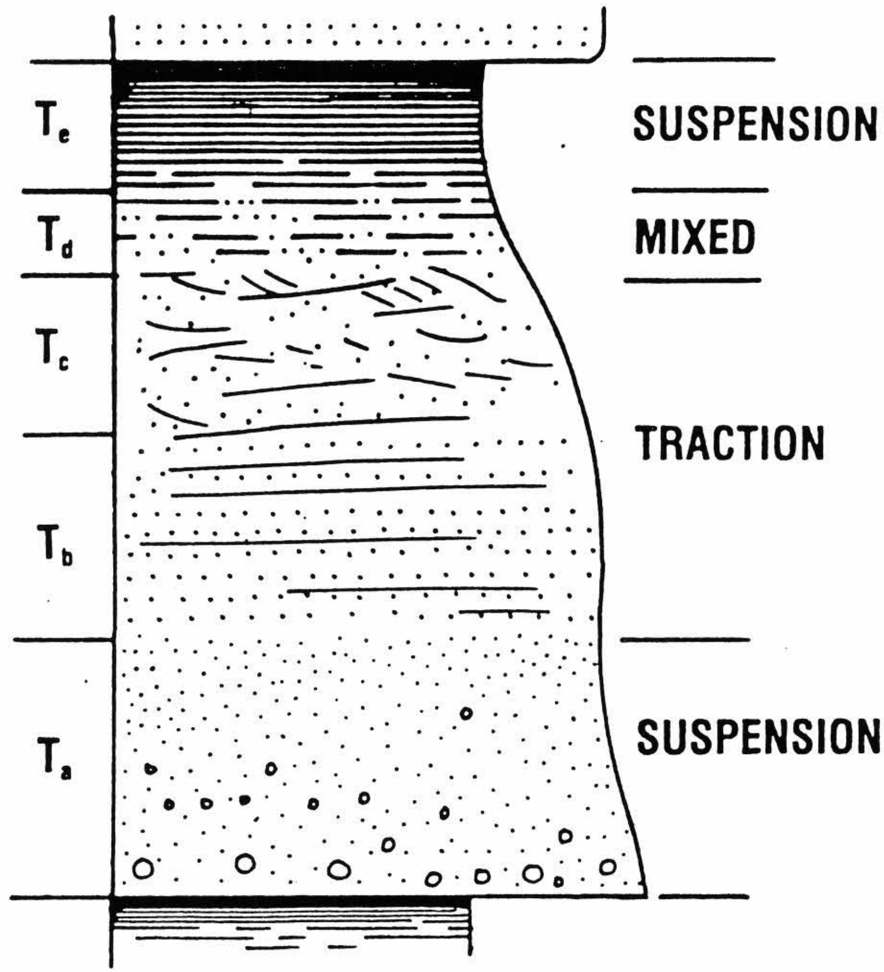
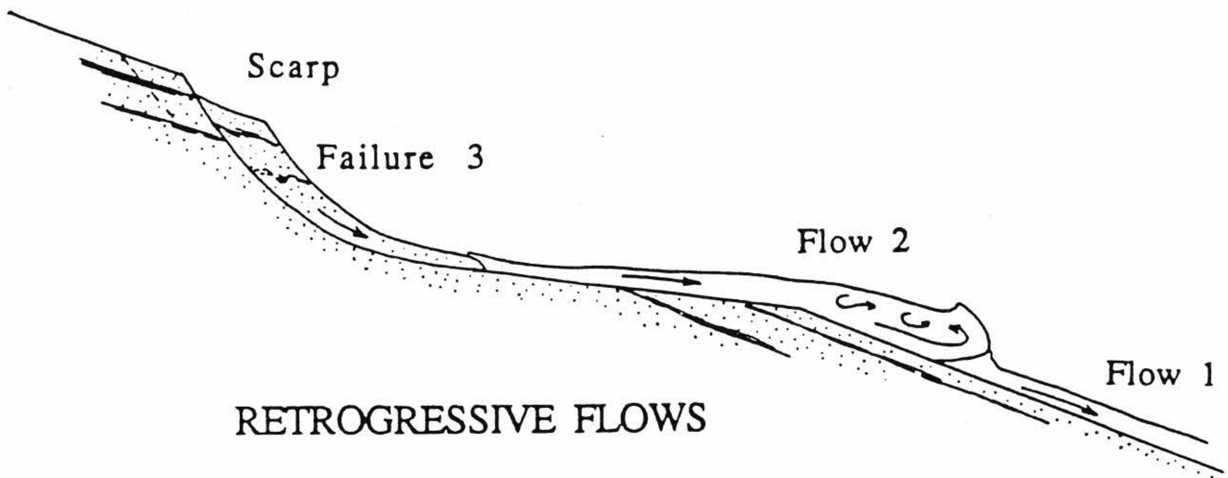


Figure **28.** The Bouma sequence and style of sedimentation represented by each division. Ta represents the waning stages of high-density sedimentation whereas Tb-e represents deposition from low-density currents.



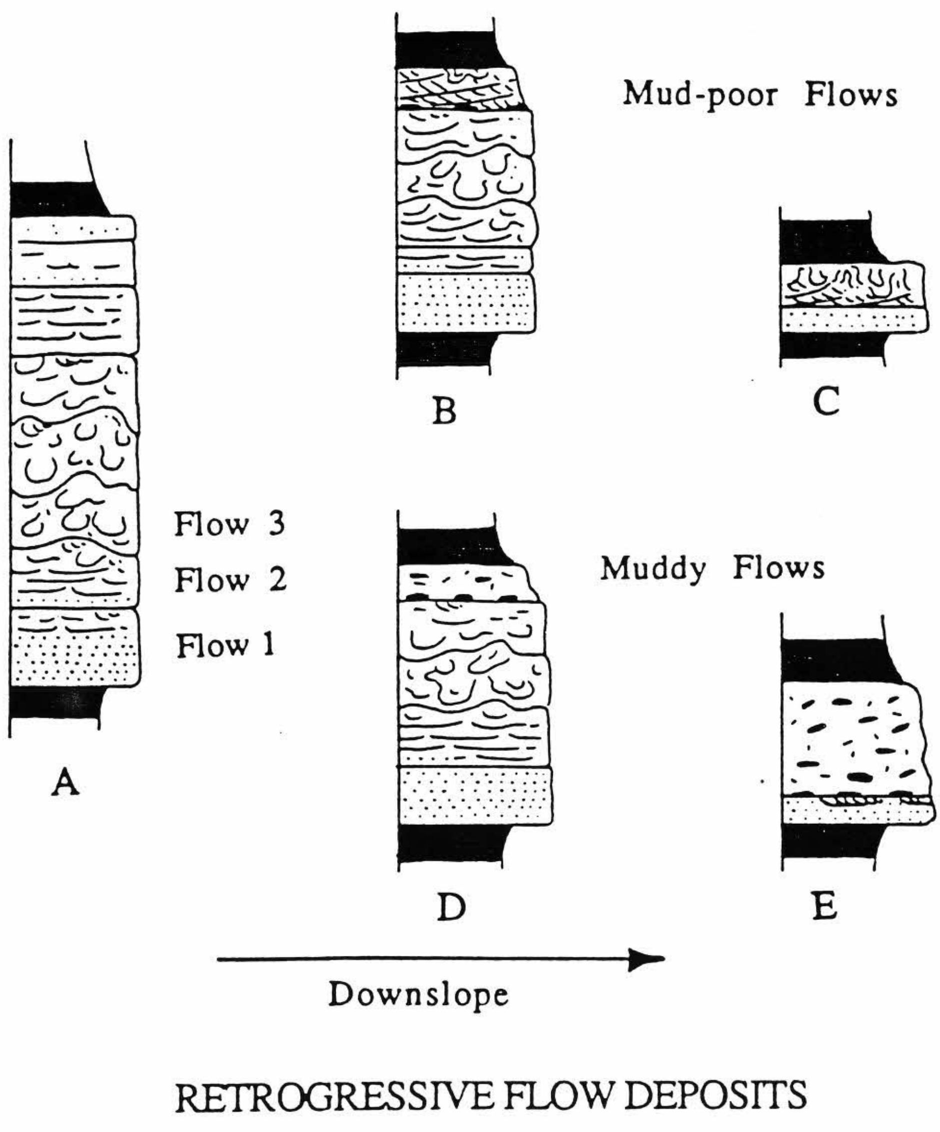


Figure **29.** (Caption on following page).

Figure 29. Retrogressive flows (top) and retrogressive flow deposits (bottom).

Top. Retrogressive flows form when an initial failure forms a scarp. Successive failures along the scarp result in scarp retreat and the generation of a series af flows that move downslope, accelerate, become turbulent, and evolve into turbidity currents.

Bottom. As seen in the Jackfork Group, many sandstone units are composite layers made up of a number, commonly 3 to 12, of individual sandstone beds, each probably representing an individual flow. Individual flow units in the Ouachitas are between 5 and 15 inches thick; thicker flow packages range from about 3 to over 15 feet thick. "A" shows a generalized flow package in a proximal setting. It consists of several discrete flow units. In the Ouachitas, flow units toward the middle of thicker flow packages show undulating contacts representing zones of liquefaction and soft-sediment loading and foundering. Downslope, flow behavior depends on the amount of entrained mud. Purely mud flows would evolve as a series of debris flows (not shown). Mud-poor flows tend to leave packages that pass upward into cross-laminated units in intermediate (mid-fan ?) regions ("B") and into thin flat- (Tb) and cross-laminated (Tc) turbidites at their distal extents ("C"). Muddy flows develop tails that are watery, mobile debris flows. These deposit slurried muddy sand beds above the current deposits in intermediate ("D") and distal ("E") regions. Some distal deposits consist largely of sandy debris flows separated by fine, organic-rich, laminated, hemipelagic shales.

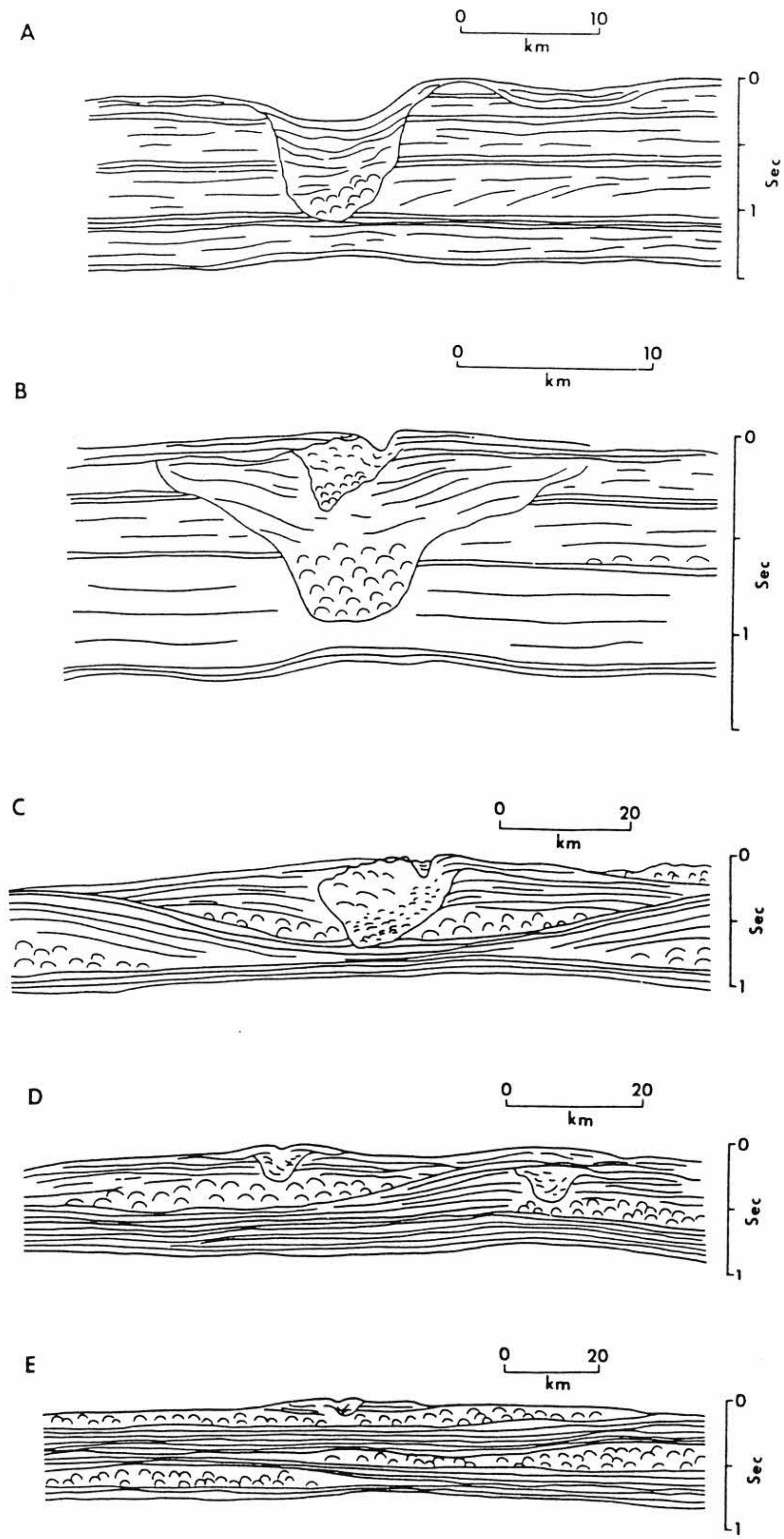


Figure 30. Schematic cross sections, based on seismic records, across the youngest fanlobe of the Mississippi Fan.

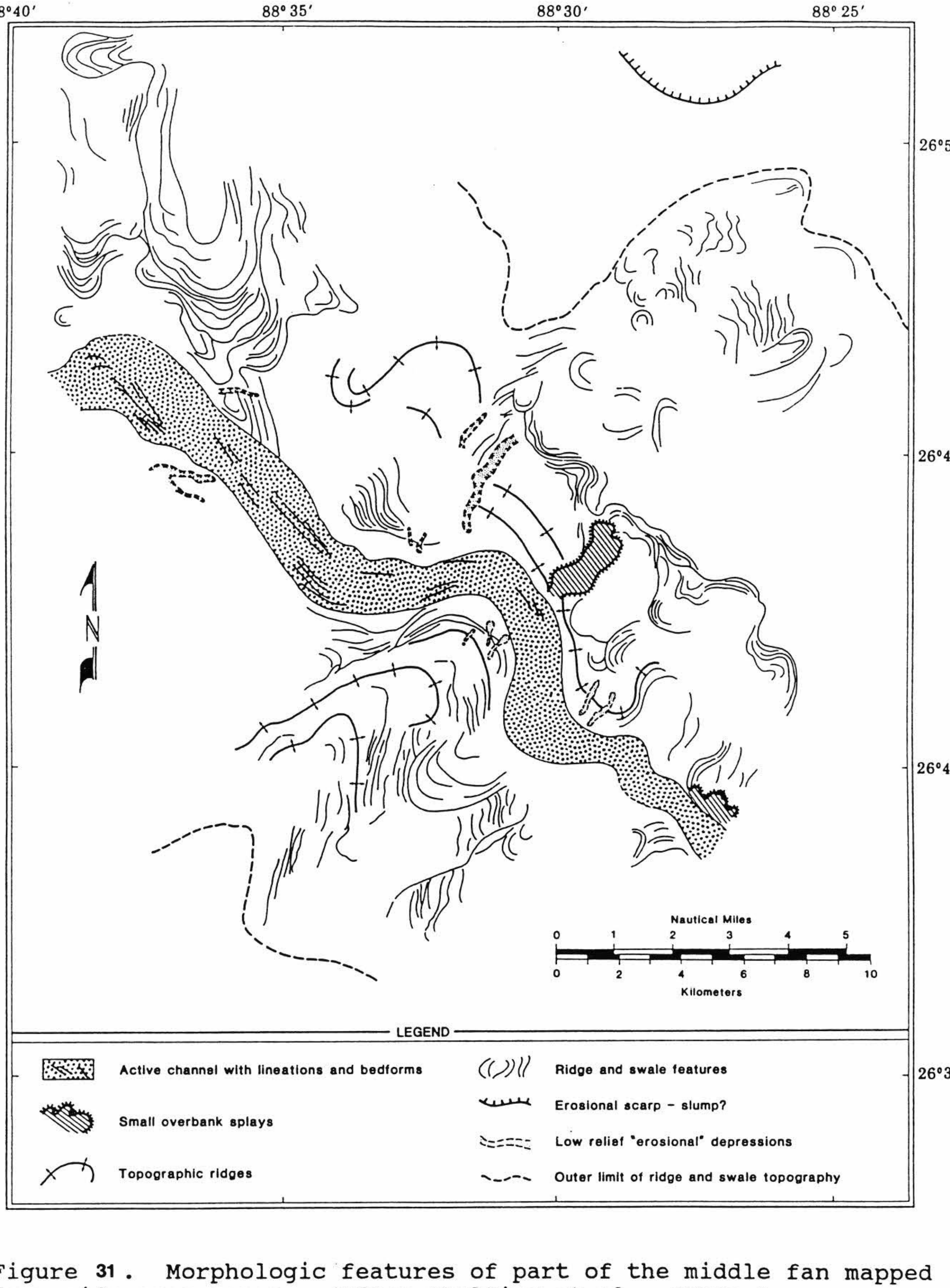


Figure 31 . Morphologic features of part of the middle fan mapped from side-scan sonar. (After Stelting et al., 1985a.)

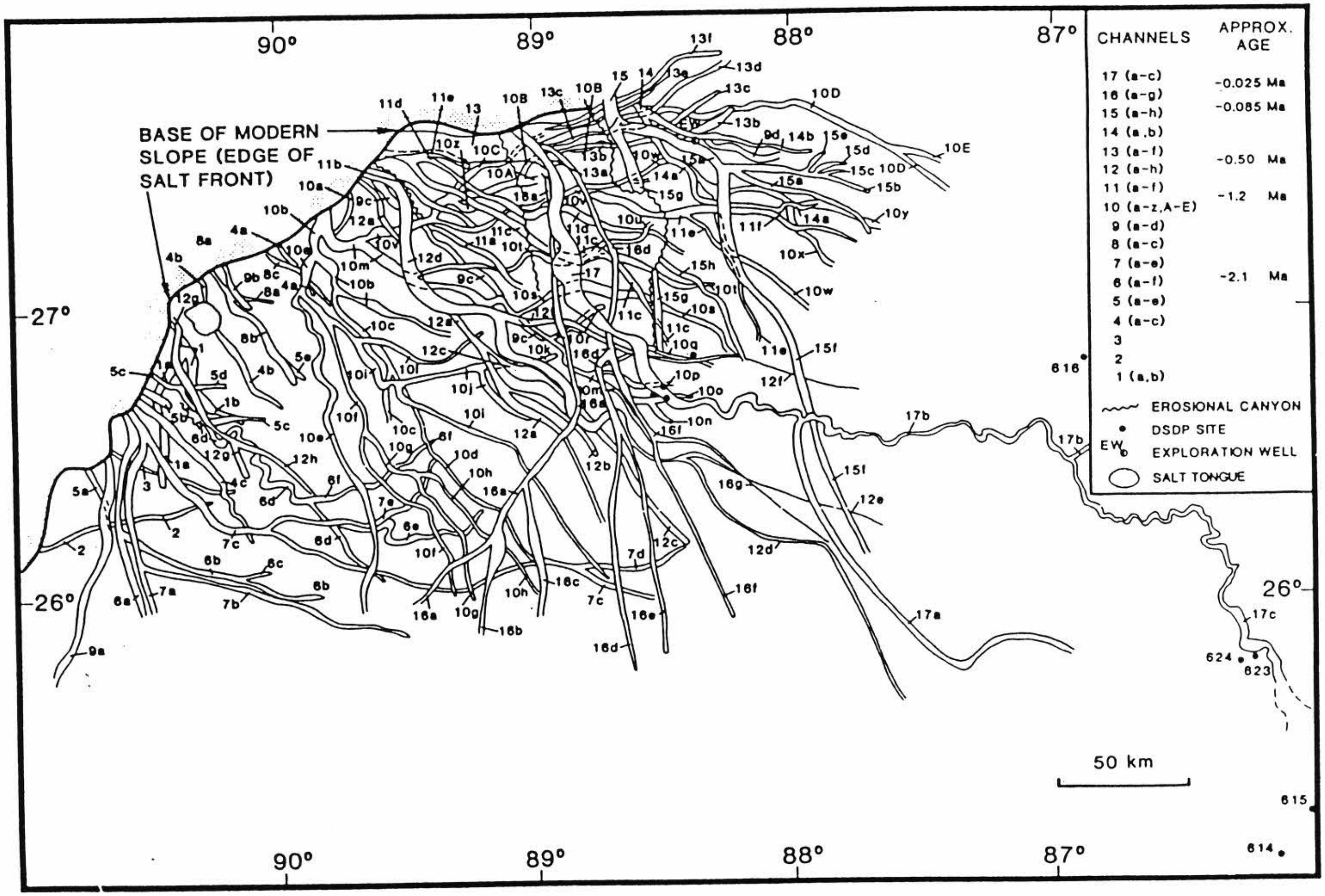


Figure 32 . Composite map of the 17 channel-levee systems identified in the Mississippi Fan. From Weimer (1989).

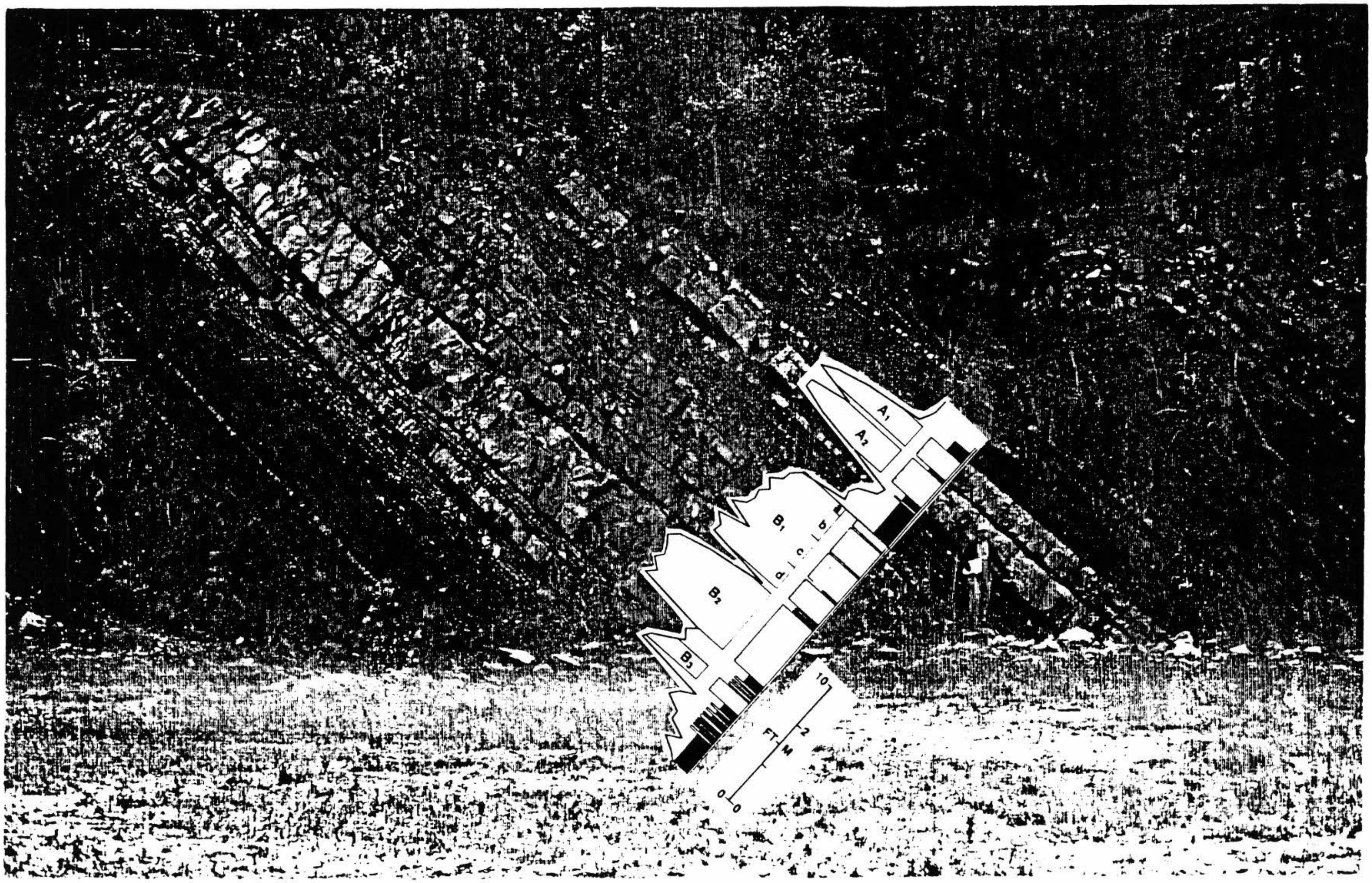


Figure **33.** Gamma-ray log obtained with a hand-held scintillometer (0.5 foot sample spacing) and detailed measured section along the east wall of DeGray Lake Spillway (STOP **16**).

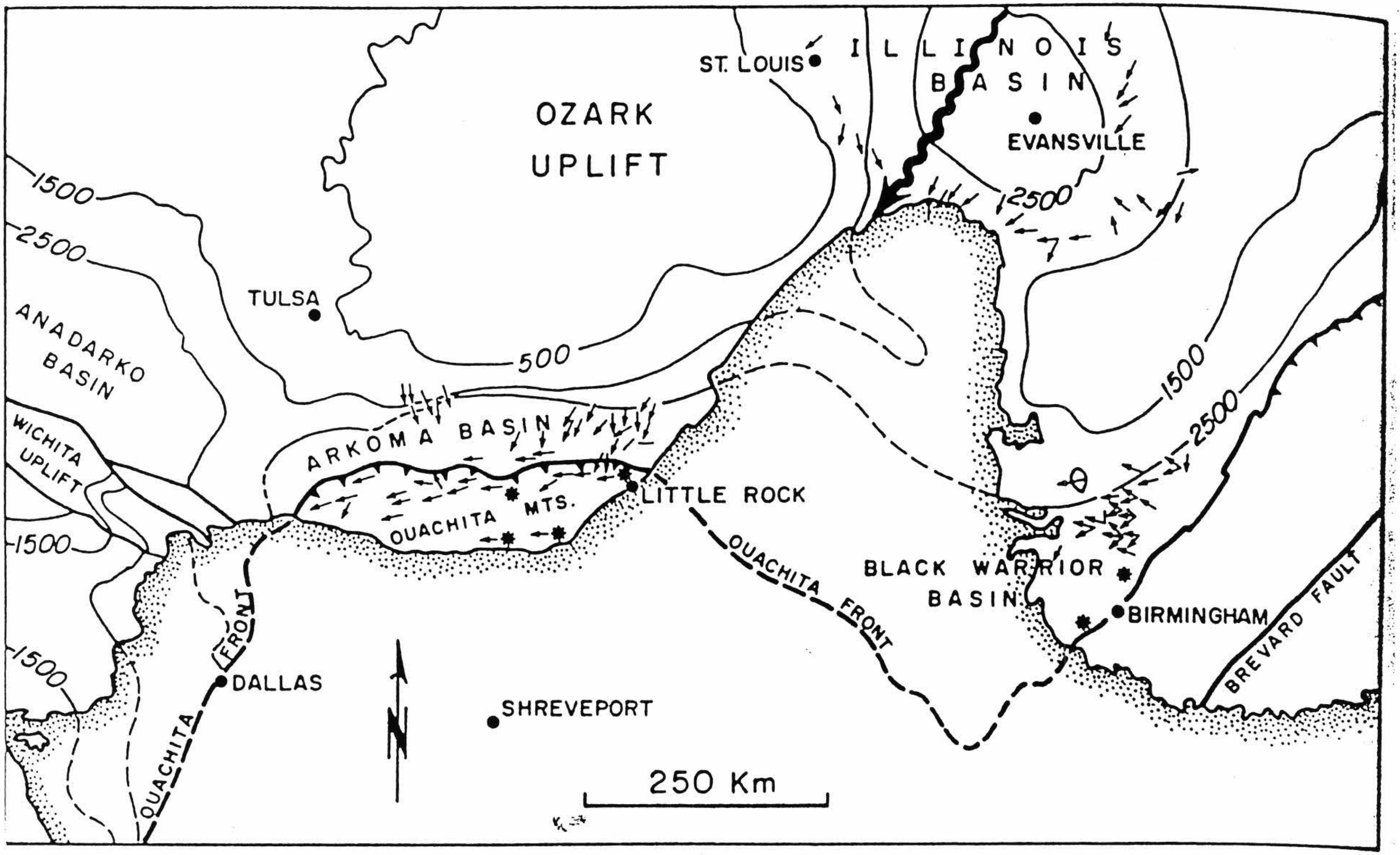


FIG. 34.—Regional tectonic sketch map showing geographic relations between Ouachita Mountains (and adjacent Arkoma basin), Illinois basin, and Black Warrior basin. Basement contours at 500 m, 1,500 m, and 2,500 m after King (1969). Stippled line is outcrop edge of Mesozoic-Cenozoic sequence of coastal plain. Carboniferous paleocurrents (indicated by arrows) from Morris (1974b) for Ouachita Mountains, Potter and Pryor (1961) and Pryor and Sable (1974) for Illinois basin, and Metzger (1965) for Black Warrior basin. Stars in Ouachita Mountains and Black Warrior basin indicate main collecting sites for this study.

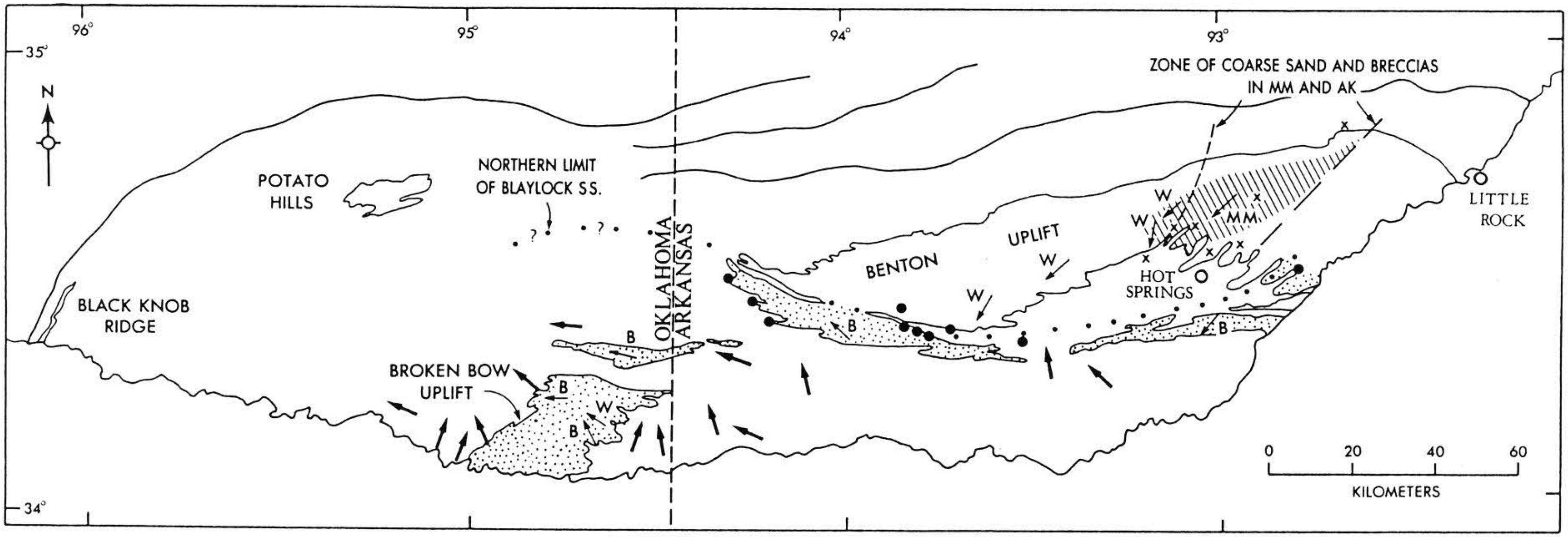


Figure **35.** Paleocurrents and inferred facies of pre-orogenic strata, Ouachita Mountains. Arrows indicate paleocurrents for Womble (W), Blaylock (B), and Missouri Mountain (MM) formations and general paleocurrent trends for Stanley Group in southern areas (heavy arrows). Paleocurrent measurements for Womble in Benton uplift from Markham (1976), Blaylock from Worrell (1984), and Stanley from Morris (1974). The existence of a major northeast-trending paleochannel in the vicinity of Hot Springs throughout the early and middle Paleozoic is indicated by the localization of boulders in the Crystal Mountain Sandstone (diagonally ruled area), paleocurrent patterns in coarse Womble limestones (W in the Benton uplift), the distribution of coarse quartz sand and chert breccia in the Missouri Mountain Shale and lower member of the Arkansas Novaculite (AK), and the localization of thick (>5 m) units of debris-flow breccia in the lower part of the Stanley Group (x). The basin axis is marked by the convergence of northerly and southerly paleocurrents, northern limit of Blaylock Sandstone, and localization of bedded barite deposits (large solid circles; from Hanor and Baria, 1977). From Lowe (1985).

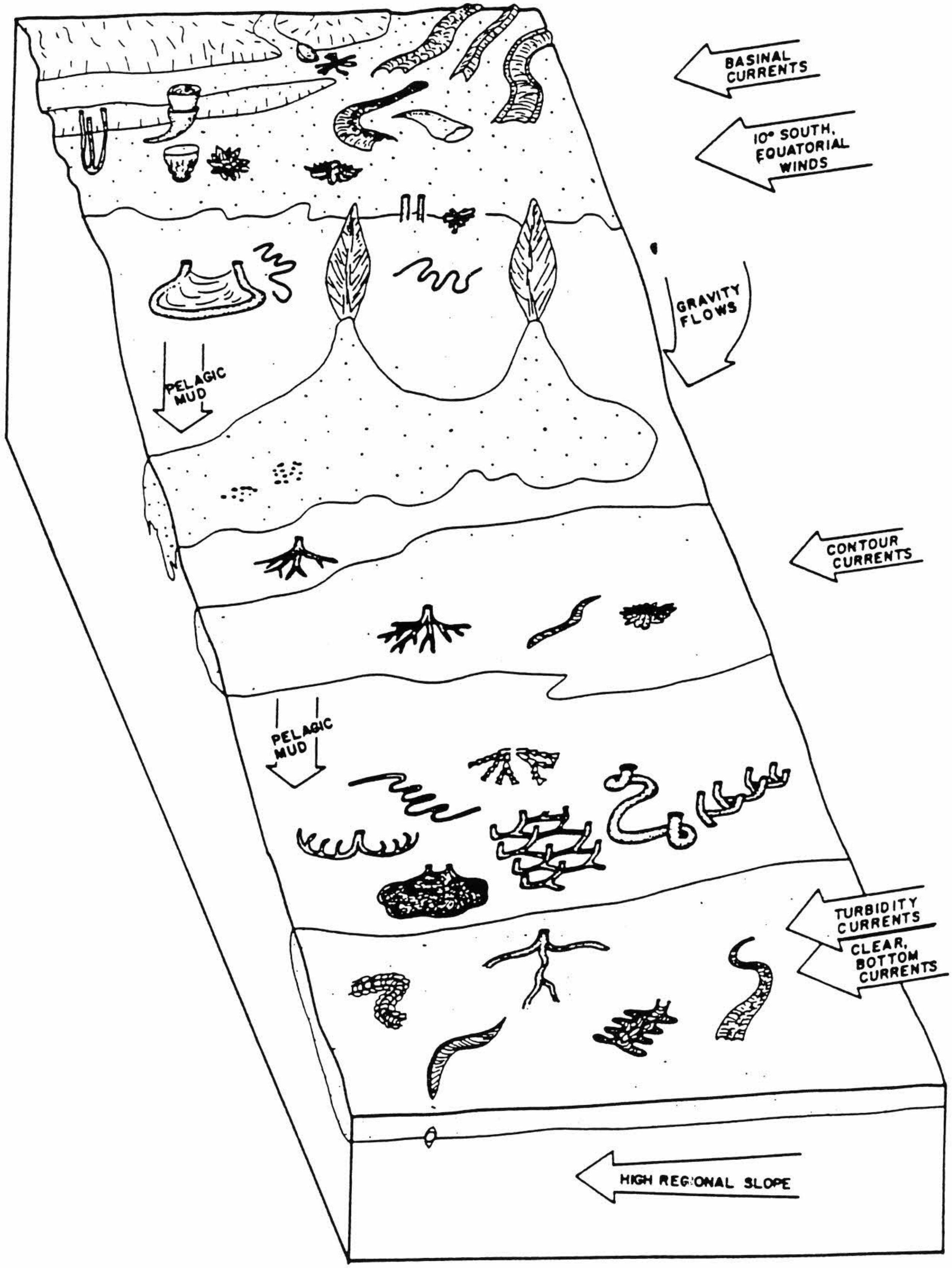


Figure 36

(from Chamberlain and Basan, 1978)

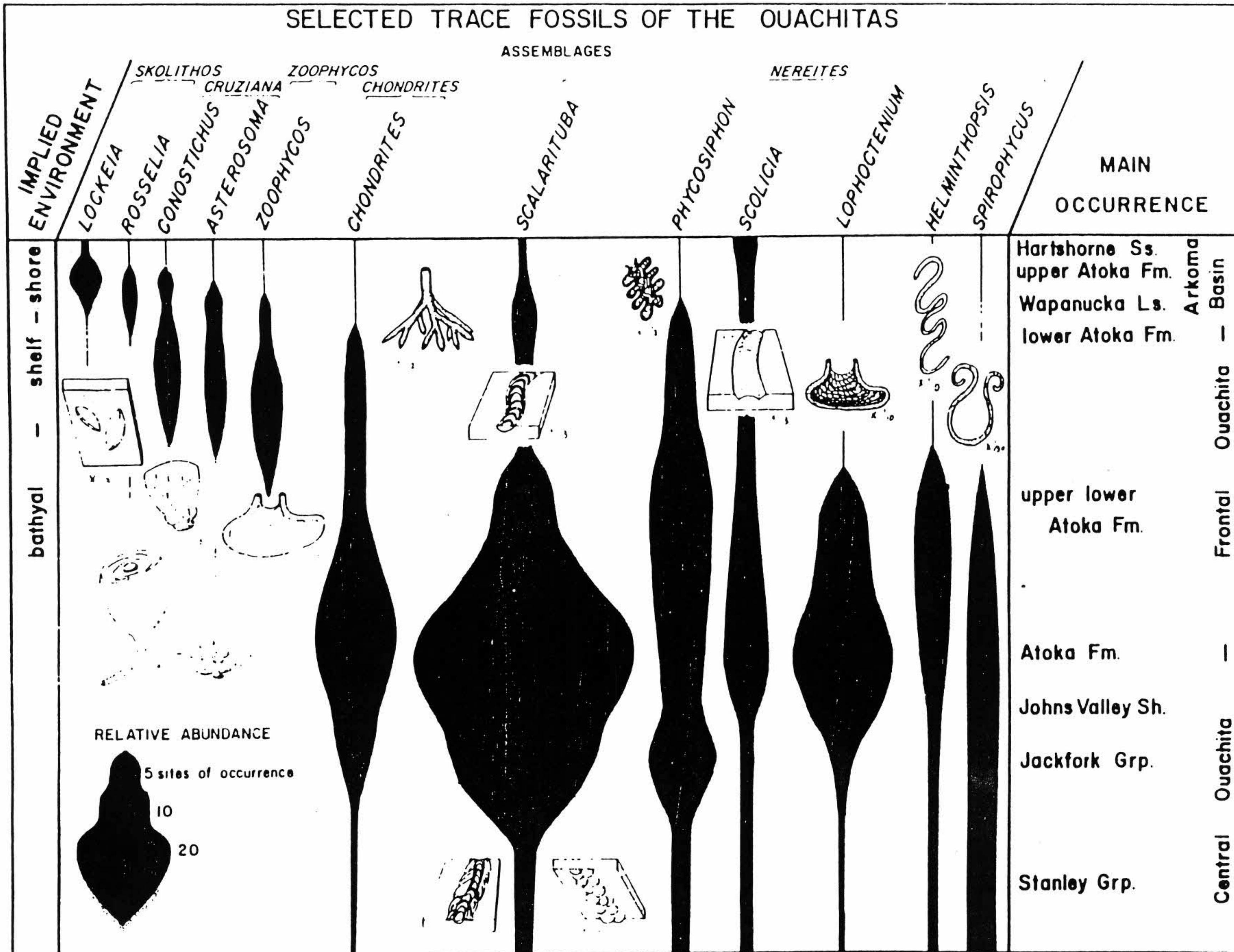


Fig. 37 (from Chamberlain and Basan, 1978)

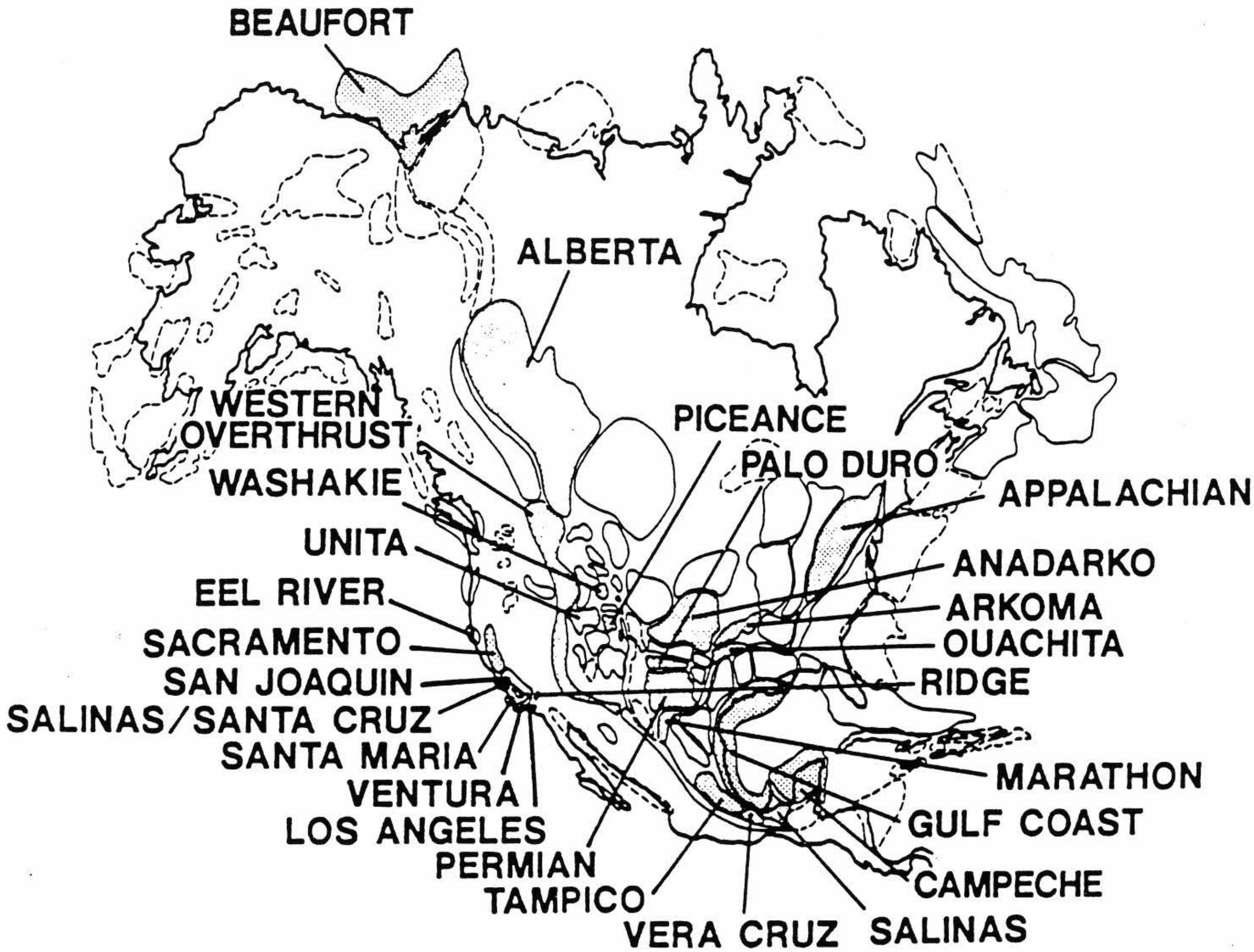


Figure 38. North American sedimentary basins showing: (a) petroleum productive basins with turbidite fields (dotted infill pattern), (b) petroleum productive basins (solid line), and (c) non-petroleum productive basins (dashed line). Base map after St. John (1984).

The following introduction to turbidity currents was extracted from:

Nelson, C.H., and Nilsen, T.H., 1984, Modern and Ancient Deep-Sea Fan Sedimentation: SEPM Short Course No. 14, 404 p.

CHAPTER 3. TURBIDITY-CURRENT OBSERVATIONS AND EXPERIMENTAL STUDIES

C. Hans Nelson

Observations

Many of the basic problems about mode of generation and physical properties of turbidity currents are no better understood now than they were 10 to 20 years ago. Even within the past 10 years, there have been few new laboratory experiments on the phenomena of turbidity currents because of the basic scaling problem and inability to relate laboratory studies to events in nature. Thus, a review of the older work remains pertinent to provide a framework for understanding the processes depositing deep-sea-fan sediment.

In both modern and ancient deep-sea-fan deposits, graded beds that are coarser-grained than the interbedded muds are observed to contain displaced shallow-water plant and animal remains. They are obviously the product of resedimentation of shallow-water debris to basin depths. These graded beds in both modern and ancient deposits often occur in cone-shaped sediment deposits with channel and interchannel areas. (Nelson and Kulm, 1973; Walker and Mutti, 1973). The resedimented deposits of deep-sea fans are inferred to be the product of various types of turbidity currents that occur on a wide variety of scales and densities of flows. These inferences are based on actual observations of some types of turbid flows, although many types of the family of turbidity-current events have not been measured in natural events.

Observations of actual turbidity currents have been most commonly made in lakes. The studies began in the late 19th century and continue to the present modern events. Movements of turbid water traced in Lake Mead revealed that 12 major turbidity currents flowed the 140-km length of the lake during a 14-year period (Gould, 1951). The densities of these turbidity currents were as low as 1.006 g/cm^3, and the flows had average velocities of 25 cm/s.

Crater Lake's unusually large summer storm events cause sheet wash off the caldera walls, resulting in visible nearsurface turbid flows that sink and travel out toward the center of the lake over a day-long period at speeds about half those observed in Lake Mead. Sediment stratigraphy from the basin floor contains interbedded mud and graded sand beds (Nelson, 1967; Nelson et al., in press). Thus, turbidity currents and their depositional products have been measured and documented in several deep lakes.

The evidence for turbidity currents on a larger scale is inferential in the marine environment. Resedimented graded sand layers in the modern deep-sea fans associated with channeled topography already have been mentioned. The most direct observations in the marine environment consist of recorded slope failure, particularly near harbors, and sequential breaks of deep submarine cables lying across canyon and channel pathways (Heezen and Ewing, 1952; Menard, 1964). In several well-documented cases, harbor areas or structures have been modified or destroyed by slope failure monitored by bathymetric surveys before and after the slump scar was created. Direct evidence of down-slope movement of sediment, perhaps evolving into a true turbidity current, is provided by successive cable breaks on the flat bottom in deep water.

The classic example cited as the best observation of turbidity current generation is the Grand Banks earthquake and series of submarine cable breaks that occurred off Canada in 1929. The largest historical earthquake in Atlantic Canada, with an estimated magnitude of 7.2 centered near the Laurentian Channel, it resulted in 5 documented cable breaks over a distance greater than 500 km and a period of 13.3 hours (Fig. 39) (Heezen and Ewing, 1952; Menard, 1964; Piper and Normark, 1982b). Although the precise path, velocity and sedimentary processes that caused the cable breaks have been disputed (Menard, 1964), there is good seismic evidence that a debris flow more than 200 m thick followed the eastern Laurentian Channel for 100 km (Piper and Normark, 1982), and that a surface turbidite with a distinctive red color was laid down over the abyssal plain for a distance of up to 1,200 km from the source (Fruth, 1965). Presence of the debris flow in the eastern Laurentian Channel and the red-colored turbidite only on the channel floor indicate that some of the flows and subsequent turbidity currents were channelized (Piper and Normark, 1982a and b; Menard, 1964). The widespread cable breaks and occurrence of the red-colored turbidite on the lower fan and abyssal plain, however, suggest that unchannelized turbidity currents evolved and spread as sheet flow in the lower fan. Because of the complexity of slump scars in the generation area and the occurrence of both channelized and sheet-flow processes, accurate calculation of the velocities of the debris flow and turbidity current event is impossible. Various estimates from 10 to nearly 30 m/s have been made based on timing of cable breaks in different areas (Heezen and Ewing, 1952; Menard, 1964).

Detailed study of a recent mass movement and probable ensuing turbidity-current event off Nice, France is now in progress. At 1400 on October 19, 1979, a 300 m^2 area of the Nice airport collapsed, created an area with 50 m water depth and "tsunami-like" waves of several meters for 100 km along the coastline (Groupe ESCYANICE, 1982). No earthquake occurred, thus, this is purely a sedimentological event caused by unstable and oversteepened slopes. SEABEAM data shows a herringbone or badlands-type bathymetry in the Nice area caused by active headward erosion of mass movement; submersible observations reveal gravel waves on the canyon floor (Group ESCYANICE, 1982; Pautot, 1981).

Following the shoreline mass movement event, two telephone cables across a lower canyon valley were broken at 17h45 and 22h00, at a distance of 80 and 110 km, respectively, (Group ESCYANICE, 1982). These data suggest that turbidity currents traveled at a rate of about 6 m/s to the first locality and 4 m/s to the second locality over gradients of 1:30 and 1:300, respectively (Gennesseau et al., 1980). If the faster rate to the first locality is subtracted from the second locality downstream on the same path, then the average speed is 2 m/s for the distal path of the apparent turbidity current. This event, much smaller in scale than the Grand Banks, created slower speeds for the entire sediment transport event.

Generation of Turbidity Currents

A fitting introduction to this topic may be to paraphrase a statement by Middleton and Hampton (1976) that still remains accurate:

> There are three main aspects to the problem of initiation of turbidity currents: (1) What causes mass movement of sediment to begin? (2) How do submarine mass movements, such as slides and

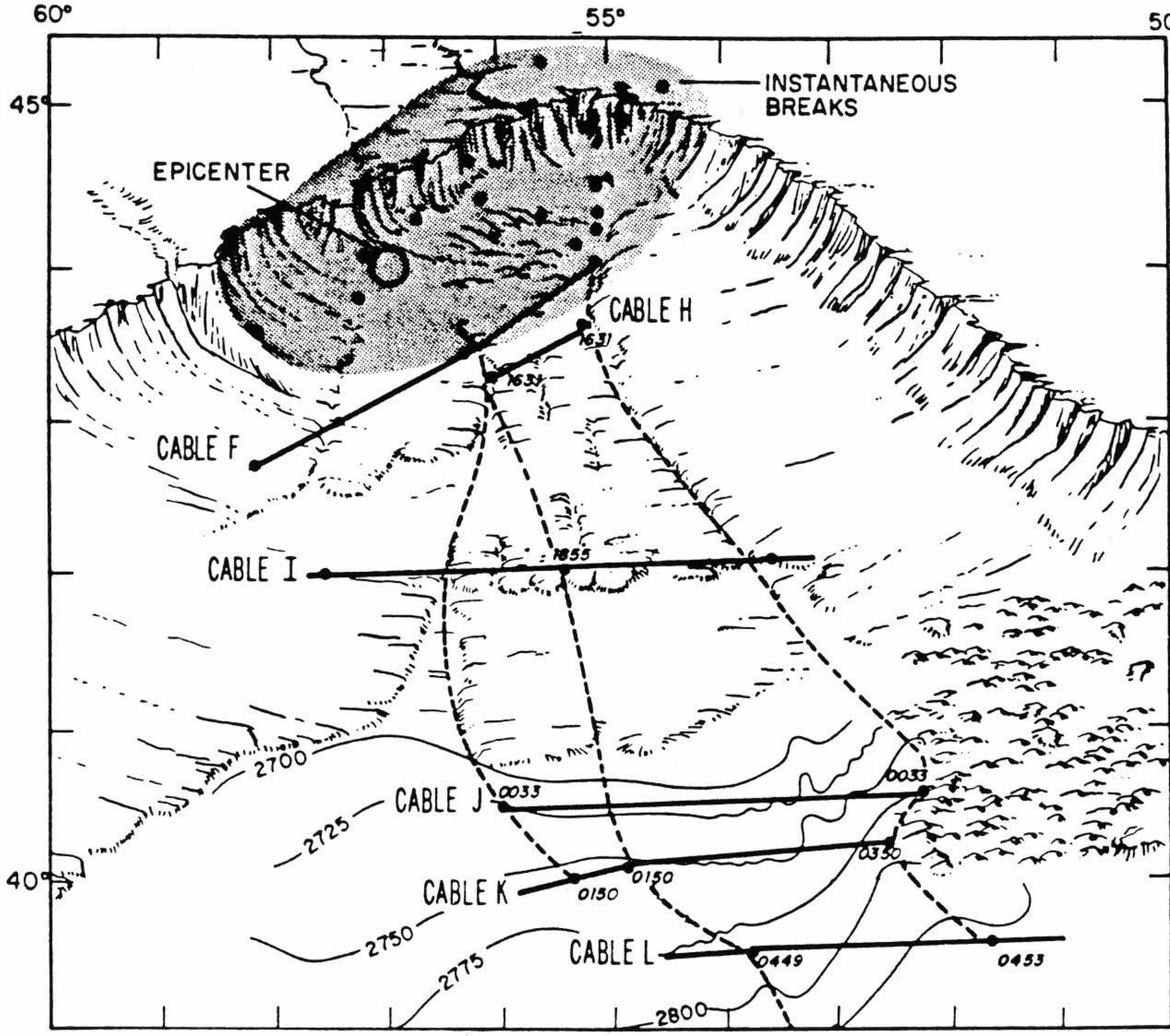


Figure **39** · The Grand Banks earthquake of 1929 and subsequent cable breaks (from Menard 1964, Fig. 97).

slumps, turn into a variety of sediment gravity flows of which turbidity currents are one type? (3) What transitions are possible between one type of sediment gravity flow and another, and what causes them to take place and result in turbidity currents? Answers to these questions can only be tentative at present. The whole question of initiation remains one of the least studied and least understood aspects of turbidity currents."

Menard and Ludwick (1951) made the general suggestion that turbidity currents may be generated by rapid introduction of sediment, agitation of bottom sediment, and mass movement of unstable bottom sediment. Rapid sediment introduction can be caused by flash floods and seasonal outflow of rivers (Drake et al., 1972), windstorms (Sarnthein and Diester-Haass, 1977), volcanic eruptions (Nelson et al., 1968) and mass flows or sheet wash off subaerial slopes into the water. In lake water ($\rho = 1.0$), river inputs may result in direct generation of turbidity currents; this is seldom if ever true for river plumes entering marine water, because of the high density of saline water and the more energetic mixing processes of the marine environment compared to the lacustrine environment. Rapid introduction of sediment into the marine environment results in rapid deposition on the continental terrace (Drake et al., 1972) from which other processes remove and transport the sediment down-slope to depositional sites such as deep-sea fans on basin floors.

The ocean floor of the continental terrace may be agitated by storm waves, tidal currents, ebb flow from storm surge, tsunamis (Kastens and Cita, 1981), or major oceanic currents like the Gulf Stream. The storm- or seismic-related events, especially in regions of rapid sedimentation from rivers, are quite likely to generate turbidity currents. Intense sediment resuspension from storm-wave oscillatory currents, liquefaction from storm-wave loading (Clukey et al., 1983) and sediment entrainment by wind-driven and storm-surge ebb currents (Nelson, 1982) may all combine to generate turbidity currents, especially in submarine canyon heads.

The mass movement of sediment, then, is commonly triggered by earthquakes, storm-wave processes, or oversteepening of slopes caused by rapid deposition of unstable sediment. The synergistic effect of rapid deposition of organic-rich, gas-charged deposits agitated by cyclic loading from storm waves or seismicity results in common slope failure (Carlson and Molnia, 1977), particularly on delta fronts (Coleman et al., 1983; Carlson and Molnia, 1977) or offshore from actively glaciated terrain (Carlson, 1978) such as was widely prevalent during glacial epochs.

Slope failure as a group of processes caused by any number of the above-mentioned factors generates turbidity currents; the recurrence of turbidity currents constructs deep-sea fans and is well-documented (Menard, 1964). The association of earthquake and turbidity-current events also is well substantiated (Cita and Ricci Lucchi, 1984) but does not explain the frequency of turbidite deposition on deep-sea fans. The recurrence interval of the high-magnitude Grand Banks earthquake (Fig. **40**) is estimated to be approximately 100,000 years (Piper and Normark, 1982b). A similar-sized slump in the Sagami Wan area off Japan generated by the great 8.2 magnitude Kwanto

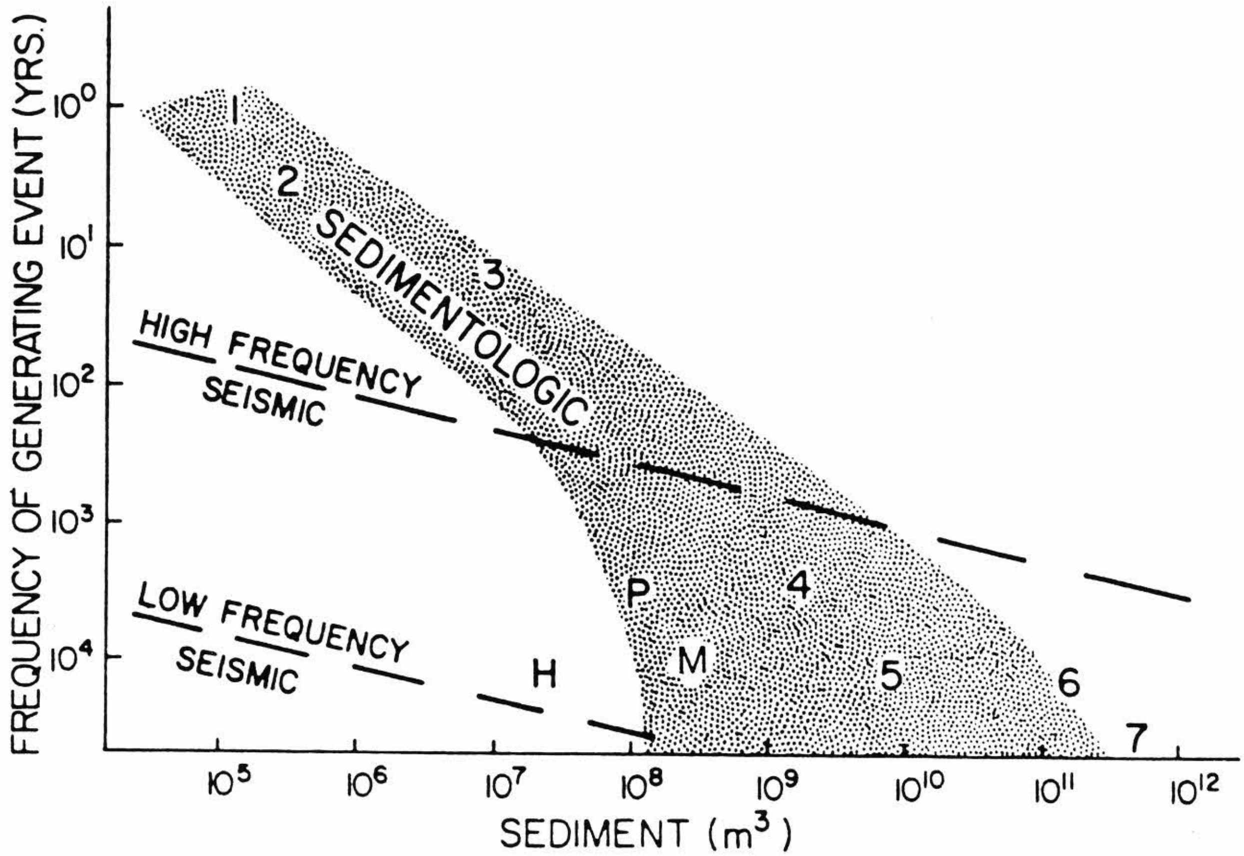


Figure **40**. Speculative plot of total volume of turbidite against frequency of initiating event (modified from Piper and Normark 1983, Fig. 10). The sizes of certain well-known turbidites are indicated. M = Mt. Mazama tuffaceous turbidites on Astoria Fan (Nelson 1968); H = Navy Fan Holocene (Piper and Normark 1983); P = Navy Fan Pleistocene (Piper and Normark 1983); 1 = La Jolla canyon (Chamberlain 1964); 2 = Magdalena delta slumps (Morgenstern 1967); 3 = Mississippi delta slumps (Morgenstern 1967); 4 = Hispaniola-Caicos basin turbidites (Bennetts and Pilkey 1976); 5 = East Alpine flysch (Hesse 1974); 6 = Black Shell turbidite (Elmore et al. 1979); 7 = Grand Banks turbidite (Piper and Normark 1982). Turbidites resulting primarily from rapid supply of sediment are shown stippled. Sedimentologic types result from normal slope failure and storm generation in canyon heads. Dashed lines show frequency relationships for turbidites primarily triggered by seismic shaking, in areas with a high and low frequency of earthquakes.

earthquake in 1923 (Menard, 1964) has a periodicity of about 100 years, similar to that of major earthquakes on the San Andreas fault system.

The frequency of large earthquakes (for example every 1500 years in the Mediterranean; Kastens, 1984) and consequent turbidity currents is not high enough to account for the much greater number of observed turbidity current events on modern (Nelson and Kulm, 1973) or ancient fans (Mutti, et al., 1984). For example, on Astoria Fan during the late Pleistocene, the minimum rate of deposition of overbank turbidites in the upper-fan-valley region was greater than one event every 7 years (Nelson, 1976). The frequency of sediment flushing from submarine canyons or slumps off delta fronts is in the range of once every 1 to 10 years (Piper and Normark, 1983). These data indicate that earthquakes or other exceptional circumstances such as the Mt. Mazama eruption (7000 BP) and consequent deposition of tuffaceous turbidites in Astoria Fan (Nelson et al., 1968) are too infrequent to generate most of the turbidites in fans. Except in volcanic caldera lakes, where numerous volcanic eruptions and precursor earthquakes definitely influenced the rate of turbidite formation (Nelson et al., 1983/84), "normal" but intermittent sedimentological processes with a higher frequency than earthquakes or volcanic eruptions appear to be responsible for generation of most turbidity currents.

The "sedimentological" turbidity currents generally occur in regions of rapid sedimentation, that are prone to overloading of slopes (Piper and Normark, 1983), particularly in areas where storm-wave energy is focused, such as submarine canyon heads or lobate delta fronts (Reimnitz, 1971). Periodic sediment flushing every year to several years, usually associated with storm waves, removes the fine-grained sand of longshore drift that rapidly accumulates in canyon heads (Chamberlain, 1964). The relatively high frequency of storm-flushing of this sediment funneled down canyon floors and through fan valleys appears to cause the majority of turbidity currents that build deep-sea fans. The "storm generation" process is enhanced by lowered sea level that allows direct access of river sediment into canyon heads. As a result, there is a much more rapid input of unstable canyon sediment, during lower sea-level regimes. The stratigraphic record in modern fans verifies that most fan growth occurs during turbidite regimes of lower sea-level stands (Nelson and Kulm, 1973). Evidence is accumulating that the same is true for ancient deep-sea fans in the rock record (Stow et al., 1983).

Major storm events that flush the sediment from canyon heads often persist for several days (Menard, 1964). This results in long periods of turbidity-current generation and may help explain the sinuous fan-channel morphology recently observed on sonographs. Sonographs (see Chapter 2,Marine Techniques) from middle and outer fans show well-developed channel meanders (Garrison et al., 1982; Damuth et al., 1983a). This sinuosity closely approximates that of river channels and suggests conditions of steady flow for some length of time during generation and deposition of deep-sea-fan valleys; it also points towards a high frequency of intermittent depositional events to maintain channel form.

All of these channel-building conditions are most likely to be caused by frequent storm-generated turbidity currents of many hours duration, rather

than instantaneous earthquake-generated events that are infrequent (Kastens, 1984). Not only channel sinuosity, but also the lower frequency, the larger size (Mutti et al., 1984), the widespread loci (Cita et al., 1984) and the different character of deposits from earthquake-generated events suggest that sedimentological or storm-generated events are the main processes responsible for development of channelized deep-sea fans. Large earthquake-generated events usually have a low frequency of every 100-1,000 years in very seismically active regions (Fig. 40) (Piper and Normark, 1983; Kastens, 1984). Large quantities of sediment typically are displaced or disrupted over a broad rather than a focused generating region (Fig. 39) (Menard, 1964; Piper and Normark, 1982; Field et al., 1982); Mutti et al., 1984). Large debris sheets also cover major portions of the surface of the Mississippi, Amazon, and Monterey Fans and the Ionian Abyssal Plain (Walker and Massingill, 1970; Damuth and Embley, 1981; Normark, et al., 1983; Cita et al., 1984; Hieke, 1984). Thus, some earthquake- and sedimentologic-generated events cause formation of major debris sheets over fans; however these are infrequent, affect widespread generating areas, and often may result in debris-sheet deposits, as opposed to storm-generated events, whose formation and deposition are focused in canyons and channels (Mutti et al., 1984).

Because the frequency of sedimentologically-generated turbidity currents is reduced in times of high sea levels, earthquake-generated mass-transport may become the dominant episodic depositional process during high sea-level time. The common occurrence of large and young (Nelson et al., 1983/1984) debris sheets on present-day fans and abyssal plains may be the result of this change in dominance from storm-generated to earthquake-generated regimes from late Pleistocene low sea levels to Holocene high sea levels.

Sedimentologically-generated turbidity currents not only occur with greater frequency and may result in different styles of deposition than those generated by earthquakes, but they also differ in the smaller amount of material and lower velocity in their flows (Fig. 40). Very little data is available on bottom-current speeds in canyons undergoing storm flushing of sediment, a sedimentologically generated event; however, rare measurements with maximum speeds up to 68 cm/s show relatively weak turbidity currents capable of transporting large quantities of sediment down canyon (Shepard et al., 1979). Diving during storm conditions in the Rio Balsas Canyon, Mexico revealed current pulses greater than 100 cm/s that transported suspended sand down the tributary canyon floor in a flow that was at least 3 m thick (Reimnitz, 1971). In contrast, the turbidity currents from the Grand Banks earthquake must have had velocities of at least 10 m/s to cause the sequence of cable breaks observed throughout the fan system (Fig. 39) (Heezen and Ewing, 1952; Menard, 1964). Thus, the few observed current speeds in sedimentological events appear to be an order of magnitude less than those calculated from the large-scale earthquake-generated events.

Classification of Sediment Gravity Flows and Turbidity-Currents

General

The previous discussion indicates that the ephemeral group of turbidity-current processes forming deep-sea fans has been extremely difficult to observe and measure in nature and that they exhibit a wide variation in characteristics of flow and deposits. As a result, a series of theoretical calculations and experimental studies have been undertaken to provide conceptual models of the depositing flows and sedimentary beds. A new terminology has evolved to describe turbidity-current processes and deposits based on these theoretical and experimental studies.

The entire spectrum of sediment flows driven by gravity, including turbidity currents as one type, has been labeled "sediment gravity flows" by Middleton and Hampton (1973, 1976). They note that in a fluid gravity flow, the fluid is moved by gravity and drives the sediment along, but in a sediment gravity flow, it is the sediment that is moved by gravity, and the sediment motion moves the interstitial fluid. Mechanisms such as suspension (by turbulence), saltation (by hydraulic lift forces and drag) and traction (by dragging or rolling of particles on the bottom) may all operate in some types of sediment gravity flows, as they do in some types of fluid gravity flows.

Middleton and Hampton (1973, 1976) have outlined a series of end members in the continuum of sediment gravity flow processes (Fig. 41) that take place from the time sediment movement initiates until it ceases at some depositional site downstream in the system. (Fig. 42). In all the processes, sediment particles move downslope parallel to the bed in response to gravity as long as there is sediment support in the flow to keep particles from settling out. The four end members are defined on the basis of the grain support mechanism as follows: (i) turbidity currents, in which the sediment is supported mainly by the upward component of fluid turbulence, (ii) fluidized sediment flows, in which the sediment is supported by the upward flow of fluid escaping from between the grains as the grains are settled out by gravity, (iii) grain flows, in which the sediment is supported by direct grain to grain collisions or close approaches and (iv) debris flows in which the larger grains are supported by a "matrix;" in other words, a mixture of interstitial fluid and fine sediment, that has a finite yield strength (Fig. 41).

The basic classification scheme of Middleton and Hampton has been slightly modified by Nardin et al. (1979b) and Lowe (1979, 1982). They base this modified classification scheme is based on flow rheology (fluid versus plastic behavior) as well as particle support mechanisms. Rock falls and landslides are included as initial phases of mass transport processes that affect submarine canyons and continental slopes feeding sediment into deep-sea-fan depositional sites.

Rock falls and landslides

Rockfall as a process in deep-sea fan sedimentation is mainly confined to canyons. We will not consider the theoretical aspects of this process because

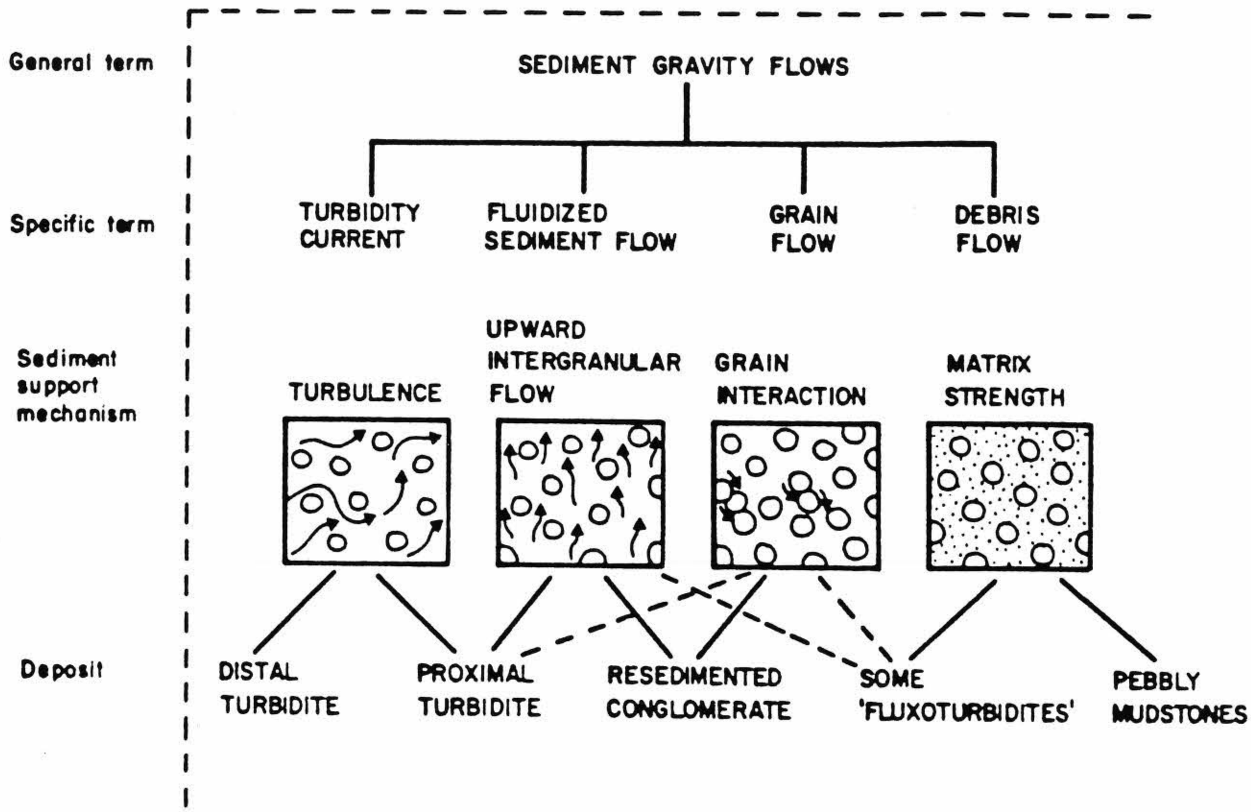


Figure **41** . Classification of subaqueous sediment gravity flows (from Middleton and Hampton 1976, Fig. 1).

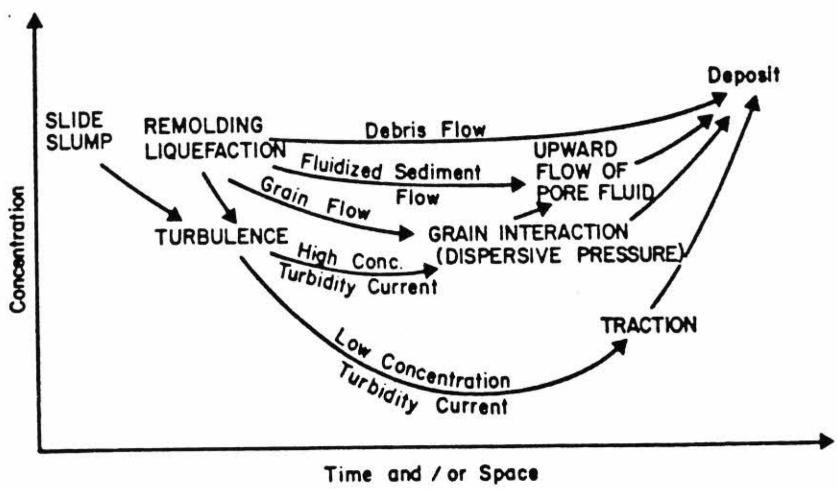


Figure 42. Hypothetical evolution of a single flow, either in time or in space (modified from Middleton and Hampton 1976, Fig. 10).

it is peripheral to most deep-sea-fan sedimentation. Like rockfall, landslides of both the translational gliding and the rotational slump type occur along the high-relief submarine canyon walls and continental slopes that border deep-sea fan deposits (LePichon and Renard, 1982; Coleman et al., 1983). Slumps also have been observed on sonographs, seismic profiles and during submersible dives over fan-valley walls (D.G. Moore, 1965; Nelson et al., 1978). Extensive translational-glide landslides may be possible over large fan areas (Embley and Jacobi, 1977; Bellaiche et al., 1981; Kolla et al. 1984). The translational landslides are distinguished by shear failure along discrete planes that can be observed in seismic profiles (Table 1). The rotational landslides or slumps are defined by discrete concave-up shear planes accompanied by rotation of the slide that also can be defined on profiles (Nelson et al., 1978). Because landslide processes are relatively well understood and have been extensively reviewed (Nardin et al., 1979b; Cook et al., 1982), we will not discuss them further in these notes.

Debris flows (mud flows)

Along open continental slopes and particularly in delta-front valleys bordering deep-sea fans (Coleman et al., 1983), landslide movement may often transform into debris flows if the landslide debris contains clay and is vigorously agitated during sliding. Agitation breaks down the physical structure of the debris and incorporates water into the sediment (Hampton, 1972). Breakdown of structure and incorporation of water decrease the strength of the landslide material and thereby causes it to move by the internal shear of debris flow mass movement rather than by the rigid sliding of landslides. Debris flows thus are characterized by shear throughout the sediment mass and clasts that are supported by the internal strength of the mud matrix and clast buoyancy (Hampton, 1972).

Debris-flow movement of mixtures of clasts, clay minerals and water on land essentially resembles flows of wet concrete; there have been no observations of such events under water and only their apparent deposits have been identified (Middleton and Hampton, 1976). Debris flows are episodic events and move in a series of waves or surges. Debris flows can transport large objects while moving slowly and have been observed to travel over slopes as low as 1-2° on land. Movement of a submarine debris flow depends on strength, submerged unit weight, and thickness of the debris, as well as the angle of the local slope (Hampton, 1972). Within a debris flow, this movement occurs where shear stress exceeds the shear strength of the debris; where shear stress is less than the strength, material is rafted along as a rigid plug. Granular solids in a debris flow are more or less "floated" during transport (Midddleton and Hampton, 1976). The clay minerals in water combine to act as a single fluid, and this fluid has finite cohesion "strength", a property not possessed by pure water. In addition, clay-water fluids have greater densities than pure water; thus, they provide greater buoyancy.

Hampton (1979) has shown that buoyancy increases with concentration of clasts. This can increase the competency and mobility of the flow because of reduced friction. Consequently, debris flows with high concentrations of clasts and low concentrations of clay (as little as 2%; Hampton, 1975) have a great deal of competence and mobility, and can move extensively over low slopes under water. Such characteristics may help explain the presence of (a)

	TYPES OF MASS TRANSPORT	INTERNAL MECHANICAL BEHAVIOR	TRANSPORT MECHANISM AND DOMINANT SEDIMENT SUPPORT	ACOUSTIC RECORD CHARACTERISTICS	SEDIMENTARY STRUCTURES AND BED GEOMETRY
	ROCKFALL		FREEFALL AND ROLLING SINGLE BLOCKS ALONG STEEP SLOPES.	STRONG HUMMOCKY BOTTOM RETURN, HYPERBOALE AND SIDE ECHOES COMMON. WEAK, CHAOTIC INTERNAL RETURN; STRUCTURELESS.	GRAIN SUPPORTED FRAMEWORK, VARIABLE MATRIX, DISORGANIZED. MAY BE ELONGATE PARALLEL TO SLOPE AND NARROW PERPENDICULAR TO SLOPE
SLIDE	TRANSLATIONAL (GLIDE)	ELASTIC	SHEAR FAILURE ALONG DISCRETE SHEAR PLANES SUBPARALLEL TO UNDERLYING BEDS. SLIDE MAY BEHAVE ELASTICALLY AT TOP; PLASTICALLY AT BASE AND THIN LATERAL MARGINS.	INTERNAL REFLECTORS CONTINUOUS AND OFTEN UNDEFORMED; ABRUPT TERMINATIONS. STRATA OF GLIDE BLOCKS MAY BE UNCONFORMABLE OR SUBPARALLEL TO UNDERLYING SEDIMENT.	BEDDING MAY BE UNDEFORMED AND PARALLEL TO UNDERLYING BEDS OR DEFORMED ESPECIALLY AT BASE AND MARGINS WHERE DEBRIS FLOW CONGLOMERATE CAN BE GENERATED. HUMMOCKY, SLIGHTLY CONVEX-UP TOP, BASE SUBPARALLEL TO UNDERLYING BEDS; 10'S TO 1000'S OF METERS WIDE AND LONG.
	ROTATIONAL (SLUMP)		SHEAR FAILURE ALONG DISCRETE CONCAVE-UP SHEAR PLANES ACCOMPANIED BY ROTATION OF SLIDE. MAY MOVE ELASTICALLY OR ELASTICALLY AND PLASTICALLY.	INTERNAL REFLECTORS CONTINUOUS AND UNDEFORMED FOR SHORT DISTANCES WITH DEFORMATION AT TOE AND ALONG BASE. CONCAVE-UP FAILURE PLANE AT HEAD AND SUBPARALLEL TO ADJACENT BEDDING AT TOE. SURFACE USUALLY HUMMOCKY.	BEDDING MAY BE UNDEFORMED. UPPER AND LOWER CONTACTS OFTEN DEFORMED. INTERNAL BEDDING AT ANGULAR DISCORDANCE TO ENCLOSING STRATA. SIZE VARIABLE.
SEDIMENT GRAVITY FLOW	DEBRIS FLOW OR MUD FLOW	PLASTIC	SHEAR DISTRIBUTED THROUGHOUT THE SEDIMENT MASS. CLASTS SUPPORTED ABOVE BASE OF BED BY COHESIVE STRENGTH OF MUD MATRIX AND CLAST BUOYANCY. CAN BE INITIATED AND MOVE LONG DISTANCES ALONG VERY LOW ANGLE SLOPES.	SEA FLOOR REFLECTORS MAY BE HYPERBOLIC, IRREGULAR, OR SMOOTH. COMMONLY ACOUSTICALLY TRANSPARENT WITH FEW OR NO INTERNAL REFLECTORS. MOUNDED OR LENS SHAPED WITH BLUNT TERMINATION AT HEAD. MAY BE CHAOTIC INTERNALLY.	CLASTS MATRIX SUPPORTED; CLASTS MAY EXHIBIT RANDOM FABRIC THROUGHOUT THE BED OR ORIENTED SUBPARALLEL, ESPECIALLY AT BASE AND TOP OF FLOW UNITS; INVERSE GRADING POSSIBLE. CLAST SIZE AND MATRIX CONTENT VARIABLE. OCCUR AS SHEET TO CHANNEL-SHAPED BODIES CM'S TO SEVERAL 10'S OF METERS THICK AND 100'S TO 1000'S (?) OF METERS LONG; WIDTHS VARIABLE.
	GRAIN FLOW		COHESIONLESS SEDIMENT SUPPORTED BY DISPERSIVE PRESSURE. USUALLY REQUIRES STEEP SLOPES FOR INITIATION AND SUSTAINED DOWNSLOPE MOVEMENT.	INDIVIDUAL FLOW DEPOSITS VERY THIN; MAY NOT BE RESOLVABLE WITH PRESENT SEISMIC-REFLECTION TECHNIQUES. REPEATED FLOWS MAY PRODUCE A SEQUENCE OF THIN, EVEN, REFLECTORS.	MASSIVE; CLAST A-AXIS PARALLEL TO FLOW AND IMBRICATE UPSTREAM, INVERSE GRADING MAY OCCUR NEAR BASE.
	LIQUEFIED FLOW		COHESIONLESS SEDIMENT SUPPORTED BY UPWARD DISPLACEMENT OF FLUID (DILATANCE) AS LOOSELY PACKED STRUCTURE COLLAPSES; SETTLES INTO A TIGHTLY PACKED TEXTURE. REQUIRES SLOPES $>3°$		DEWATERING STRUCTURES, SANDSTONE DIKES, FLAME AND LOAD STRUCTURES, CONVOLUTE BEDDING, HOMOGENIZED SEDIMENT.
	FLUIDIZED FLOW	FLUID	COHESIONLESS SEDIMENT SUPPORTED BY UPWARD MOTION OF ESCAPING PORE FLUID. THIN (<10 CM) AND SHORT-LIVED.		
	TURBIDITY CURRENT FLOW		CLASTS SUPPORTED BY FLUID TURBULENCE. CAN MOVE LONG DISTANCES ALONG LOW ANGLE SLOPES.	THIN, EVEN, CONTINUOUS, ACOUSTICALLY HIGHLY REFLECTIVE UNITS; ONLAPS SLOPE OR RAISED TOPOGRAPHY. DISCONTINUOUS, MIGRATING AND CLIMBING IN CHANNEL SEQUENCES.	BOUMA SEQUENCES. MM'S TO SEVERAL 10'S OF CM THICK. 10'S TO 1000'S OF METERS IN LENGTH; WIDTHS VARIABLE.

Table 1. Major Types of Submarine Mass Transport on Slopes and Suggested Criteria for their Recognition (modified from Cook et al. 1982, Table 1; modified from Nardin et al. 1979).

debris sheets extending for 10's to 100's of km that have recently been observed on deep-sea fans and (b) sandy debris flow deposits that may closely resemble massive sand beds deposited by grain and fluidized flows (Middleton and Hampton, 1976).

Solid grains in debris flows are mainly transported by the support matrix, but they also may be carried by traction as bedloads (Middleton and Hampton, 1976). Deposition from debris flow occurs by mass emplacement when the driving stress of gravity decreases below the strength of the debris and "freezing" occurs. Textures of muddy debris-flow deposits resemble those of tillites, but sandy debris-flow deposits with low clay content (Fig. 43) (Middleton and Hampton, 1976) can have textures similar to other sandstones. Sandy debris flow deposits may contain lamination perhaps with normal or inverse grading--attributes that also may characterize grain flows (Hampton, 1975). Internal sedimentary structures may include pull-aparts and other features observed in natural subareal and experimental debris flows (Fig.43) (Middleton and Hampton, 1976).

Grain flows

Grain flows as avalanching of sand have been observed by divers in the upper reaches of submarine canyons (Shepard and Dill, 1966). These flows are believed to transport canyon-floor sediment to upper-fan valleys and to erode the bedrock walls and to erode of canyons (LePichon and Renard, 1982). Identification of grain-flow deposits in deep-sea cores or outcrops, however, is equivocal because the originally defined type locality of grain-flow deposits (Stauffer, 1967) more recently has been interpreted as a high-density turbidity-current deposit (Lowe, 1982).

Middleton and Hampton (1976), in their conceptual model of sediment gravity flows, defined grain flows on the basis of grain-to-grain interaction or dispersive pressure counteracting the tendency for the grains to settle out of the flow. In the modified scheme of Lowe (1979) and Nardin et al. (1979b) based on mechanical behavior, grain flows are considered to represent a transition between plastic behavior and fluid behavior. High-concentration grain flows have frictional strength and plastic behavior. At low volume concentrations, grain flows are probably better described by fluid rather than by plastic behavior (Table 1).

Theoretical calculations suggest that grain-flow transport would require slopes from 18° up to more than 30° (Middleton and Hampton, 1976). Because such slopes are rare on a regional scale on the ocean floor, pure grain flows probably do not account for significant long-distance transport of marine sediment and are volumetrically unimportant in deep-sea-fan deposits (Hampton, personal communication, 1984). Bagnold (1956) and Middleton (1970), however, suggested that grain flows may move on lower slopes if the density of the interstitial fluid is greater than pure water. This would increase the buoyancy and reduce the need for dispersive pressure to support grains. An interstitial fluid density of 2.0 g/cm^3 would reduce the slope angles to 10.5° (Middleton, 1970). In addition, turbulence may be present under high-energy flow conditions, and this would also reduce the need for grain support by dispersive pressure (Middleton and Hampton, 1976). In natural sediment gravity flows with high energy, turbulence, and interstitial fluid density, modified grain flows may be important in depositing the massive basal part

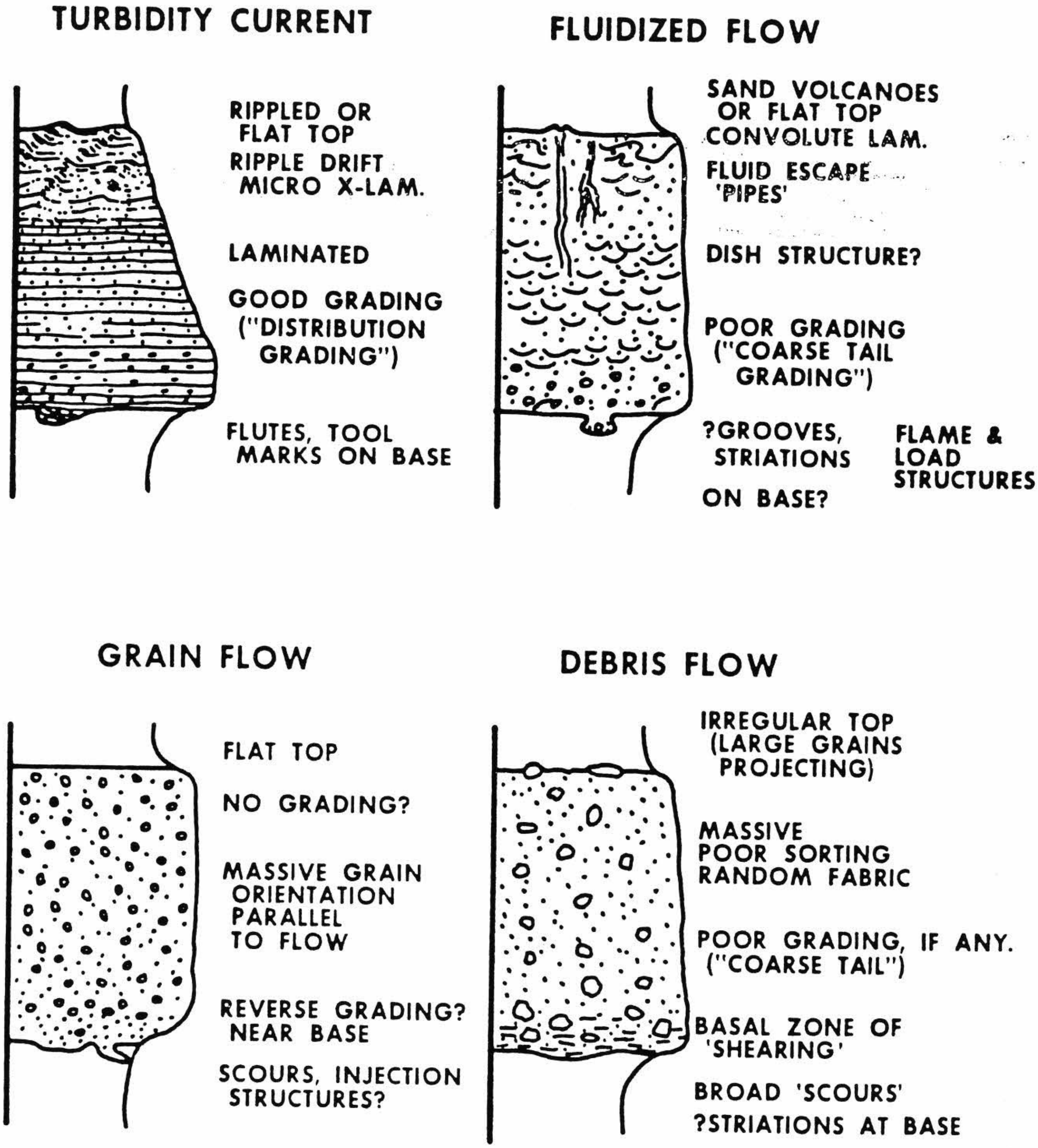


Figure **43**. Idealized sequences of sedimentary textures and structures in hypothetical single-mechanism deposits of deep-water coarse clastic sediments (from Siemers and and Tillman 1981, Fig. 2; as modified from Middleton and Hampton 1976).

(Bouma T_a) of turbidites (Fig. 44) (Lowe, 1979, and 1982; Nardin et al., 1979b).

Deposition from grain flows is fundamentally by mass emplacement (Middleton and Hampton, 1976). In contrast to the particle-by-particle deposition from suspension and traction in fluid flows, mass emplacement theoretically involves a sudden freezing of the flow, resulting in quick deposition of a layer several grains thick. Freezing occurs when the driving stress becomes less than that necessary to move the sediment.

Deposits from experimental grain flows show both lateral and vertical inverse grading, particularly near the base (Middleton and Hampton, 1976). An analogy would be shaking a box of different-sized sand and noting that the finer grains will fall sieve-like to the bottom and produce inverse grading (Middleton and Hampton, 1976). Although the mechanics of pure grain flows are very distinct and inverse grading is characteristic, their identification in natural deposits remains problematic (Lowe, 1982), perhaps because most natural deposits of this type result from modified grain flows.

Fluidized and liquefied flows

Fluidized and liquefied flows, like grain flows, may represent a transient phase between the initiation of sediment movement by slides and the eventual mature turbidity-current flow in deep-sea fan channels. Fluidized and liquefied flows require that cohesionless sediment be supported by the upward motion of escaping pore fluid; in the former case, full support for grains is provided by pore fluid and in the latter case, only partial support is provided as grains settle through pore fluid (Lowe, 1979). Loosely packed sediment, and fine sand to coarse silt particularly, are subject to liquefaction (excess pore pressures) because of cyclic loading due to storm waves or seismicity (Clukey et al., 1980; Clukey et al., 1983). As long as the grains are supported by pore fluid, the sand has little strength, behaves as a fluid, and can flow rapidly down slopes of 3-10° (Middleton and Hampton, 1976).

Storm-wave surfbeat (Reimnitz, 1971) or storm-wave cyclic loading on fine sand to coarse silt appear to be able to create a liquefied state to several meters depth (Clukey et al., 1980; Clukey, 1983) for an extensive storm period during which lengthy sediment transport (>100 km) and resuspension may persist even on nearly flat shelf slopes (Nelson, 1982). When sediment liquefaction is combined with storm-driven downwelling bottom currents of the inner shelf (Nelson, 1982; Swift et al., 1983) particularly in submarine canyon heads (Shepard et al., 1979), significant sediment movement and flushing from submarine canyons to deep-sea fans may be initiated. Similarly, widespread liquefaction structures caused by an earthquake and observed on the shelf (Field et al., 1982) suggest liquefaction may be initially important in earthquake-induced sediment gravity flows.

Fluidized and liquefied flows deposit sediment because of loss of pore fluid that is expelled upwards (Middleton and Hampton, 1976). As a result, fluid escape pipes and dish structures are characteristic (Fig. 43) (Lowe and Lo Piccolo, 1974). Post-depositional liquefaction of poorly consolidated sediment, resulting from sudden loading or shock, may also cause development of these same structures (Nardin et al., 1979b).

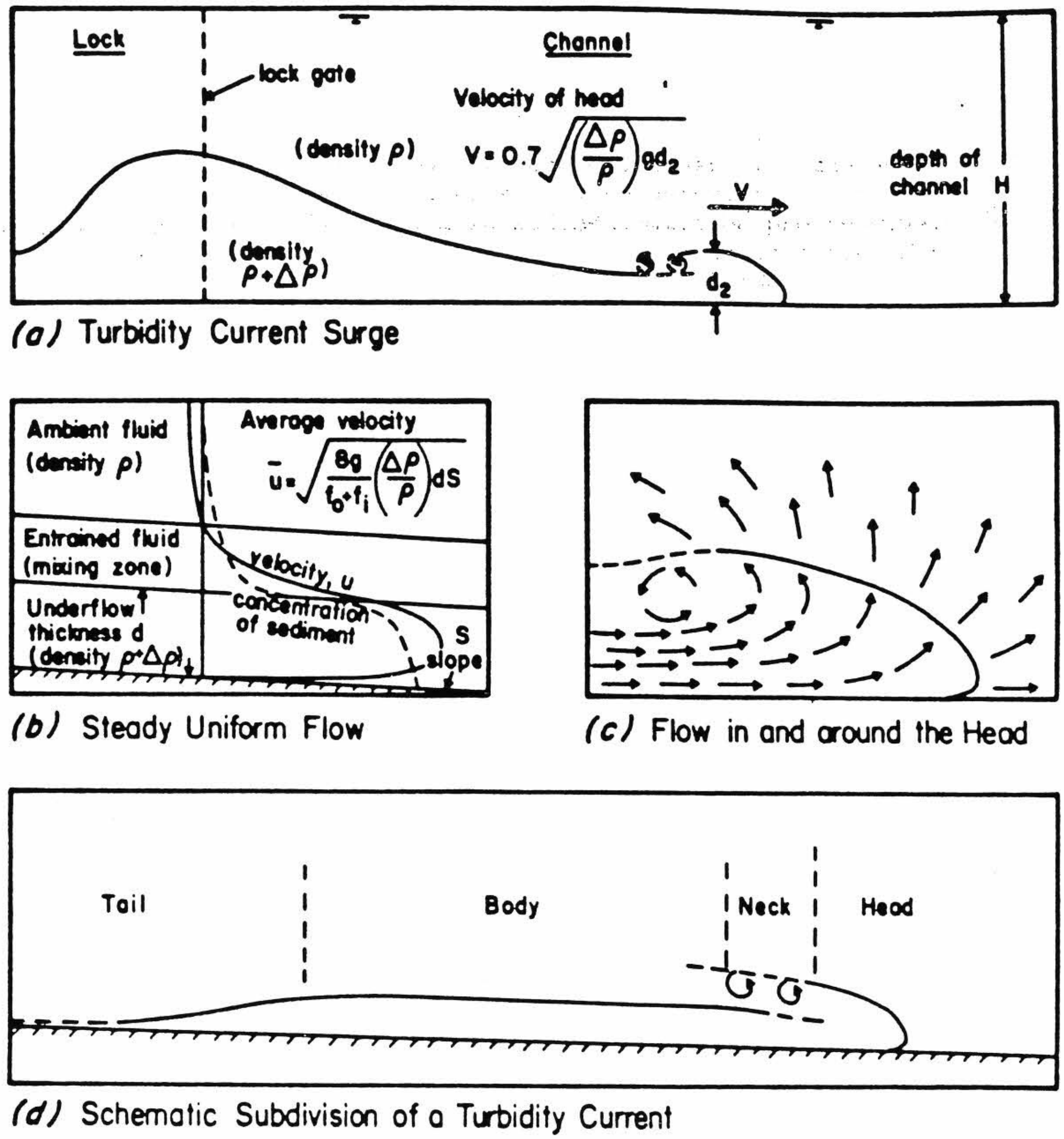


Figure **44** . Hydraulics of turbidity currents (from Middleton and Hampton 1976, Fig. 3). (a) Turbidity-current surge, as observed in a horizontal channel after releasing suspension from a lock at one end. The velocity of the head $\underline{v}$, is related to the thickness of the head $\underline{d_2}$, the density difference between the turbidity current and the water above $\underline{\Delta\rho}$, the density of water $\underline{\rho}$, and the acceleration due to gravity $\underline{g}$. (b) Steady, uniform flow of a turbidity current down a slope $\underline{s}$. The average velocity of flow $\underline{u}$ is related to the thickness of the flow $\underline{d}$, the density difference, and frictional resistance at the bottom $\underline{f_0}$ and upper interface $\underline{f_i}$. (c) Flow pattern within and around the head of a turbidity current. (d) Schematic division of a turbidity current into head, neck, body, and tail.

Turbidity currents

Turbidity currents depositing on deep-sea fans appear to have two typical sites of generation and pathways of flow to depositional sites on the deep-sea fans. Most turbidity currents appear to originate in submarine canyons, follow a confined pathway through upper- and middle-fan valleys, and spread as sheet flow across the outer fan and basin or abyssal-plain floor (Nelson and Kulm, 1973). Other turbidity currents are generated by earthquakes or sediment-failure events on slopes surrounding the basin floor and move down the open slope and across the fan surface as unchannelized sediment gravity flows.

Two basic modes of flow also occur in turbidity currents. The first is a constant uniform flow of a denser fluid below a less dense fluid, for example the well-documented case of the Colorado River at flood stage entering Lake Mead (Gould , 1951).The second mode of turbidity-current flow is a surging flow usually generated by episodic storm and earthquake events. Such currents last only as long as there is a continued supply of the sediment/water mixture produced by the initial catastrophe (Middleton and Hampton, 1976). The surge-type turbidity currents have been modeled in the laboratory by numerous investigators who released denser sediment/water mixtures into less dense fresh water (Kuenen and Migliorini, 1950; Middleton, 1970; Luthi, 1981).

Surging turbidity currents vary from: (1) those with a low density (approximately 1.01 g/cm^3) that travel with generally low speeds (approximately less than 25 cm/s) to (2) those with high density (typically 1.1 g/cm^3 or greater) that travel at high velocities of approximately 1 to greater than 10 m/s. The low-density types typically develop from storm-wave resuspension on the upper continental margin that creates turbid-layer flows or nepheloid layers movingacross the open slope or down canyons and fan valleys (Duncan et al., 1970b) to basin floors (D.G Moore, 1969). High-density turbidity currents also generate low-density and velocity suspension clouds that spread beyond channels and lag behind the main channelized high-density flow. The high-density flows are produced by the larger, episodic events already discussed.

High-density turbidity currents generally begin from an initial sediment gravity flow, most likely debris flows (Fig. 40 ; Table 1) (Middleton and Hampton, 1976). Experiments indicate that the turbidity-currents are generated by erosion of sediment from the front of the debris flow, where it is abruptly thrown into turbulent suspension (Hampton, 1972; Middleton and Hampton, 1976).

Once the initial input of energy from debris or other sediment gravity flows form the turbidity current, it may continue almost indefinitely (Middleton and Hampton, 1976). A hypothetical state of dynamic equilibrium or autosuspension (Bagnold, 1962) may result because there is a feedback loop of sediment maintained in suspension by turbulence. This turbulence is generated by downslope flow, which is caused by gravity acting on the more dense suspension, which in turn is caused by the turbulence. To maintain flow, the gravitational pull downslope must compensate for energy lost to friction. Consequently, turbidity currents can move over extremely low slopes for great

distances once they are generated, particularly when their movement is confined in a channel.

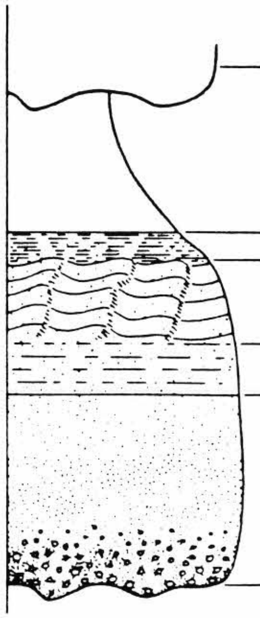
Turbidity currents, once traveling away from their source, divide into three main parts: the head, the body, and the tail (Fig. 44) (Middleton and Hampton, 1976). As Middleton and Hampton (1976) state, this flow morphology has several important consequences: (1) the head may be a region of erosion, forming scour marks even when deposition is taking place from the body and tail; (2) the coarse sediment will be progressively concentrated in the head; (3) the head will generally be at least twice as thick as the body, except on very low slopes, where Komar (1972) suggests the body may be as thick as the head and spill over channel levees.

Grain-by-grain deposition by gravity from a turbulent suspension results in the characteristic vertical grading of texture and the sequence of sedimentary structures (Bouma T_a-T_b) observed in turbidites (Fig. 45) (Middleton and Hampton, 1976). Because most turbidity currents are the result of surge-type generation, the turbidite beds are characterized by pulses of sedimentation that are identified by varying breaks in textural gradation separating the vertical sedimentary structures (Lowe, 1982). Bedding-plane structures in turbidites, such as flute casts, groove casts, and so on, can be explained by the erosion caused by the head of the turbidity current, whereas the vertical sequence of sedimentary structures can be attributed to the rapidly waning deposition from the head, the body, and the tail of the current as it passes over a specific location with varying flow regimes (Middleton and Hampton, 1976).

In the idealized, complete turbidite bed, initial deposition takes place rapidly from a high flow regime that gives rise to a quick bed deposition without structures (T_a) (Fig. 45) (Middleton and Hampton, 1976). Recent experiments show deposition of T_a beds from a thin layer with high-density bedload flow that does not represent settling out from suspension, as do the overlying intervals T_b to T_e (Sanders, 1965) (Luthi, 1981). The overlying T_b layer is developed from traction-influenced sedimentation (Fig. 47) (Lowe, 1982) during plane-bed deposition of a high flow regime (Sanders, 1965). The remainder, the T_c through T_e sequence, deposits from low flow regimes resulting in T_c ripple drift or convolute lamination overlain by T_d flat lamination and then structureless T_e muds. We divide the T_e division into the T_{et} turbidite mud capped by the T_{ep} pelagic mud that deposits after all the final suspended sediment deposition of the turbidity current event is completed (Fig. 41). Development of vertical sedimentary structures is modified by the location within the deep-sea-fan system (Nelson and Kulm, 1973) and by varying the density of the individual turbidity-current events. These aspects can be demonstrated experimentally for sheet-flow turbidity-current events (Fig. 46) (Luthi, 1981) and will be described more completely in later discussions of turbidite facies.

Summary

Because some sediment-gravity-flow types evolve from one another, and because deposition may occur at any time during the flow, an individual bed may represent deposition by a combination of sediment-gravity-flow processes (Figs. 42 and 47) (Middleton and Hampton, 1976). Other mechanisms, such as traction, may operate during the last stages of deposition and produce or modify textures and structures observed in the sediment bed that is finally



Grain Size		Bouma (1962) Divisions	Interpretation
Mud	T_{ep}	Pelite	Pelagic sedimentation
	T_{et}	Massive or graded Turbidite	fine grained, low density turbidity current deposition
Sand-Silt	T_d	Upper parallel laminae	? ? ?
	T_c	Ripples, wavy or convoluted laminae	Lower part of Lower Flow Regime
Sand (to granule at base)	T_b	Plane parallel laminae	Upper Flow Regime Plane Bed
	T_a	Massive, graded	? Upper Flow Regime Rapid deposition and Quick bed (?)

Figure **45** Idealized sequence of sedimentary textures and structures in a classical turbidite, or Bouma sequence (modified from Middleton and Hampton 1976).

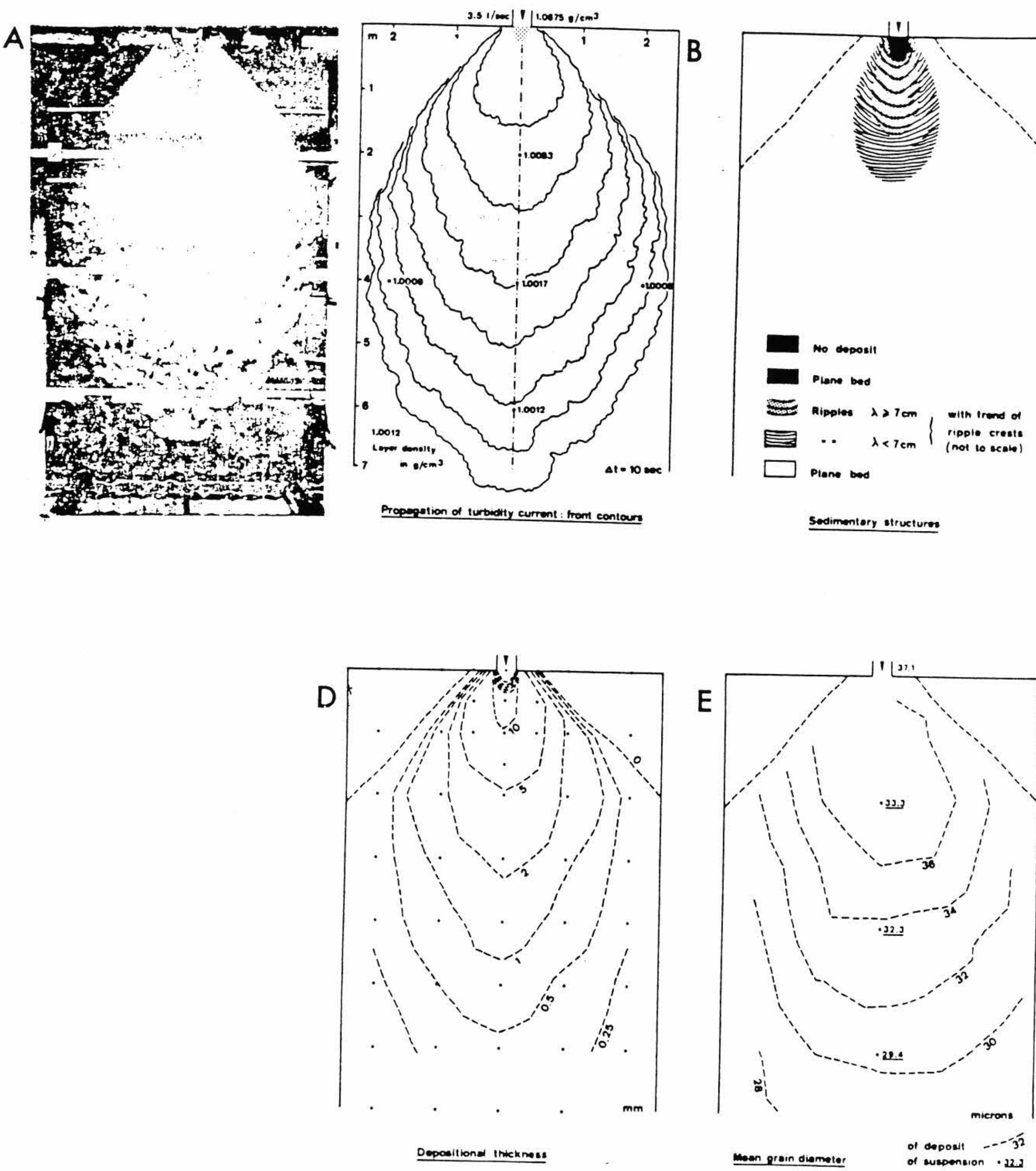


Figure **46** A - Photograph of a turbidity current after 70 seconds generated in an experimental tank (from Luthi 1981, Fig. 2). B to E - Anatomy and characteristics of a nonchannelized quartz silt turbidity current and the resulting turbidite generated in an experimental tank.

deposited from the flow. Thus, it is very difficult to distinguish between deposits formed from the different types of flows, except for the distinctive vertical sequence of structures created by deposition from turbulent suspension of turbidity currents (Fig. 3-9) (Lowe, 1982).

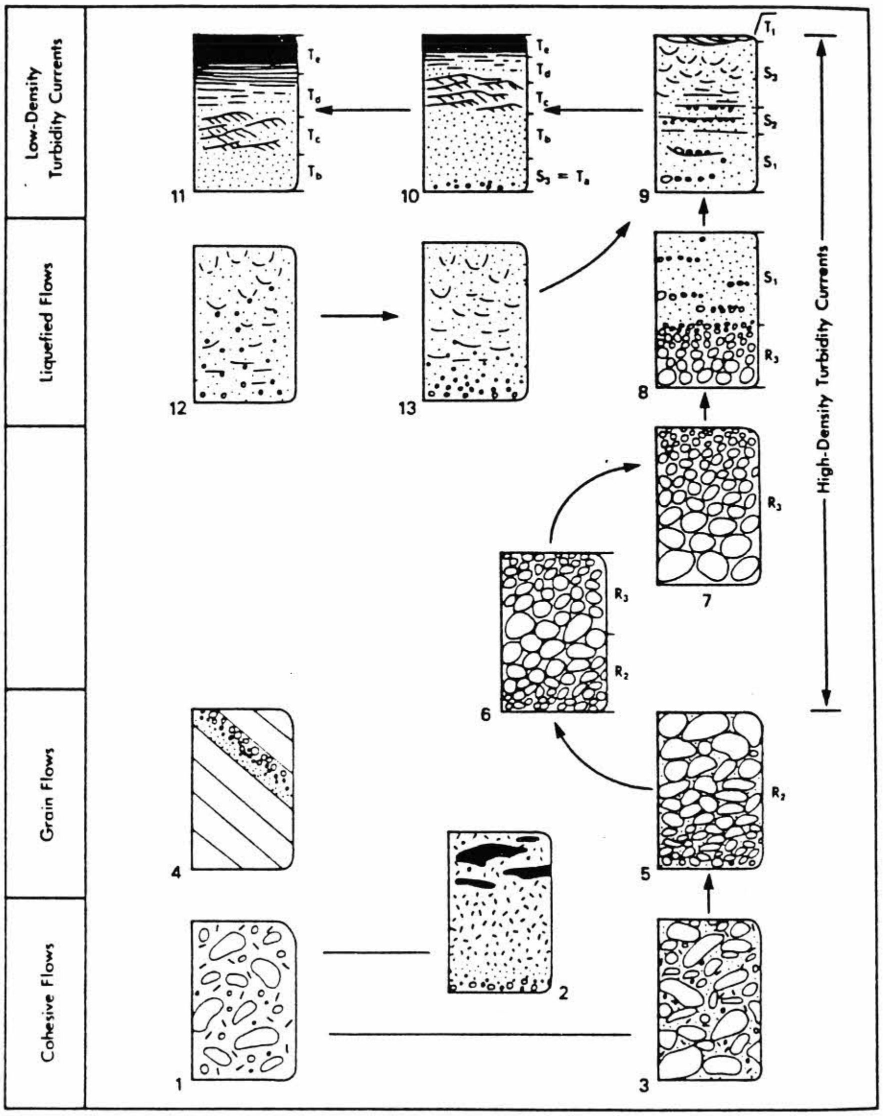


Figure **47** . Summary of the main deposit types formed during deposition from sediment gravity flows (from Lowe 1982). Lines without arrows (i.e., 1-2 and 1-3) connect members between which there probably exists a continuous spectrum of flow and deposit types but which are not parts of an evolutionary trend of single flows. Arrows connect members which may be parts of an evolutionary continuum for individual flows. The transition from disorganized cohesive flows (1 and 3) to thick, inversely graded density-modified grain flows and traction carpets (5) and to turbulent gravelly high-density turbidity currents (6) is speculative but may occur.

The following paper on submarine fan models was extracted from:

Nelson, C.H., and Nilsen, T.H., 1984, Modern and Ancient Deep-Sea Fan Sedimentation: SEPM Short Course No. 14, 404 p.

CHAPTER 10. MODELS OF ANCIENT TURBIDITE DEPOSITS

Tor H. Nilsen

Introduction

The Mutti and Ricci Lucchi (1972) model, which has been discussed in detail, including both facies and facies association, is only one of many models that have been published (Fig. 48). It has gained much acceptance in most ancient deep-sea fan systems. However, like most models, it does not apply everywhere or apply to all deep-marine clastic systems; it has been modified extensively since 1972 and is now known to be but one end-member of a spectrum of different types of models that are applicable to different types of deep-marine clastic systems. The Mutti and Ricci Lucchi (1972) model does provide, in my opinion, a well-documented and easily understood framework to which other models and proposed departures from the model can be compared. I will summarize herein the evaluation of fan models and concepts regarding the application of models.

Early Models

Many workers, both in modern and ancient, attempted in the 1960s to model the sedimentation of deep-water clastics. Jacka et al. (1967, 1968) proposed a simple model that divided fans into proximal, intermediate, and distal parts (Fig. 49). Simple modifications of this system, such as the model by Nilsen and Simoni (1973) for the Butano Sandstone of northern California (Fig. 50) served only to emphasize the observation that the more proximal parts of deep-sea fans contained coarser sediment and larger channels compared to the more distal parts.

Nelson et al. (1970) studied the modern Astoria Fan system and provided a useful model for some studies because Astoria Fan clearly showed a downfan decrease in the size of its channels, abandonment of channels, bifurcation and division of channels, and transition from channeled to nonchanneled fan (Fig. 51). Haner (1971) proposed a model for the modern Redondo Fan of offshore southern California that involved a downfan change in channel character from straight to meandering to braided (Fig. 52); she also attempted on the basis of piston cores from various parts of the fan to relate the fan deposits to variations in the Bouma sequence.

Fundamental Models

One of the most important models proposed for modern deep-sea fans is that of Normark (1970, with modifications in 1974, 1978). On the basis of study of several relatively small deep-sea fans in the Southern California Borderland, he proposed that a topographic bulge on the middle part of these fans, which he referred to as the suprafan, was the locus of fan sedimentation (Fig. 53). The suprafan bulge, located directly downfan from a larger leveed inner-fan channel, was thought to have a radiating system of channels on its upper part and a smooth surface on its lower part. By lateral shifting of the suprafan lobes over a period of time, a fan-shaped body of sediment would be constructed (Fig. 54).

Walker (in Walker and Mutti, 1973) proposed a model which brought together the chief elements of both Normark's (1970) suprafan model and the Mutti and Ricci Lucchi (1972) model with outer-fan lobes (Fig. **55**). One result of this combining of a model for small modern deep-sea fans with one for ancient deep-sea fans has been some confusion in the literature about the term "lobe" (see Nilsen, 1980 and reply by Walker, 1980 for a discussion of some aspects of this problem).

Nelson and Nilsen (1974) attempted to generate a model for deep-sea-fan deposits through a detailed comparison of the modern Astoria Fan and the Eocene Butano Fan of the California Coast Ranges. Their comparison showed a major difference in scale, types of data, and methodology (Table **2**).

Mutti and Ghibado (1972) emphasized the similarities between channelized versus nonchannelized sedimentation on fans and deltas (Fig. **56**). Mutti and Ricci Lucchi (1974, 1975) added to their original model, emphasizing the fundamental aspects of channeled versus nonchanneled parts of the fan (Figs. **57** and **58**). In this model, they have shown the outer-fan lobes detached from the feeding channels. However, Ricci Lucchi (1975), in a paper that emphasized the cyclic nature of deep-water clastic deposits, used a model (Fig. **59**) similar to the original model of Mutti and Ricci Lucchi (1972).

Van Vliet (1978) proposed a modification of the Mutti and Ricci Lucchi (1972) model for a lower Tertiary basin in northern Spain in which the outer-fan lobes shift in the same direction, probably as a result of a plunge to the basin floor (Fig. **60**). His model combines the concepts of outer-fan lobes and suprafan lobes, although his representative stratigraphic sections (Fig. **61**) are very similar to those proposed by Mutti and Ricci Lucchi (1972).

Walker (1978) suggested a model for deep-marine clastics that has been widely accepted and used in many areas (Fig. **62**). This model, like his earlier model, combines the concepts of the outer-fan lobe of Mutti and Ricci Lucchi (1972) and the suprafan lobe of Normark (1970) into a single model. Walker (1978) showed in cross-section how the lobes on loci or deposition shifted laterally through time, with deposition of muds during intervals of inactive sedimentation (Fig. **63**).

Models Constrained by Sediment Grain Size

Workers in California and Italy recognized two distinct types of deep-sea fans that had different characteristics. The first type is similar to the fans described initially by Mutti and Ricci Lucchi (1972), which are small to large in size, are composed of a mixture of grain sizes, have well-developed outer-fan lobes and fan-fringe deposits, and are surrounded by a basin plain on which abundant sandstone turbidites are deposited. Mutti (1979) referred to these fans as highly efficient, implying that turbidity currents were efficient in transferring sediment to the fans and were capable, because of the mixture of sediments, of transporting sediments for long distances out on to the basin floor (Fig. **64**).

The second type of fan is small, composed mostly of sand, does not have well-developed outer-fan lobes or fan-fringe deposits, and is surrounded by a basin plain on which few sandstone turbidites are deposited. Mutti (1979) referred to those fans as inefficient, implying that the highly concentrated,

sand-laden sediment gravity flows were not capable of transporting the sediments for long distances out onto the basin floor (Fig. **65**). Instead, the sand was dumped rather abruptly at the base of the slope, producing a small, very sandy fan.

We now refer to those two types of fans as mixed-sediment fans and sand-rich fans. It is increasingly clear that the Mutti and Ricci Lucchi (1972) model is applicable to mixed-sediment fans, which form the majority of fans in the ancient record, and that Normark's (1970) suprafan model is applicable to sand-rich fans. If this is the case, then the Walker (1978) model, which combines these two into a single model, may not be applicable to any particular fan. Ancient sand-rich fans have been described in California (Nilsen, 1979b, 1981; Link and Nilsen, 1980; Link, 1981; Link and Welton, 1982) and elsewhere. Models for the Eocene, the rocks sandstone and cretaceons Chatworth Formation in California are shown on Figures **66** and **67** , respectively.

The lower Tertiary Cantua Sandstone of central California is an excellent example of a sand-rich fan (Fig. **68**). It consists almost wholly of channelized thinning- and fining-upward cycles (Fig. **69**) surrounded by basin-plain deposits that are composed of mudstone with a few interbeds of sandstone (Fig. **70**). It is a small deep-sea fan of arkosic composition deposited in a small restricted basin.

As additional knowledge has been gained from the study of both modern and ancient deep-sea fans, it has become clearer that there is a complete spectrum of different types of fans whose internal geometry and characteristics are dependent upon the grain-size distribution of the sediments deposited on the fans. In Figures **71** and **72** , I have attempted to show the character of two other types of accumulations. Debris wedges represent the assumulation of variably sized debris at the base of slope, without the development of channelized and nonchannelized facies. Debris wedges can develop at the base of siliciclastic slopes (Nardin et al., 1979) and carbonate slopes (Cook et al. 1981); instead of true submarine canyons, chutes develop on the slope and carbonate or mud-rich breccias slide down these chutes to accumulate as a poorly sorted wedge of debris at the base of slope.

Deep-sea fans composed mostly of fine-grained sediment are not well known in the sratigraphic record, but are abundant in modern oceans. Fans of this type can be extremely large and appear to be dominated by long channel-levee complexes that may extend for the entire length of the fans.

Summary

Work on ancient turbidite systems has yielded a number of useful models for basin analysis. Fans of different types can be constructed in the same basin where the types of sediment supplied have different grain-size distributions. We can probably confidently recognize five types of deep-sea fan/wedge accumulations in the ancient record on the basis of sediment supply: (1) breccia-supplied debris wedges, (2) gravel-supplied very small conglomeratic fans, (3) sand-supplied small sandstone-rich fans, (4) mud-supplied mudstone-rich fans, and (5) mixed-sediment supplied fans of the type originally described by Mutti and Ricci Lucchi (1972). Many other factors control the geometry and morphology of deep-sea fans, and these will be addressed in the concluding chapter of these notes.

Characteristic	Submarine canyon	Physiographic regions of fans				Depositional environments of fan regions		
		Upper fan	Middle fan	Lower fan	Fan fringe	Channel	Levee	Interchannel
A Physiography	Generally straight, V-shaped, and deep (hundreds of meters); commonly steep-walled with cliffs and overhangs	Few main channels (generally one or two) that are very deep (100+ m in large fans), steep-walled and wide (several kilometers); levees are well developed; most irregular channel topography of any fan region	Few main channels branching into tens of distributaries; (up to tens of meters deep); width variable (tens of meters to tens of kilometers); steep to gently sloping walls; channels may meander	Numerous shallow, narrow, braided, randomly shifting, distributary channels having gently sloping walls; limited or lack of levee development; size intermediate between middle and fan-fringe channels	Flat surface having a limited number of gentle channels of few meters width and depth in open basin fans or may lack channels in fans of restricted basins	U-shaped, leveed, flat-floored; typically depositional-erosional features on the fan surface	Sediment wedges up to tens of meters thick flanking channels in upper to middle fan areas; range from tens to hundreds of kilometers wide depending on fan size	Generally smooth, flat, gently sloping, and concave upward surfaces except in the middle fan where convex suprafan bulges may occur
B Stratigraphy and sedimentary facies	Thick (up to tens of meters) coarse-grained beds having highly variable and poorly developed bedding; fining upward to fewer and thinner coarse-grained beds in muds	Abrupt vertical and lateral distribution of channel and interchannel facies; channel bodies enclosed within interchannel facies, and isolated blocks of levee facies enclosed within channel facies	Like upper fan region evolving to distinctly bedded and structured channel facies having thin well-developed mud interbeds associated with limited interchannel facies	Dominant channel facies throughout section evolving to dominant interchannel facies containing fewer and thinner coarse-grained beds; channel beds as much as 5 to 10 m thick	Thin (5 to 30 cm) coarse-grained beds of even thickness over great lateral extent and a greater amount of mud beds	Like canyon but thick coarse-grained, lenticular beds dominate; mud interbeds cut out or incompletely developed except in fining upward sequence	Thin (<15 cm) well-bedded lenticular coarse-grained layers alternating with equal amounts of fine-grained beds in rhythmic sequence	Very thin and evenly bedded coarse-grained layers alternating with greater amounts of interbedded mudstone
C Sediment types	Dominantly sandstone conglomerate, or pebbly mud; silty mud important on canyon walls or in fining upward parts of canyon fill	Conglomerates and sandstones in channel beds flanked by fine sandstone and siltstone in levee and interchannel areas; silty mudstones limited in channel environments, but dominant in interchannel	Sandstone in channels; interchannel having sandstone and siltstone; limited silt and clay-sized mudstone	Medium- to fine-grained channel sandstone and interchannel sandstones and siltstones; clay-sized mudstone most common in interchannel areas	Clay-size mudstones dominant but numerous fine-grained sandstones and siltstones throughout area	Dominantly sandstones, conglomerates, or muddy conglomerates and limited mudstones	Fine-grained sandstone, siltstone, and mudstone in equal amounts	Sandstones, siltstones, and mudstones in variable amounts in different regions of the fan
D Sedimentary structures (see also table 1)	Bouma ae; typically lacking or poorly developed internal structure; may be highly bioturbated in muds; commonly very highly contorted; slump structures common; beds typically amalgamated in lower part of valley fill sequence	Channel is like canyon; Bouma ae predominant; some ab possible; coarsely laminated beds having dish structure, mudstone ripups and clasts, very highly contorted beds and slump structures common; channel sands and gravels injected into interchannel beds; interchannel like levee environment	Full Bouma abcde sequences common in channels; typically less complete Bouma sequences in interchannels; muds most bioturbated in interchannel	Bouma bcde sequences common in channels and Bouma cde in interchannels; base of beds flat and typically lacking sole marks; mud most bioturbated in interchannel	Bouma cde grading downfan to de; lamination dominant; cross lamination limited; muds most highly bioturbated of any fan environment	Bouma cde, typically poorly developed if present; beds displaying well-developed sole marks and channeled basal contacts typical	Some Bouma cde or de units convoluted and typically cross laminated, showing rippling and starved ripples characteristically at top of sandstone and siltstone beds	Similar to levee but generally thinner bedded and laminated in upper fan; in midlower fan, like channel deposits but thinner bedded and more base cutouts
E Grain-size distribution in coarse-grained beds	Very poorly sorted, very positively skewed	Very poorly sorted, very positively skewed; size grading vertically is generally slight or lacking	Poorly to moderately sorted, positive to neutral skewness, size graded vertically	Moderately sorted, slightly positive to negative skewness, size graded vertically	Positively skewed, poorly sorted, size graded vertically	Varies downchannel; see description from upper to middle fan areas	Moderately sorted, not highly skewed	Moderately to poorly sorted, not highly skewed
F Sandstone/shale ratio	Generally very high except in upper part of total channel fill	Very high in channel facies, very low in interchannel facies, and intermediate in levee facies	High in channel facies, low to intermediate in interchannel facies	High in channel, intermediate in interchannel facies	Low to very low throughout facies	High throughout but may become lower in upper part of channel fill	Intermediate and higher than interchannel and distal fringe areas	Low except in suprafan areas
G Conglomerate/sandstone ratio	Very high to 0	May be very high to 0 in channels	Generally low to 0 in channels	Generally 0 in channels		High to 0	Generally 0	Generally 0
H Maximum possible size	Boulders	Boulders in channels; sand in levee	Pebbles in channels; coarse sand in interchannel	Fine pebbles to coarse sand in channels; medium sand in interchannel	Fine sand in channel or interchannel	Boulders	Coarse sand	Medium sand
I Composition and textural maturity	Displaced shelf fauna characterize coarse-grained beds; intense in situ species characterize hemipelagic mud; wackes dominate	See channel and levee; high % heavy minerals; high terrigenous content in sand fraction of hemipelagic mud; wackes typical	Composition of minerals and fauna graded vertically in each sand layer; wackes and arenites present in channel sands	Sand composition and maturity like middle fan; high pelagic content in sand fraction of hemipelagic beds	High content of mica and plant fragments and some displaced biota in coarse-grained beds but in situ fauna dominates; hemipelagic like lower fan, but pelagic oozes possible; sands typically wackes	Dominance of displaced biota and minerals from shelf occurs in coarse-grained beds; in situ pelagic and terrigenous debris forms hemipelagic mud	In coarse-grained beds high mica and plant fragments and fewer, smaller sized, and deeper environment displaced biota; hemipelagic like channel; wackes dominate	Coarse-grained beds like levee; hemipelagic muds like channel
J Depositional processes	Slumps, slides, grain flow, fluidized sediment flow, debris flow, and bottom currents dominant; turbidity currents limited; turbid layer flow dominant for hemipelagic deposition	Dominant downfan grain flow, fluidized sediment flow, and debris flow; some turbidity-current bed load in channels; suspension load typical in levees and interchannel areas; turbid layer flow most important in interchannel hemipelagic deposition	Turbidity-current bed load dominant in channels and suspension load throughout channel and interchannel areas; continuous particle by particle fall most important in interchannel hemipelagic deposition	Turbidity-current bed load and suspension load throughout channel and interchannel areas; hemipelagic like middle fan	Turbidity-current suspension load throughout entire area; continuous particle by particle hemipelagic deposition most important throughout area	Slumps, slides, grain flow, fluidized sediment flow, debris flow; turbidity currents and bottom currents; see canyon for hemipelagic deposition	Mainly turbidity-current suspension load; hemipelagic deposition mostly from continuous particle by particle fall except near continental margin where turbid layer flow may dominate	Mainly turbidity-current suspension load; hemipelagic deposition like levee

[1] The characteristics have been synthesized from our research and the following studies: Haner (1971), Mutti and Ricci Lucchi (1972), Normark (1970), Normark and Piper (1972), Piper (1970a, b), and Stanley and Unrug (1972).

Table 2. Characteristics of Modern and Ancient Deep-Sea Fans and Criteria for Recognition of Ancient Deep-Sea-Fan Deposits (from Nelson and Nilsen 1974, Table 4).

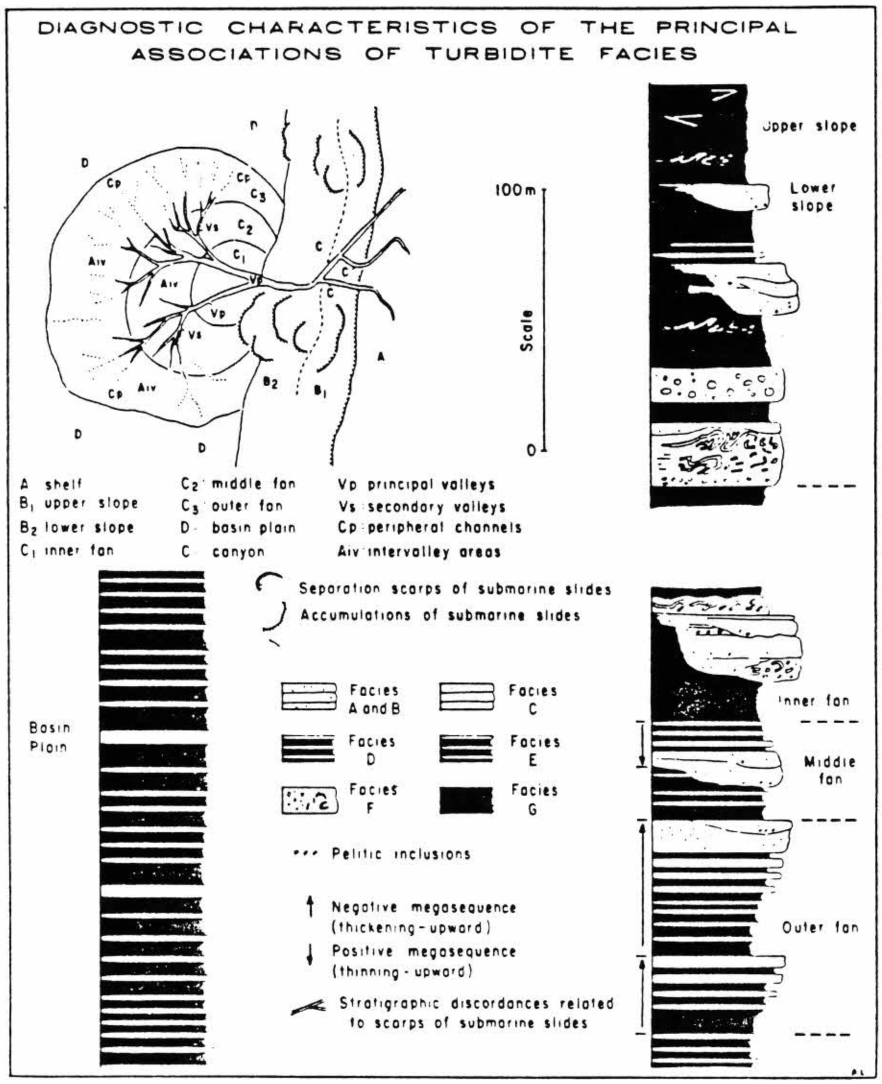


Figure **48** . Depositional framework of turbidite facies and representative stratigraphic sections of turbidite facies associations (from Mutti and Ricci Lucchi 1972, Fig. 14, translated by T.H. Nilsen).

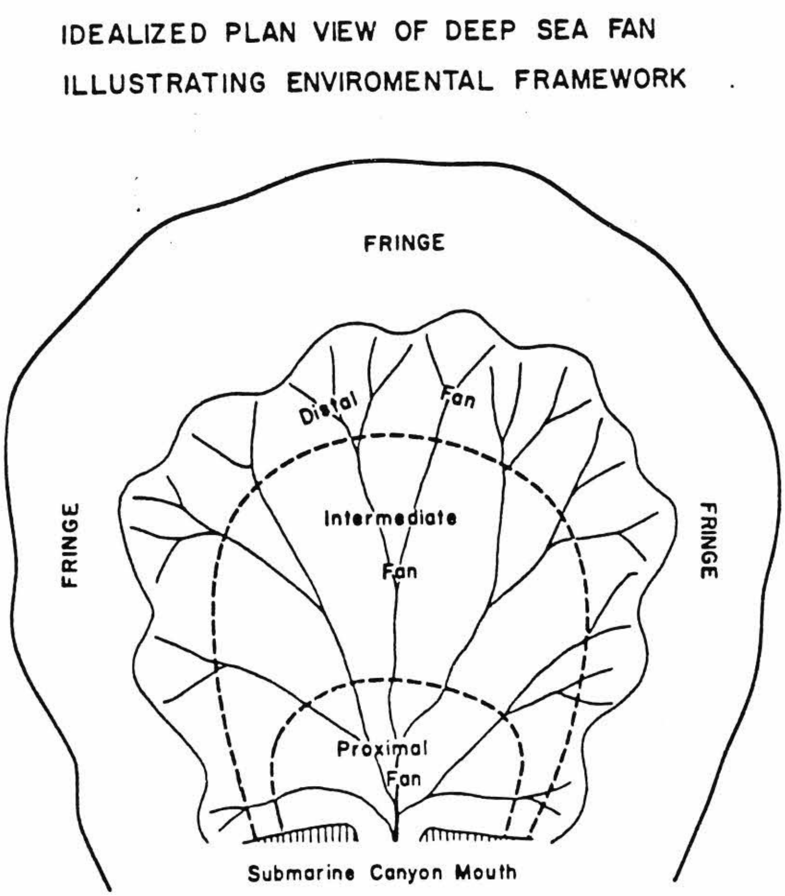


Figure **49** . Generalized model of deep-sea-fan deposition (Jacka et al. 1968, Fig. 8)

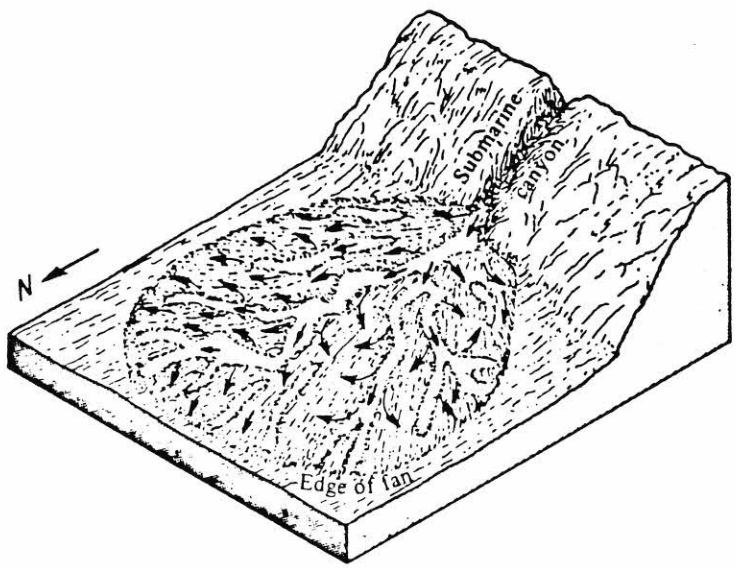


Figure 50. Depositional model for deep-sea-fan deposits of Eocene Butano Sandstone, Santa Cruz Mtns., California (from Nilsen and Simoni 1973, Fig. 12).

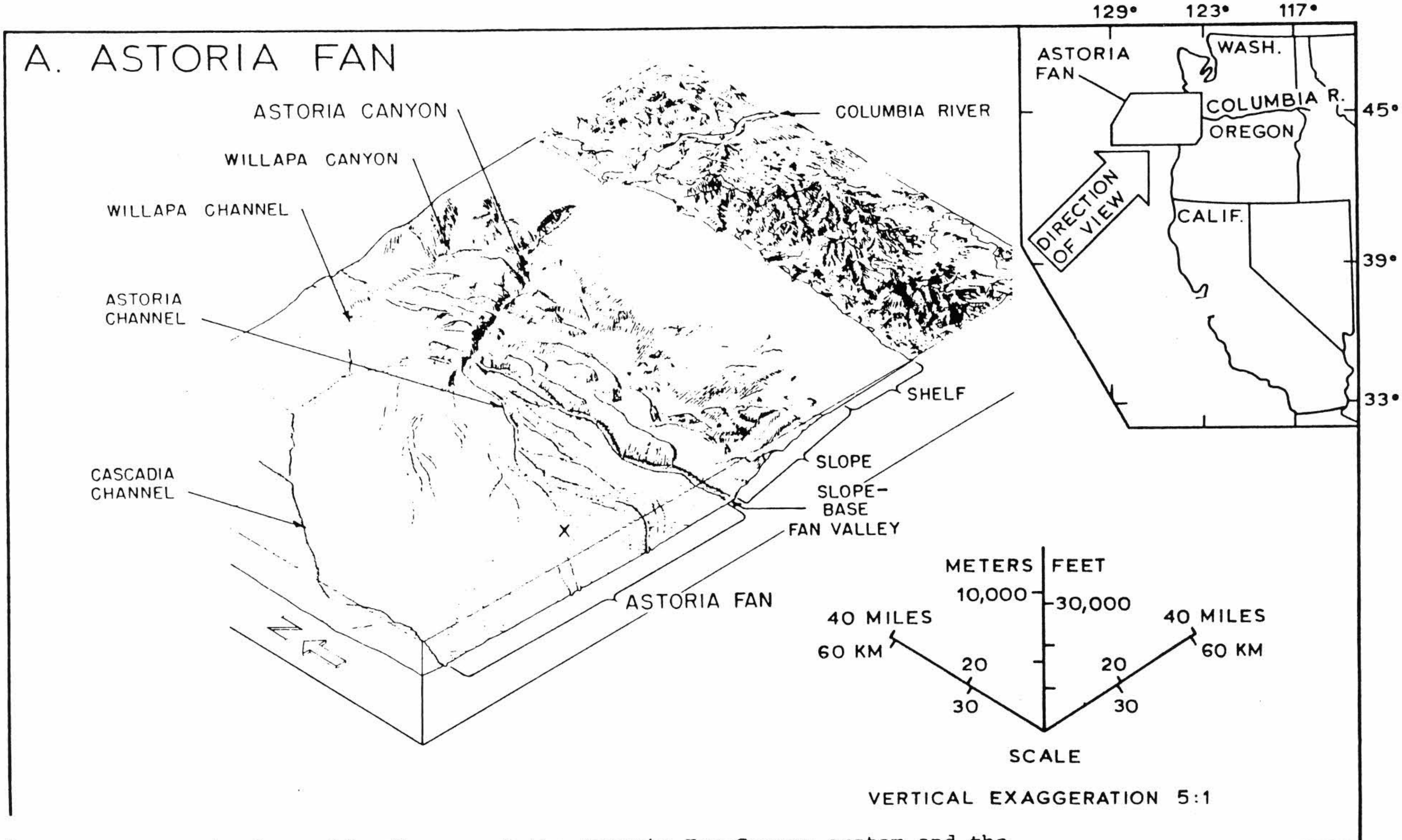


Figure **51** . Physiographic diagram of the Astoria Fan-Canyon system and the adjacent continental margin of the northern Oregon and southern Washington (from Nelson et al. 1970, Fig. 1).

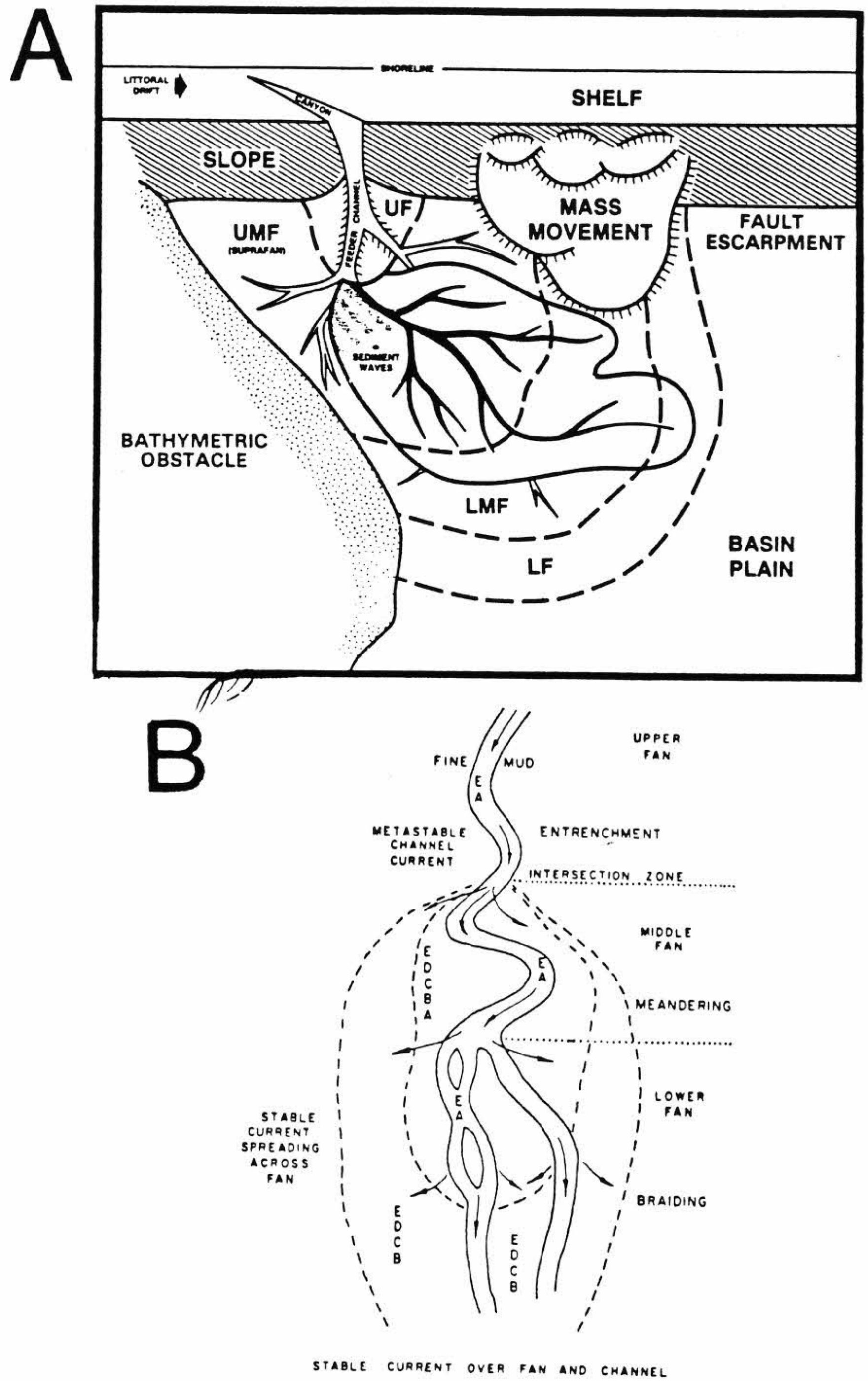


Figure **52** . A - Composite of salient features of late Quaternary submarine fan development in Santa Monica and San Pedro basins. Stippled area is active depositional lobe (from Nardin, 1983). B- sketch of the distribution of fan facies with reference to current dispersion and Bouma sequences (from Haner 1971, Fig. 15).

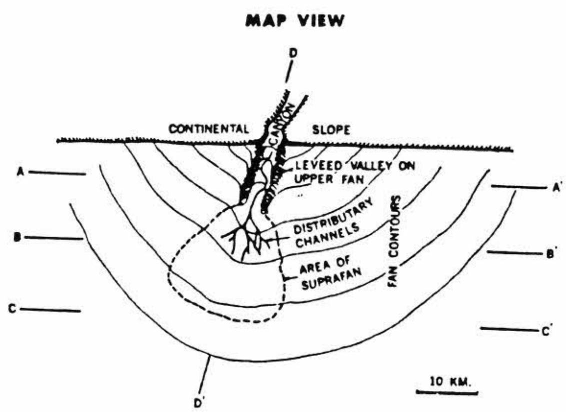


Figure 53 . Depositional model for modern deep-sea fans showing location of suprafan (from Normark 1970, Fig. 22).

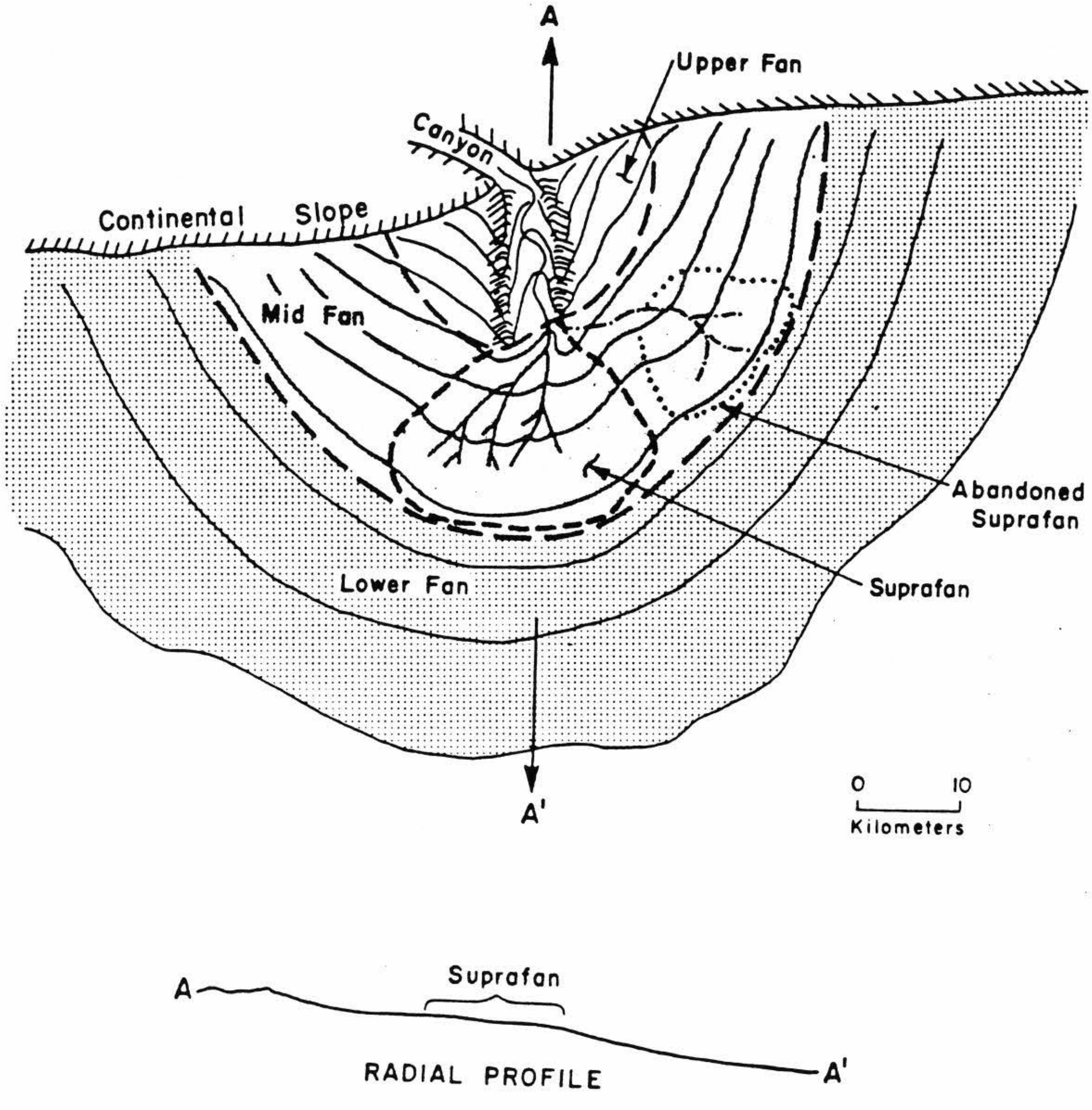


Figure **54** . Schematic representation of model for submarine-fan growth emphasizing active and abandoned depositional lobes termed suprafans (modified from Normark 1970a) (from Normark 1978).

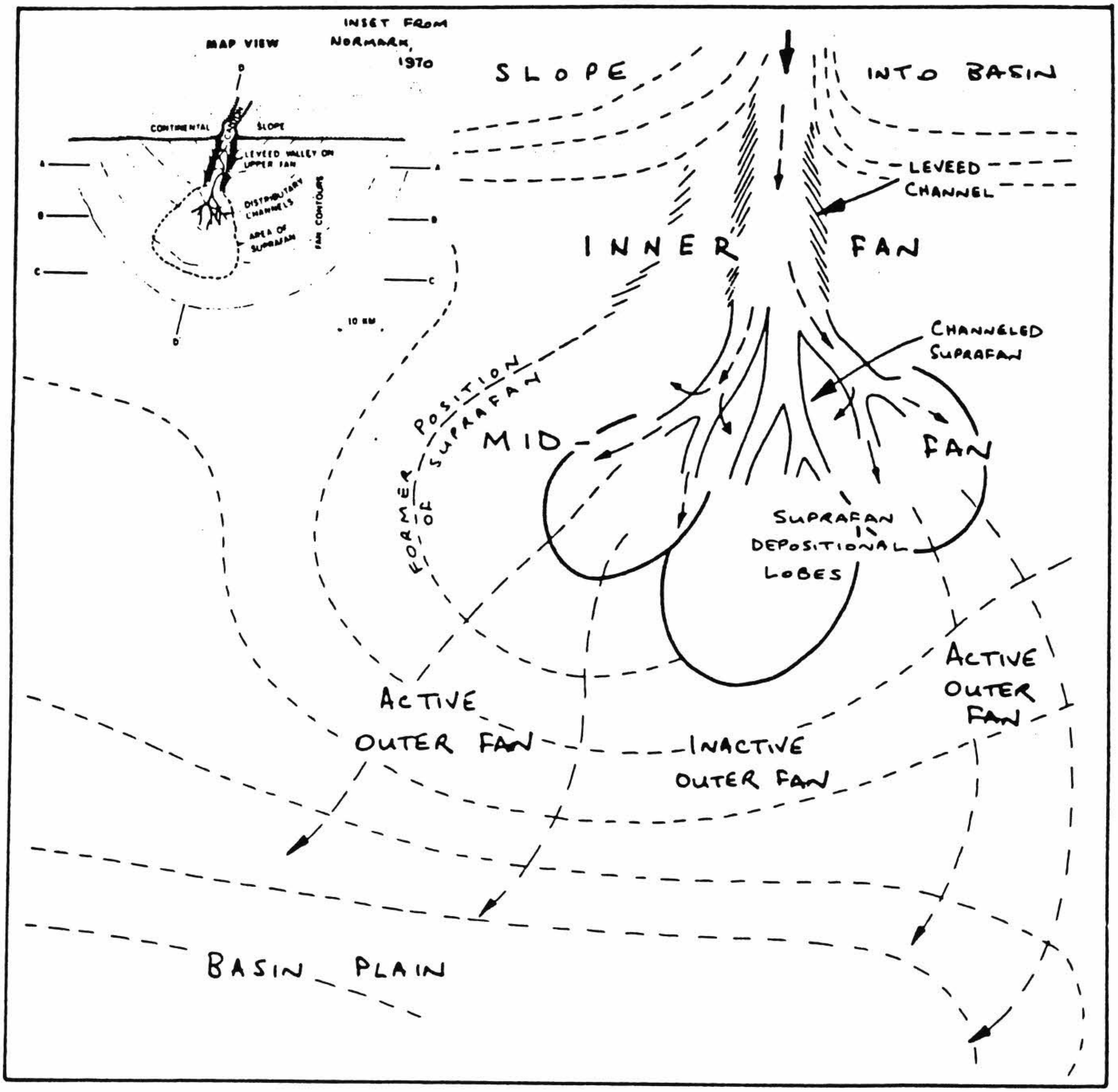


Figure 55 . Idealized slope-fan-basin floor system, showing relationships between the various facies associations, and possible paleocurrent directions (from Walker and Mutti 1973, Fig. 11). Zonation of the fan into upper (inner), middle and lower (outer) segments is shown, with a suprafan developed in the middle-fan area. The suprafan itself is subdivided into a upper channelized segment and a lower segment consisting of depositional lobes. Inset diagram of suprafan is reproduced from Normark (1970, p. 2189).

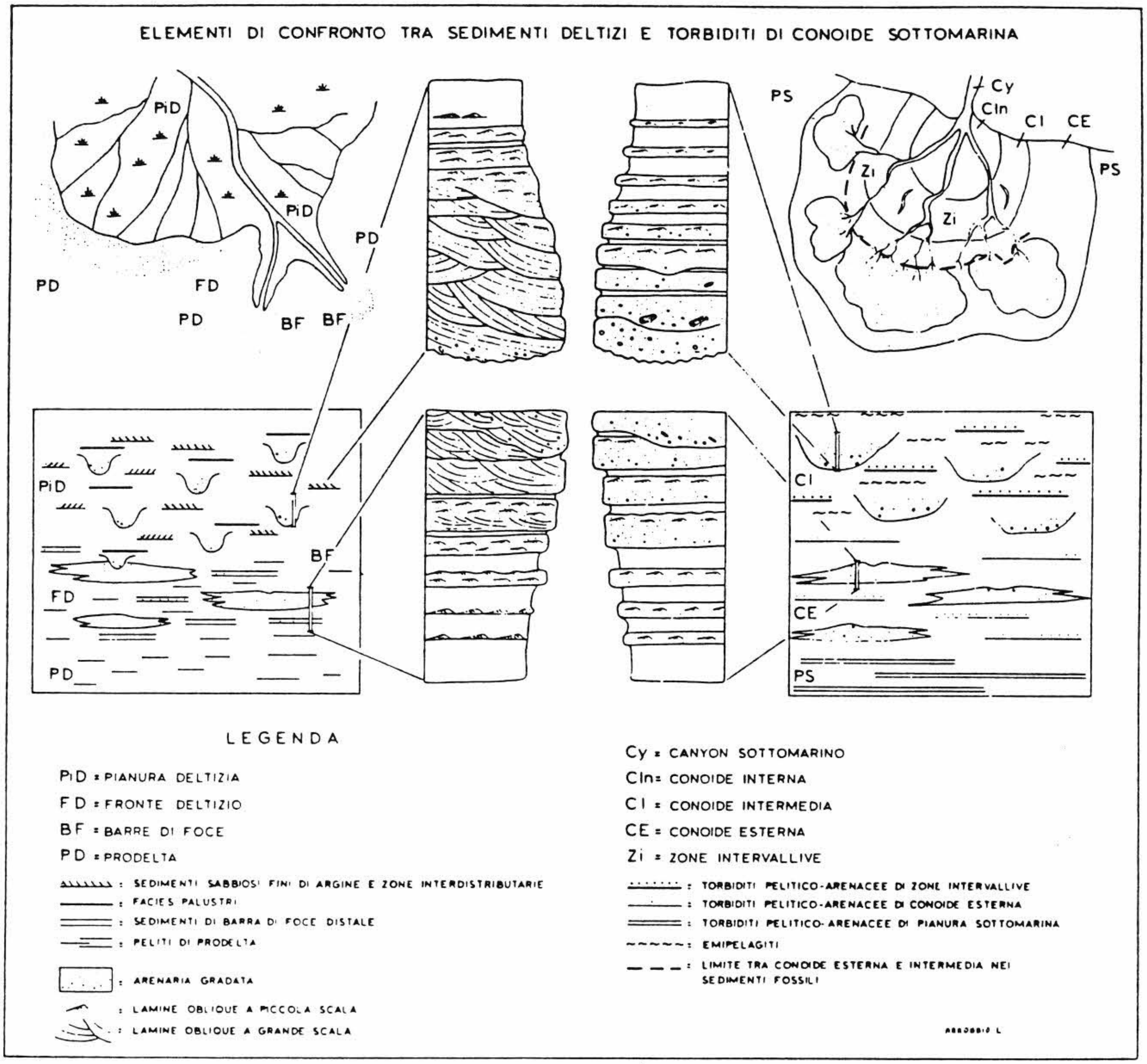


Figure **56** . Comparison of chief characteristics of deltaic and deep-sea-fan depositional systems (from Mutti and Ghibaudo 1972, Tav. 2).

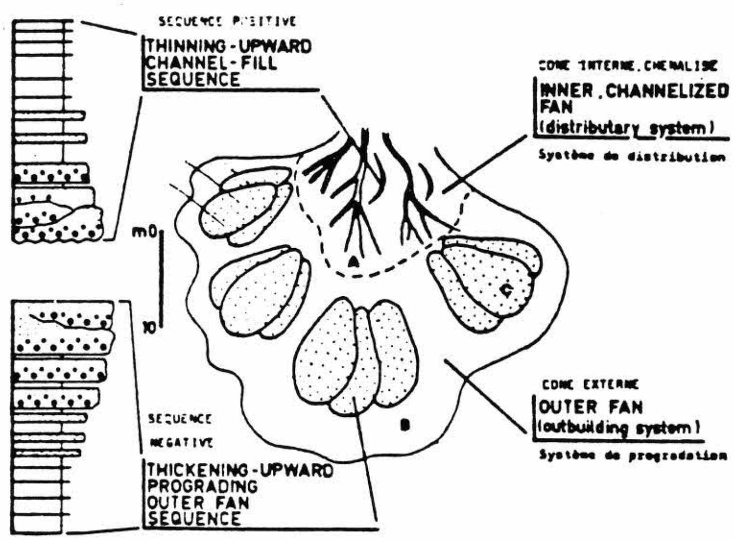


Figure **57** . Simplified model of deep-sea-fan deposition emphasizing channelized vs. nonchannelized sedimentation and the development of characteristic vertical sequences in these parts of the fan (from Mutti and Ricci Lucchi 1974, Fig. 2)

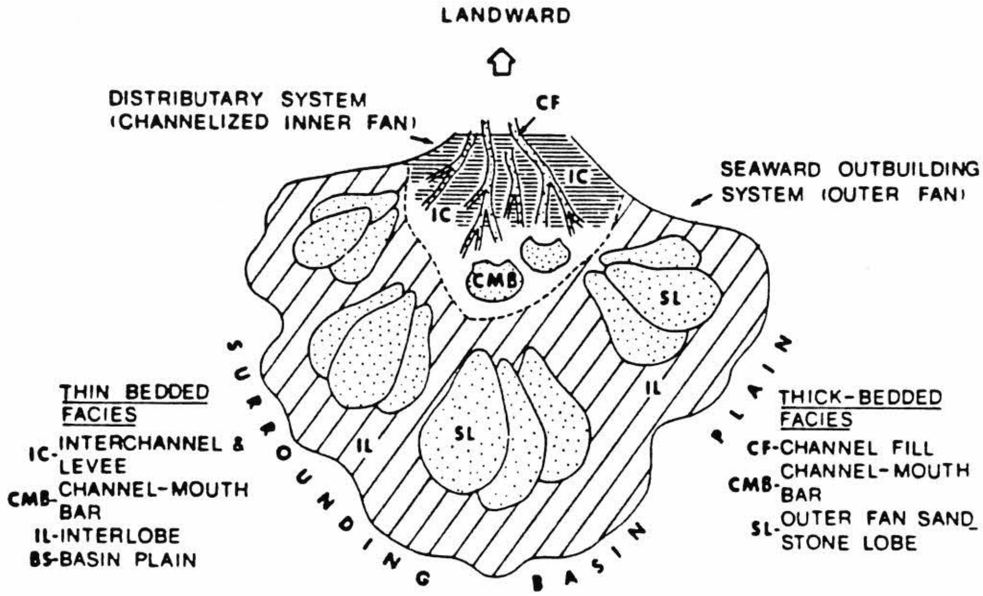


Figure **58** . Depositional model for ancient deep-sea fans (from Mutti and Ricci Lucchi 1975, Fig. 9).

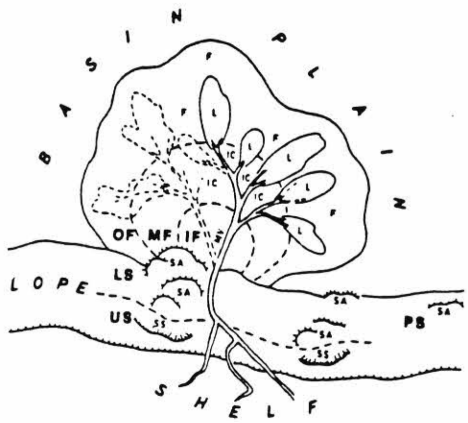


Figure **59** . Model of deep-sea-fan sedimentation (from Ricci Lucchi 1975, Fig. 4). US = upper slope, LS = lower slope, OF = outer fan, MF = midfan, IF = inner fan, PS = passive slope, SS = slump scar, SA = slump accumulation, MV = main valley, IC = interchannel area, L = depositional lobe, F = fringe.

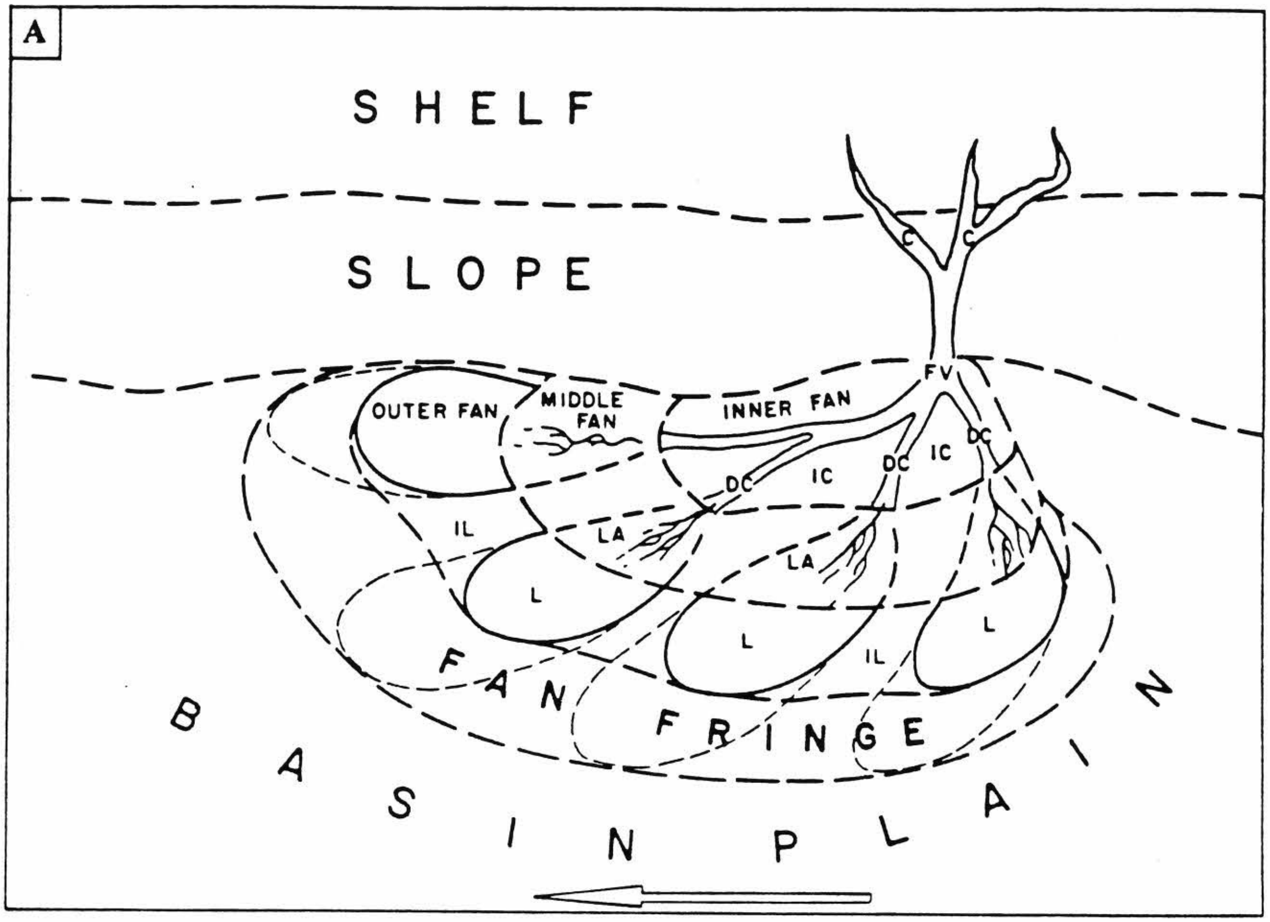


Figure 60 . Model for deep-sea fan of early Tertiary age, San Sebastian area, northern Spain (from Van Vliet 1978, Fig. 14-2A). C = feeder canyon of fan; FV = fan valley; DC = distributary channel; IC = interchannel area; LA = apex of depositional lobe ("suprafan"); L = depositional lobe; and IL = interlobe depression. Arrow indicates plunge of basin axis.

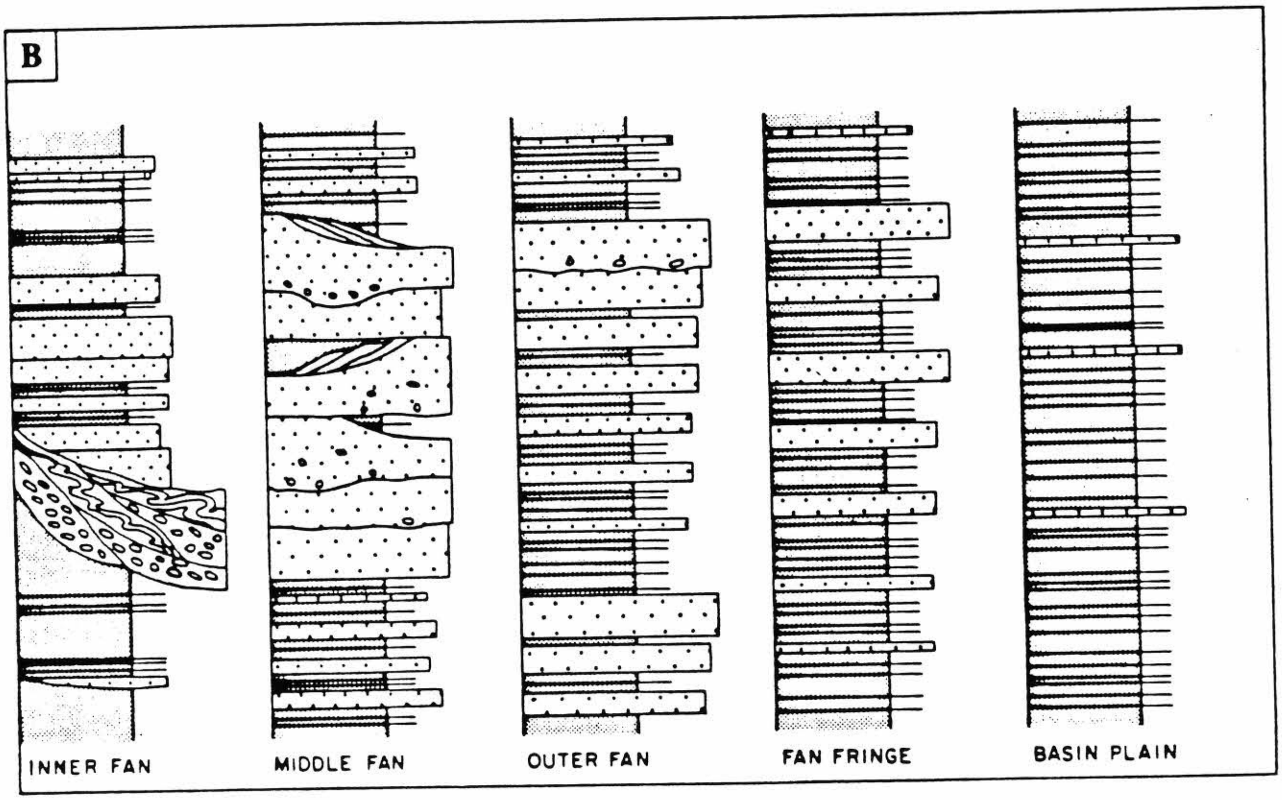


Figure **61** . Schematic representation of depositional sequences in various deep-sea-fan subenvironments (from Van Vliet 1978, Fig. 14-2B).

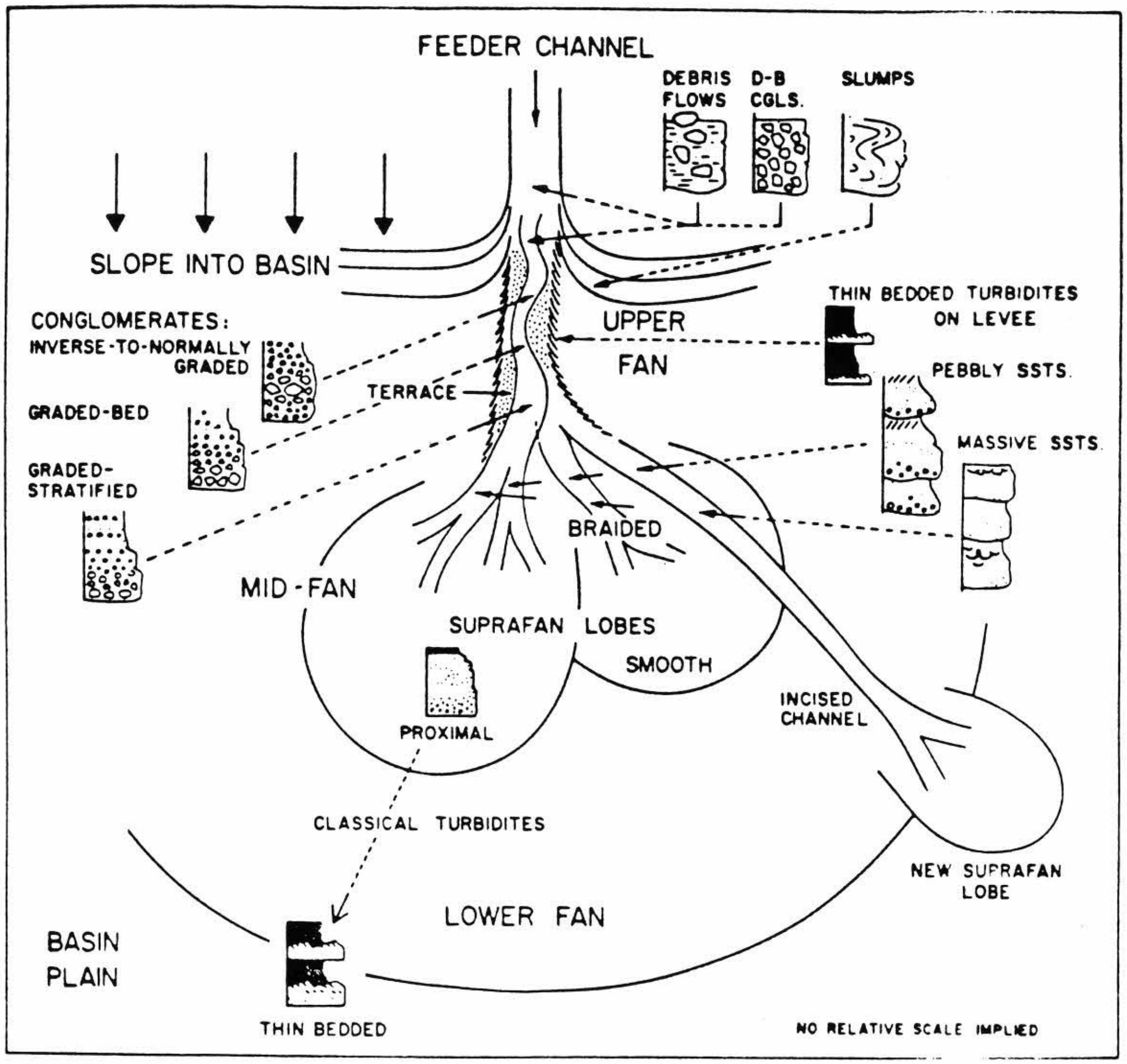


Figure **62** . Model of deep-sea-fan deposition, relating facies, fan morphology, and depositional environments (from Walker 1978, Fig. 13). D-B indicates disorganized-bed conglomerates.

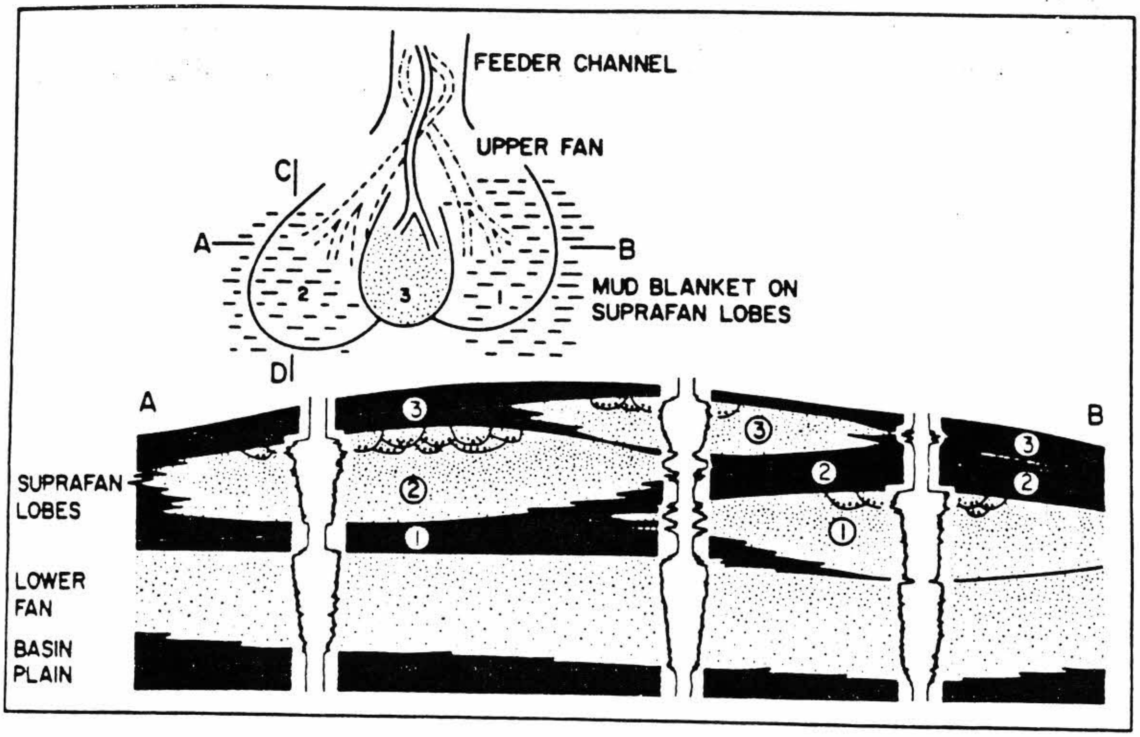


Figure 63 . Hyopthetical vertical cross section across prograding lower- and middle-fan system (from Walker 1978, Fig. 18). Lower-fan sequence is coarsening-upward, and each suprafan lobe has coarsening-upward sequence, from thin-bedded turbidites to channelized massive and pebbly sandstones. Each suprafan-lobe sequence also shales out laterally into mudstone drape that covers adjacent parts of fan. Sequence shown in cross section is result of suprafan lobe switching from position 1, to 2, to 3 (see plan view, upper part of diagram). Lobe switching may be related to changing meander patterns in upper-fan channel. Hypothetical electric logs illustrate coarsening-upward prograding-lobe sequences, fining-upward channel-fill sequences, and highlight problems of correlation in such a system.

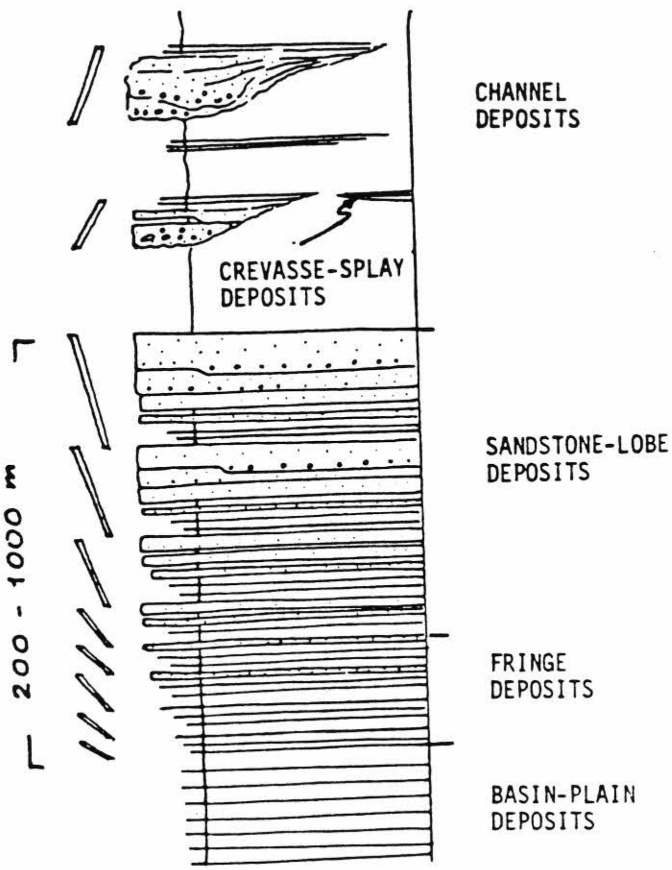


Figure **64** . Vertical sequence of mixed-sediment deep-sea fan (from Mutti 1979, Fig. 18).

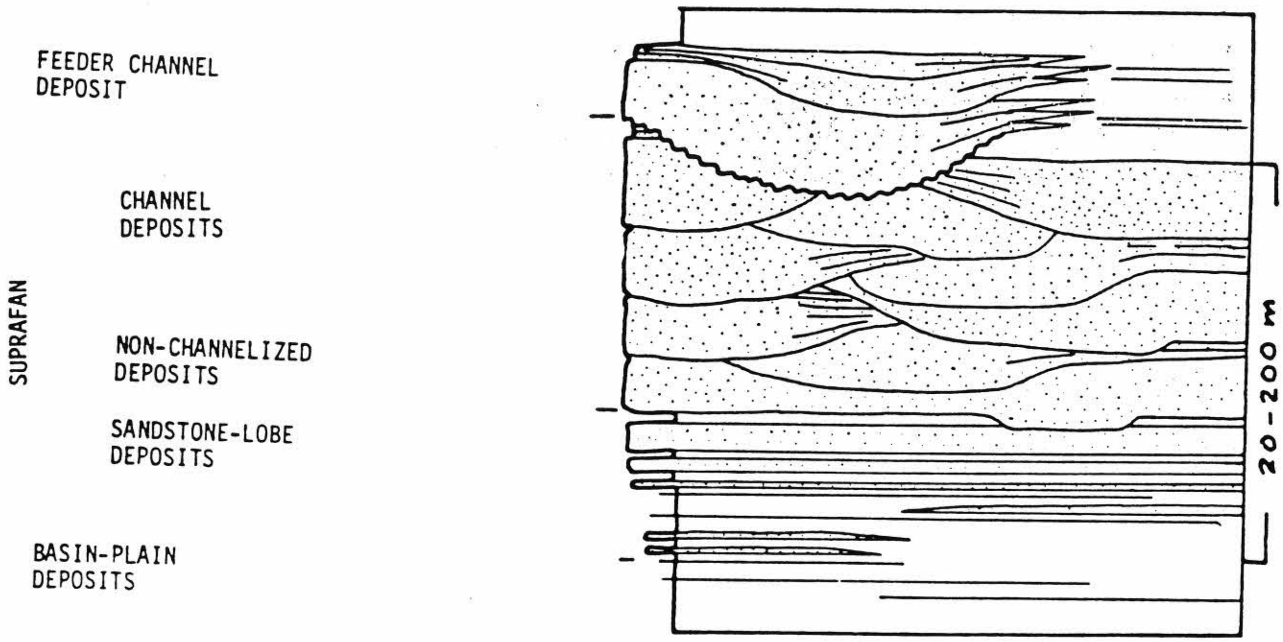


Figure **65** . Vertical sequence of sand-rich deep-sea fan (from Mutti 1979, Fig. 16).

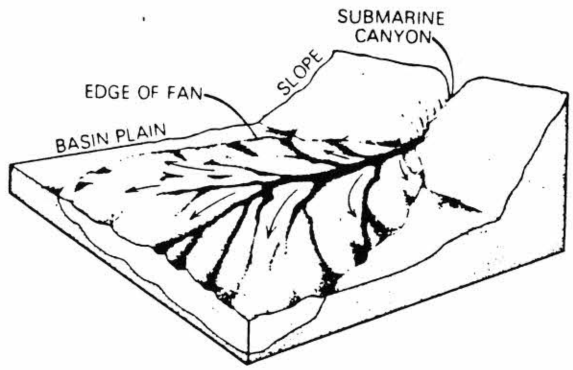


Figure **66** . Deep-sea-fan model for The Rocks Sandstone showing abundant distributary channels, lack of differentiation of inner-, middle-, and outer-fan components, and rapid and abrupt dumping of sand at base of slope (from Link and Nilsen 1980, Fig. 13).

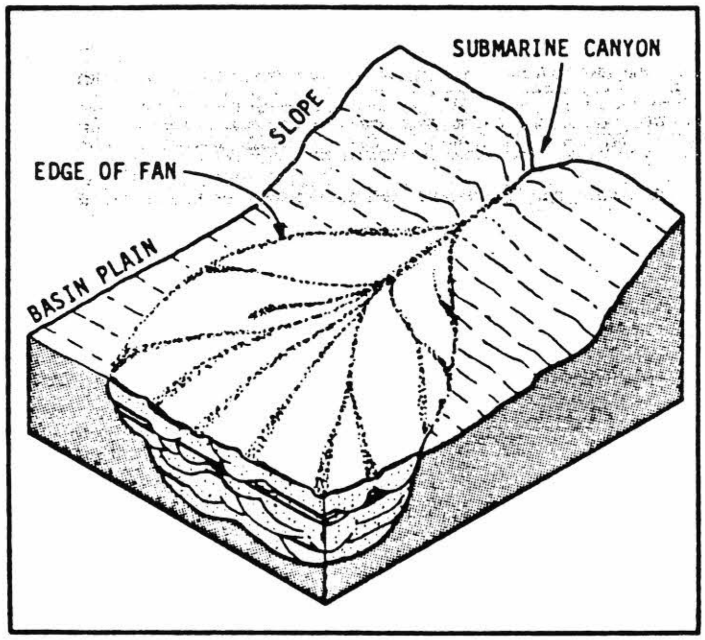


Figure 67 . Depositional model for the Cretaceous Chatsworth Formation of southern California showing a sand-rich system composed most of channel and interchannel deposits with minimal and outer-fan facies association (from Link 1981, Fig. 9).

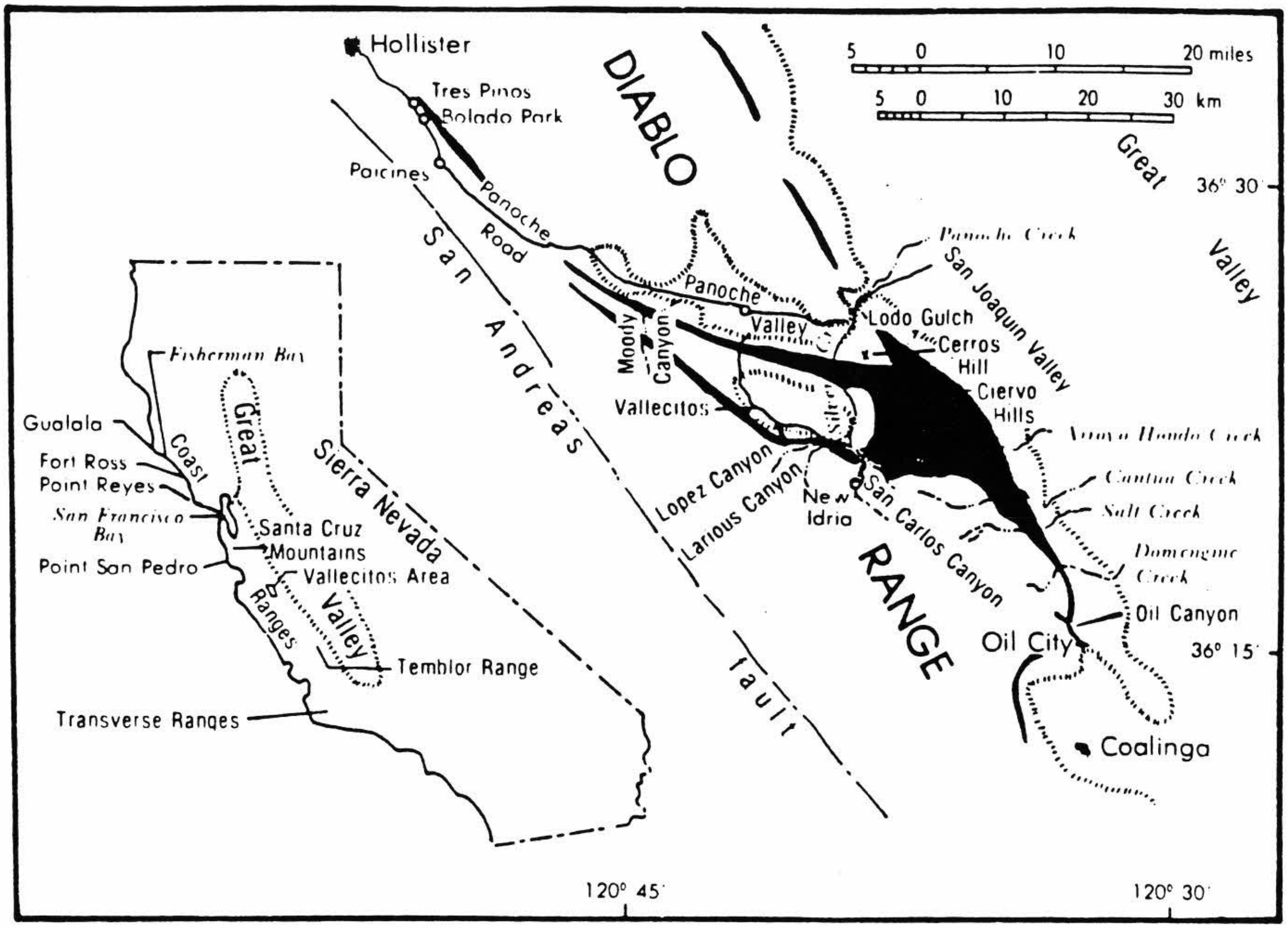


Figure 68 . Index map of central Diablo Range, California, showing outcrop extent of sand-rich lower Tertiary deep-sea-fan deposits of the Cantua Sandstone and related units of the Vallecitos area and the Tres Pinos Sandstone of the Paicines area (from Nilsen 1981, Fig. 9).

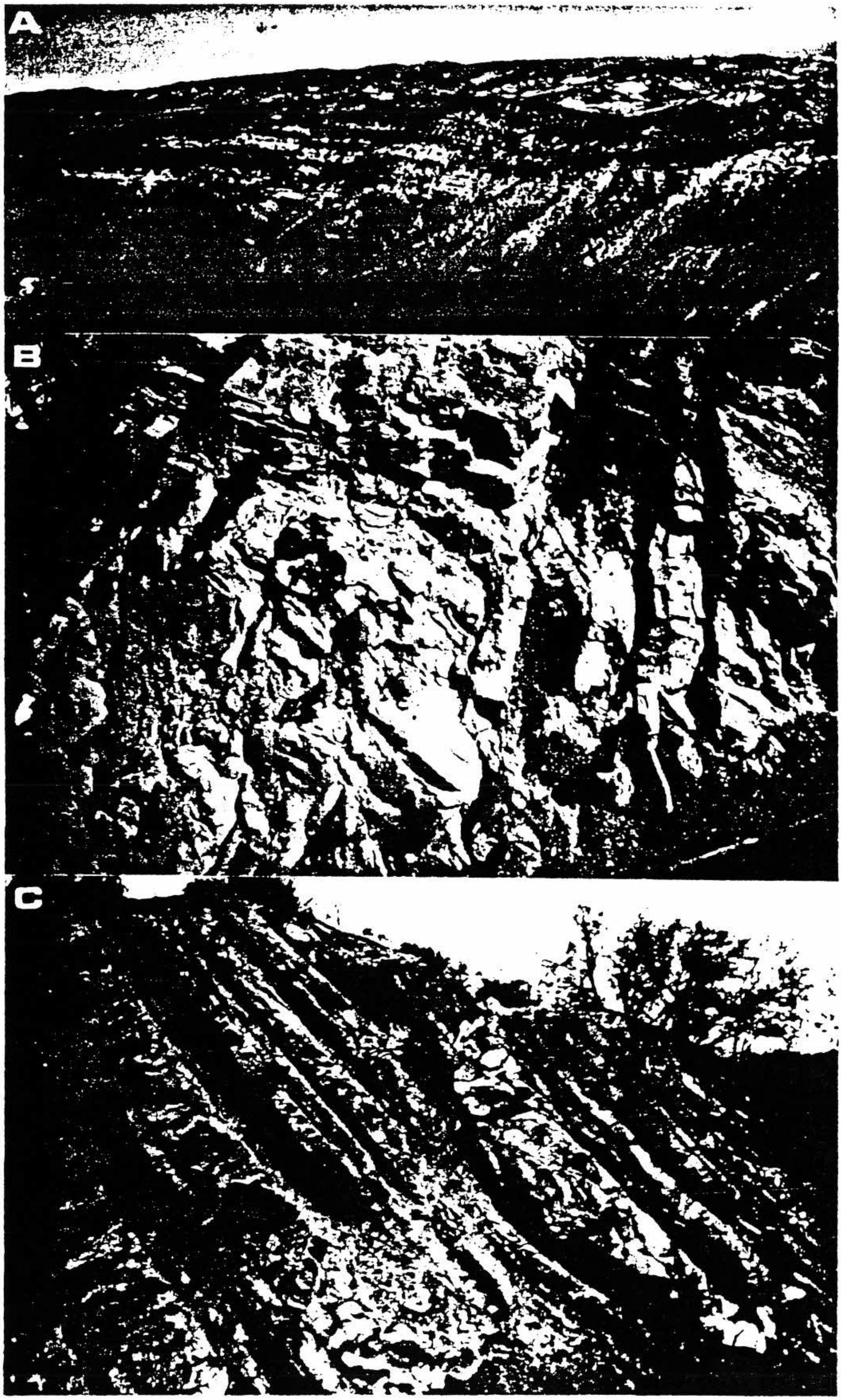


Figure **69** . Photographs of the lower Tertiary Cantua Sandstone member of the Lodo Formation, central Diablo Range, California. A - View north of channelized deposits exposed in cross section. B - Repetitive thinning- and fining-upward channelized cycles of sand-rich fan. Stratigraphic top to left. C - Repetitive thinning- and fining-upward channelized cycles of sand-rich

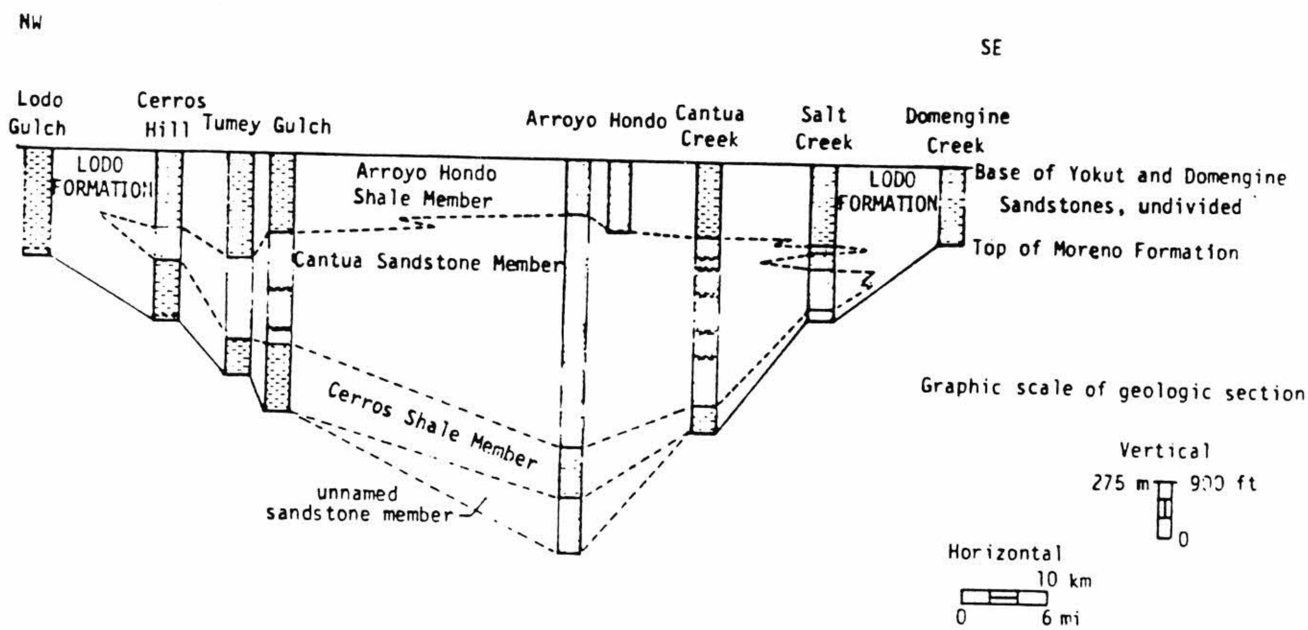


Figure 70 . Restored northwest-to-southeast cross-section across the Vallecitos syncline showing stratigrahic relations and thickness variations in the Lodo Formation and its members (from Nilsen 1981, Fig. 15, modified from White 1940).

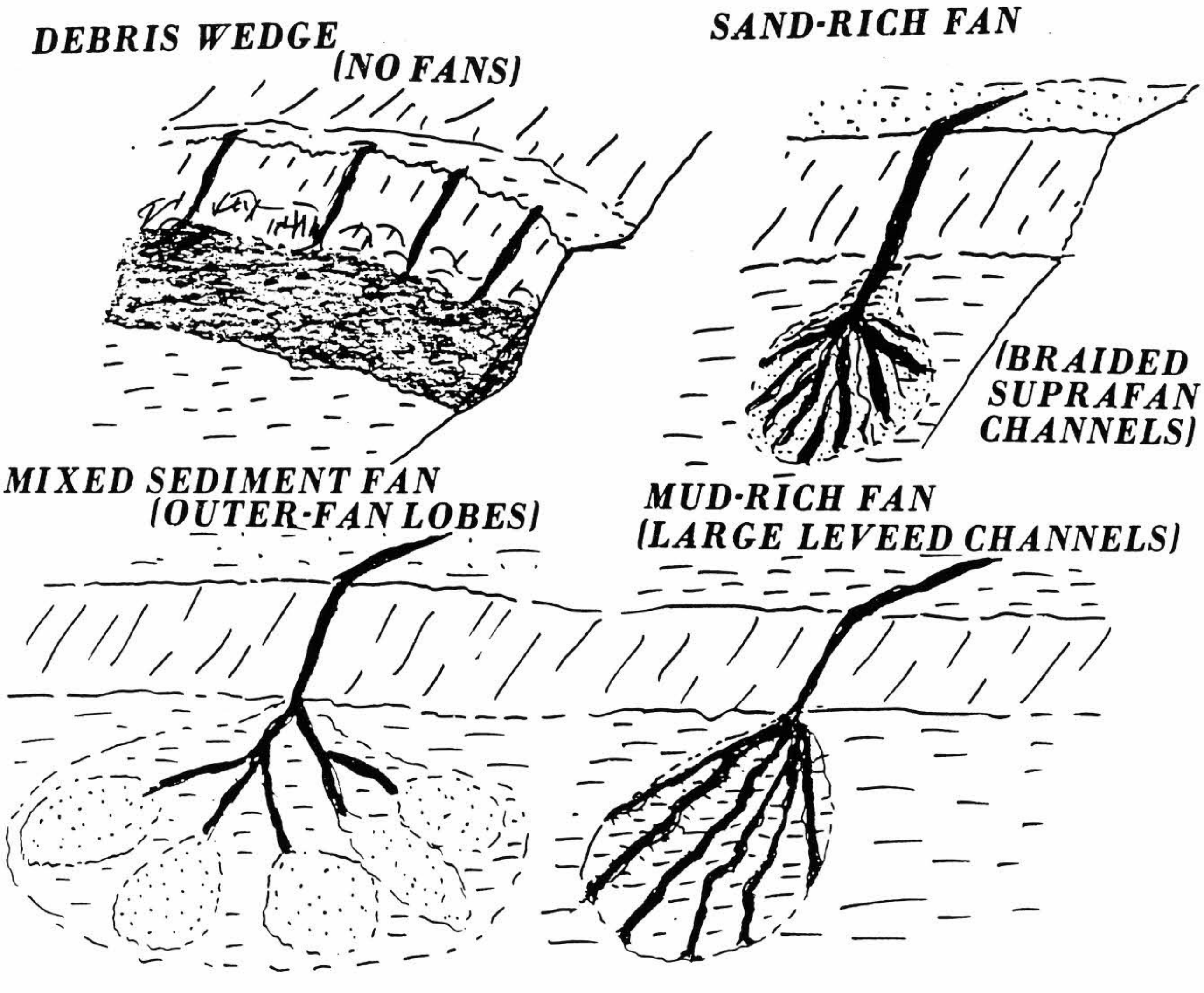


Figure 71 . Morphology of sand-rich, mixed sediment, and mud-rich deep-sea fans and debris wedges (drawn by T.H. Nilsen 1979).

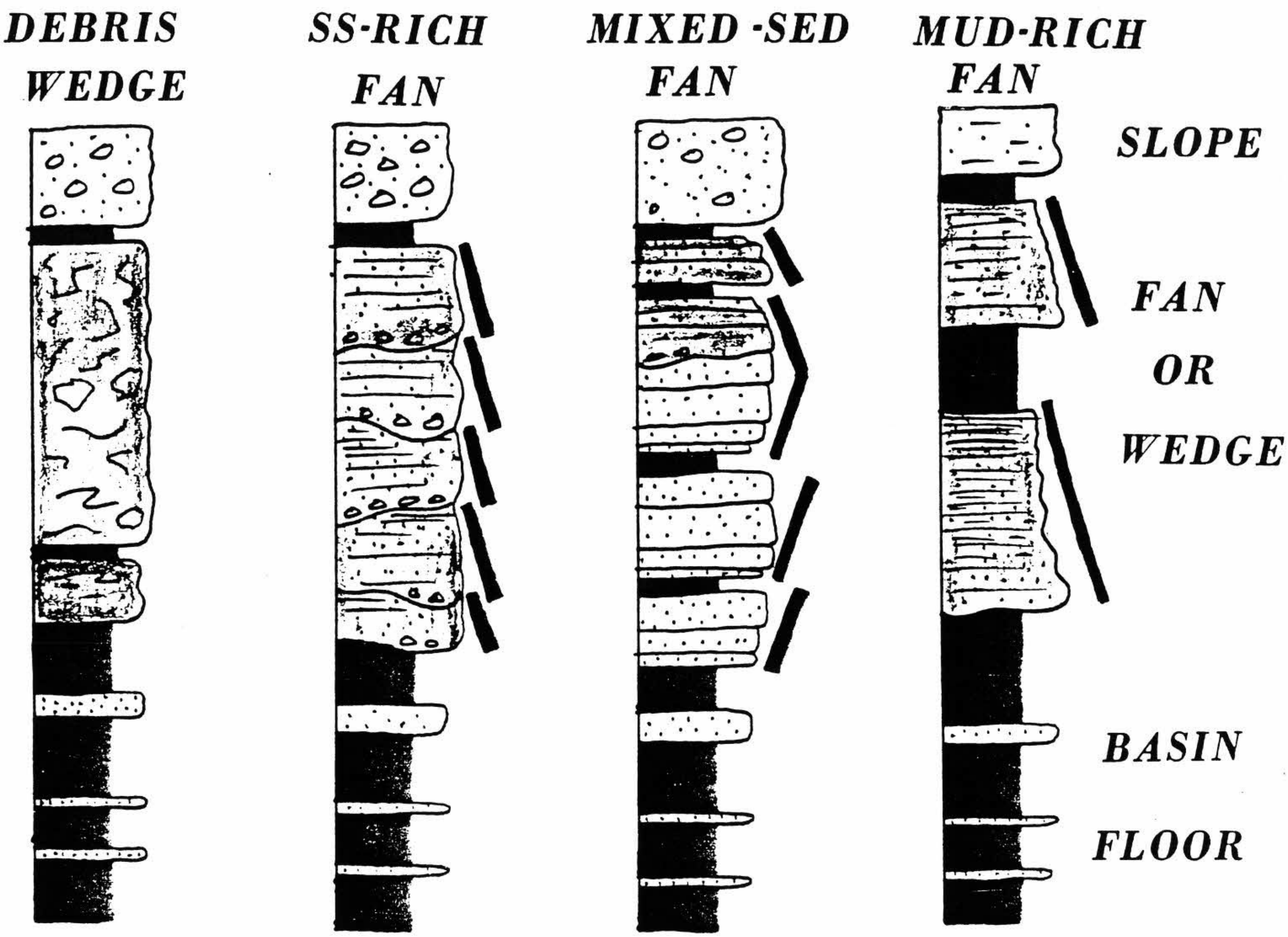


Figure **72** . Vertical sequences of sand-rich, mixed sediment, and mud-rich deep-sea fans and debris wedges (drawn by T.H. Nilsen 1979).

FIELD TRIP DESCRIPTION -- DAY ONE

On the first day of the field trip we will examine several of the classic Lower Pennsylvanian Jackfork Formation deposits of slope and canyon -- upper submarine fan facies in the vicinity of Little Rock. We will proceed then northward through upper to middle fan sequences in the Middle Pennsylvanian Lower Atoka Formation on the Bayou Meto anticline then into a series of shallow water - deltaic deposits in the middle -- upper Atoka Formation primarily near Morrilton. Then we continue back to the south through more proximal to mid-fan deposits of the Lower Atoka and Jackfork Formations eventually reaching our destination in the scenic Mazarn Basin - Zigzag Mountains in the central Ouachita Mountains at Hot Springs.

(Description from
JORDAN et al., 1991)

STOP 1

Visitor Center, Pinnacle Mountain State Park, Arkansas

Purpose:

The purpose of this stop is to view a melange interval that is overlain by massive-appearing, matrix-free sandstones, and discuss possible origin of the sedimentary package.

Description of Locality

Extending west of STOPS 2 and 3 is an east-west trending chain of steep hills that locally reach 1,000 feet in elevation (Fig. 73). These hills are made up of thick, but apparently discon-tinuous units of moderately to steeply-dipping sandstone and shale of the upper part of the lower Jackfork ('middle' Jackfork according to local terminology) (Stone and McFarland, 1981). We will examine these rocks in an abandoned quarry near the Visitor Center of Pinnacle Mountain State Park (Figs. 73, 74).

The Jackfork sequence in and around the Visitor Center consists of fine- to medium-grained quartzose sandstone and intervening shale units that probably total 200-400 feet thick. In general, the exposed sequence dips south at 15 to 45 degrees, but mapping (Fig. 74) suggests that much and perhaps all of the exposure is a melange of large blocks of stratified sandstone, some including as much as 200 feet of section, in a matrix of mudstone and diamictite.

From the parking area southward along the quarry pond, three general lithologic units can be recognized. A "basal" unit or units of stratified quartzose sandstone is exposed on both sides of the parking area. It totals about 50 to 60 feet thick and is underlain and overlain by stratified mudstone (Fig. 75). A "middle" unit, exposed only on the west side of the pond, includes about 70 to 90 feet of diamictite composed of disarticulated and contorted blocks and beds of sandstone encased in unstratified mudstone. Invertebrate fossils of shallow-water organisms have been reported from sandstone blocks in this and similar melanges just west of this area.

The uppermost unit exposed at the southern end of the quarry consists of over 100 feet of resistant, evenly-bedded, finely laminated, clean, quartzose, fine- to medium-grained sandstone (Fig. 76).

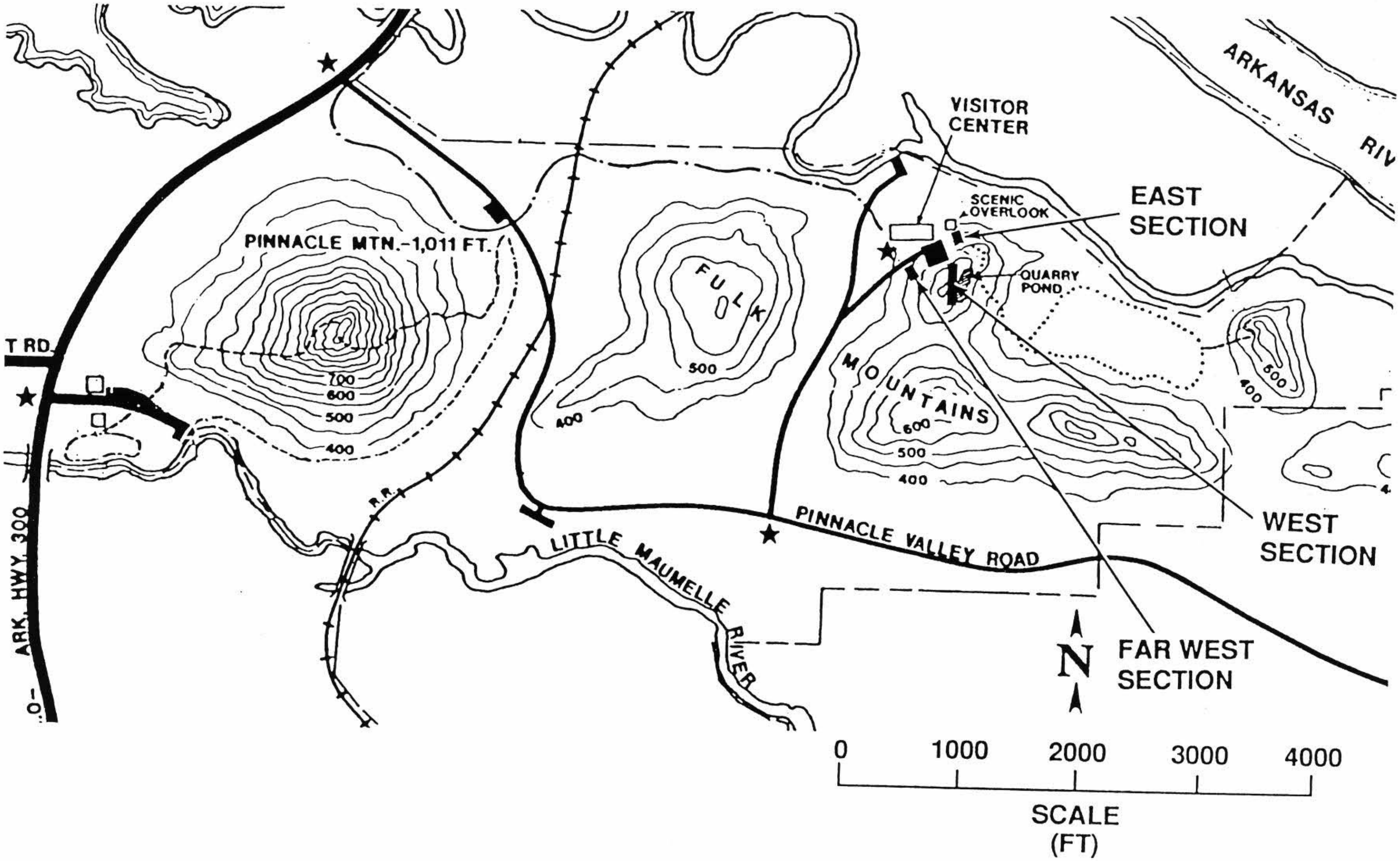


Figure 73. Location of Pinnacle Mountain State Park Visitor Center and outcrop sections near quarry pond.

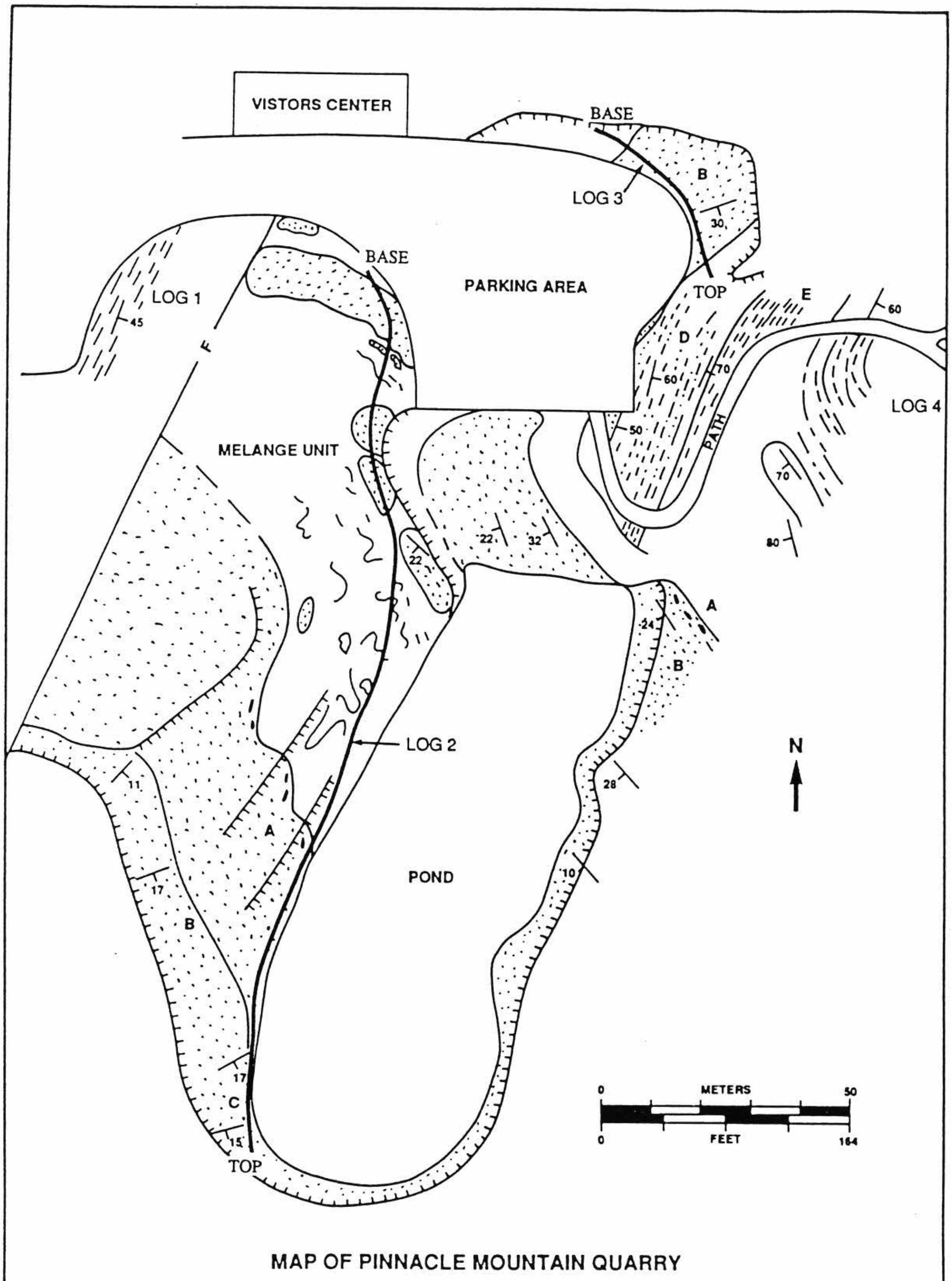


Figure **74.** Geologic sketch map of the Visitor's Center area at Pinnacle Mountain State Park, Arkansas. Large blocks of medium- to fine-grained quartzose Jackfork sandstone (stippled) are interbedded and mixed with units of muddy diamictite and laminated mudstone and shale, respectively. The letters (A, B, C, D, and E) refer to localities where sub-facies shown in the stratigraphic column (Fig. **77**) can be seen. Location of gamma-ray log paths are shown.

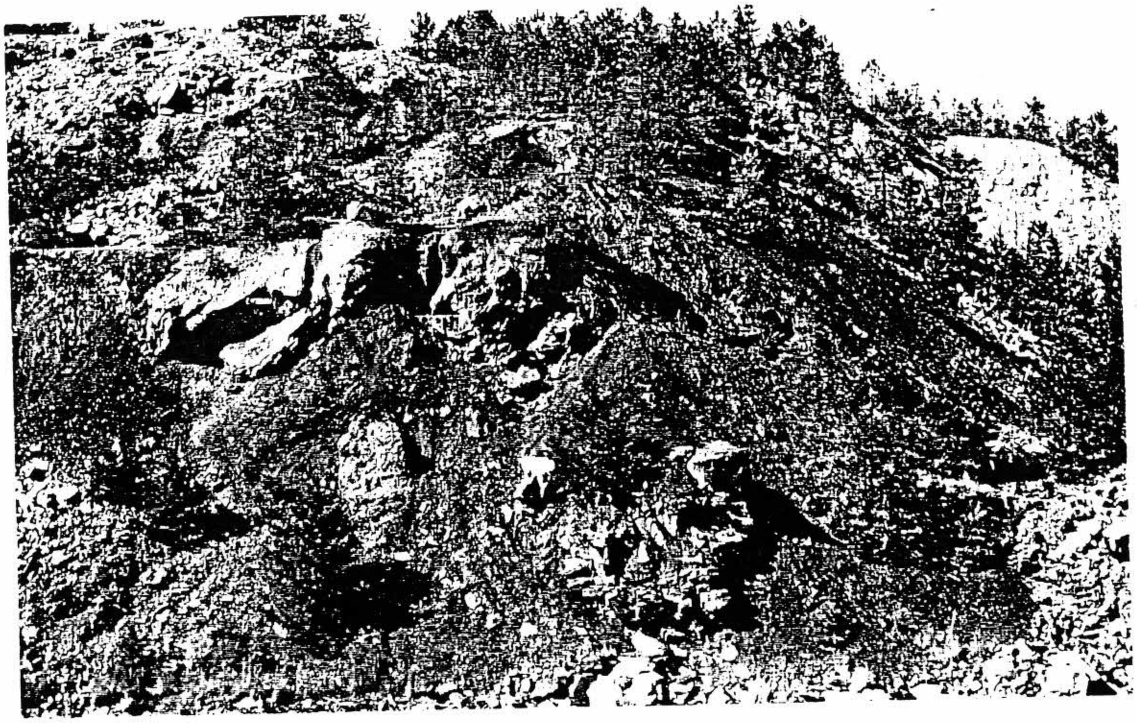


Figure **75.** West side of the quarry pond, melange interval comprising over 100 feet of contorted blocks and beds of sandstone encased in unstratified mudstone.

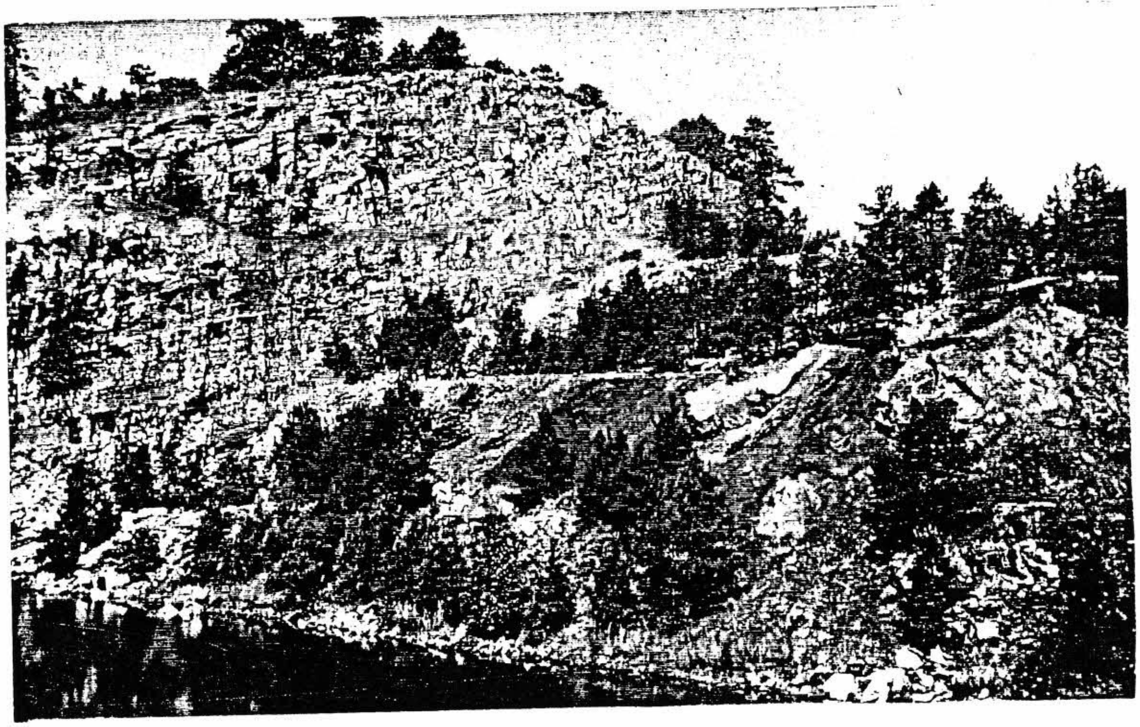


Figure **76.** The uppermost interval exposed at the southern end of the quarry consisting of over 100 feet of resistant, evenly-bedded, finely laminated, clean, quartzose, fine- to medium-grained sandstone overlying a melange unit.

Interpretation

None of the sandstone units at this stop can be traced continuously for over a few hundred (up to 600) feet. Some adjoining lenses may exceed three-quarter of a mile in length. Most terminate abruptly against shales or sandstone units with different orientations, suggesting that all are blocks. Stone and McFarland (1981) suggest that these sandstone units could represent (1) resistant beds caught up and mixed with intervening mudstones, between thrust faults, (2) remnants of individual submarine fans (favored interpretation), or (3) sandstone slabs that have slid downslope into deeper parts of the basin.

Where relatively complete sections of sandstone units can be identified, they show an internal succession of textures and structures suggesting that they formed as erosive, fining- and thinning-upward, channel-fill units (Fig. 77). In all probability, they represent the fill of large slope channels, eroded into underlying slope mudstone, that have been detached and incorporated into large downslope slides.

Reservoir Implications

Another hypothesis is that the lowest "unit" of mainly sandstone consisting of 55 feet of sandstone exposed on the eastern side of the parking lot is the same sandstone interval that correlates to the lower sandstone unit (25 feet) exposed on the western side of the parking lot.

Correlations shown in Figure 78 might then be valid. It is not clear if this sandstone was deposited within a channel elsewhere, only to have been detached and incorporated into a large downslope slide, or if this sandstone is in fact a lenticular body deposited within an 'in situ' channel. It would be difficult to interpret this sandstone body as a detached channel sandstone within a hydrocarbon reservoir based on a correlation of similar logs.

Note on the log the presence of several sandstones in the diamictite interval. these sandstones are not bedded, as might be interpreted from the logs alone, but are actually detached and convoluted blocks of sandstone that are a few to greater than 10 feet in diameter ("glumps and gloops" by local terminology). Using only the character of the logs for correlation purposes and in the construction of cross-sections for modelling, a possible oil field reservoir may be doomed if these sandstone blocks were interpreted as laterally continuous sandstone beds!

IDEALIZED CHANNEL-FILL SEQUENCE

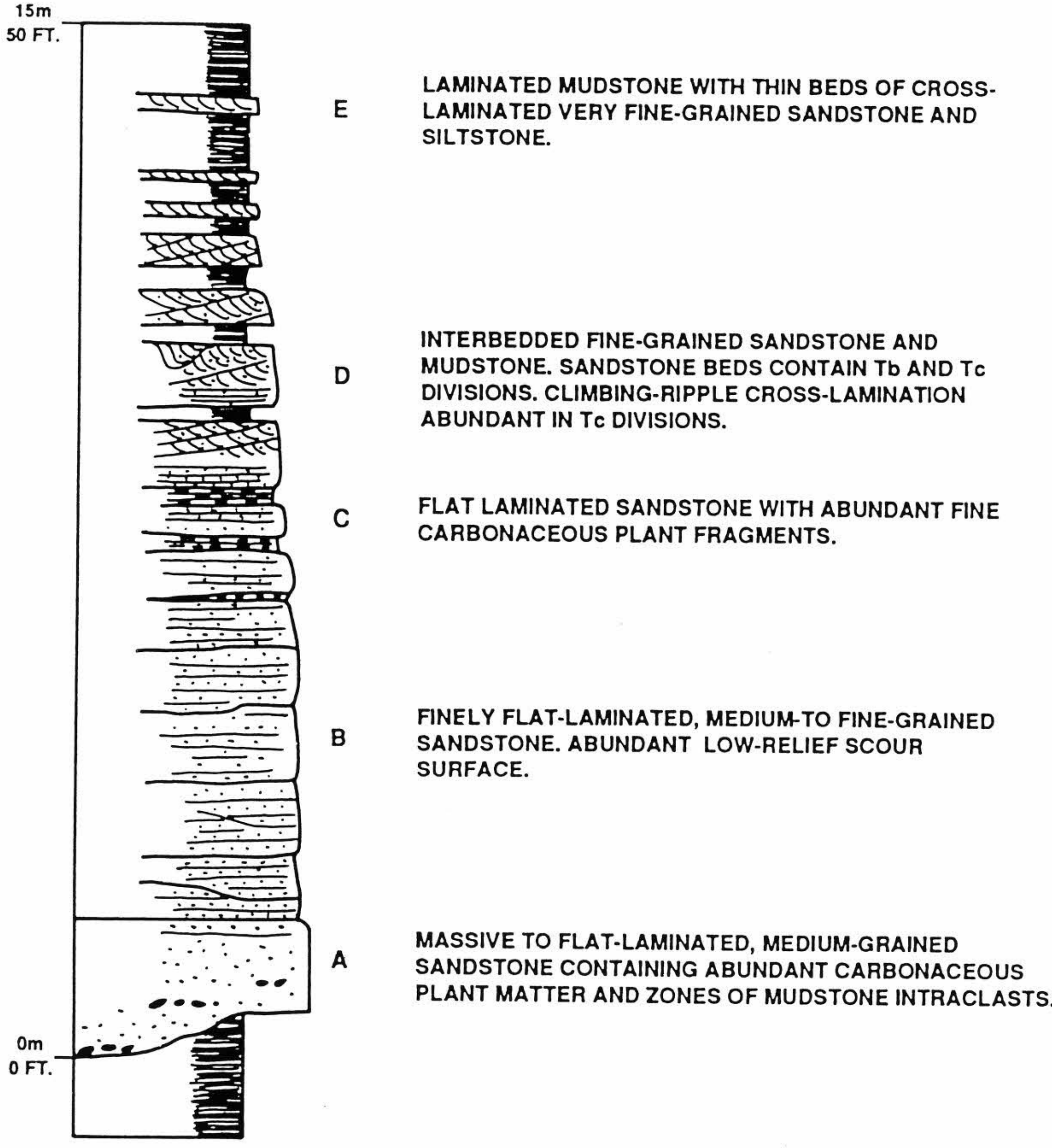


Figure 77 Schematic composite stratigraphic column of an idealized channel-fill sandstone and interbedded mudstone and shale sequence present at Pinnacle Mountain State Park. Letters denote main sub-facies located in Figure 74.

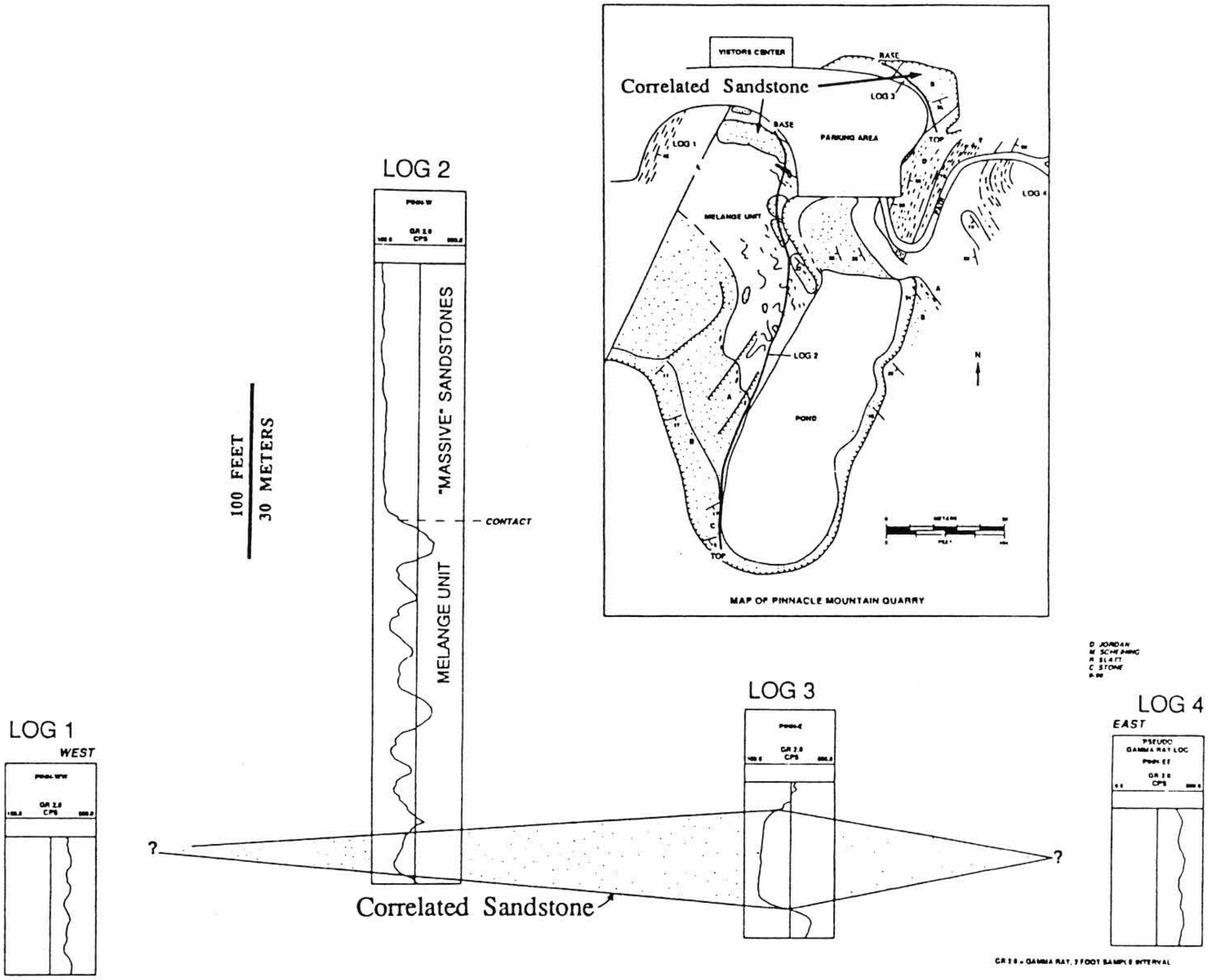


Figure **78.** A possible correlation of lower sandstones across the parking lot at the Pinnacle Mountain Visitors Center. This correlation suggests that the sandstones are lenticular and they pinchout to the east and west due to faulting, because the sandstone body is a slide block (part of the overlying melange interval), or the sandstone body is an <u>in situ</u> channel-fill deposit.

(Description from JORDAN et al., 1991)

STOP 2

I-430 Roadcut, Little Rock, Arkansas

Purpose:

The purpose of this stop is to view a downslope section equivalent to that exposed at Big rock Quarry (approximately 5 miles to the east) and to view sedimentary features that could not be studied at STOP 3 because of the steepness of the quarry face. The location of the roadcut (Fig. 79) is on the west side of Interstate 430 south in northwestern Little Rock, just north of the Arkansas Highway 10 exit ramp (Fig. 80A).

Description of Locality

This section is absolutely sliced to pieces by faults. Any inference regarding vertical trends, contacts, etc., at this outcrop may be regarded as suspect. The sequence lies along the flank of the Big Rock Syncline and beds generally dip steeply to the north. The lower Jackfork consists of a sequence of black carbonaceous shales with siltstone interbeds and laminae that are intensely sheared. The upper Jackfork consists of interbedded thin and thick-bedded sandstone and thinner shale and siltstone. The lower part of the sandy sequence consists of highly fractured and faulted quartzitic sand-stones (Fig. 80B), many which may be exotic sandstone blocks.

Interpretation

This section is interpreted as the same stratigraphic sequence as at Big Rock Quarry; lower Jackfork shales overlain by upper Jackfork sandstones (Fig. 81). The thinning and fining-upward sequences of sandstone beds (Fig. 82) were interpreted to represent submarine fan-channel deposits that were deposited farther downslope than equivalent strata at Big Rock Quarry to the east (Stone and McFarland, 1981; Link and Stone, 1986b; McFarland, 1988). Link and Stone (1986b) suggest a succession of events led to the deformation observed at this locality; (1) Southward slumping down the continental slope, (2) Northward stacking of several thrust-fault slices, (3) Folding during several episodes, and (4) Backfolding and faulting caused, in part, by the piling up and crowding at the toe of the larger thrust sheets. The last 3 events took place in late Paleozoic time as the Ouachita Mountains were being formed. Hydrothermal quartz veins of late Paleozoic age also cut through the lower part of the section near the base (Stone and McFarland, 1981).

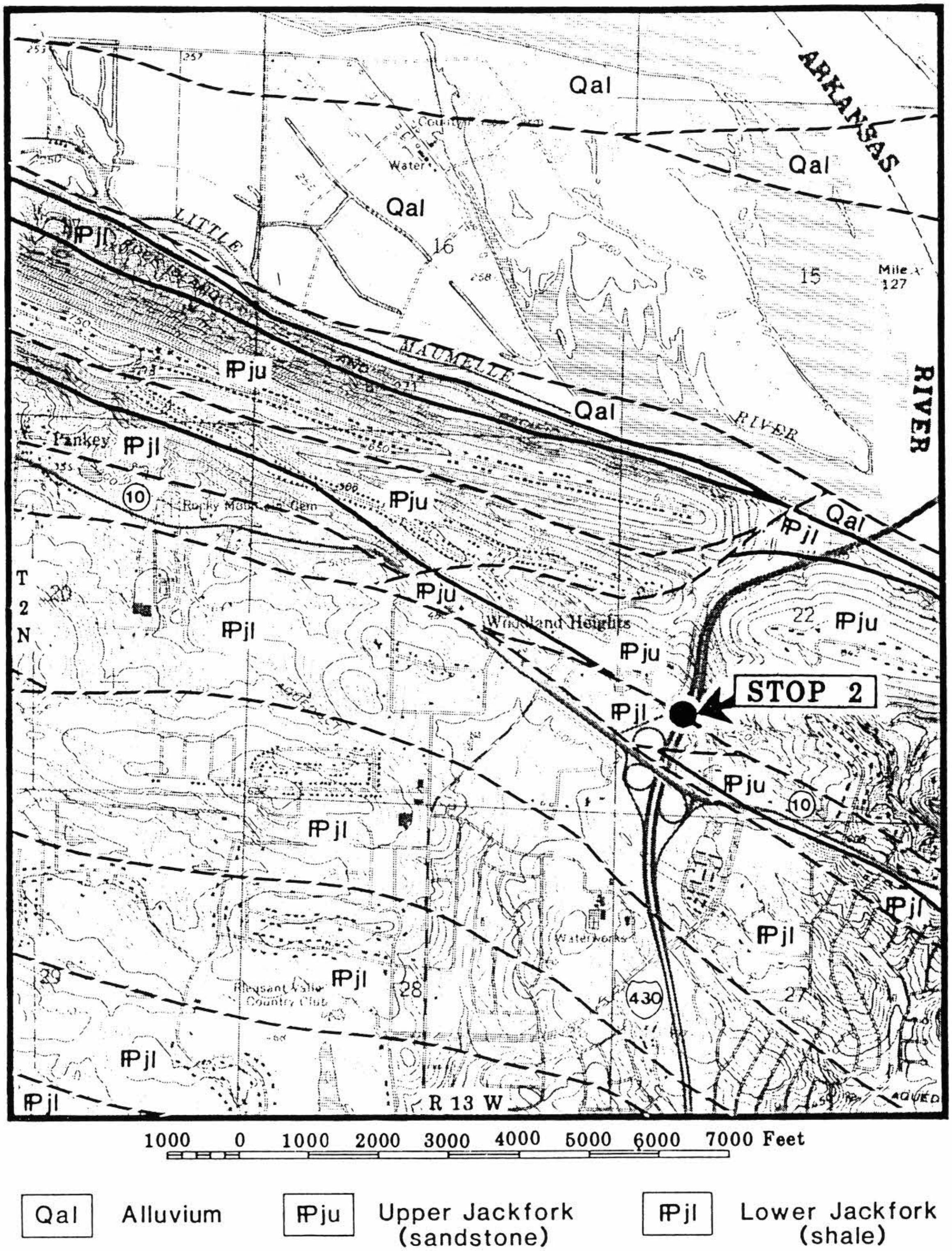
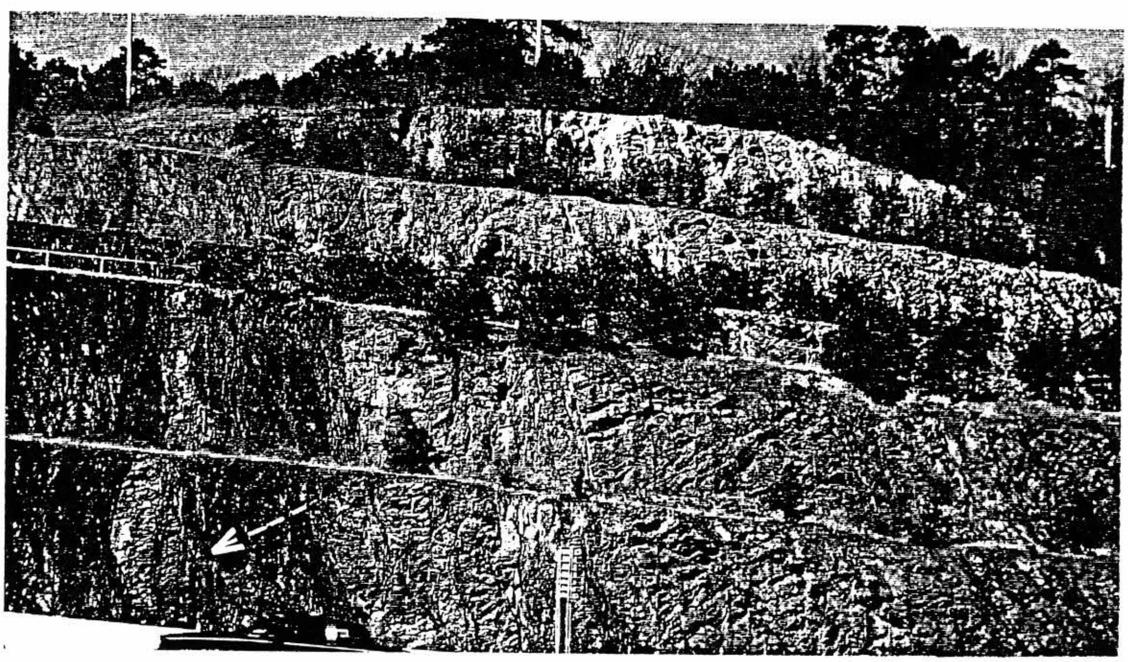


Figure 79. Geologic map of northwestern Little Rock in vicinity of Stop 2 along Interstate 430.

B



A

Figure **80.** (A) West side of Interstate 40 roadcut at the State Route 10 exit ramp. Basal part of sandy section showing faulted (arrow) Jackfork sandstone and shale with possible sedimentary slide features. (B) Close-up of photograph shown in A displaying interbedded nature within pod of sandstone and shale, and northward-dipping fault plane (arrow). Photograph courtesy of Arkansas Geological Commission.

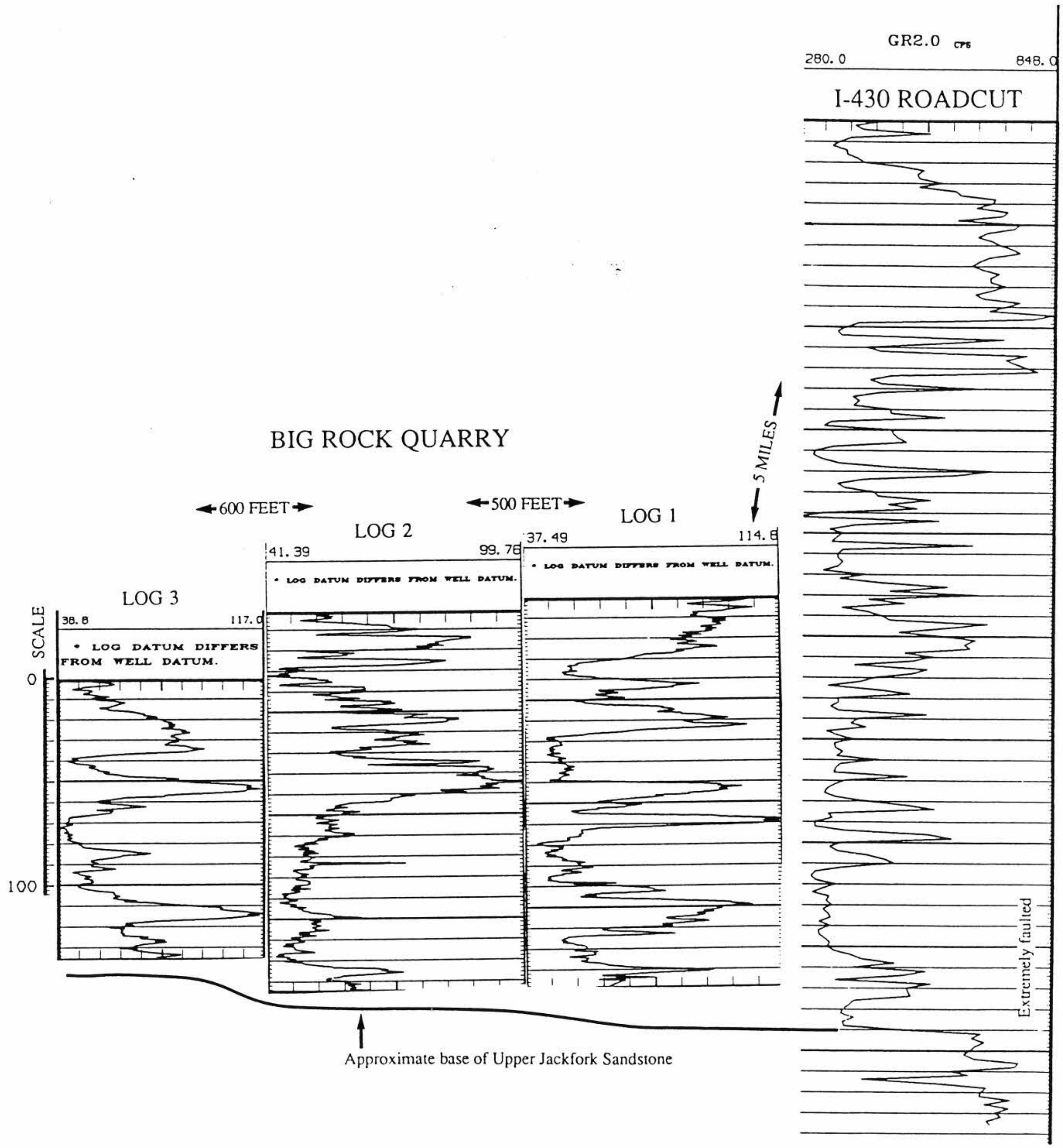


Figure **81**. Gamma-ray logs hung on the approximate (erosional?) base of the upper Jackfork Sandstone. The section at the Interstate 430 roadcut is inferred to be stratigraphically equivalent to the section at Big Rock Quarry. Note changes in horizontal scale.

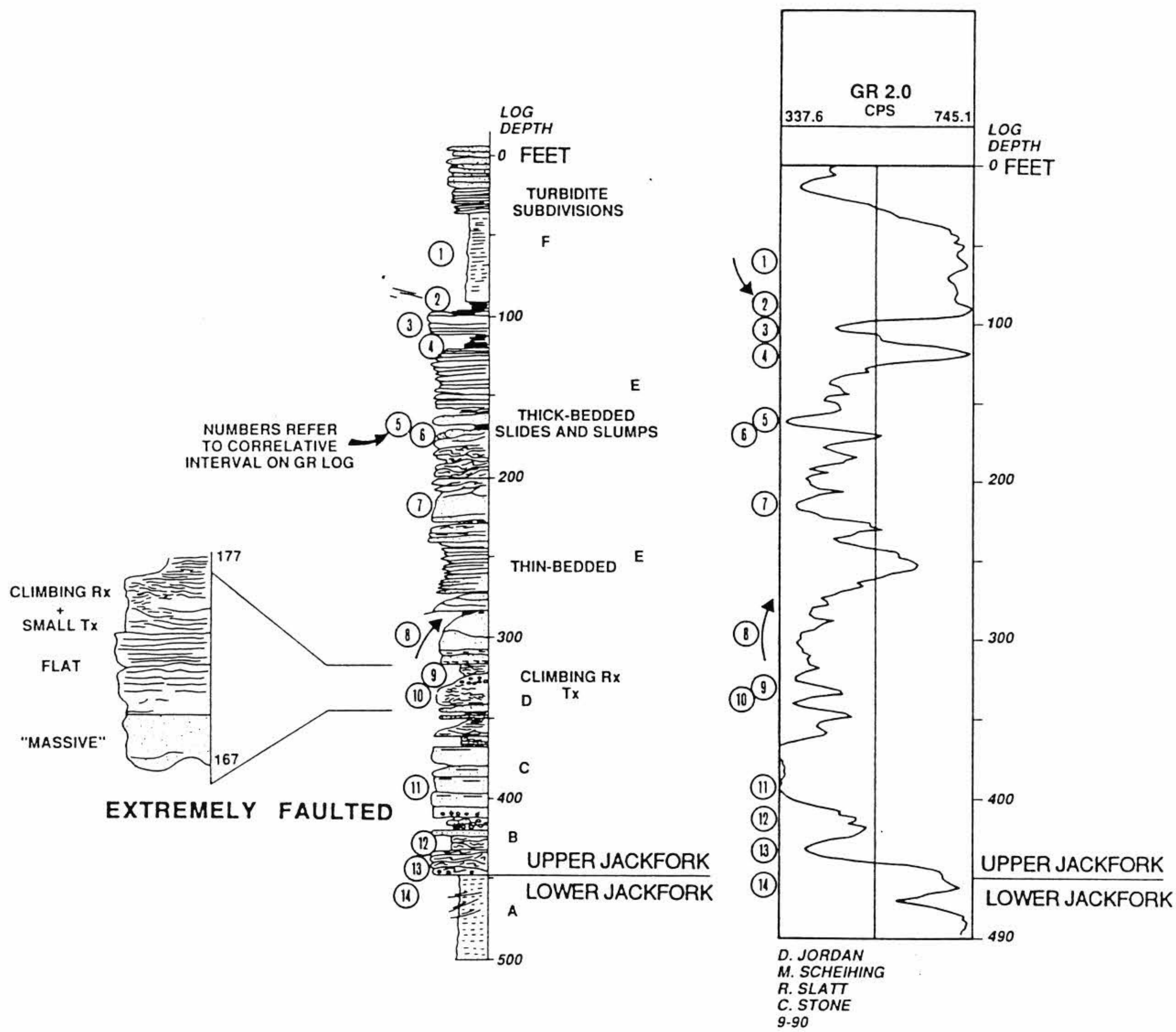


Figure **82.** Stratigraphic column and gamma-ray log character of the lower and upper Jackfork exposure along Interstate 40.

Several intervals in the highly-contorted lower Jackfork shale beds contain soft-sediment-deformation, but many of the beds have been offset and contorted by tectonic processes. Can you tell the difference? What is the evidence? What is the depositional setting of the shale? Look for trace fossils, especially *Chondrites* (thin, branching, siltstone-filled tubes). One or two types of trace fossils reflecting specialized activity may assist in offering clues to the nature of the depositional substrate and/or overlying water column. Within the sandstones, look for the mudstone-clast conglomerate and breccia similar to that found in the float block on the quarry floor. In addition, an interesting interval lying about 125 feet above the base of the upper Jackfork sandstone contains massive-appearing, 'clean' sandstone overlain by flat- to small scale cross-laminations, capped by an interval with climbing ripples within a whitish-gray sandstone and shale sequence. What are some of the possible scenarios for deposition of these beds? Which parts of the sequence suggest deposition from high-density and low-density flows? could contour currents be involved?

STOP 3

"Old" Big Rock Quarry

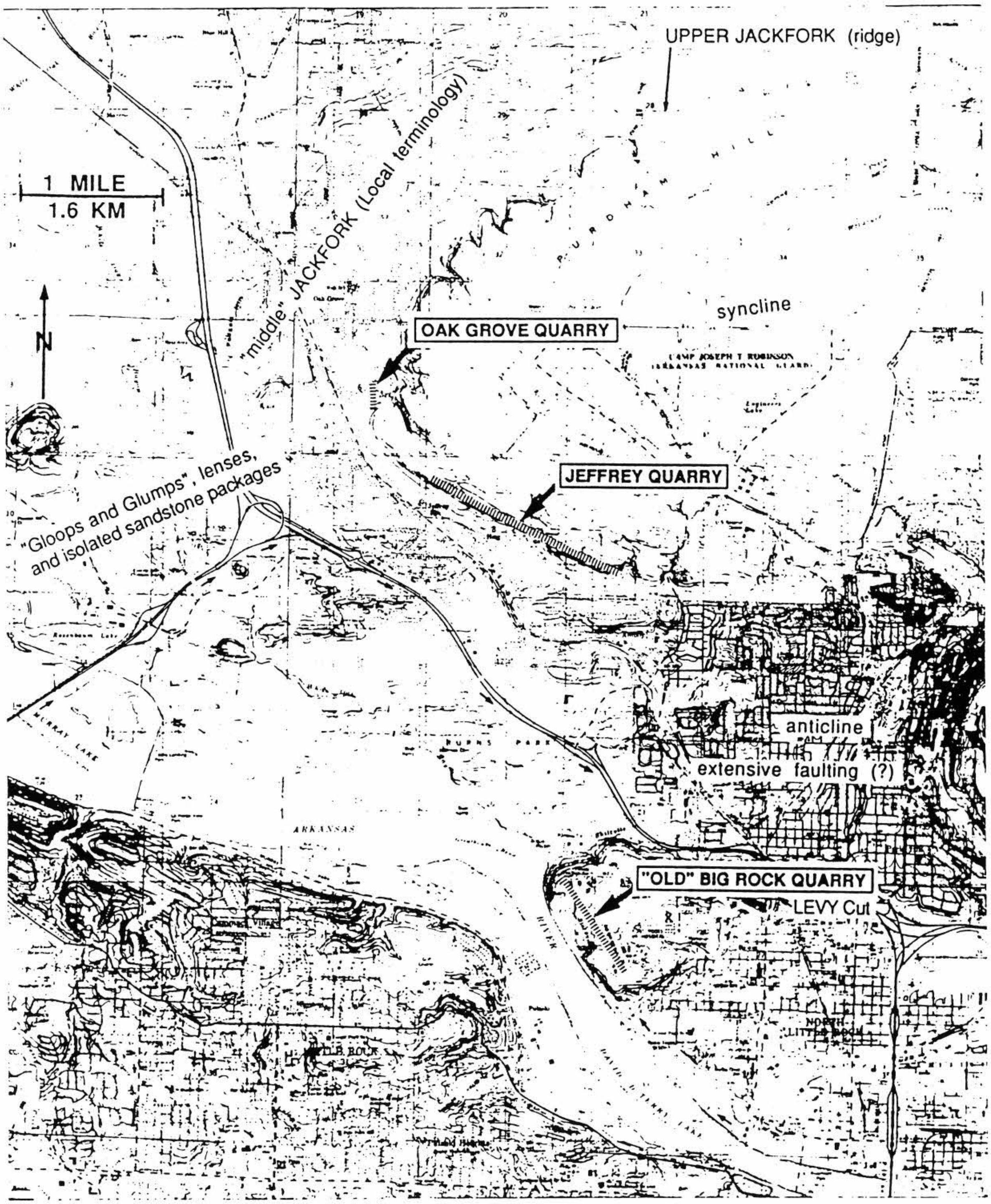
This important stop shows reservoir and outcrop-scale heterogeneities and the different ways correlations of a set of well logs can be performed. This large exposure will first be viewed from a high-located overview point which will provide a generalized impression of the complicated channel complex likely located originally close to the downdip side of the base-of-slope. Its environmental location likely was slightly downdip from the Pinnacle Mountain site.

We will drive around and enter the quarry for some close up observations and a discussion of the different correlations that can be performed using the three well log sites occupied by ARCO. The text following comes from their AAPG fieldtrip guide (Jordan et al., 1991).

Big Rock Quarry, located along the north bank of the Arkansas River in North Little Rock, Arkansas (Fig. 83), provides perhaps the largest continuous exposure of proximal deepsea flysch deposits in North America. The quarry was a source of crushed stone in the 1950;s but is used today mainly for storage of sand dredged from the Arkansas River.

The vertical faces of Big Rock Quarry expose a two-dimensional view of the lower part of the upper Jackfork Group that is up to 200 feet high and nearly 3000 feet long. Lower Jackfork shales are exposed in the floor of the northern part of the quarry. Regional paleocurrent analysis (Morris, 1977) and facies studies (Link and Roberts, 1986) suggest that this area lay close to the northeastern edge of the deep-water Ouachita Trough during Jackfork deposition. Sediments from sources in the Appalachian Mountains, more southern orogens, and possibly the Canadian Shield, moved across alluvial and deltaic plains, poured onto the fringing slopes leading into the deep-water trough, and flowed generally westward within the basin. Rocks exposed within the quarry are thought to represent some of the most proximal Jackfork units in the Ouachita Mountains.

Pervious interpretations have suggested that this quarry contained fourteen upper submarine fan channel deposits (and the sequence pinches out about 2.5 miles away) (Stone and McFarland, 1981); at least fourteen coalescing and stacked channelized packages of an inner (upper) fan valley (Moiola and Shanmugam, 1984); and channel-fill and levee deposits in a submarine canyon or upper part of a submarine fan channel system (inner fan valley system) (Link and Stone, 1986a; Link and Roberts, 1986). The lateral dimensions of the major sandstone bodies are unknown but may have reached several miles.



UPPER JACKFORK (ridge)
1 MILE
1.6 KM
N
"middle" JACKFORK (Local terminology)
syncline
OAK GROVE QUARRY
JEFFREY QUARRY
"Gloops and Glumps", lenses, and isolated sandstone packages
anticline
extensive faulting (?)
"OLD" BIG ROCK QUARRY
LEVY Cut

Figure 83.

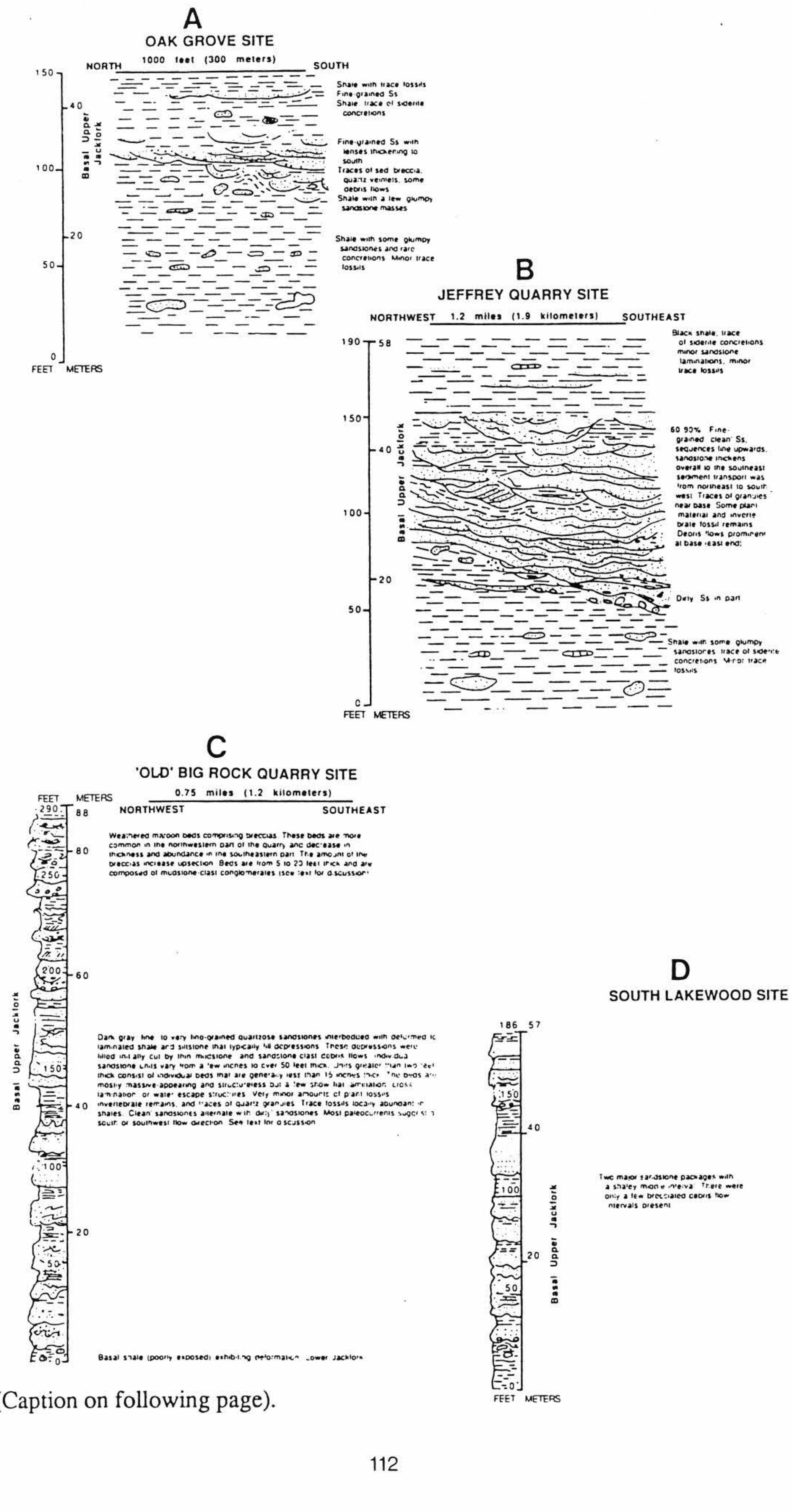


Figure **84.** (Caption on following page).

Figure 84. Schematic field sketches depicting lithology and lateral extent of northern part of sandy submarine canyon fill near Little Rock. Refer to Figure 83 for locations of sections oriented in a generally north to south direction which may have been the orientation of the canyon opening. Sediments were fed generally from east to west. (A) Basal upper Jackfork sandstones near the axis of the MarcheSyncline at the Oak Grove site. The strata strike north-south and dip some 10° to the east. (B) Laterally equivalent to sequence in A, but sequence contains more sandstone (it is presumably closer to the axis of the ancient submarine canyon); south flank of Marche Anticline. Strata dip 23° to the north. (C) 'Old' Big Rock Quarry within the axis of the Big Rock Syncline. Dips range from 0 to 10° across the entire exposure. The section averages 80% sandstone. (D) Column representing a series of partial exposures that were mostly destroyed during the construction of I-40 and U. S. Highways 67-167 along the southern border of Lakewood in North Little Rock, approximately 6.1 miles due east of 'Old' Big Rock Quarry. Located on the south flank of the Big Rock Syncline, the strata strike east-west and dip 30° to the north.

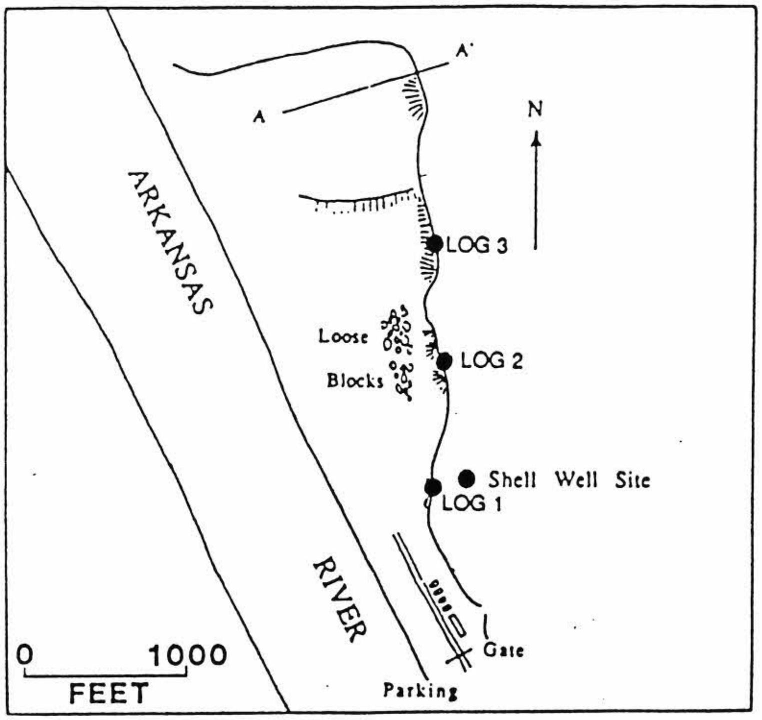
The quarry section can be divided into three main lithologic units (Fig. 85: (1) a basal unit, representing the upper portion of the lower Jackfork Group, (2) a section of sandstone and shale, 75 to 150 feet thick, at the base of the upper Jackfork Group, and (3) a capping unit of interbedded shale-clast breccia and sandstone.)

The lower Jackfork mudstone underlies much of the quarry floor, but is well exposed only toward the northwestern end of the quarry. The erosive nature of the lower to upper Jackfork contact is well exposed at the Levy outcrop section and at the I-430 outcrop. At the Levy outcrop, about 3 miles northwest of Big Rock Quarry, upper Jackfork sandstones fill a large channel cut into underlying lower Jackfork mudstone. The mixing and deformation exhibited by the basal shale here and in adjacent outcrops suggest that it may constitute either a large slump sheet, emplaced before deposition of the upper Jackfork sandstone and shale, or that the upper Jackfork is a large soft-sediment glide or tectonic sheet emplaced by shearing along the contact.

Within the sandstone and shale section exposed in Big Rock Quarry, virtually every major and minor lithologic unit is discontinuous. Individual units vary from a few inches to over 50 feet thick. Units greater than two feet thick invariably consist of individual beds or layers that are generally less than 15 inches thick, and are composed of dark gray, fine- to very fine-grained quartzose sandstone. The beds are mostly massive-appearing and structure less but a few show flat lamination, cross lamination, or water escape structures. Wavy and undulatory bed contacts are common and were probably produced by liquefaction and sediment foundering. Most beds were emplaced by turbidity currents. The thinness of individual beds and the paucity of ripped up mudstone clasts and clearly scoured bases suggests, however, that the currents were not highly erosive.

Thicker sandstone units, from 5 to 50 feet thick, are lenticular and interrupted by inclined accretionary bedding, truncation surfaces, or internal scour surfaces when traced across the outcrop. When viewed from the northwest end of the quarry (Fig. 86), a number of the thick sandstone units can be seen to lens out from southeast to northwest against the overlying breccia. These units could represent sheets that moved downslope as large slides.

In the southeastern part of the quarry (Fig. 87), near the base many small, scoured depressions are eroded into sandstone units. These depressions were filled initially by thin mudstone- and sandstone-clast debris flows or drapes of laminated mudstone. The common association of scours, mud drapes, and small debris flows suggests that after the scours were cut, they were draped by fine muddy sediments. These observations are supported by the core description of Link and Stone (1986a), discussed later. An excellent example of



Outline Map of Big Rock Quarry area, North Little Rock, Arkansas. Location of shelf core hole and gamma-ray logging sites 1-3 are indicated (Jordan et al., 1991).

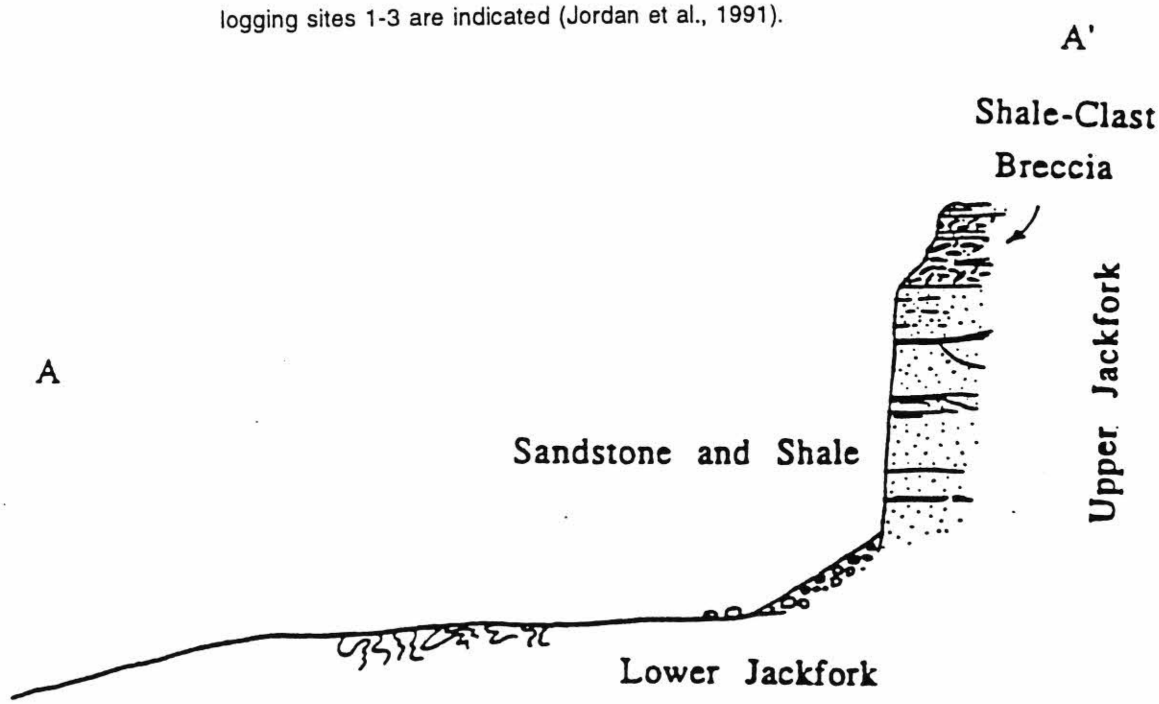


Figure **85.** Cross section of Big Rock Quarry.

Figure **86.** View from top of northern face of Big Rock Quarry showing pinching out of large sandstone packages beneath mudstone-clast breccia. Note abundance of surfaces inclined at angles to bedding and perhaps defining accretion surfaces or channel margins.

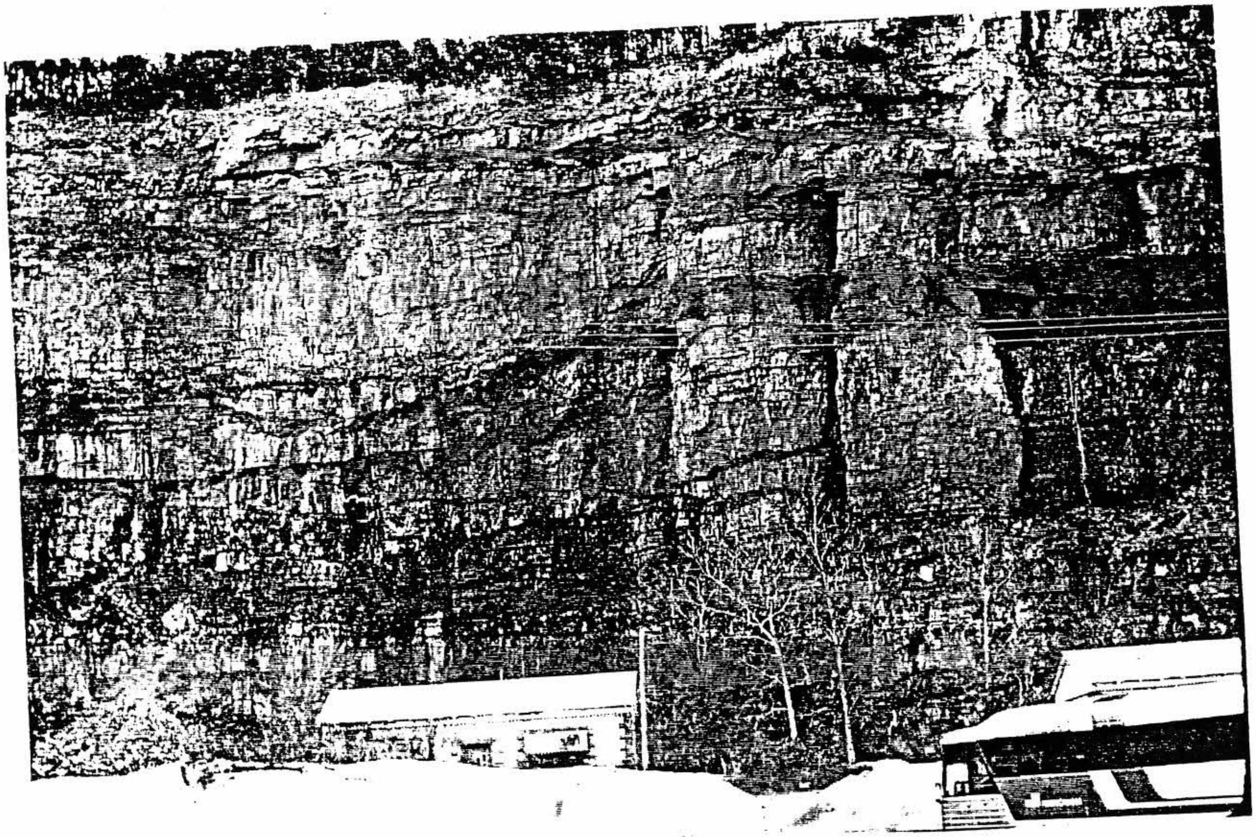


Figure 87 Southeastern part of Big Rock Quarry showing many small, scoured depressions eroded into sandstone units.

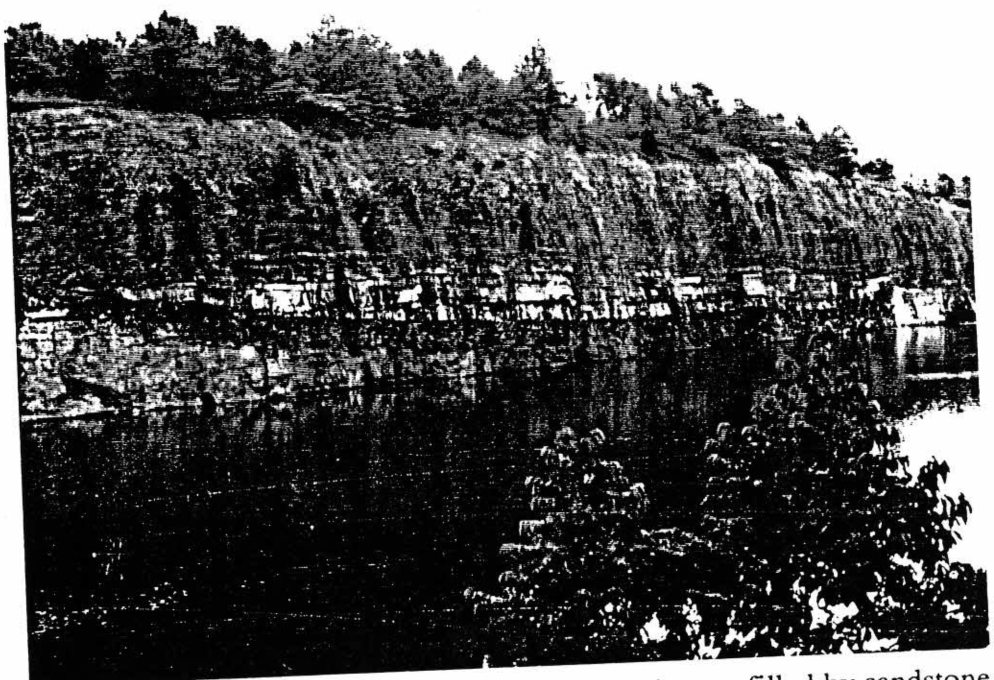


Figure 88 An excellent example of a large scale, shale-draped scour filled by sandstone. Jeffrey Quarry (see Fig. 83 for location and Fig. 84 for description). These scours are unusual because they are not filled with coarser sediment or lag deposits left by energetic erosive currents that would be required to cut the depressions if they are channels. More likely, the channel-like truncation surfaces represent excavations or scars left by large slope failures and slides.

a larger scale, shale-draped scour filled by sandstone occurs approximately 7 miles north of Big Rock Quarry at Jeffrey Quarry (Fig. 88).

These scours are unusual because they are not filled with coarser sediment or lag deposits left by energetic erosive currents that would be required to cut the depressions if they are channels. It seems unlikely that the small debris flows at the bases of the channel-fill were effective in eroding the depressions. Possibly these features are channels eroded by turbidity currents that bypassed the local area without leaving deposits. If so, it seems unusual that the channel-fill consists largely of fine- to very fine-grained sandstone that should also have bypassed the area. More likely, the channel-like truncation surfaces represent excavations or scars left by large slope failures and slides. Such failures should have been abundant on slopes leading into the deep-water trough. The contained sandstone might then represent sediments trapped in the scar and locally left on the slope during large-scale floods of coarse sediment into the basin. However, the dominance of fine-grained sandstone in the quarry implies that this represents the coarsest sediment being transported across this portion of the slope. Coarser sediments may have moved through slope channels in other areas, but it seems unlikely that Big Rock Quarry provides an example of proximal feeder systems.

Downslope depositional sites (such as those we will see on Day 2) actually show substantially coarser-grained sediments. The uppermost unit exposed along the quarry face consists of from 5 to 20 feet of mudstone-clast conglomerate and breccia (Fig. 89). The matrix varies from clean fine-grained sandstone to muddy sandstone. The unit is unusual for several reasons: (1) it contains an extremely high percentage of mudstone clasts (over 50%) in a clean sandstone matrix, making it unlikely that it represents a cohesive debris flow; (2) it contains thin, undisturbed layers of sandstone from less than an inch to over 2 feet thick enclosed in breccia; (3) breccia blocks fallen to the quarry floor commonly show a crude alignment of mudstone clasts that outline original layers which have not been completely obliterated, and (4) the mudstone clasts are commonly blocky and some are oriented at nearly right angles to bedding, suggesting little shearing and squeezing during flowage. These units probably formed as dense sand slurries formed by local slope failures. Most have not travelled far. More turbulent flows moved downslope beyond the present area and may have deposited the interlensing, intact units of clean sandstone. Sandstone beds within the breccia define large low-angle cross stratification or dipping accretionary bedding that has been interpreted by some to have formed on natural levees. Mapping of these beds suggests they accreted laterally toward the north.

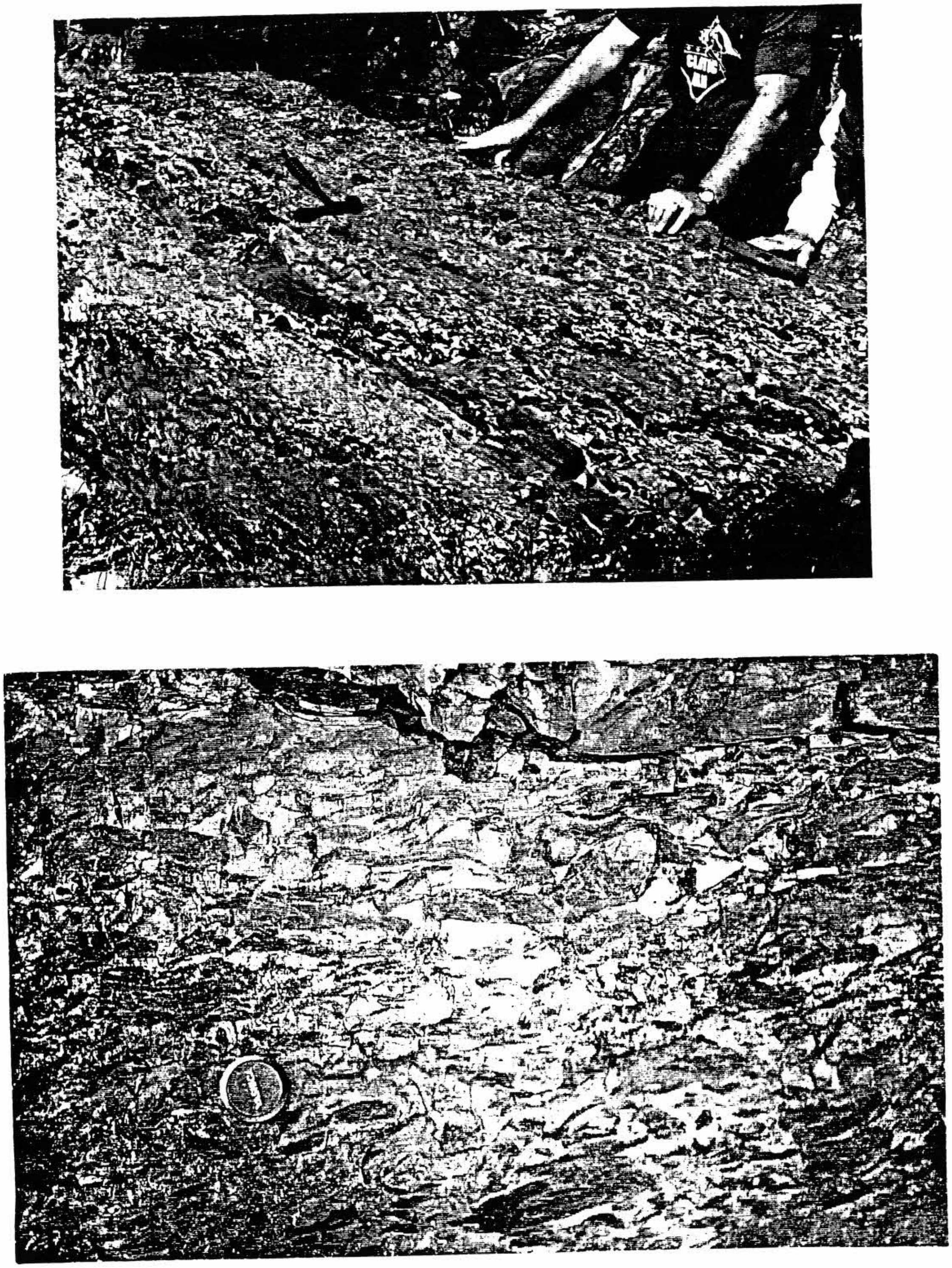


Figure **89** Fallen block on floor of quarry derived from upper part of quarry section. (A) Crude alignment of mudstone clasts that outline original layers which have not been completely obliterated. (B) Close-up of fallen block shown in A. Mudstone clasts are commonly blocky and oriented at nearly right angles to bedding, suggesting little shearing and squeezing during flowage. These units probably formed as dense sand slurries formed by local slope failures.

Overall, Big Rock Quarry exposes a slope or base-of-slope succession of the type that might be expected on the proximal parts of deep-sea fans within slope channels or canyons. The lateral variability of individual sedimentation units and thicker sediment packages reflects the activity of localized movements, such as slumps and slides, as well as the importance of depressions of all scales in localizing flows. Scour and successive slope failures have cut out many originally more continuous units. Sedimentation was dominated by turbidity currents, possible liquefied flows, debris flows, and slides and slumps (Fig. 90).

The overall thickness of the sandy units exposed at Big Rock Quarry decreases in a northerly and easterly direction, and suggests that what is present within the upper Jackfork in this area represents the coarser clastic fill of a submarine (slope?) canyon. Field relationships suggest that the Big Rock Quarry and several exposures in the immediate area are related and represent a submarine canyon-fill sequence. This canyon was probably fed from the east and sediments were transported to the west. The present northwestern edge of the submarine canyon fill is possibly near the "dying" Jackfork ridge to the northeast of the Oak Grove site. The basal upper Jackfork Sandstone is thicker at the Big Rock Quarry site and to the west (at the I-430 roadcut) than inexposures in the northwest and southeast. This suggests proximity to the axis of the sandy submarine canyon fill.

Three gamma-ray logs, obtained with a logging truck and spaced 500 and 600 feet apart (Fig. 91), were measured along the northeast quarry wall. Description of a detailed lithologic section was not possible because of the steepness of the quarry wall. However, Figure 92 compares log #1 with a subsurface core of the upper part of the same interval taken approximately 50 feet behind the quarry wall several years ago by Shell Oil company and described by Link and Stone (1986). The correlation between the log response and gross lithology is good. The exception to this occurs in the thinly-bedded interval between 20 and 40 feet. The log response depicts this interval as a thicker sandstone sequence than is actually the case.

The three logs (Fig. 93), taken at this site illustrate potential problems in well log correlation of laterally discontinuous strata. On several field trips, participants have correlated these logs prior to examining the outcrop (Fig. 94). Typically four or five distinct stratigraphic intervals are subdivided on the logs (Fig. 94A). However, when the quarry is examined, most beds are not laterally continuous at the 500 and 600 foot spacing of the logs. A better correlation, made after studying the quarry face, is shown in Figure 95.

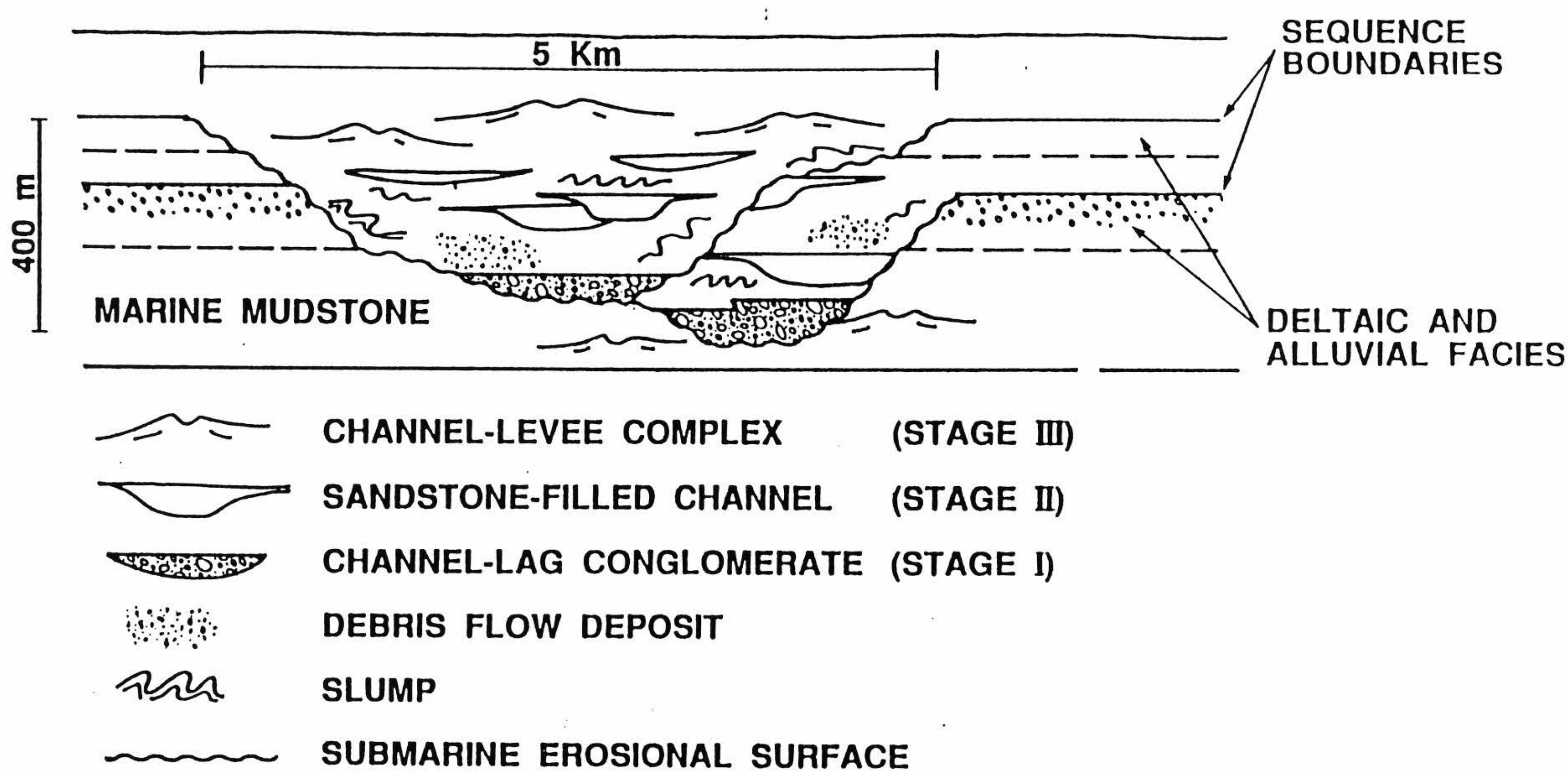


Figure 90. Fill of large-scale submarine erosional channels. Note different stages in filling of the channels (after Mutti, 1984). Sedimentary deposits comprising the sedimentary fill within the large submarine erosional feature at Big Rock Quarry include debris flow deposits, slumps, channel-lag conglomerates, sandstone-filled depressions, and possible channel-levee accretionary beds at the top of the quarry.

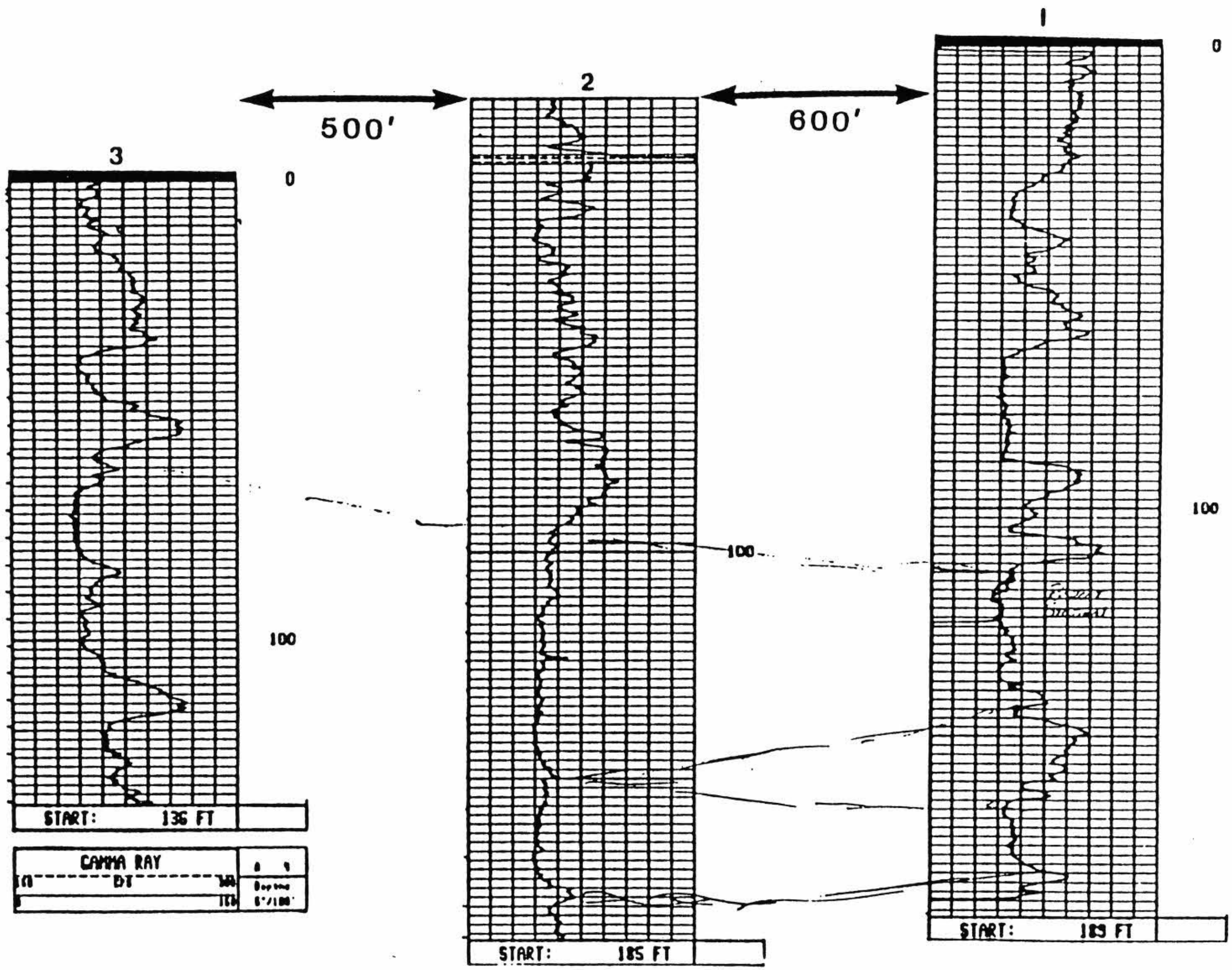


Figure **91.** Gamma-ray logs obtained with a logging truck along the northeast wall of Big Rock Quarry. Datum is top of quarry.

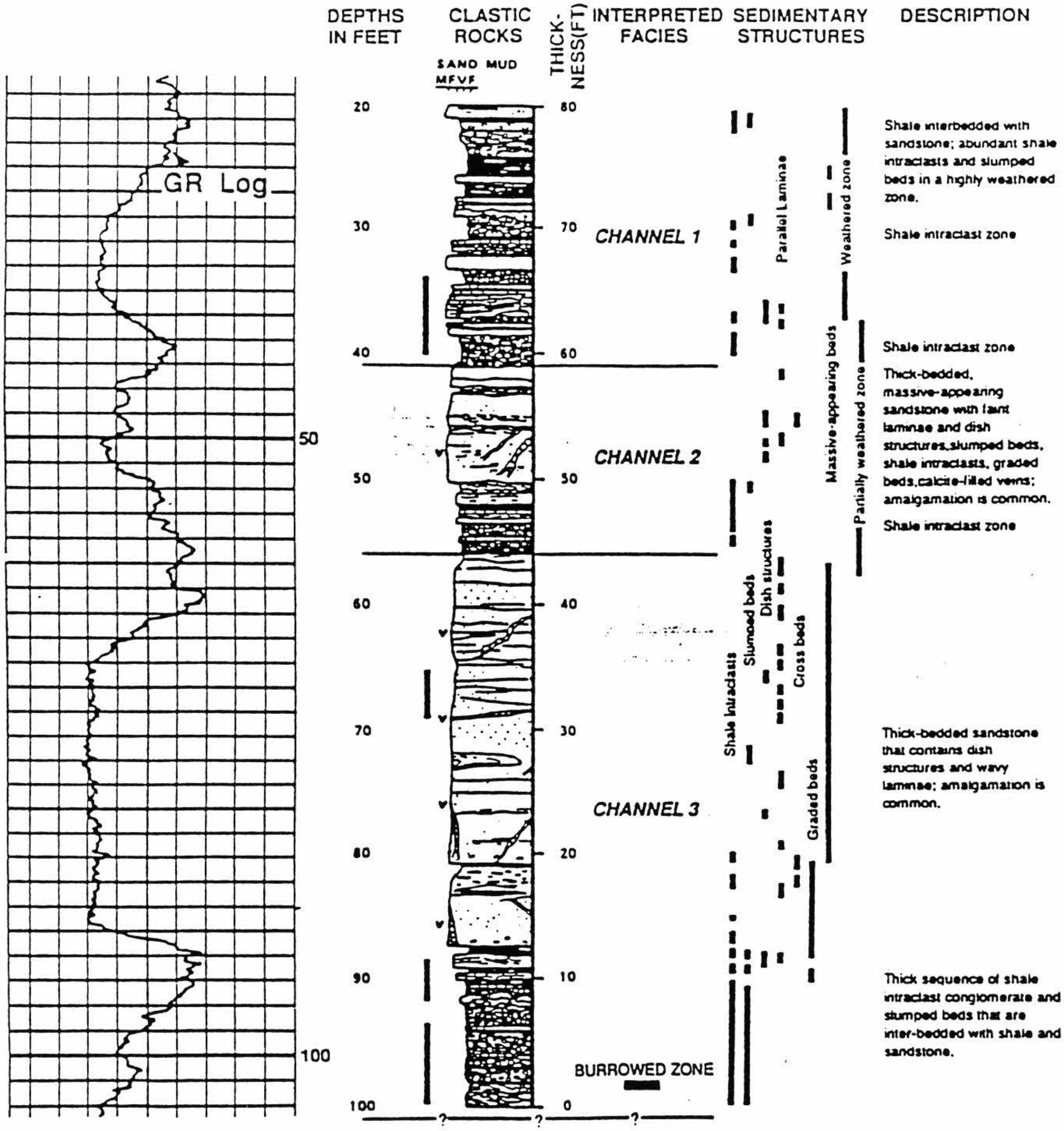


Figure **92.** Comparison of gamma-ray log #1 at Big Rock Quarry with a Shell Oil Company core taken 50 feet behind the quarry face at the logging site. Core description is after Link and Stone (1986).

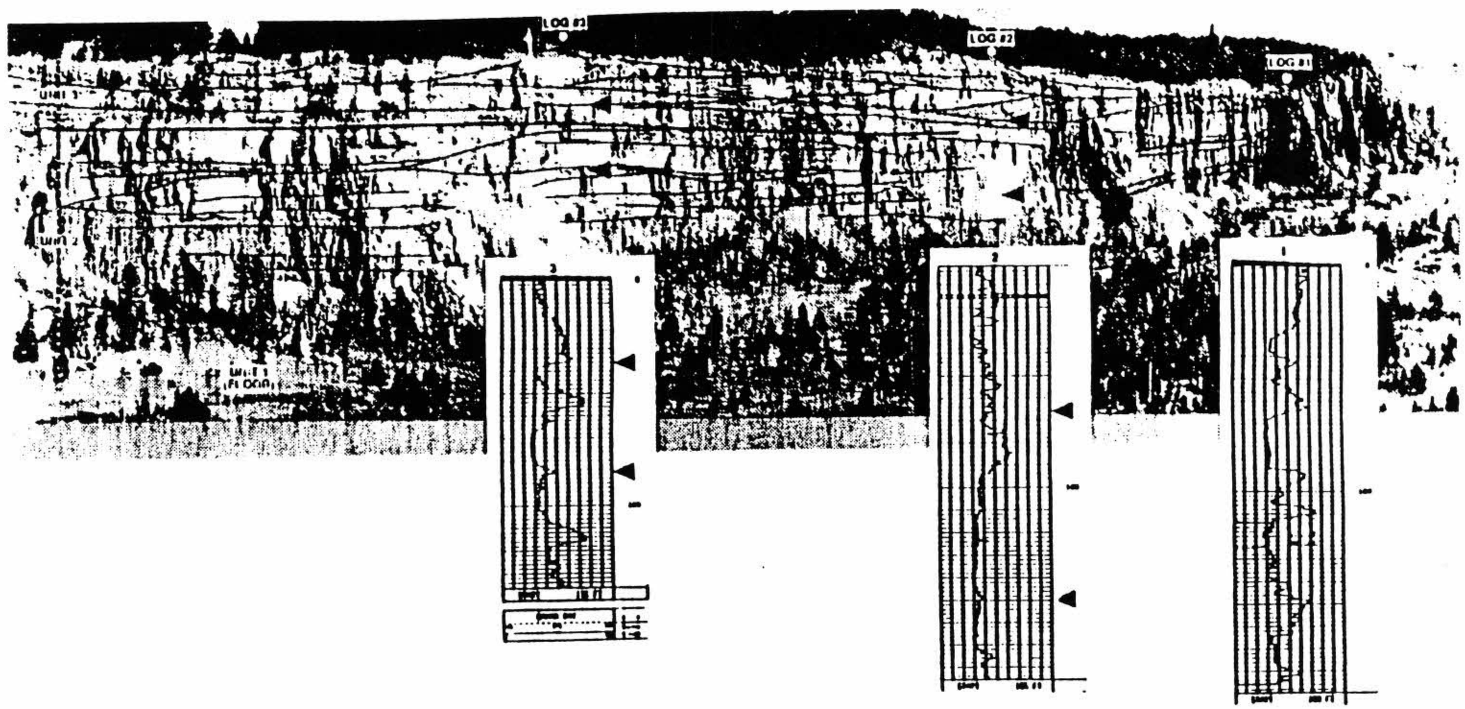
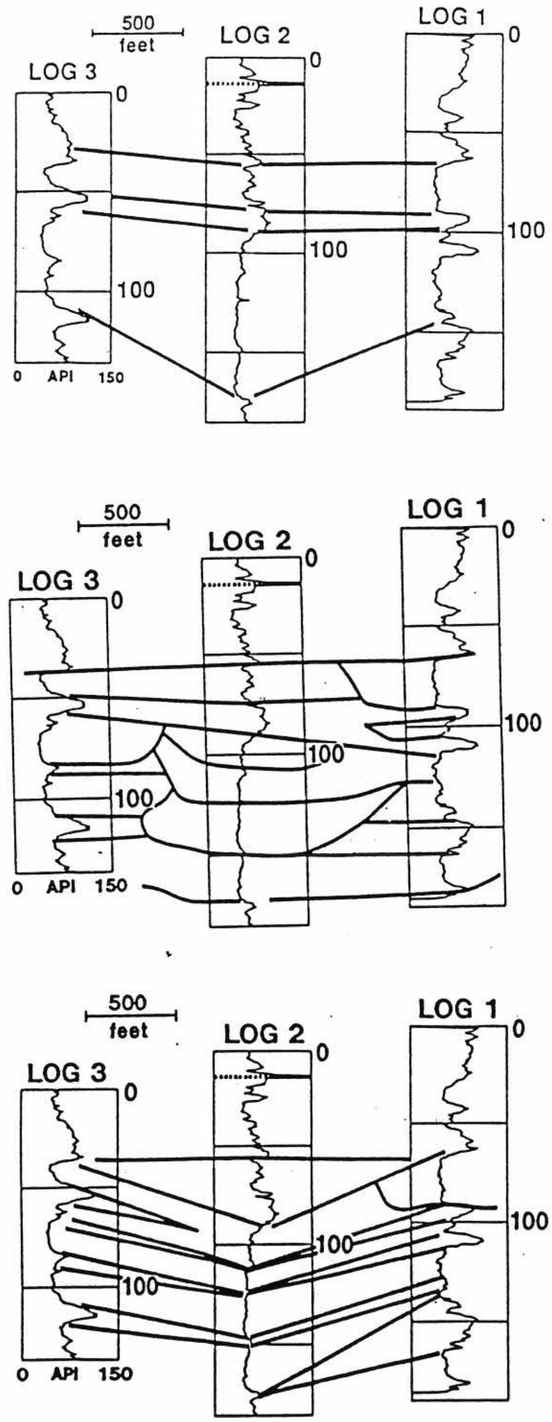


Figure **93** Three gamma-ray logs positioned at log sites on quarry face. Key bedding boundaries are outlined in black on outcrop wall. Correlation points between the outcrop and the well logs are shown by black triangles. Note extreme lenticularity. Compare with Figures 12 and 13.



"LAYER-CAKE" CORRELATION

"LENTICULARITY" CORRELATION

"MIXED" CORRELATION

Figure 94. Three possible, but incorrect, correlations of gamma-ray logs from Big Rock Quarry, illustrating (A) layer-cake, (B) lenticular, and (C) mixed correlations. Correlation shown in A is probably the most common correlation made before studying the outcrops. Correlation in B invokes numerous channels and erosion of section. Correlation C document changes in thickness of interbedded shale

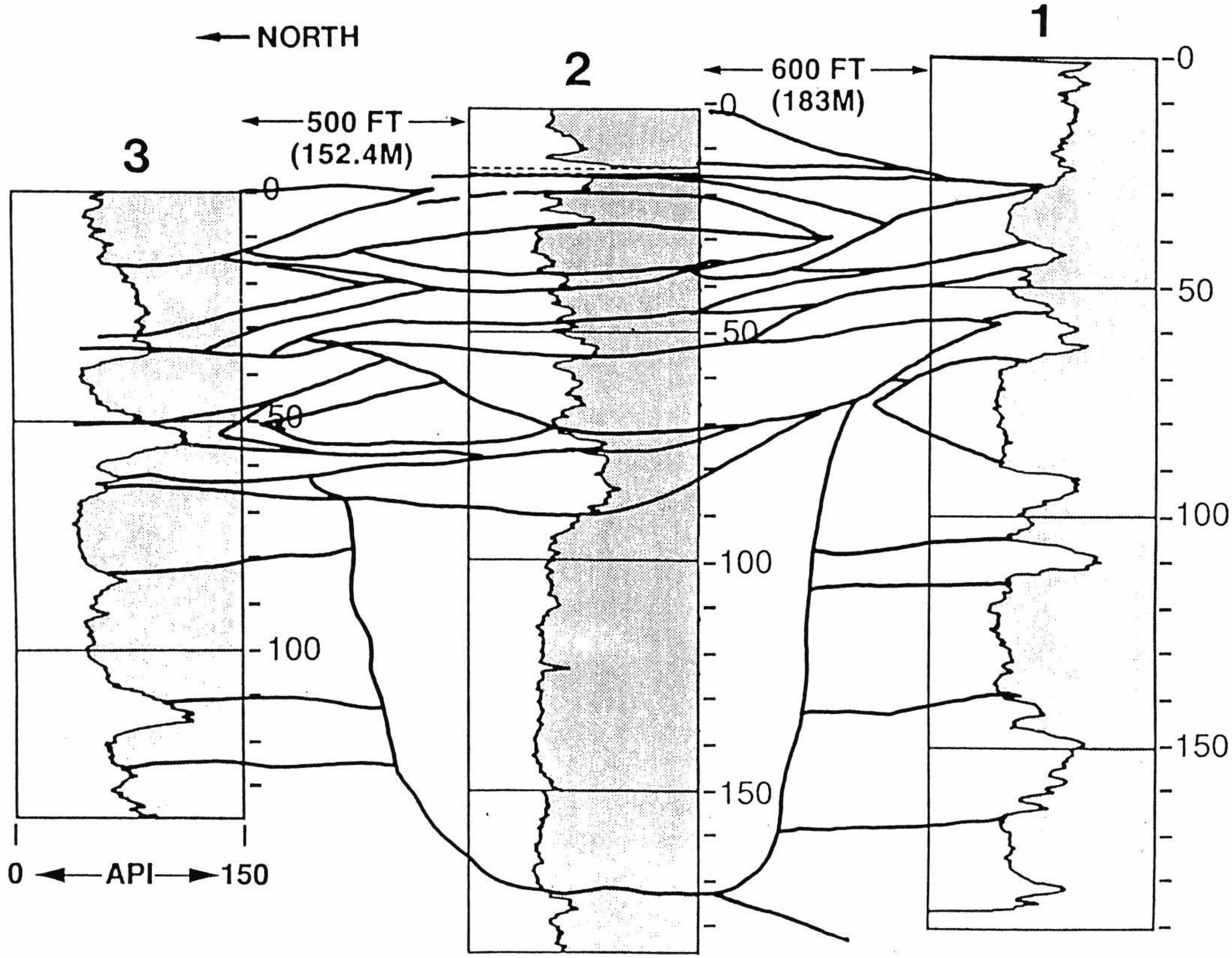


Figure **95.** Stratigraphic correlations of gamma-ray logs #1, 2, and 3 at Big Rock Quarry, based of field sketches (From Jordan et al., 1991).

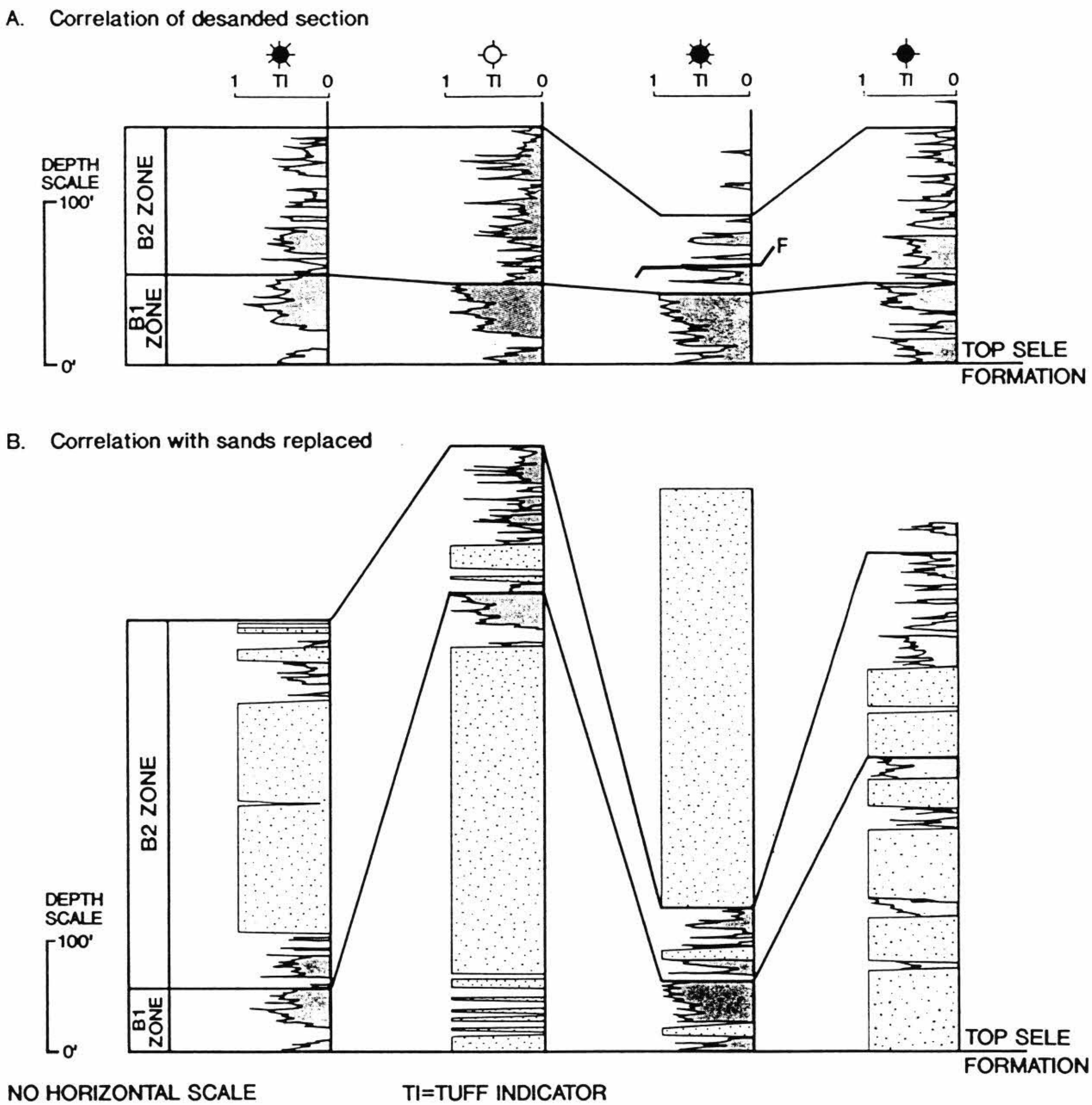


Fig.95aCorrelation of Gryphon Field well logs. **(Newman et. al 1993)**

In comparing this correlation to that shown in Figures 93 and 94, it is particularly noteworthy that the thick sandstone from 88-168 feet on Log #2 is a discontinuous sandstone slump upon which younger sandstone beds onlap, and (2) the cleaning upward sandstone unit at 0-30 feet on Log #3 is a laterally discontinuous sandstone bed near the top of the quarry face. In the subsurface, these beds, though thick, would not extend between typically-spaced wells (40 acre spacing), so knowledge of the expected depositional geometries would help clarify the degree of uncertainty associated with well log correlations.

Sandstone/shale ratios for the Jackfork exposed in Big Rock Quarry were calculated in an effort to quantify lateral and vertical variability of the deep-water facies. Ratios were determined by counting sandstones and shales within equally-sized rectangular cells superimposed on enlarged outcrop photographs. A cell contains an area that is 500 feet wide and 50 feet high. Each cell was overlain by a 0.1" X 0.1" grid which defined the actual point that was counted. Three outcrop sections were examined.

SECTION 1

Three equally-sized cells were outlined on the outcrop photograph of the southeastern end of Big Rock Quarry. Sandstone/shale ratios were calculated for each cell.

Comparing these results with visual observations on photographs points out: (1) overall lithologic differences between units are not as large as would be estimated by simple visual observation, and (2) the presence of thin-bedded units drastically alters interpre-tation of sandstone/hale ratios. For instance, cell number 3 contains lenticular sandstone bodies overlain by a combination of thick-bedded sandstones and thinly-interbedded sandstones and shales. One might visually estimate a sandstone/shale ratio of 1.5 (60% sand); however, counting the sandstones and shales within the cells produces a much higher value of 4.65 (approximately 82% sand). Based upon these counted data, this unit actually proves to be the most sand-rich of the three cells.

SECTION 2

This outcrop section is divided into 4 cells with each cell further subdivided into 2 sub-cells (a & b). Results show that sandstone/shale ratios are vertically and laterally highly variable. The lower portion of the section contains large lenticular sandstone bodies which have high sandstone/shale ratios (3b = 5.30 and 4b = 4.64); whereas, the upper portion of the section contains thinner sandstone bodies which grade laterally into shaley deposits with lower sandstone/shale ratios (1b = 2.76 and 2b = 2.55).

Results also indicate that sandstone/shale ratios are highly variable depending upon: (1) the lateral continuity of major units, and (2) the relationship of these units to the cell pattern established for point-counting. For example, units 3b and 3a display the most dramatic change as interbedded sand-tones and shales of unit 3b (ratio = 5.3) grade into massive sandstones within unit 3a (ratio = 93.7). However, if considered as a single cell (cell 3a + 3b), the overall ratio is 9.67. This is of the same magnitude as the value for cell 3b; however, it is not of the same magnitude as cell 3a. Grid cells must be assigned with a consistent set of criteria (lithology, facies, etc.) and a constant sense of scale. for example, if a cell encompasses a uniform lithology, then the larger cell size may be just as accurate as the smaller (sub-cell) size. If a cell encompasses highly variable lithologies, then a smaller grid may be necessary to accurately reflect the true sandstone/shale ratio.

SECTION 3

The central and northwest portions of Big Rock Quarry were also gridded into 12 cells. Sandstone/shale ratios are presented on the bottom of the photograph.

Sandstone/shale ratios cannot be used as isolated data. For instance, a large sandstone-rich slump feature occurs in the lower right corner of the outcrop photograph over which log #2 was run. This sandstone mass creates abnormally high sandstone/shale ratios within the cells. However, this slump is limited extent, and the sandstone/shale ratios are not representative of the entire cell.

Sandstone/shale ratios were also measured from the gamma-ray logs at location sites #1, #2, and #3 to compare and calibrate log responses to lithologic properties. The logs were subdivided into vertical units that corresponded to the cell boundaries drawn on the outcrop photographs. Sandstone/shale ratics were calculated by determining (1) the shale base line on the gamma-ray log (assumed to represent a clean, 100% shale), and (2) the line of maximum gamma-ray deflection (assumed to represent a clean, 100% sandstone). The interval between these two end points was divided into thirds. The portion of the curve remaining to the left of the 33% shale cut-off line (drawn one third left of the clean shale base line) was counted as sandstone, and the portion of the curve to the right of this line was counted as shale. Sandstone/shale ratios were calculated, and comparisons of log and outcrop derived ratios are presented in Table 3.

Sandstone/shale ratios calculated from gamma-ray logs at log site #1 closely match the corresponding values obtained from computations within the outcrop grid cells. Log and outcrop data are comparable (82% log versus 79% outcrop for cell 3 and 68% to 70% for cell 1) in this geologic setting

Log Location #1		Log Location #2		Log Location #3	
Outcrop	Log (33% Shale Cut-Off)	Outcrop	Log (33% Shale Cut-Off)	Outcrop	Log (33% Shale Cut-Off)
2.29 (70%)	2.11 (68%)	1.78 (64%)	7.71 (89%)	0.469 (32%)	23.8 (96%)
3.84 (79%)	4.65 (82%)	1.16 (54%)	0.71 (41%)	4.53 (82%)	4.0 (80%)
–	–	1.59 (61%)	– (100%)	10.52 (91%)	– (100%)
–	–	3.25 (76%)	– (100%)	20.0 (95%)	3.33 (77%)

Table 3 Big Rock Quarry sandstone/shale ratios. Comparisons of outcrop and log-derived values.

where thinly-bedded units are minimal. The 33% shale cut-off used in the log-measuring technique appears appropriate in this case.

Log and outcrop ratios at log site #2 match poorly. The slump-feature appears on the gamma-ray log as nearly 100% sandstone, but the surrounding beds observed on the outcrop face consists of interbedded sandstones and shales within the remaining portions of the cell. Also, thin-bedded sandstones and shales that can be observed on outcrop photographs are included in the computation, but most of these units are too thin to be resolved by the gamma-ray logging tool and are not represented on the log curves. A gamma-ray logging sonde with a shorter crystal would improve thin bed resolution; however, this shorter, lighter tool is less rugged and more likely to damage or failure during outcrop logging.

Results at logging site #3 are mixed due to problems of thin-bed resolution by the logging tool and poor photographic coverage of the lower section of the outcrop.

This exercise demonstrates that (1) Sandstone/shale ratios are sensitive to how grid cells are defined in relationship to major lithologic units and (2) Outcrop derived ratios and log derived ratios are similar in simple geologic situations where thinly bedded intervals are minimal.

STOP 4

Old Park Hill Quarry

What is now a shopping center was once the old Park Hill rock quarry (Fig.96). The Upper Jackfork Sandstone outcrops in this quarry are typical sandy turbidites. The sandstones are interpreted to be channel fill and levee deposits in a submarine canyon or more likely in the upper part of a submarine fan complex.

A slump and debris flow deposit is present at the base of the sequence. Of particular significance is the apparent presence of laterally accreting cross bed sets (Fig. 97) indicative of traction current bedding. This section is stratigraphically slightly above the Big Rock section, but appears similar to beds in the upper portion of that quarry's face, near the center of the quarry. The underlying shales and siltstones are significantly bioturbated at a nearby locality.


STOP 4
STOP 5
STOP 3
RIVER
Big Rock Mountain
Golf Course
Fort Roots VA Hospital
Edgewood Cem
Junior High Sch
Shilcotte
Golf Course
Pulaski
22
21
28
26
300
350
250

Figure 96


CKFORK FAN
upper fan traction current cross bed
No. Little Rock, A

Figure 97

(Description from Stone and Haley, 1986)

STOP 5

'Wildflysch" Slope Facies in the Middle Jackfork Formation Near Levy, Arkansas

The rocks exposed in this roadcut consist of black and reddish-brown shale with thin bedded layers to a few angular to rounded olistoliths of light-gray quartzose sandstone, very similar lithologically to the typical sandstone of the Jackfork. When comparable olistoliths are partially or completely weathered from the shale and are exposed on the surface, they resemble the "knockers" described by Alpine geologists. Fossil-bearing sandstone olistoliths have not been found at this outcrop, but are present in nearby outcrops. This locality is in the Aly belt of Haley and Stone, 1981. Thrust faults are present and are marked by slickensides and quartz veins. Some of the quartz veins contain minor amounts of lead, zinc, copper, and silver minerals.

Chaotic and olistolithic zones of the type observed in this outcrop, with their characteristic rubbly bedding, are common in the middle part of the Jackfork Sandstone in Pulaski County and are present, but less abundantly, for a distance of about 130 miles to the west. Similar zones are also present in the lower part of the Atoka Formation, are very common in the Johns Valley Shale, and common in the Stanley Shale. Tectonic deformational indicators such as the faults, slickensides, incrustations of dickite, quartz veins, and cleavage present here are not unique to this outcrop, but are present in many of the other olistolithic zones.

Invertebrate fossils have been collected from sandstone olistoliths at two localities in the middle Jackfork west of this stop. Fossil-bearing sandstone olistoliths appear to have been limy and are thought to have slid southward from an area of shallow-water deposition. The following ammonoid cephalopods of early Morrowan (Hale) age have been identified: *Syngastrioceras globusum* (Easton), *Bisatoceras (Schartymites) paynei* (Gordon), *Reticuloceras tiro* (Gordon) *Retites semiretia* (McCaleb), *Cymoceras adonis* (Gordon), and *Stenopronorites quinni* (Gordon, 1968 and Gordon and Stone, 1977).

Viele (1973, 1979) named the belt in the vicinity of this stop the "Maumelle chaotic zone". He suggested that it formed as part of a subduction complex upon an overriding plate to the south when the plate was widely separated from the shelf-slope zone of North America.

Voicing another opinion, Morris (1985, p. 29) noted that from all appearances the rubble bedding in this area is

indistinguishable from that in the Jackfork in other parts of the frontal Ouachita Mountains. He further offered the following in rebuttal: "1) The rocks of the "Maumelle Zone" are an integral part of the Jackfork Group. 2) The Jackfork had a clear connection with the shelf/slope area of North America, having received from there the limestone, chert, and non-feldspathic sandstone clasts in rubbly flows and the rounded, mature mineral suite for the sands forming the slope fans. 3) Point 2 may be repeated for the underlying Stanley as well as the overlying Johns Valley. 4) The rubbly zones, many in number, can be mapped, although little of this has been done."

It has been our opinion for quite some time that these rocks in the lower Jackfork Formation represent a slope facies and that they have many examples of successive submarine slumps and resultant chaotic - appearing deposits. That subsequently a high tectonic imprint and low-grade metamorphism was imparted to them during late Pennsylvanian and early Permian Ouachita Mountain orogenic events.

(Description from Arnold Bouma, personal communication)

STOP 6

Jeffrey Quarry

This is an abandoned quarry with most of the section under water. The entire section consists of rather massive sandstone, comparable to Big Rock Quarry, overlain by a more shaly section with a number of small channel fills in it. Only the small channels are visible across the artificial lake.

Two different types of fill can be observed, one type is very simple (Fig. 98), the other shows an irregular upward thinning of layers and fining of the grain size with the sides of the top of the erosional cut barely visible. The top fill of the channel does not differ from the overlaying shales.

It is important to observe both types of fill because those architectures will never be visible on seismic reflection, even in channels with a size that is visible on seismic records. The nature of the fill, the terminations of the channels (thinning and downlapping), and the connectivity blockages can well be observed.

Detailed correlation with Big Rock Quarry is currently impossible. It is suggested that these small channels in Jeffrey Quarry either can be compared to the smaller, more shalier, large-scale inclined layers in Big Rock Quarry, and/or that they represent the abandonment stage.

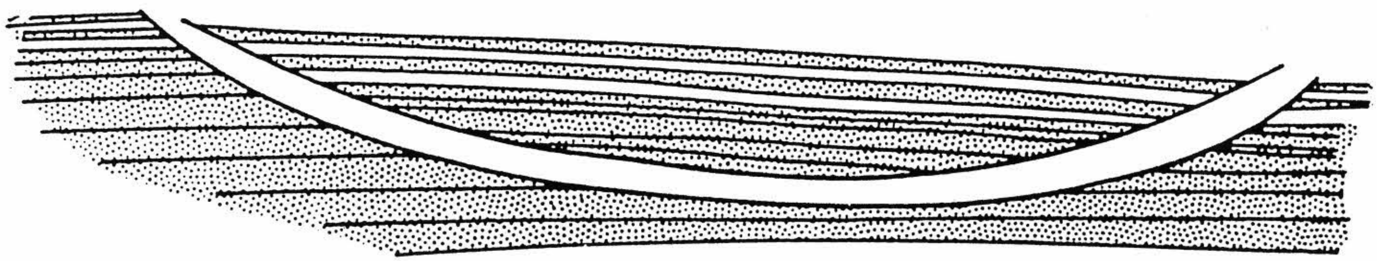


Figure **98.** Drawing from a photograph showing an undulating erosional sandstone surface, capped by shale, and followed by a channel fill restricted to one of the depressions. The channel fill shows a thinning-upward sequence. Abandoned Jeffrey Quarry near Haig, Arkansas, Jackfork Formation.

(Description from Stone,1968 etc. and Joslin, 1979)

STOP 7

Bayou Meto Anticline Roadcut

About 1,100 feet of the lower part of the Lower Atoka Formation is exposed along roadcuts on Arkansas Highway 5, on the south flank of the Bayou Meto anticline west of Cabot, Arkansas.

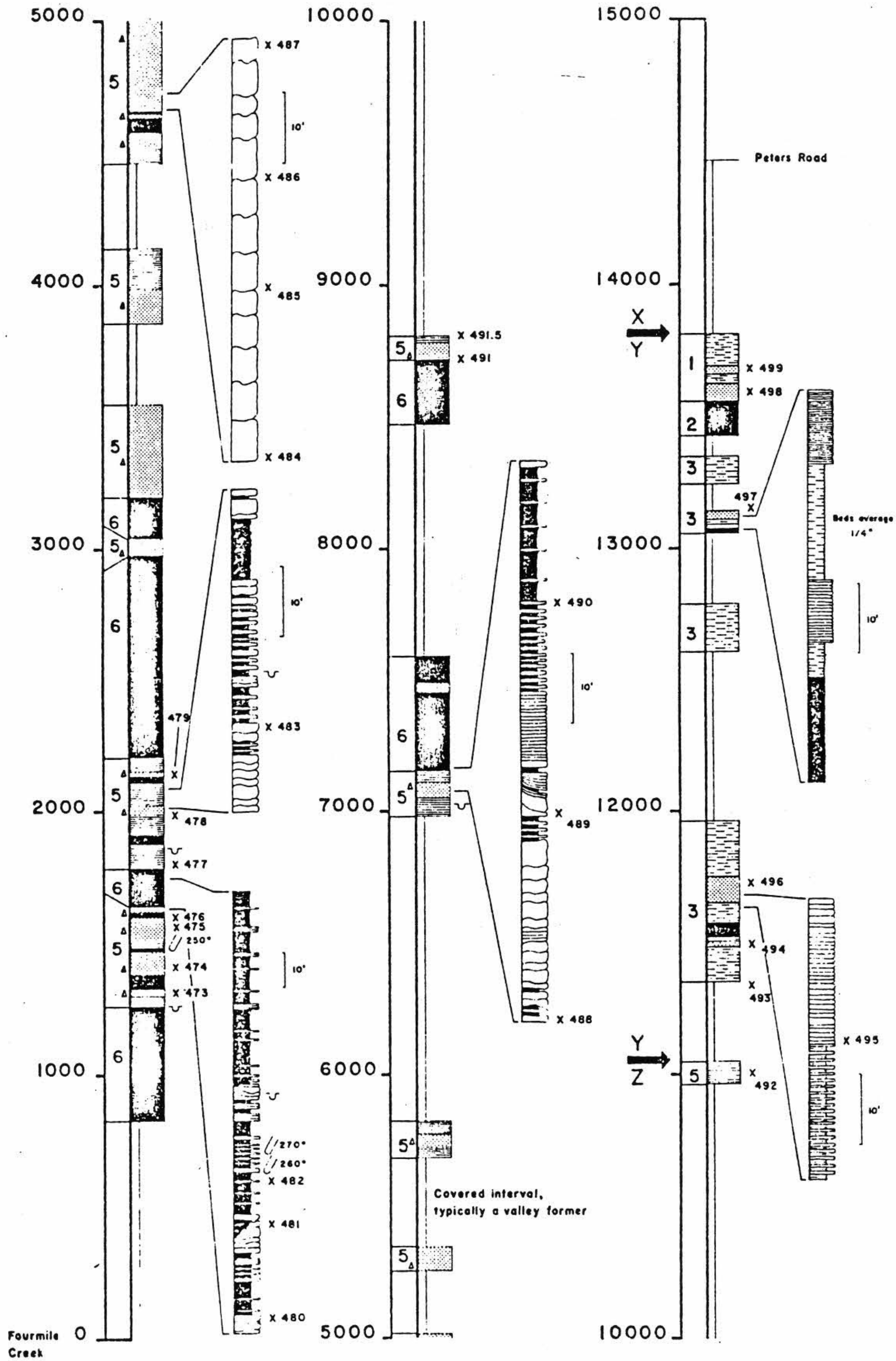
A general correlation from one side of the road to the other is not difficult. Details, however, have often changed significantly over that short distance which is close to the paleocurrent direction.

This section is composed of proximal-appearing, amalgamated turbidites, thin overbank or lobe intervals of sandy and shaly strata, and an occasional debris flow with ripup clasts. Some sideritic-clay ironstone concretions and layers occur in the hemipelagic and other shales. The apparent thick proximal submarine channels, near the base of the exposure, consist of medium to dark gray, micaceous, very fine- to medium-grained, friable to hard, medium- to thick-bedded sandstones belonging to the AB divisions of the Bouma sequence. Contacts are generally somewhat scoured and free of interbedded shales. The overbank or lobe intervals include subequal amounts of sandstone, siltstone, and shale. Sandstones are brownish-green to medium gray, micaceous, very fine- to medium-grained, and thin- to medium-bedded. Tool marks, and flute and load casts are abundant. The turbidite intervals ABCDE, BCD or CDE are most common in the graded sandstones. Paleocourrents on most of these beds are axial, down the "trough" towards the west, although some groove casts and other markings have a more southwesterly flow. "Trashy" blue beds with coalified plant fragments, clay, silt, sand and mica occur at the top of many of these units and represent the liquified lag debris flows of turbidity currents. Most of the shales are typically black and fissile, but some are silty and sandy in part, possible more carbonaceous shales are probably hemipelagic. These rocks probably were fed by the large prograding deltas that existed both to the northeast and east-southeast during a maximum sea level lowering and basin downwarping in early Atoka time.

South of the intersection of highways 64 and 5 are a number of outcrops in deltaic deposits of the middle part of the Atoka Formation.

The highly thrust faulted axis of the Bayou Meto anticline forms the valley to the north, and a few exposure of Johns Valley Shale (late Morrowan) with minor olistoliths of sandstone occur. It is likely

that this unit represents a submarine canyon and slope facies, and that most of the clastic sediments bypassed this locality and were deposited in the (then) deeper part of the basin to the south. The Johns Valley Shale is renowned for its large quantities of platform exotics in the Ouachita Mountains of Arkansas and Oklahoma.



BAYOU METO SECTION

(➡ Lithic-Unit Boundary)

Figure 99

(Description by Stone, 1991)

STOP 8

Round Mountain Shale Pit

This stop is located in the upper Atoka Formation on the west end of Round Mountain, near the axis of the Conway syncline, about 2 miles west of El Paso, Arkansas.

The excavation exposes (Fig. 100) approximately 150 feet of black shales deposited in marine and back bay environments. Three thin coal beds, the thickest about 9", are found within the shale sequence. The thickest bed has been sporadically mined as a local energy resource. A few brackish water invertebrate fossils have been found, and thin intervals of iron carbonates and concretions occur in the interval overlying the thick coal bed near the middle of the sequence. A medium-grained, cross-bedded, fluvial sandstone occurs above the shales. The cross-beds indicate sediment transport form east to west.

A chemical analysis provided by Bostick of the U.S. Geological Survey and others of the main coal seam indicates that it is a high-rank bituminous coal.

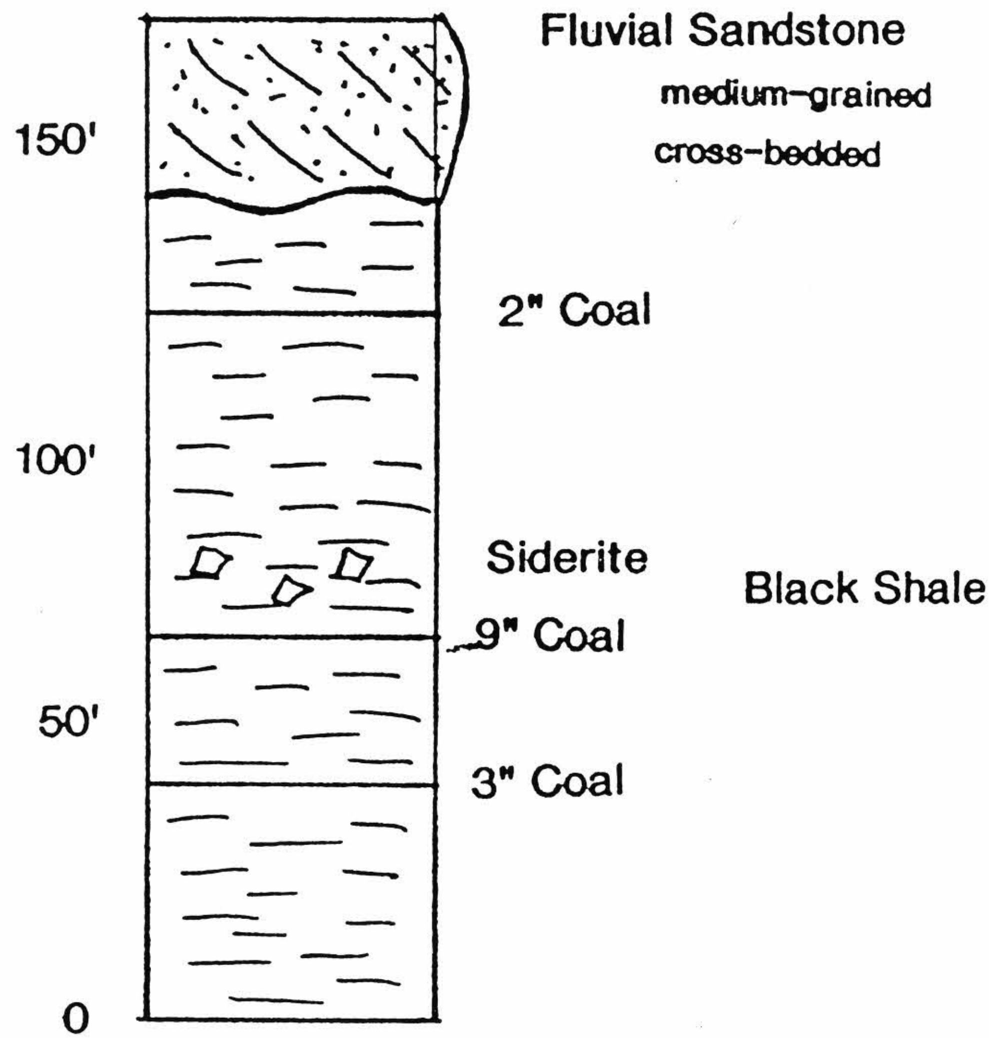


Figure 100 **Generalized Section**

Round Mountain Shale Pit

(Description from Rufus J. Leblanc, 1976)

STOP 9

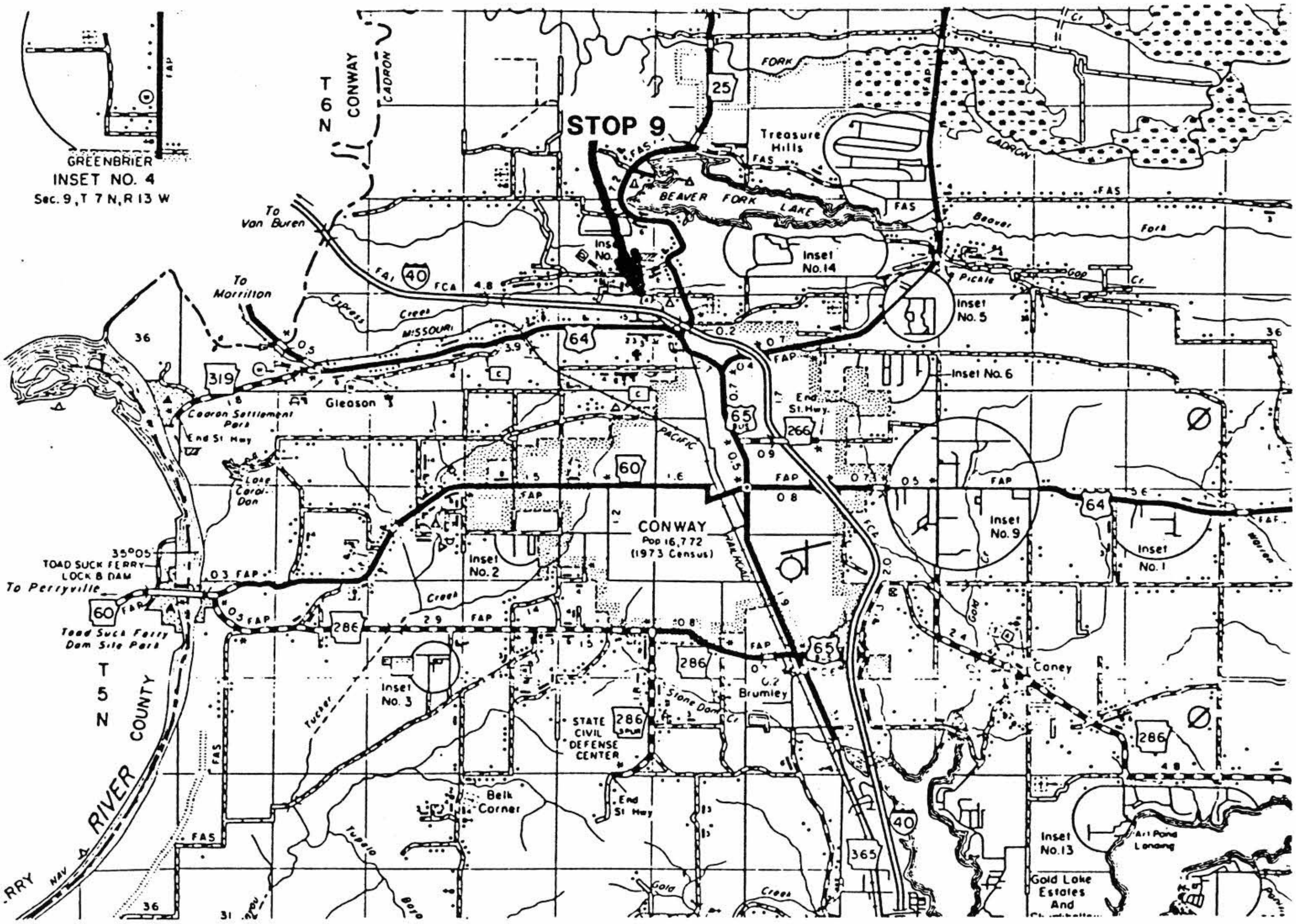
I-30 Roadcut at Conway

A thick (190') delta sequence in the Middle Atoka at Conway, Ark. (Fig. 101) Roadcut along North bound lane of Interstate 40 (Fig. 102). Up is to the left (South).

This sequence consists of thin bedded outer fringe sandstone and shale which grades upwards into inner fringe sandstone (Fig. 103). The upper part of this deltaic sequence consist of cross-bedded distributary channel sandstone.

Mapping by Stone and Haley indicates that this unit forms one of three or more clastic sequences ("traceable three") in the middle Atoka. These deltaic deposits extend westward some 100 miles (at least) to the vicinity of Heavener, Oklahoma.

Were there vigorous longshore currents that helped to transport these deltaic propagations?



STOP 9
GREENBRIER
INSET NO. 4
Sec. 9, T 7 N, R 13 W
T
6
N
CONWAY
CADRON
Treasure Hills
BEAVER FORK LAKE
Inset No. 14
Inset No. 5
Inset No. 6
To Van Buren
To Morrilton
MISSOURI
PACIFIC
Gleason
Cadron Settlement Park
End St Hwy
Lake Carol-Don
CONWAY
Pop 16,772
(1973 Census)
Inset No. 9
Inset No. 1
Inset No. 2
35°05'
TOAD SUCK FERRY LOCK & DAM
To Perryville
Toad Suck Ferry Dam Site Park
Inset No. 3
Brumley
Coney
STATE CIVIL DEFENSE CENTER
Belk Corner
End St Hwy
Inset No. 13
Gold Lake Estates And
T
5
N
COUNTY
RIVER

Figure 101

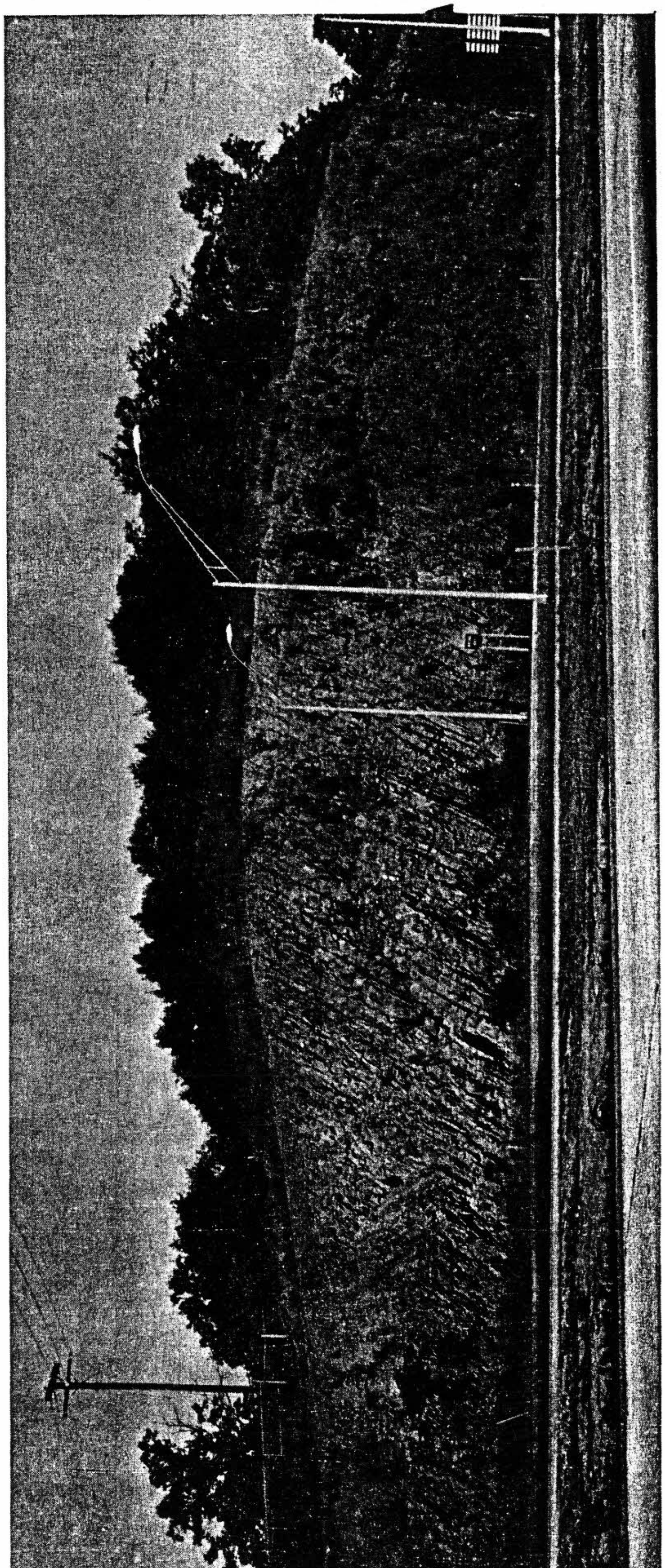
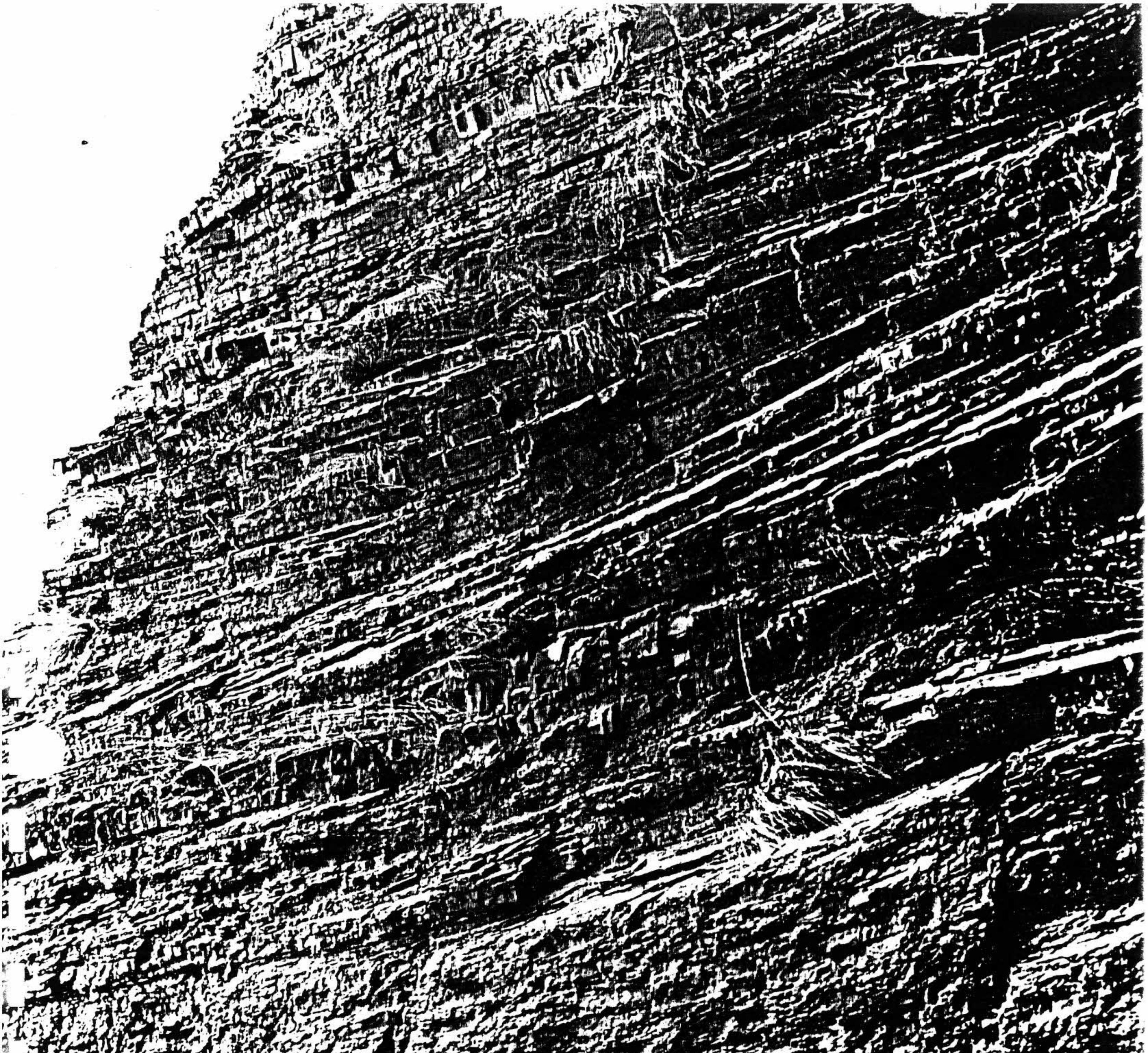


Figure 102



ARK. FIELD TRIP STOP 9 - ATOKA DELTAIC FACIES AT CONWAY, ARK. ON INTERATATE HIGHWAY 40. DELTA FRINGE FACIES OVERLAIN BY CROSS-BEDDED DISTRIBUTARY CHANNEL FACIES

Figure 103

UP

(Description from Stone & McFarland, 1981, Leblanc, 1976, Joslin, 1979 and others)

STOP 10

Upper Atoka Formation Deltaic Sequences on Arkansas Hwy. 9 Morrilton Bypass--Morrilton

Directly east of Morrilton along Arkansas Highway 9 are four road outcrops (Fig. 104 & 105) that show variations in deltaic deposits, especially with regard to the contact between pro-delta shales and the distributary sandstones.

One of the outcrops shows a good coarsening-upward sequence starting with pro-delta shales, followed by thin-bedded delta front deposits, and being topped by channel sands (Fig. 106). The other outcrops show that distributary channels can suddenly migrate laterally with the result that delta front deposits are lacking.

The convenient model of deltaic deposits with its coarsening-upward cycle is the ideal situation that occurs less frequent than is often realized.

A. UPPER ATOKA FORMATION -- MIDDLE INTERVAL

These rocks represent the middle unit of three separate deltaic sequences on the south flank of the Morrilton anticline.

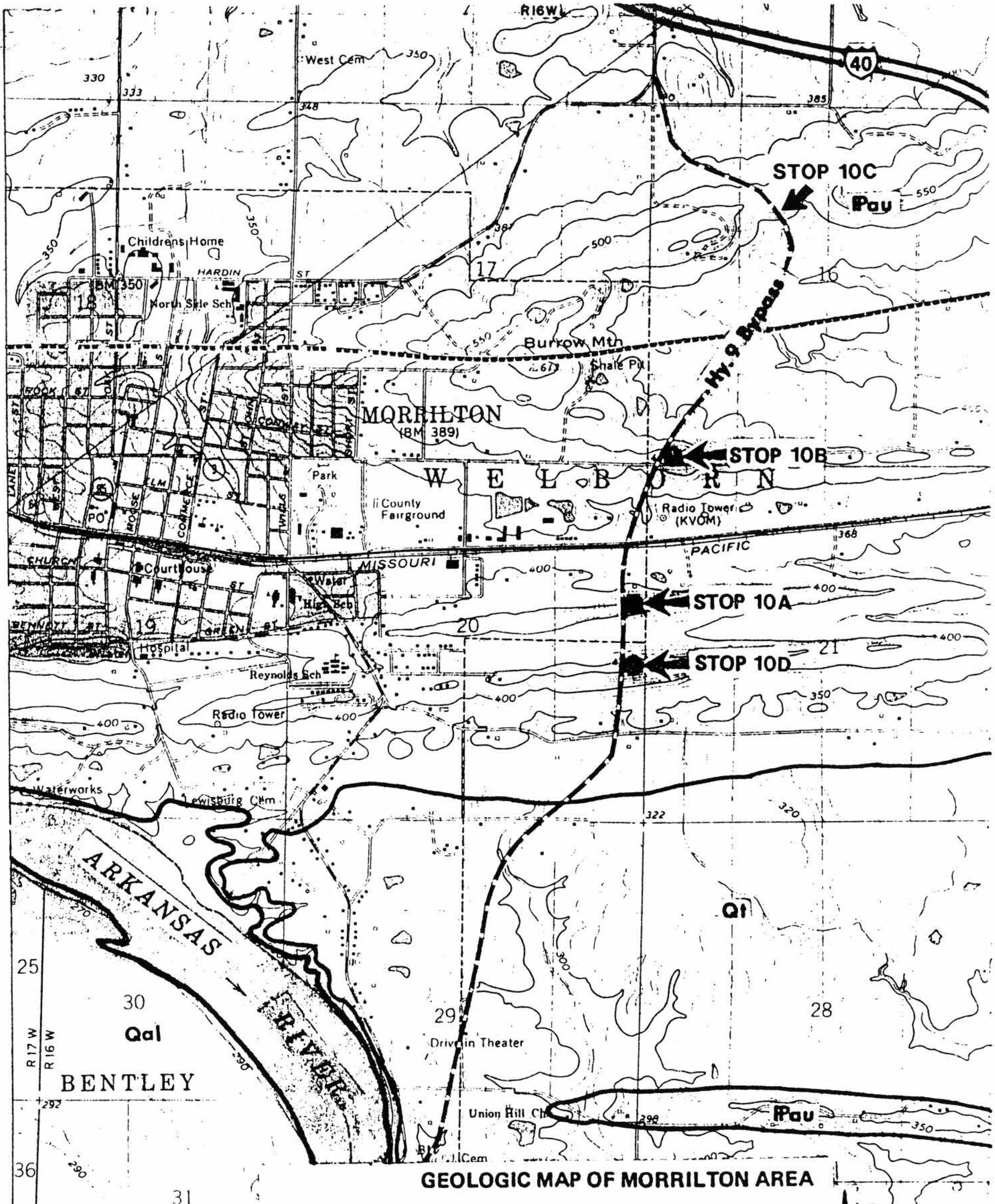
From north to south these rock units are: (1) marine black shale with a thin bed of fossiliferous limestone; (2) a pro-delta silty gray black shale; (3) outer fringe deposits of silty sandstone and shale; (4) inner fringe even-bedded sandstone; (5) a transgressive unit of fossiliferous sandstone; and, (6) pro-delta black silty marine shale.

The rocks in the outer delta fringe interval are extensively bioturbated (Fig.108)-note Conosticthus and other shallow water forms. A water expulsion feature is present on the east side of the roadcut in the sandstone of the inner fringe sequence.

B. UPPER ATOKA FORMATION -- LOWER INTERVAL

These clastic deposits are the lowermost deltaic facies exposed on the south flank of the Morrilton anticline.

The rocks exposed in this outcrop represent two major delta sequences (Fig. 109); one is quite "dirty" and the other is "clean". From north to south there is a very thick "dirty" delta sequence composed of silty gray-black shale, thin falser bedded, gray siltstone, and silty sandstone forming pro-delta



GEOLOGIC MAP OF MORRILTON AREA

– STOPS 10A,10B,10C and 10D

Figure 104

1000 0 1000 2000 3000 FEET

N

Qal	River Alluvium	Pau	Upper Atoka Formation
Qt	Terrace Deposits	·········	Thrust Faults
		———	Contacts

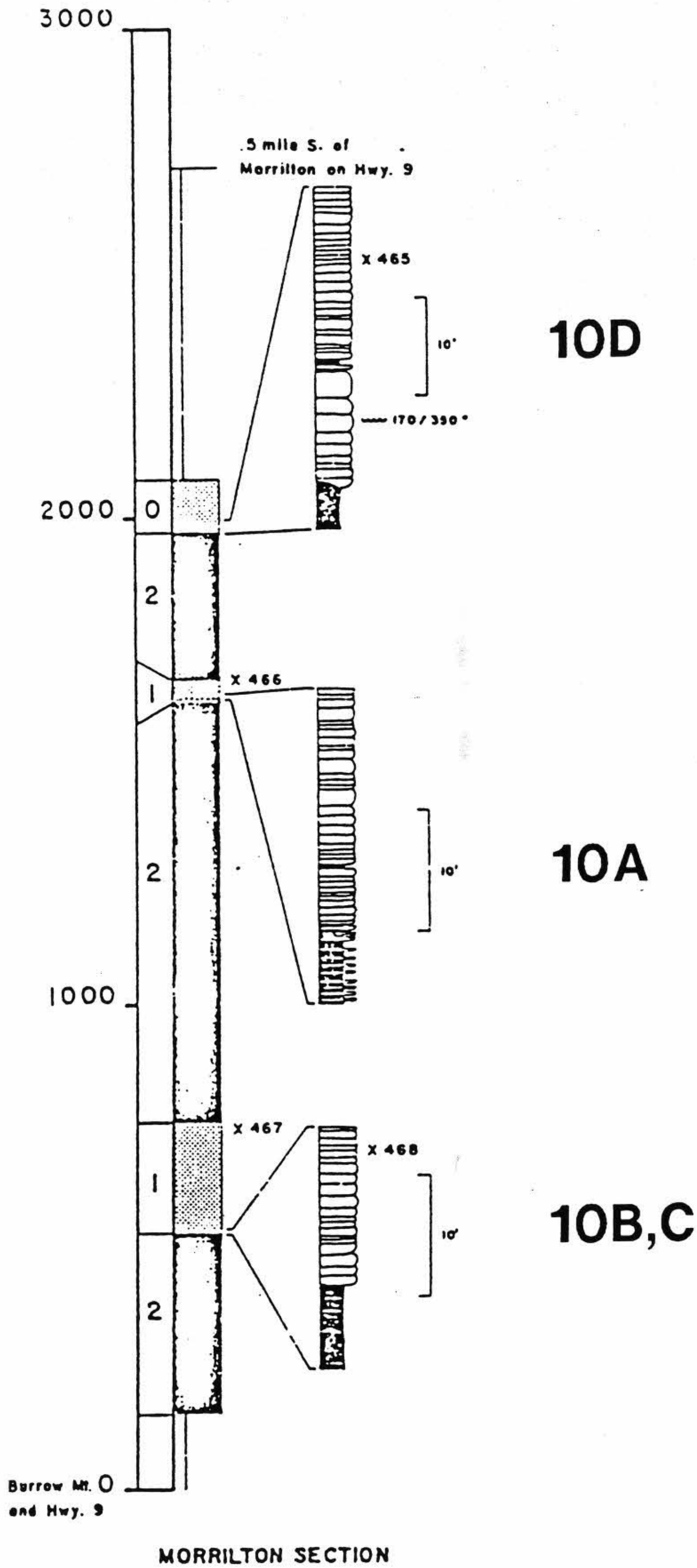


Figure 105

(from Joslin, 1980)

SP/GR Log Character

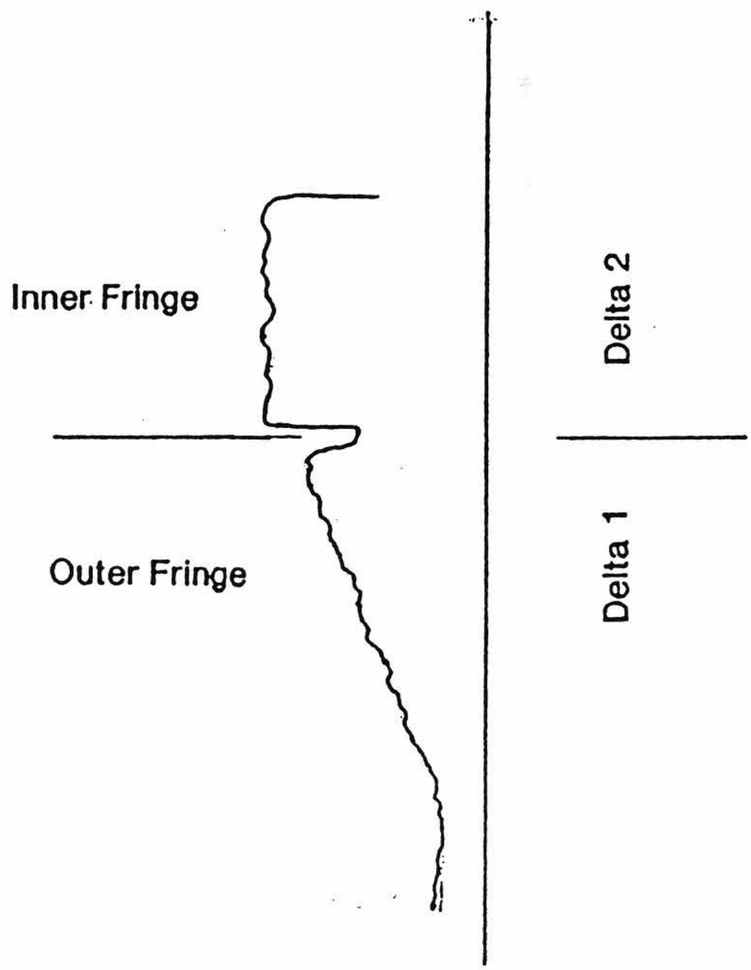


Figure 106

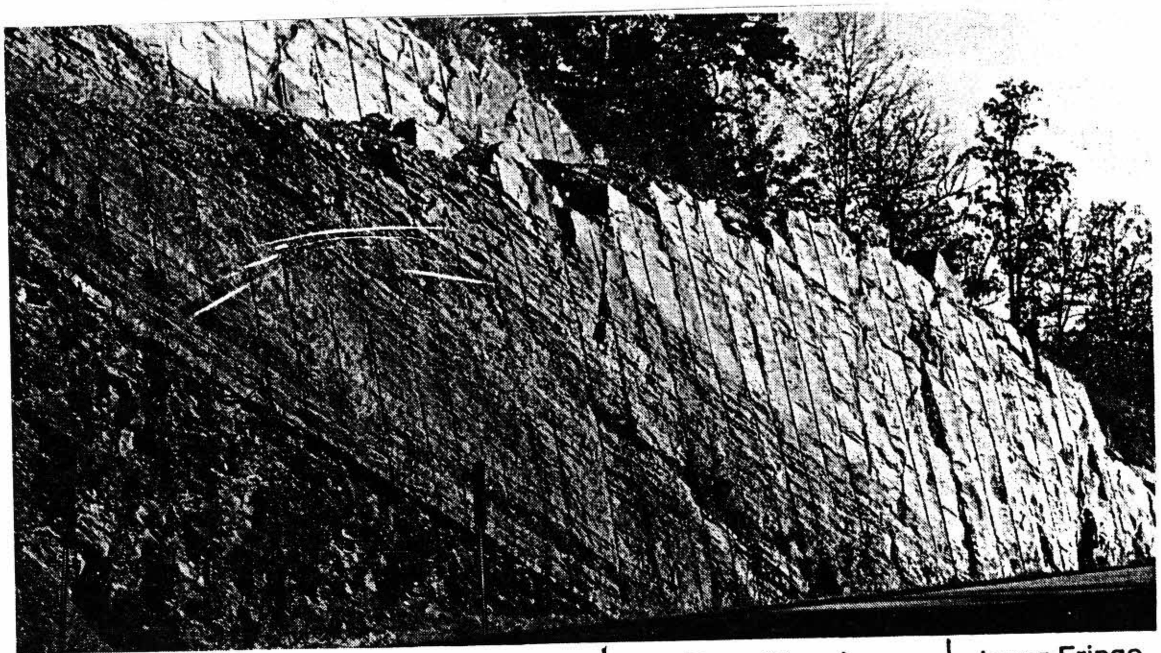


Figure 107. STOP 10A–A nearly complete delta sequence consisting of: (1) marine and pro-delta silty shale (below to left) overlain by, (2) thin bedded outer fringe sandstone, (3) massive even bedded inner fringe sandstone, and (4) a thin bed of transgressive marine sandstone. These rocks are in the upper Atoka Formation on the Morrilton Bypass.

Figure 108. STOP 10A – Bioturbated rocks (including Conostichus) in the outer fringe deposits of the upper Atoka Formation on the Morrilton Bypass.

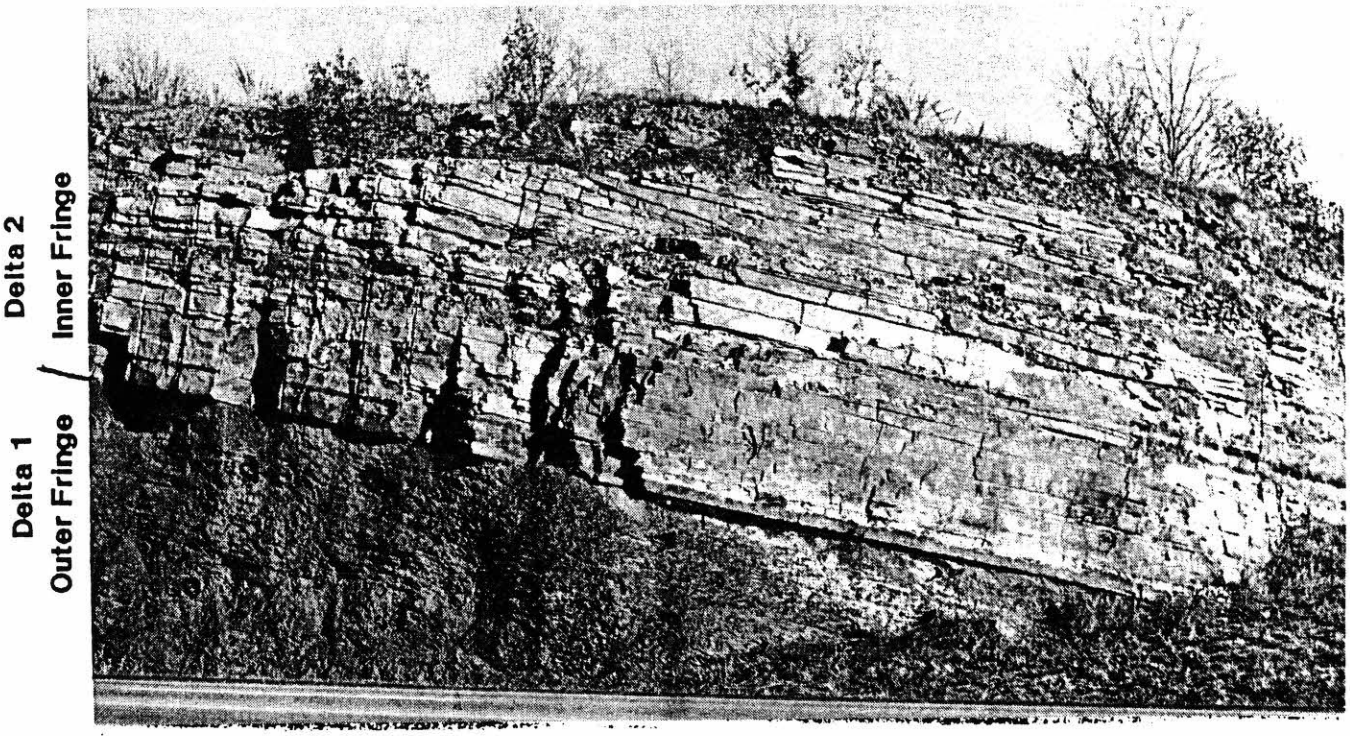


Figure 109.
STOP 10B – This sequence of rocks consists of: (1) pro-delta silty shale and very thin flaser-bedded sandstone (below) overlain by, (2) even bedded, inner fringe sandstone, and capped by (3) a cross-bedded distributary channel sandstone. These rocks are in the upper Atoka Formation on the Morrilton Bypass.

Figure 110.
STOP 10B - Distributary channel sandstone below overlain by shale (clay plug) deposited in an abandoned channel.

and outer fringe sequences. Overlying this sequence of rocks is a much cleaner delta interval composed of thin, even bedded, inner fringe sandstone, cross-bedded distributary channel sandstone capped locally by a shale lense (Fig. 110) (clay plug) and all overlain by more delta fringe sandstone. Some bioturbations are present in the 'dirty" delta outer fringe rocks. Also note the pyritic interval and its white oxidation product melanterite (iron sulphate) at the base of the inner fringe sandstone.

C. UPPER ATOKA FORMATION - LOWER INTERVAL NORTHLANK

MORRILTON ANTICLINE

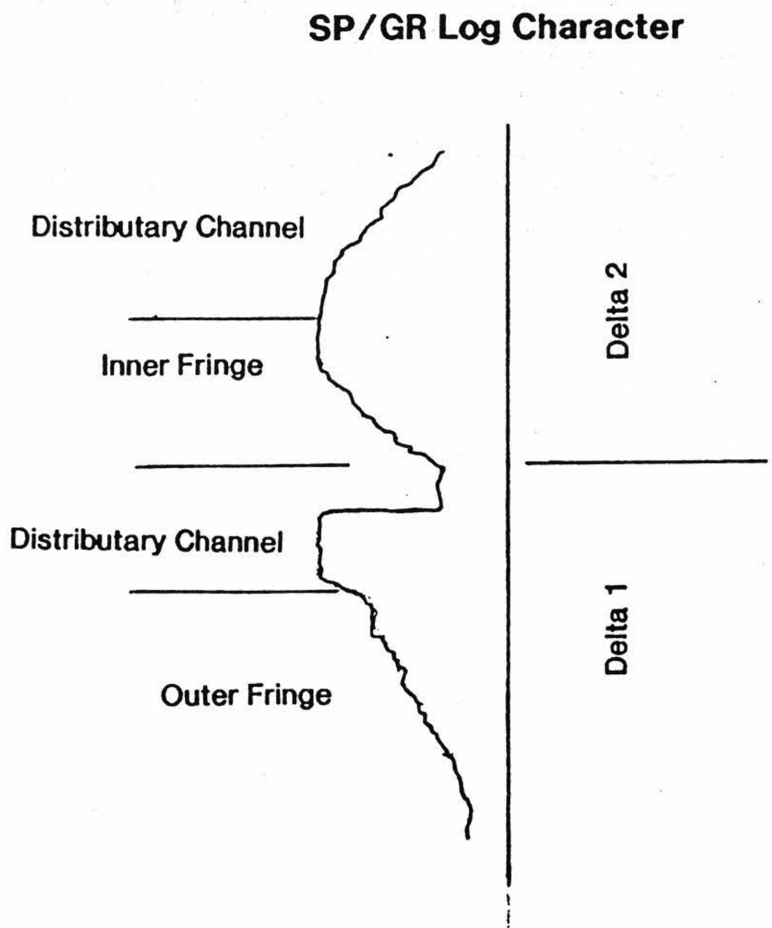
These rocks are on the north flank of the Morrilton anticline and are equivalent to the lowermost deltaic facies in the upper Atoka at Stop 10B.

These rocks are equivalent to the rocks of the upper Atoka at Stop 10B. From south to north the rocks are marine and pro-delta shale and siltstone; inner fringe sandstone; cross-bedded sandstone of a river mouth bar; even bedded inner fringe sandstone capped by distributary channel deposits at the top. Small thrust faults are present in the exposure.

D. UPPER ATOKA FORMATION -- TOP INTERVAL

The rocks exposed in this roadcut are the uppermost deltaic interval in the lower part of the upper Atoka Formation on the south flank of the Morrilton anticline. From north to south the sandstone and shale sequences represent parts of four deltaic deposi-tional cycles (Fig. 111 & 112) as follows: (1) pro-delta with overlying outer fringe strata; (2) distributary channel with a transgressive marine unit at the top; (3) inner fringe overlain by a river mouth bar interval (Fig. 111); and, (4) inner fringe capped by a distributary channel.

Falser bedding and small ripple marks are common in the rocks of the outer fringe. Thin, even bedded rocks are present in inner fringe deposits. Load features, scour channels, shale and sandstone clasts, and cross bedding mostly in a single direction are present in the rocks of the distributary channel sequences. Cross bedding oriented in many directions with some load and slump features typifies the rocks in the river mouth bar deposits. Small shale clasts, fragments of coalified plants, and invertebrate fossils are present in the rocks of the transgressive marine unit. Shallow-water trace fossils are especially numerous in the rocks of the outer fringe interval. These delta sequences had a probable source to the northeast as indicated by the crossbeds and ripple marks.



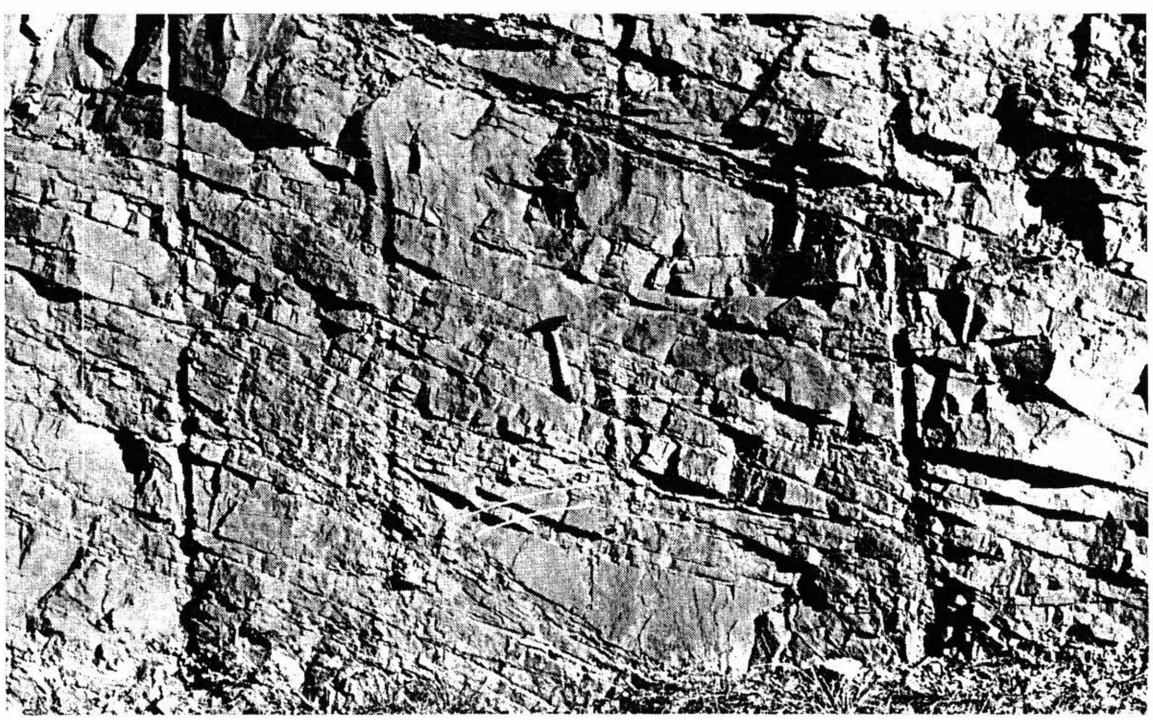


Figure 111. STOP 10D - Probable river mouth bar sandstone in the upper Atoka Fm. on the Morrilton Bypass

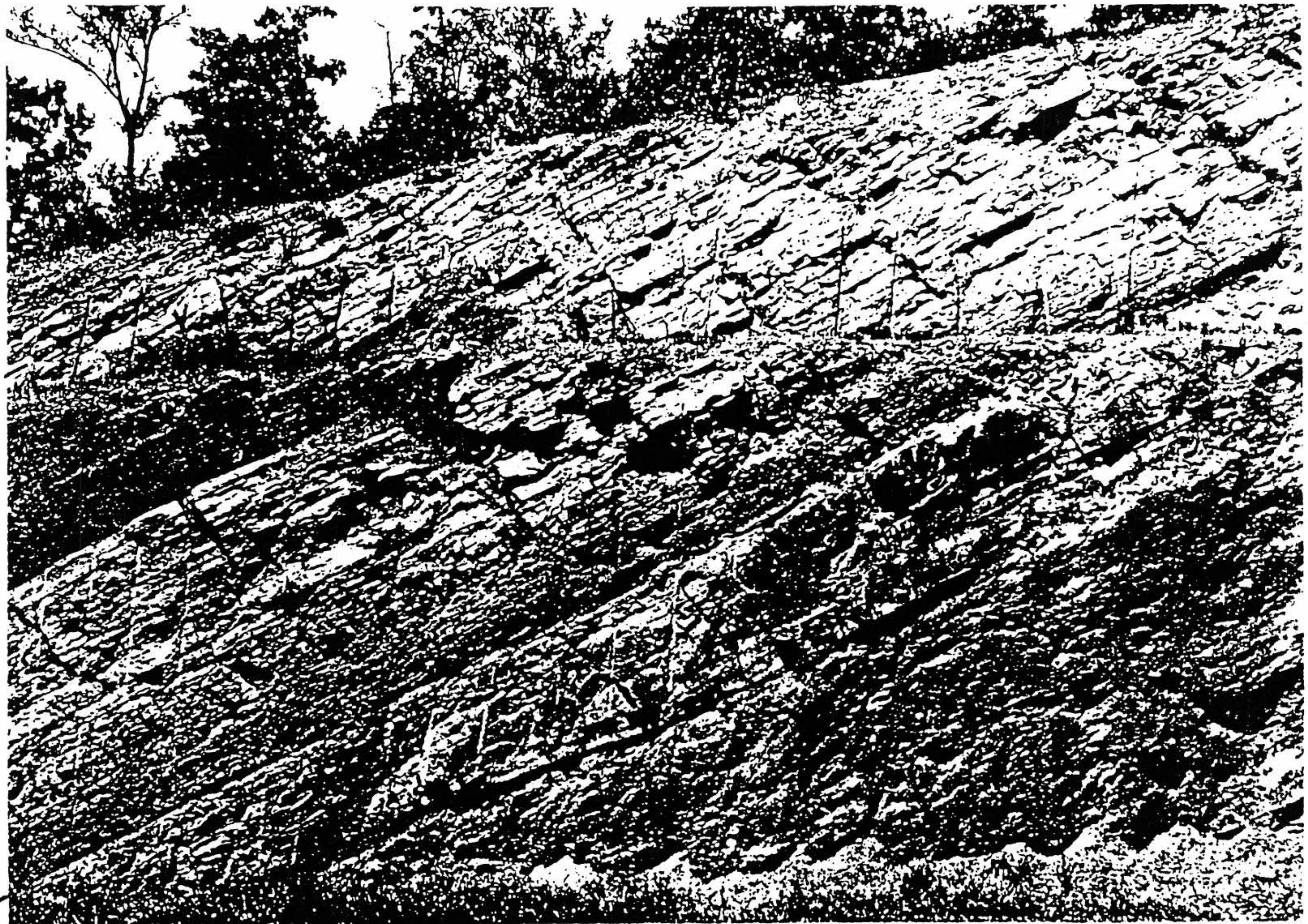


Figure 112. STOP 10D - Exposure of portions of four deltaic sequences in the uppermost interval of the upper Atoka, south flank of the Morrilton Anticline.

(Description from Stone & Haley, 1981
Joslin, 1979, Coleman and
Van Swearingen, 1991, and others)

STOP 11

Lower Atoka Formation at Tom's Mountain on Arkansas Hwy. 10 Near Perryville, Arkansas

Several road outcrops show lower Atokan submarine fan deposits. Stratigraphically the variety is tremendous with sometimes no lateral variability and other times significant variability visible in the short, tilted, layers. Interpretation of the sandstones and shales with regard to their depositional location within a submarine fan system is very dangerous unless one adheres blindly to some of the submarine fan or turbidite system models.

These outcrops are an ideal introduction to submarine fans because the outcrops reveal a lot of variability that generates lots of questions and discussions. It has to be realized that such incomplete information is very common and that the explorer, and certainly the production geologists and engineers, have to make most of that poor data source. it is, therefore, advisable to have as many good interpretation scenarios on hand from which to select the most applicable before finalizing the interpretation and guidelines to optimum exploration/production of the series.

This series of roadcuts in lower Atoka strata is exposed just south of the Ross Creek thrust fault (Fig. 113).

Displacement on the Ross Creek fault is approximately 15,000 ft. (4,600 m). Lower Atoka deep water sedimentary rocks are thrust over upper Atoka shallow water sedimentary rocks (Fig. 114) along the railroad tracks at Perry, 1.1 mi. (1.7 km) north of here.

Begin walking north along Arkansas Hwy. 10 to the bottom of the hill. Be careful - considerable traffic on this road! These exposed rocks exemplify turbidite deposition with monotonous repetitions of: southward dipping, thick to thin bedded, graded, bottom marked, and convolute bedded subgraywacke; very micaceous siltstone containing coalified plant fragments (blue beds); and, black fissile shale with some sideritic iron concretions. Deep water trace fossils are numerous in this sequence. Some minor dickite coated slickensides and minute quartz veins are present.

The Perryville igneous breccia (lamprophyre - carbonate intrusive) of probable early Upper Cretaceous age is exposed to the south of this area near the house on the east side of the highway.

Joslin (1979) found the section consisted of proximal turbidites, sandy flysch, shaley flysch, and debris flows.

Proximal turbidites consist of light to gray, very fine- to medium-grained, massive, thick- to very thick-bedded sandstones. Contacts are generally scoured and thin interbedded shales are extremely rare. Sandstones represent the AB divisions of the Bouma sequence exclusively.

Sandy flysch is a succession of unequal amount of medium- to thick-bedded sandstones and black shales. Sandstones are light to dark gray, very fine- to medium-grained, and commonly grade into light brown, thinly laminated siltstone and fissile shale. The bottoms of sandstones are usually sharp and contain numerous sole marks. The graded sandstones commonly have ABCDE, BCD, or CDE divisions of the Bouma sequence.

Shaley flysch interval are dominated by black fissile shales containing minor thinly bedded, convoluted, very fine-grained sandstone and siltstone. Bases of graded sandstones have small-scale tool marks and flute casts. Nearly complete Bouma sequences are common, including ABCD, BCD, and CDE divisions.

Debris flows comprise a meaningful portion of this section. They consist of a black to dark brown, blocky, chunky, mud matrix containing angular to well rounded pebble, cobble, and boulder clasts or sandstone and shale.

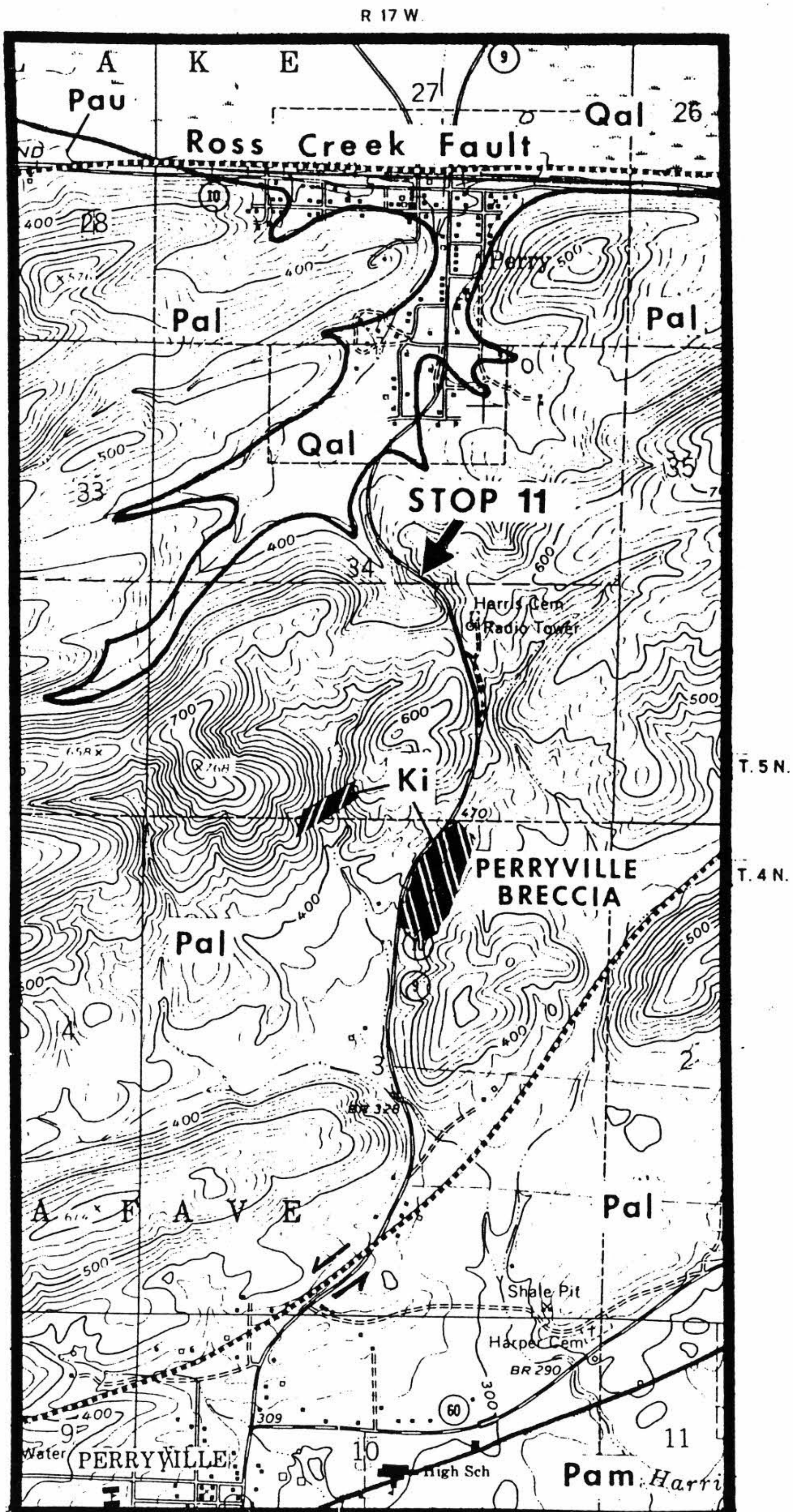
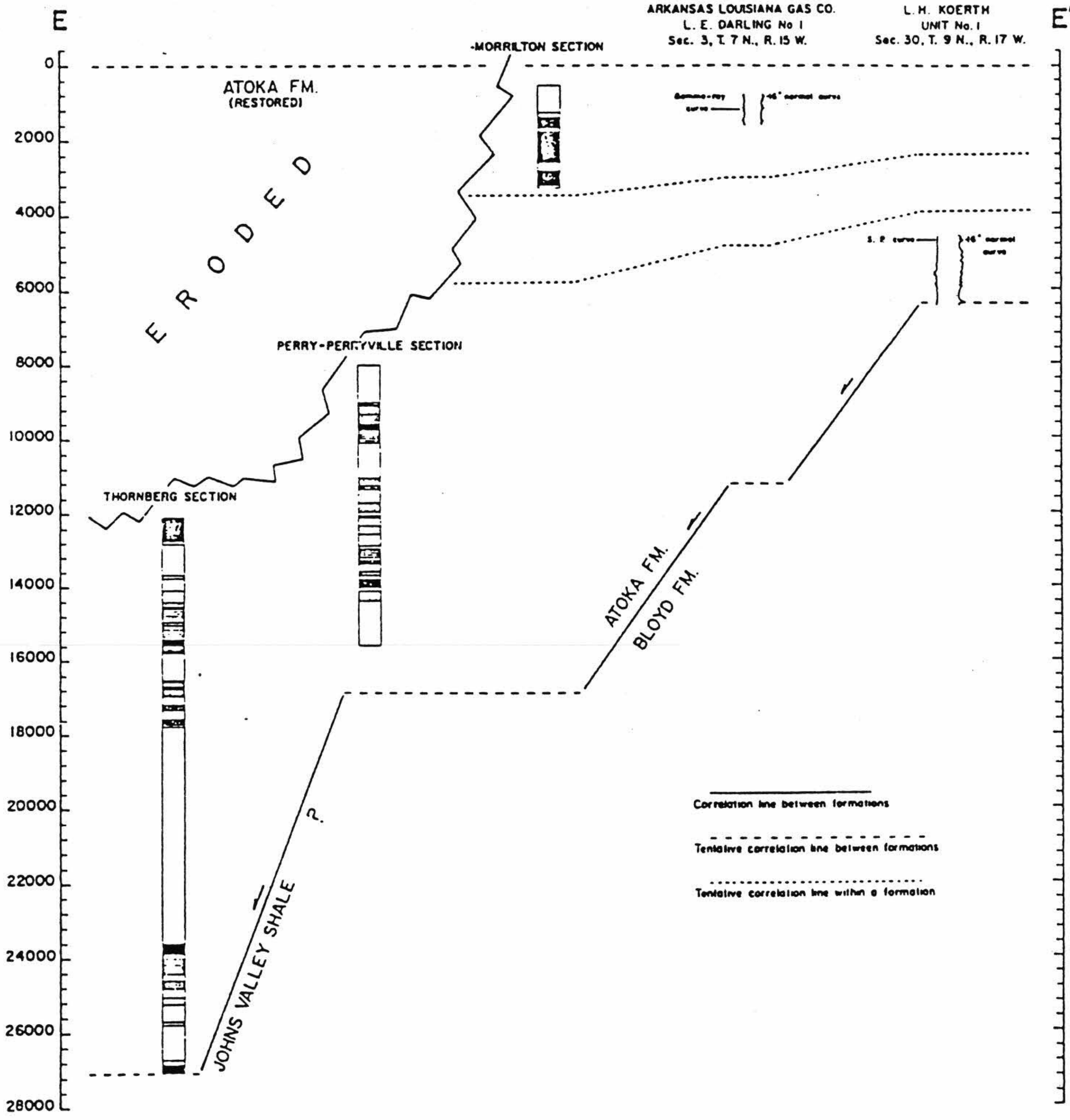


Figure 113 GEOLOGIC MAP OF PERRY-PERRYVILLE, ARKANSAS

1000 0 1000 2000 3000 FEET



E
E'
ARKANSAS LOUISIANA GAS CO.
L. E. DARLING No I
Sec. 3, T. 7 N., R. 15 W.
L. H. KOERTH
UNIT No. I
Sec. 30, T. 9 N., R. 17 W.
MORRILTON SECTION
ATOKA FM.
(RESTORED)
ERODED
PERRY-PERRYVILLE SECTION
THORNBERG SECTION
ATOKA FM.
BLOYD FM.
JOHNS VALLEY SHALE
?
0
2000
4000
6000
8000
10000
12000
14000
16000
18000
20000
22000
24000
26000
28000
Correlation line between formations
Tentative correlation line between formations
Tentative correlation line within a formation

Figure 114

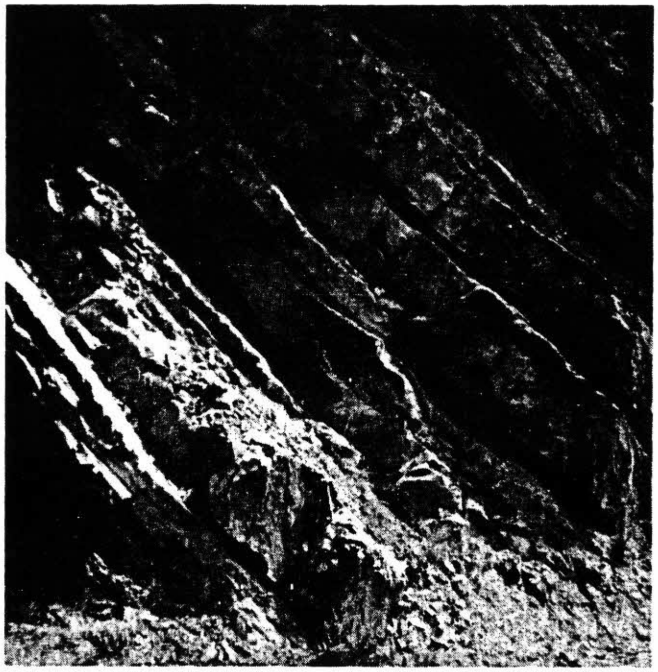


Figure 115.

Close-up of lower Atoka sandstone, siltstone and shale, with examples of convolute bedding and localized soft sediment slump and sliding **and/or scouring**

Figure 116.

Alternating sandstone and shale of the lower Atoka, an example of flysch.

Figure 117. Interbedded sandstone and shale mostly in thinning (of beds) and fining (of grains) upward sequences that may represent midfan submarine channel deposits in the lower Atoka Fm. north of Perryville on Arkansas Hwy. 10.

(Description from Stone & Haley, 1981 etc., Coleman & Van Swearingen, 1991, and others)

STOP 12

Upper Jackfork Formation at Rock Shelter on Arkansas Hwy. 9

Interbedded shale sandstone of the upper part of the Jackfork Sandstone are exposed on the west side of the road (Fig. 118 & 119). This sequence of rocks was possibly deposited in a submarine fan channel. Bottom marks, graded bedding, convolute bedding, and soft-sediment slump features are present. Several high-angle faults dip to the north and the apparent movement along these was towards the south. Slickensides coated with dickite are common. Steep northward dipping cleavage is well-developed in the shale. Three generations of quartz veins were emplaced in the following order: (1) veins that are folded and sheared and contain pyrophyllite; (2) veins of milky quartz with dickite; and, (3) veins of milky quartz with rectorite, cookeite, and pyrophyllite. As you can see, the section is dominated by shale with very thin sandstone beds, which thicken and thin across the outcrop. Even in the sand-rich interval the sands thicken and thin. To illustrate this, stand in the center of the sand package and turn around and find the high, sandstone-supported ridge across the road, in the trees. What is the apparent change in thickness? The sandstones commonly have scoured bases and are lenticular or pod-like in nature. This section, in and of itself, is not easy to fit into its niche on the Walker model. However, it does fit part of a pattern that suggests a slope environment of deposition.

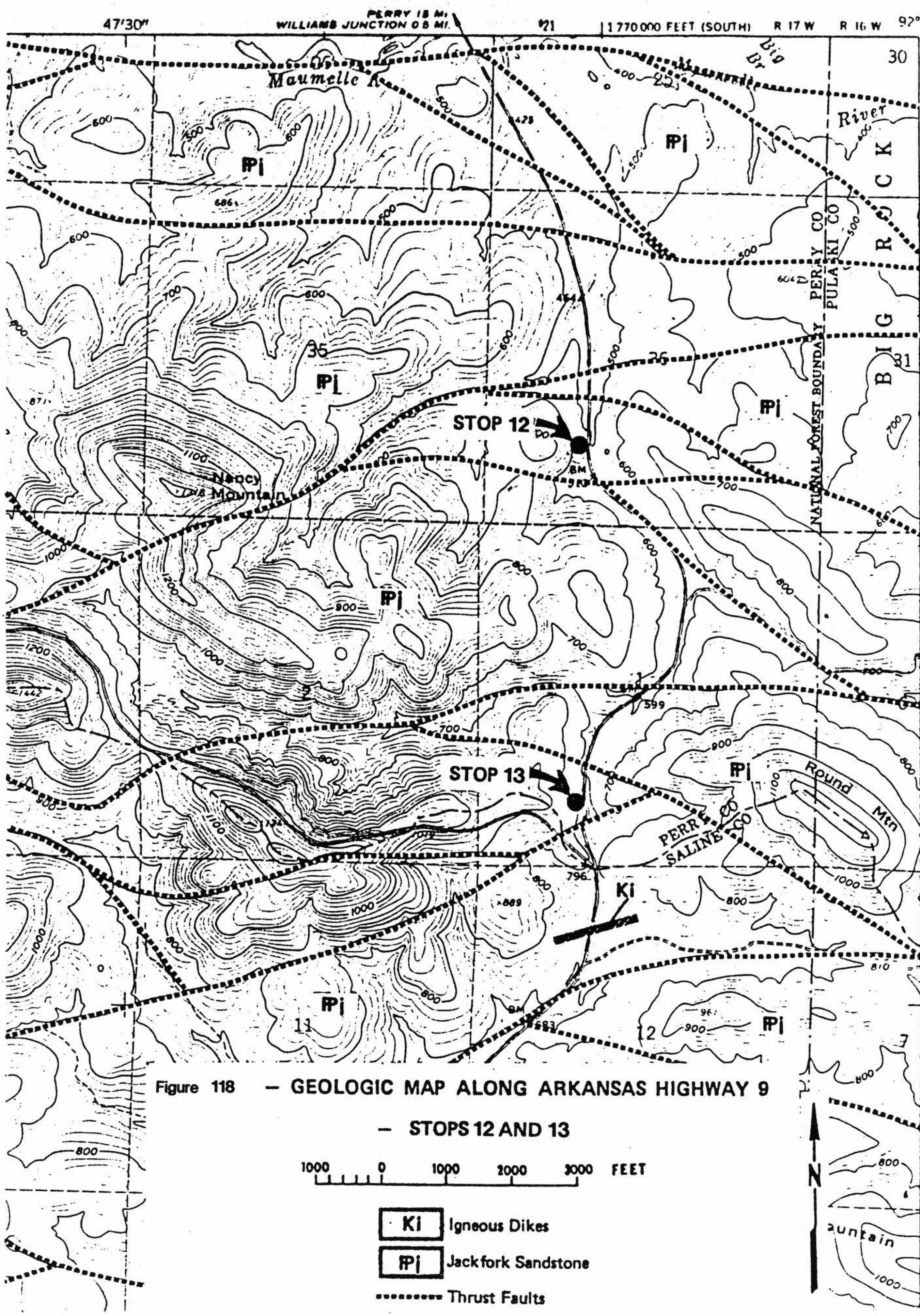


Figure 118 — GEOLOGIC MAP ALONG ARKANSAS HIGHWAY 9 — STOPS 12 AND 13

Figure 119. The interbedded sandstone and shale represents a submarine fan channel deposit in the upper Jackfork Sandstone on west side of roadcut on Arkansas Hwy. 9.

STOP 13

Lower Jackfork Formation on Arkansas Hwy. 9

Interbedded, vertically dipping sequence of light gray quartzose sandstone and reddish-black shale (bottom to north) of the lower Jackfork Sandstone are exposed at this stop. An intraformational conglomerate is evident at the base of a likely submarine fan channel sequence. A small southward inclined fold near the center of the outcrop has been related by some workers to tectonic deformation and there is a small thrust fault at the base of the fold with dickite coated slickensides. Other workers have related the fold to southward directed sedimentary slumping prior to the faulting (Fig. 120).

Quartz veins fill some fractures in the rock and contain needle quartz, rectorite, and cookeite. Quartz veins of this type are thought to have been emplaced during the latest part of the Ouachita Orogeny.

These Jackfork Stops are in the area that Viele has referred to as the "Maumelle Chaotic (Subduction) Zone", taking the name from the seemingly structural complexities of the rocks exposed (at Stops 12 & 13) along the Lake Maumelle trend. The structure and the lithology of these rocks has evoked much discussion, quite a bit of it most noisy. Does the chaos of the sandstone and shale in some of the Jackfork along this trend represent zones of submarine slumping and sliding from the north into the Ouachita trough or zones that slid from the south off the backs of northward-advancing nappes or zones of tectonic melange related to subduction faults?

Figure 120. A small inclined fold of soft-sediment slumping (?) or tectonic (?) origin in the middle Jackfork Sandstone on Arkansas Hwy. 9.

DESCRIPTIONS OF FIELD TRIP STOPS
JACKFORK FORMATION -- Day 2

The second day of this field trip will be spent in examining various upper to outer submarine fan sequences in the 6,000' thick Jackfork Formation in the southeastern Ouachita Mountains north and northwest of Arkadelphia, Arkansas. Many of the better Jackfork sequences are exposed at or near the scenic Lake DeGray (Dam site).

DeGray lake is located in the flooded valley of the Caddo River. DeGray Lake spreads out over 13,800 acres from the water gap at DeGray Dam. Construction on the dam began in 1963 and was completed in 1972 at a cost of $63,800,000. The dam contains about 7 million cubic yards of compacted earth fill. It measures 1,500 feet at the base, 50 feet at the top, and has a total length of 3,400 feet. It stands 243 feet above the river, with an elevation at the crest of 453 ft msl. The powerhouse, on the downstream (or south) side of the dam holds two turbine generators with a combined power output of 68,000 kw.

The lake and dam was originally approved in 1950 as a flood control project. Changes in 1961 and 1965 allowed water supply and recreation as authorized project purposes. And the dam was built. At optimum storage, the lake holds 427,200 acre-feet, with a maximum depth of 202 feet and an average depth of 49 feet. The shoreline is 225 miles long when filled to flood control level.

(Descriptions from Stone, Lumsden, & Stone, 1983; LeBlanc, 1978; Stone & Haley, 1981; and Bouma, 1993)

STOP 14

Friendship Roadcut

At Milepost 81 along I-30 south of Friendship and north of Arkadelphia, Arkansas (Fig.121), is an exposure of stacked submarine fan channels of the Upper Jackfork Formation (Fig. 122 & 123). Outcrops are on both sides of the highway. The best outcrops are on the west side of the road because trees and brush have been removed by the highway department. These outcrops are on the edge of the exposed Ouachita rocks near the coastal plain onlap, and in the southern thrust belt of the orogen.

Based on models (Mutti and Ricci Lucchi, 1972; Mutti and Normark, 1987) these fan channels are typical of middle to inner fan facies, but perhaps less proximal than the large channel complex at the North Little Rock quarry. Stone et al. (1983) recognized nine channels and a small lobe sequence at this outcrop. The channels range from 30 to 100 feet in thickness. Also present are debris flow deposits, load structures, ripple marks, plant fragments, and bottom marks.

The channels are characterized by fining and thinning-upward sandstone and shale bed sequences. The larger amount of shale in the thinning-upwards sequences are suggestive of a more distal position relative to the North Little Rock facies. The "blue beds", which are interpreted as debris flows, contain significant amounts of land-derived plant material. The channel bases, as is typical of most deep water turbidites, are only slightly scoured into underlying beds.

This exposure is in the southern outcrop belt of the Jackfork, in south-dipping imbricate thrust sheets. Although we are 50 miles "down-dip" from the North Little Rock channels, intervening complex structures make the paleogeographic relationship between the two areas obscure.

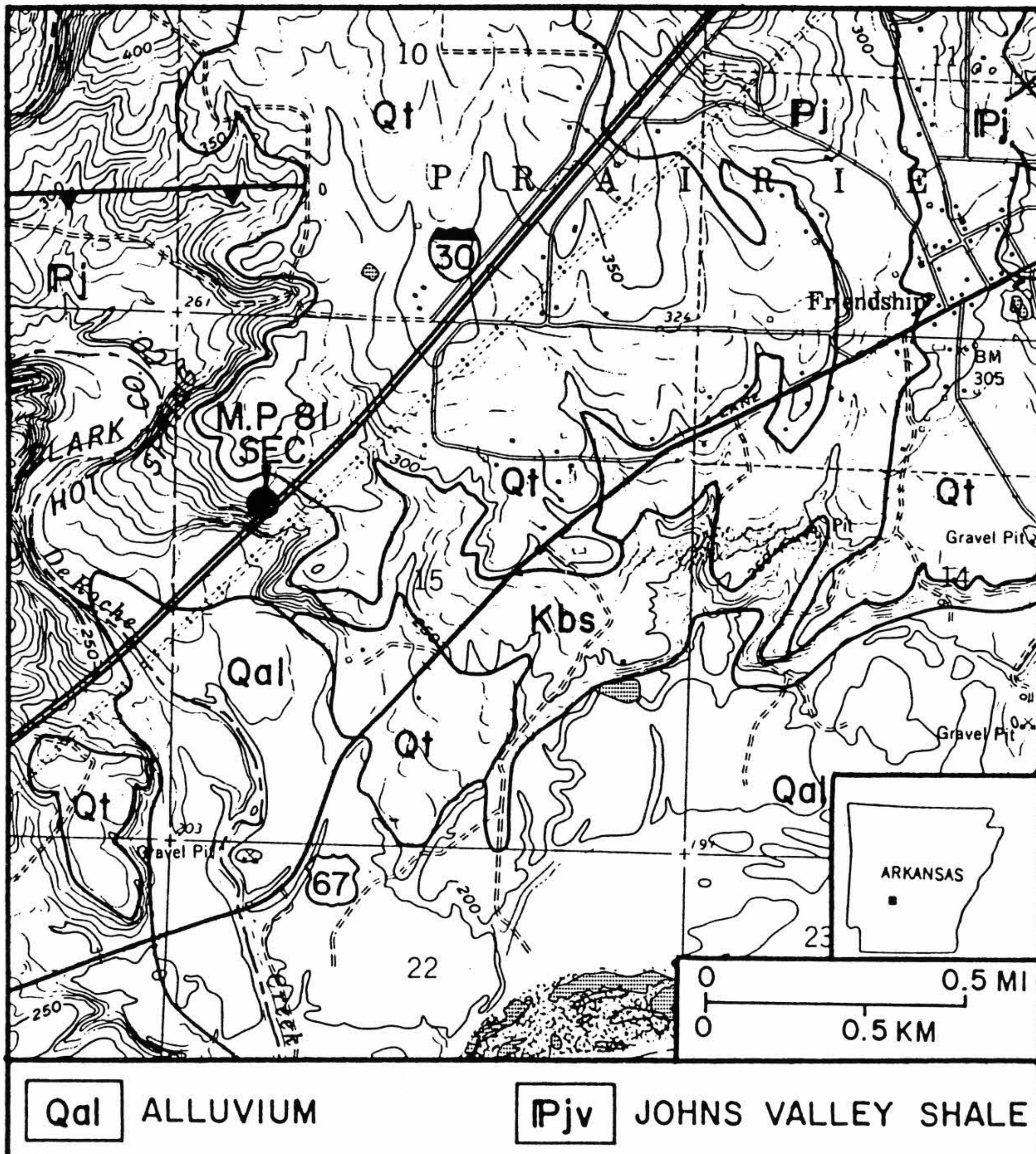


Figure 121. Map showing location of upper Jackfork sandstone outcrops at Mile Post 81, I-30, near Arkadelphia, Arkansas

FRIENDSHIP ROADCUT

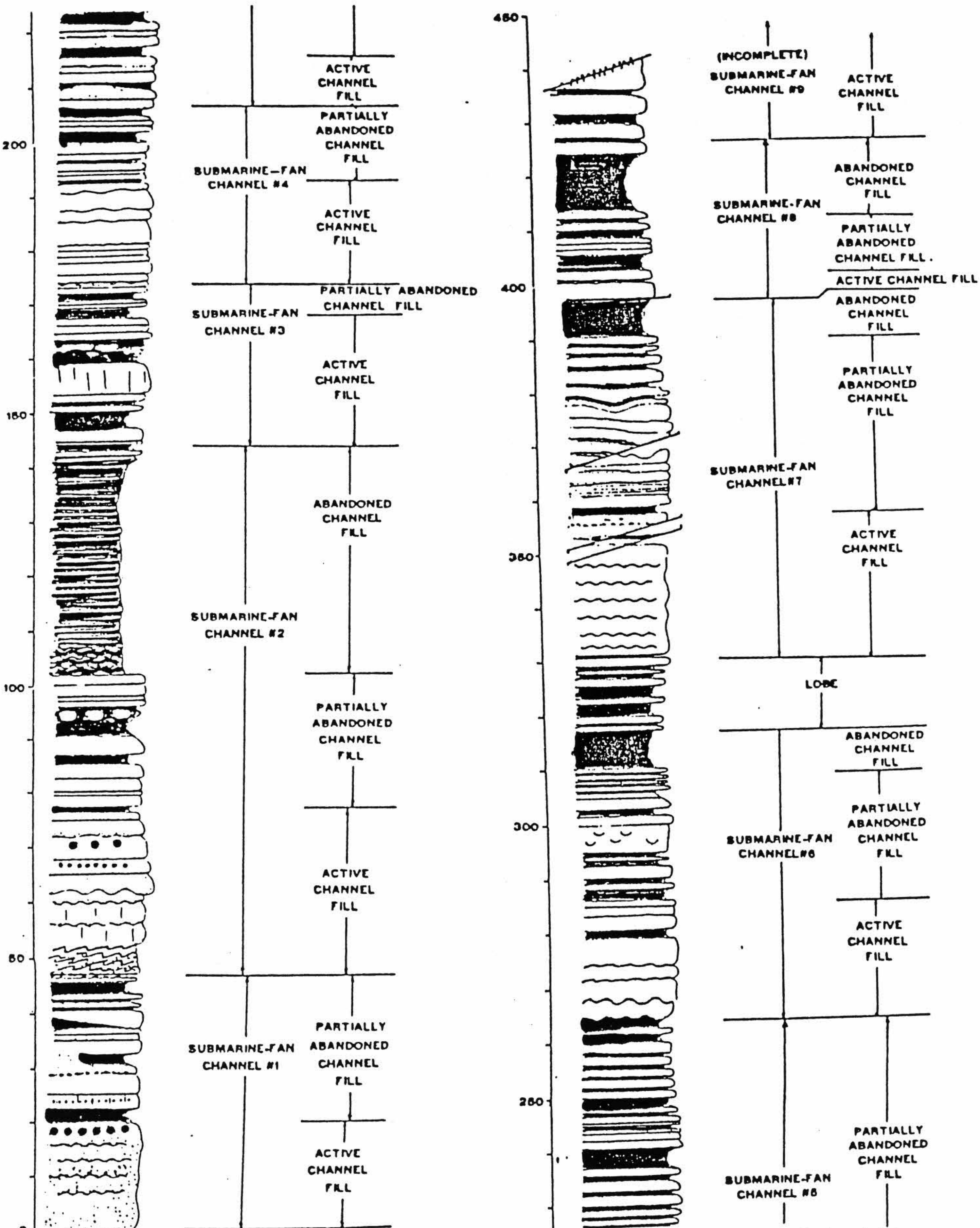


Figure 122. Measured Section of upper Jackfork Sandstone near Friendship, Arkansas.

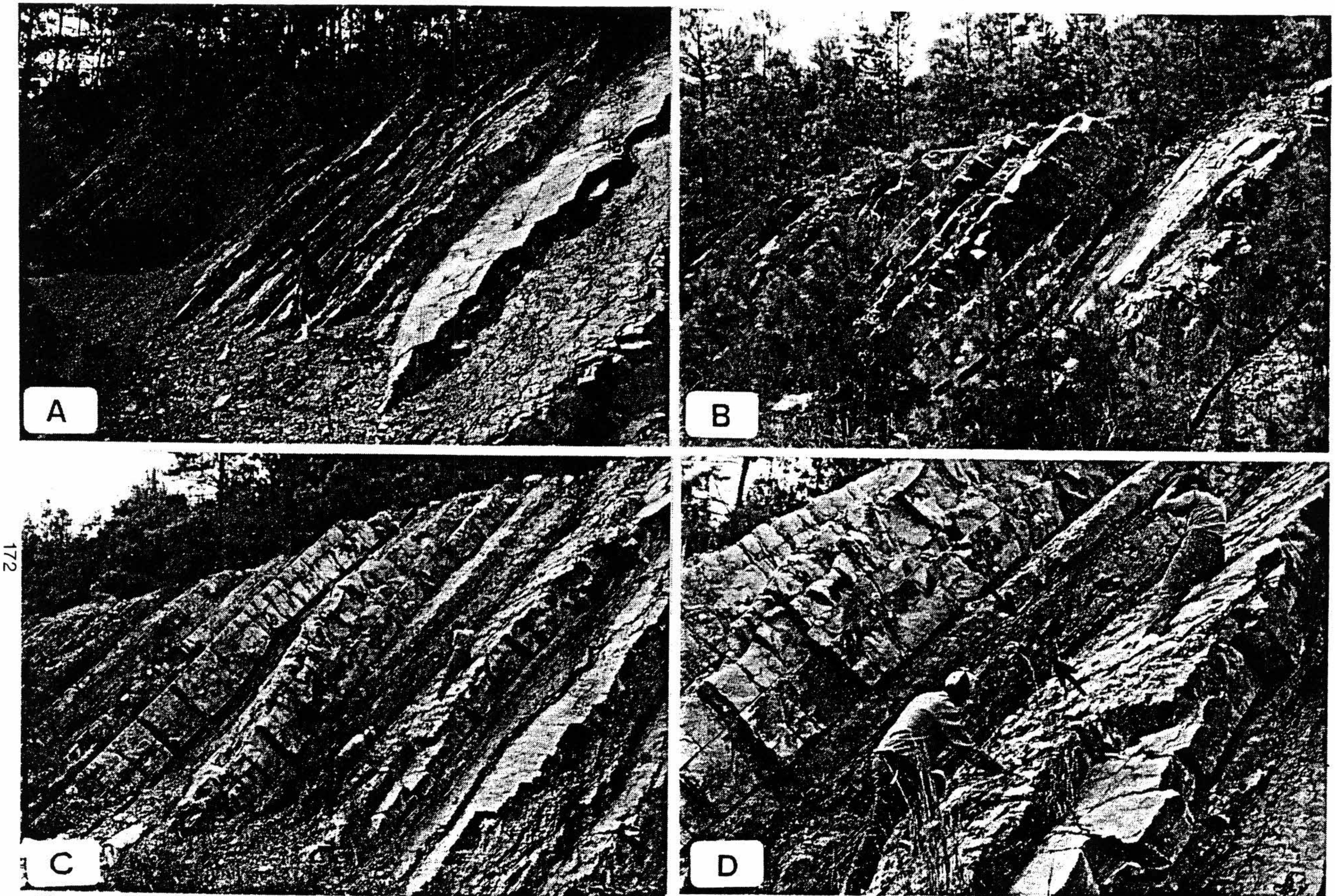


Figure 123

Friendship Roadcut submarine fan channels: (A) upper part of channel sequence, (B) base of channel sequence, (C) channel sequences with slurry bed or debris flow (near people), (D) close-up of debris bed.

(Description from Jordan, et al., 1991)

STOP 15

Murray Quarry, Near Arkadelphia, Arkansas

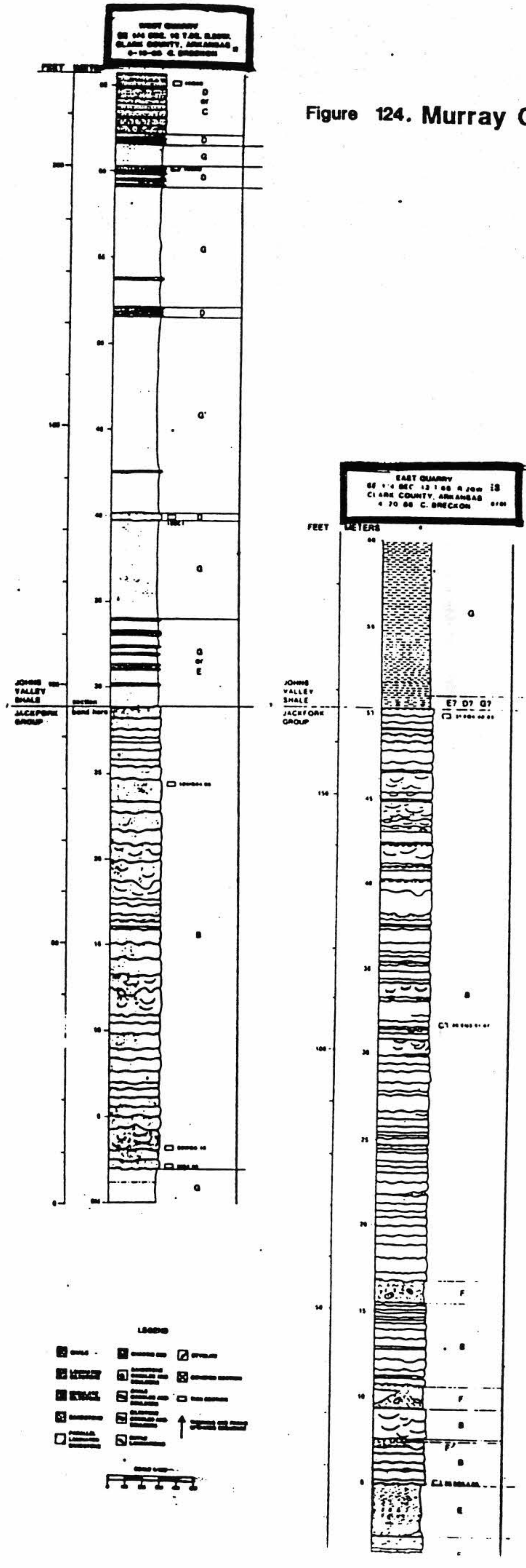
The purpose of this second stop on the second day of the trip is to view and discuss the sedimentology, lateral geometry, and heterogeneity of the uppermost part of the upper Jackfork and overlying Johns Valley Shale. The sandy deposits have much more continuity than the deposits seen on the previous day at the stops near Little Rock. This stop, having steeply dipping but strike-oriented beds, is a prelude to STOP 16 where upper Jackfork rocks are more accessible but viewed in dip-oriented sections.

Approximately 150 feet of southward-dipping upper Jackfork Formation are exposed at the abandoned Murray Quarry near DeGray Lake. Although not easily accessible, two stratigraphic sections have been measured by Breckon (1988), who interprets the succession to be a combination of levee overbank, channel-fill, and extrachannel deposits.

The upper part of the section is deeply incised and infilled with shale and thin bedded sandstones, and is interpreted by Breckon (1988) to be the stratigraphically younger Johns Valley Shale. The base of one incisement is lined with clasts and boulders (Fig. 125) which were deposited as a deep-water channel lag or it is an accumulation of rubble which slid down the channel walls after incisement. The Jackfork is interpreted to be upper fan, channel-levee complex. Is this how a deep sea fan complex dies?

Although we have not examined this section in detail, nor obtained gamma-ray logs, it is worth noting the greater number of thinner beds and greater lateral continuity of many individual beds than were present at most earlier Jackfork stops. These beds are more likely to have a three-dimensional sheet-like geometry.

Figure 124. Murray Quarry Sections



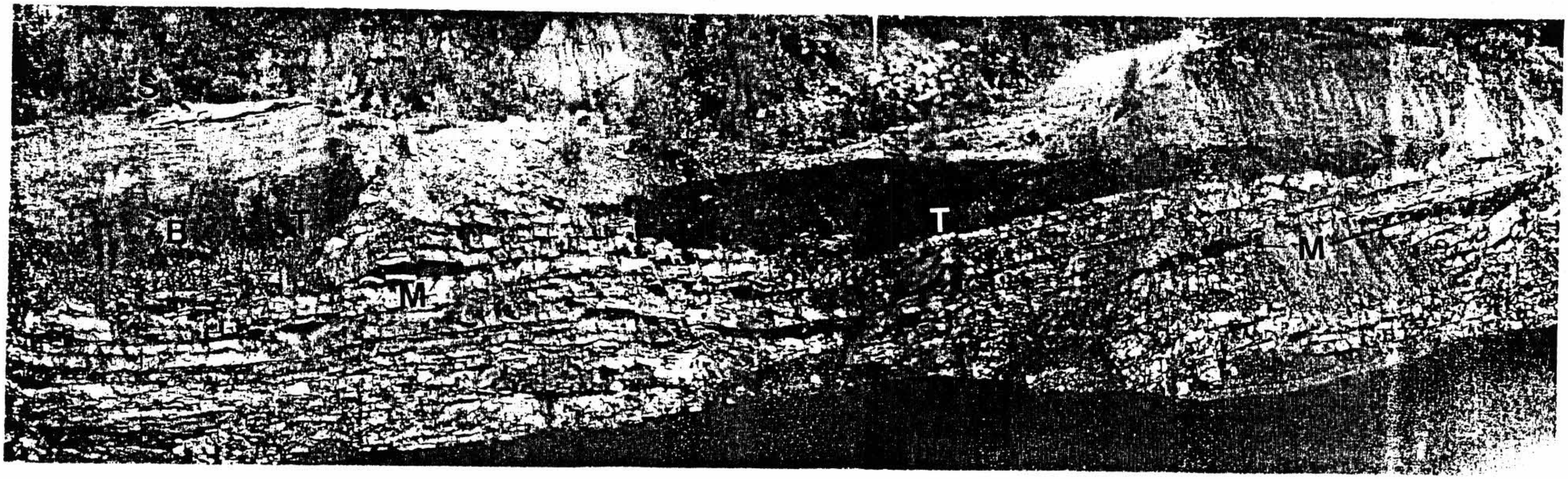


Figure **125.** Murray Quarry, near Arkadelphia, Arkansas. This section exposes the uppermost part of the Jackfork overlain by the lowermost part of the Johns Valley Shale. Note the presence of more laterally continuous sandstone beds than were apparent at the Stops of Day 1, the erosional truncation (T) of the uppermost part of the Jackfork that is overlain by sandstone boulders (B) within a mudstone matrix, and isolated lenticular mudstones (M). Some isolated sandstones occur within the John Valley Shale (S).

(Description from Jordan, et al., 1991; Stone, 1973; Morris, 1974; Tillman, 1991; Bouma, 1993; and others)

STOP 16

DeGray Dam Spillway

At DeGray Dam, an excellent exposure of the middle and upper Jackfork Formation outcrops along a spillway cut (Fig. 126). About 1000 feet of steeply-dipping strata are exposed on both sides of the north-south spillway. The section is several hundred feet lower in the Jackfork section than the exposures at Murray Quarry.

At the DeGray Dam spillway southward dipping, fine grained, quartzitic sandstone, subgraywacke, gray siltstone, and black shale occur in the middle and upper Jackfork Sandstone (Fig. 129). The sequences consist of alternating thick- and thin-bedded sandstones interbedded with shales and slumped mudstones. Most of the sandstone sequences form thinning- and fining-upward cycles interpreted to be submarine channels on the middle part of a fan. Very thick Mutti Ricci-Lucchi "R" facies sandstones, which include a single(?) 25-foot thick bed, outcrop near the top of the section (800-924'; Fig. 127 and Fig. 128). Good examples of westerly oriented sole marks, graded bedding, load and slump structures, dish and pillar structures, ripple marks, broad scour, and some fractures can be observed in these beds. These beds are mostly facies "B" and "C" of Mutti and Ricci-Lucchi. Mutti, Ricci-Lucchi "E" facies and deep-water trace fossils are also present. The shales and mudstone sequences are finely laminated to highly slump folded in character and show abundant wedge-outs of strata laterally. Determination of stratigraphic-up in this outcrop may be accomplished using diagnostic sedimentary structures.

Some very large olistoliths are found "floating" in the mudstone. These fine-grained rocks are interpreted to be interchannel deposits which formed between sand-filled channels. These sandstone olistoliths ("glumps and gloops", R. Haley, personal communication) occur in two intervals, especially in the shale sequences below the upper more massive sandstone near the south end of the spillway. Coalified plant fragments are quite common in some of the debris flow deposits represented by the siltstone "blue beds." A few invertebrate fossil remains have been found at various places along the exposure. Slumps and submarine landslides from the adjacent slope settings and/or from levees and sides of channels contributed material which make up these deposits. These deformed beds are facies "F", slumps and debris flows, of Mutti and Ricci-Lucchi.

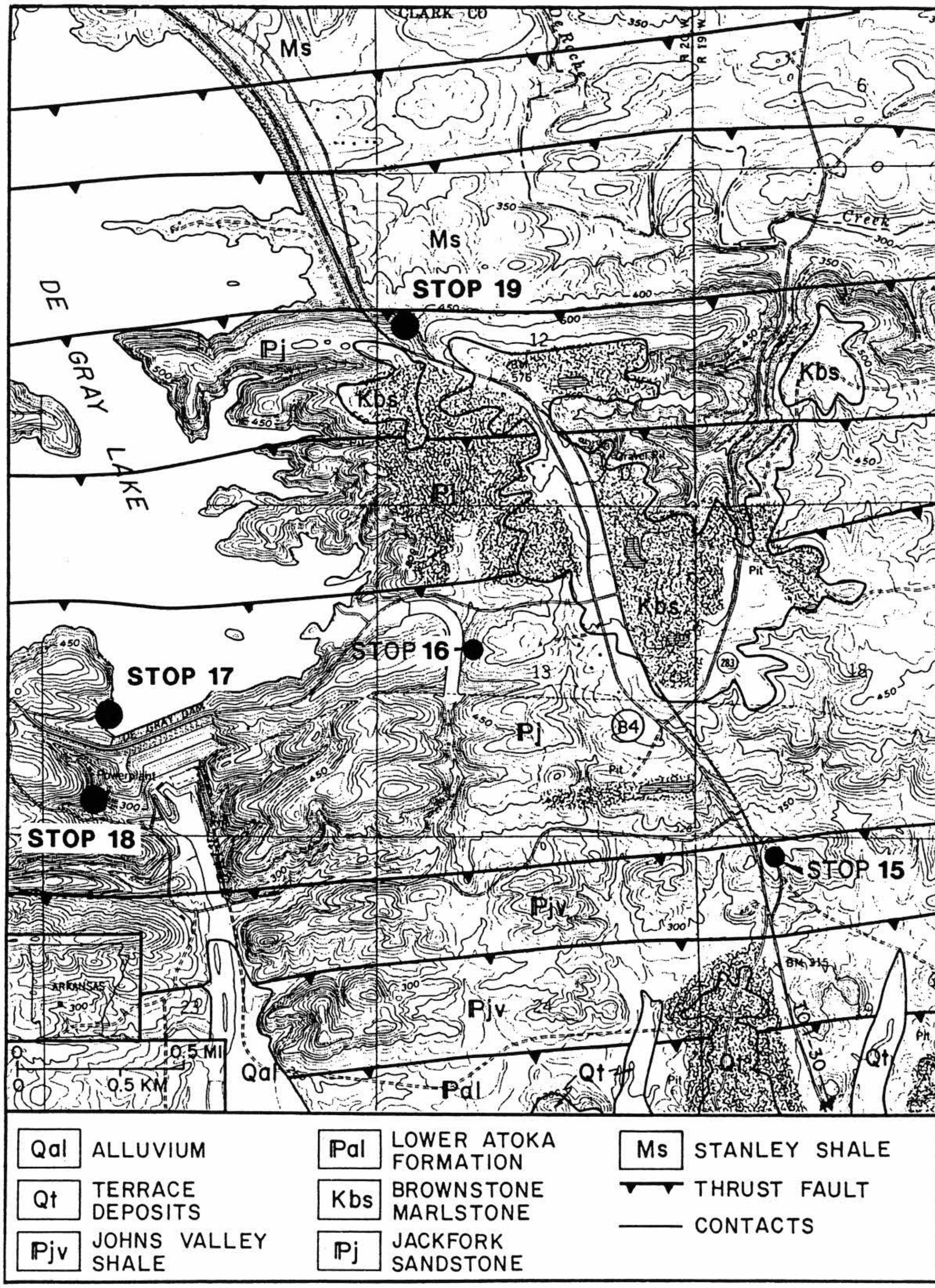


Figure 126 Map showing location of Stops 15 thru 19 at eastern Lake DeGray and vicinity.

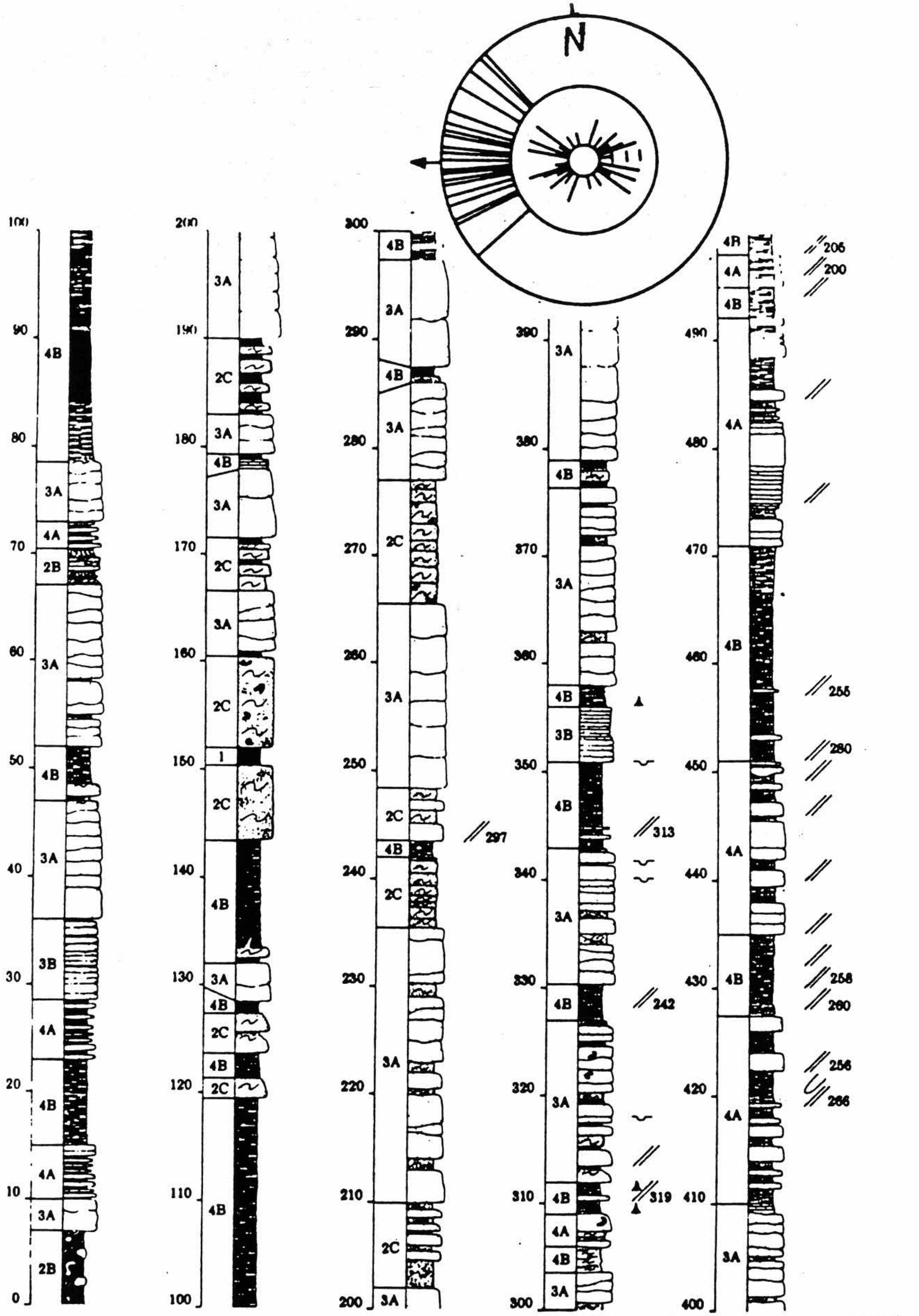


Figure 128A. Measured Section at DeGray Dam Spillway (after Morris, 1974). Paleocurrent measurements in 360 degree circle. Figure continued on following page.

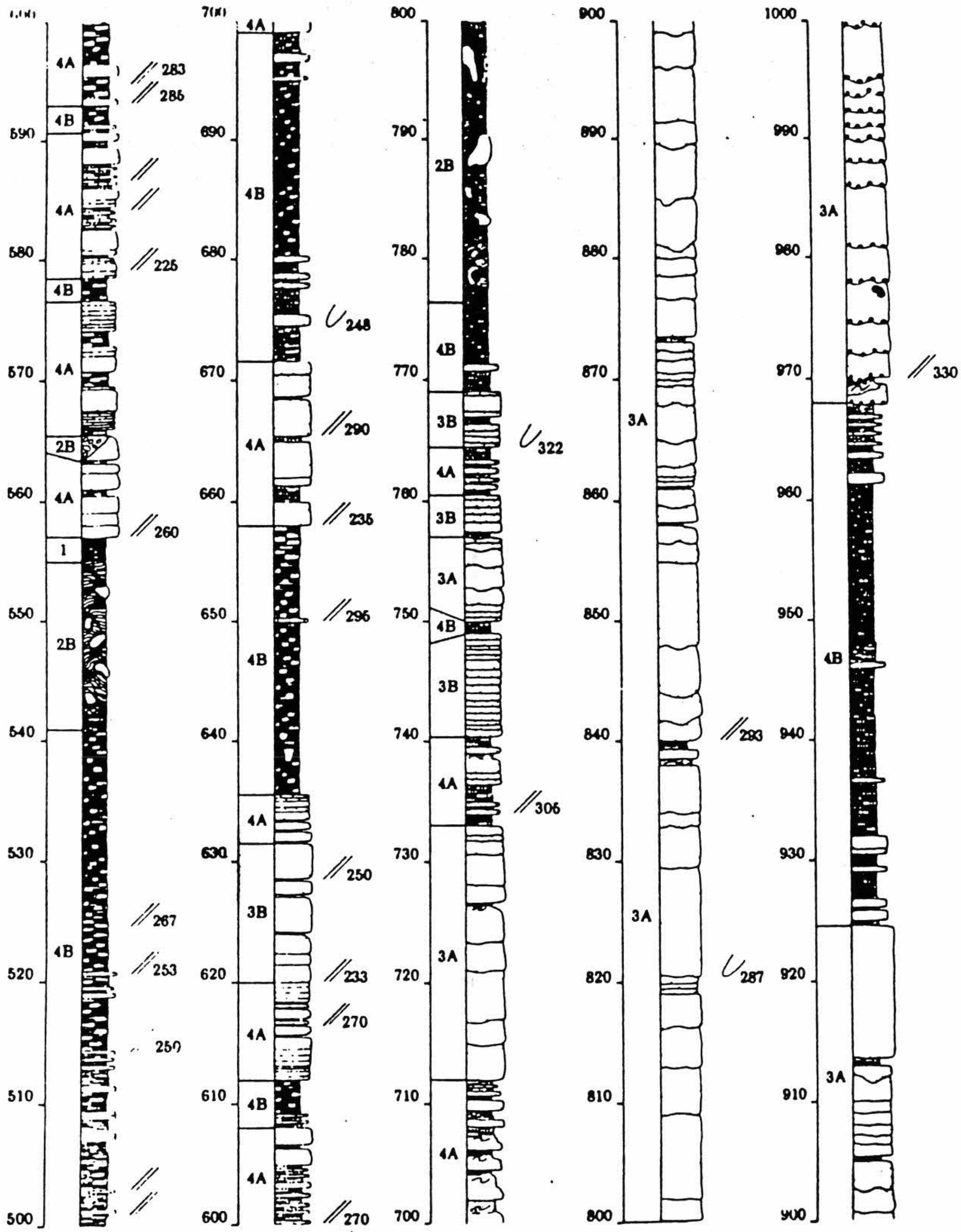


Figure 128 B. Measured section at DeGray Dam Spillway (after Morris, 1974). Refer to paleocurrent measurements and the numbers indicate the orientation of the feature based at 360° circle.

Morris (1974) has studied this spillway and his measured section is included here (Fig. 128). It would be a useful exercise for you to reinterpret at least portions of this section using the Mutti, Ricci-Lucchi nomenclature or other models as you walk the section. Locate the thinning- and thickening-upward sequences. Also see if you interpret the massive conglomeratic sands at the top of the section as obvious channel fill deposits. Decide whether Morris' "rubble bedding" (especially 410-470') is due to tectonics or to soft-sediment slumping.

At the south end of the spillway, near the top of the stratigraphic section, there are at least 15 cycles of granule or "grit" deposition in a 200-foot sequence. These "grit" beds are commonly graded and have some small to rather large scour features. The granules are composed of quartz and metaquartzite.

The conglomerates in the outcrop (Fig. 130) are somewhat unusual in the Jackfork and some have speculated that they came from the south. Their highly quartzose composition, however, may be suggestive of an eastern or northern origin.

All workers generally agree that the Jackfork is a moderately to well sorted, dominantly very fine grained quartz arenite/lithic arenite typically well cemented by quartz. All recognize the presence of some coarser grained and conglomeratic units. Calculations and percentages vary between authors depending on the method used for mineral "partitioning" an example being quartz types such as monocrystalline, polycrystal-line, mosaic, chert, etc. The Jackfork typically is shown to contain 80-95% quartz, 1-5% feldspar (plagioclase + K-spar), 2-10% lithics (MRF, VRF, PRF, SRF), and 10-25% matrix (detrital + authigenic clays + grains too small to identify petrographically). Kaolinite, illite, chlorite, and halloysite have been docu-mented. Quartz is uniformly reported as the dominant cement with minor amounts of limonite, siderite, hematite, and dolo-mite. Muscovite, biotite, zircon, rutile, apatite, and fossil fragments are the commonly reported accessories and heavy minerals.

In recent years, there have been a number of published interpretations of the depositional processes and setting represented by upper Jackfork rocks exposed in this section (Table 4). Two basic types of sediment gravity flows are represented by deposits within this section: (1) turbidity currents and (2) debris flows. Turbidity currents were responsible for deposition of virtually all of the sandstone beds within the DeGray Lake Spillway section. Many probably began as retrogressive failures of submarine slopes as suggested by the common compound character of sandstone units. Thicker sandstone units are composites of many thin, 2 to

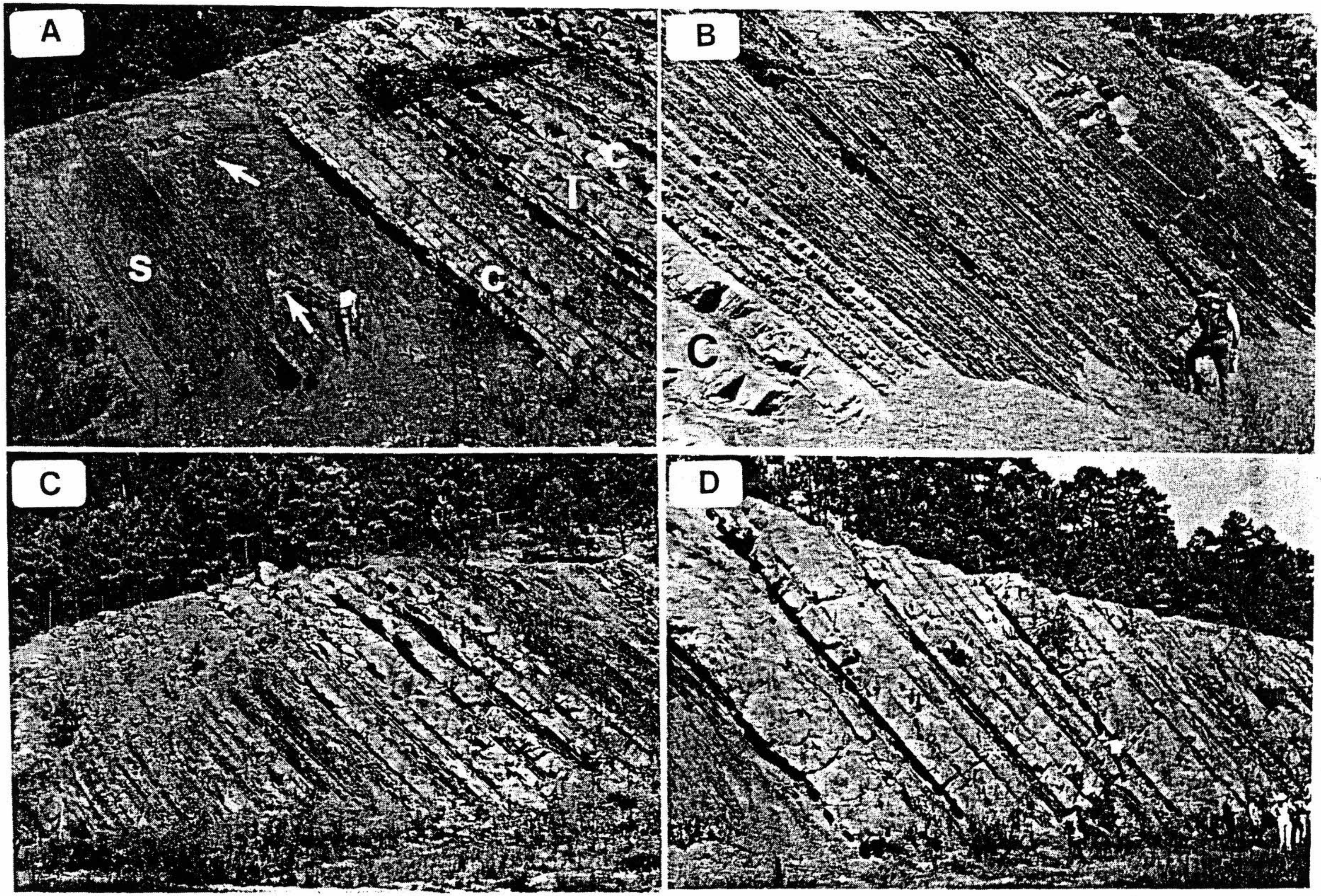
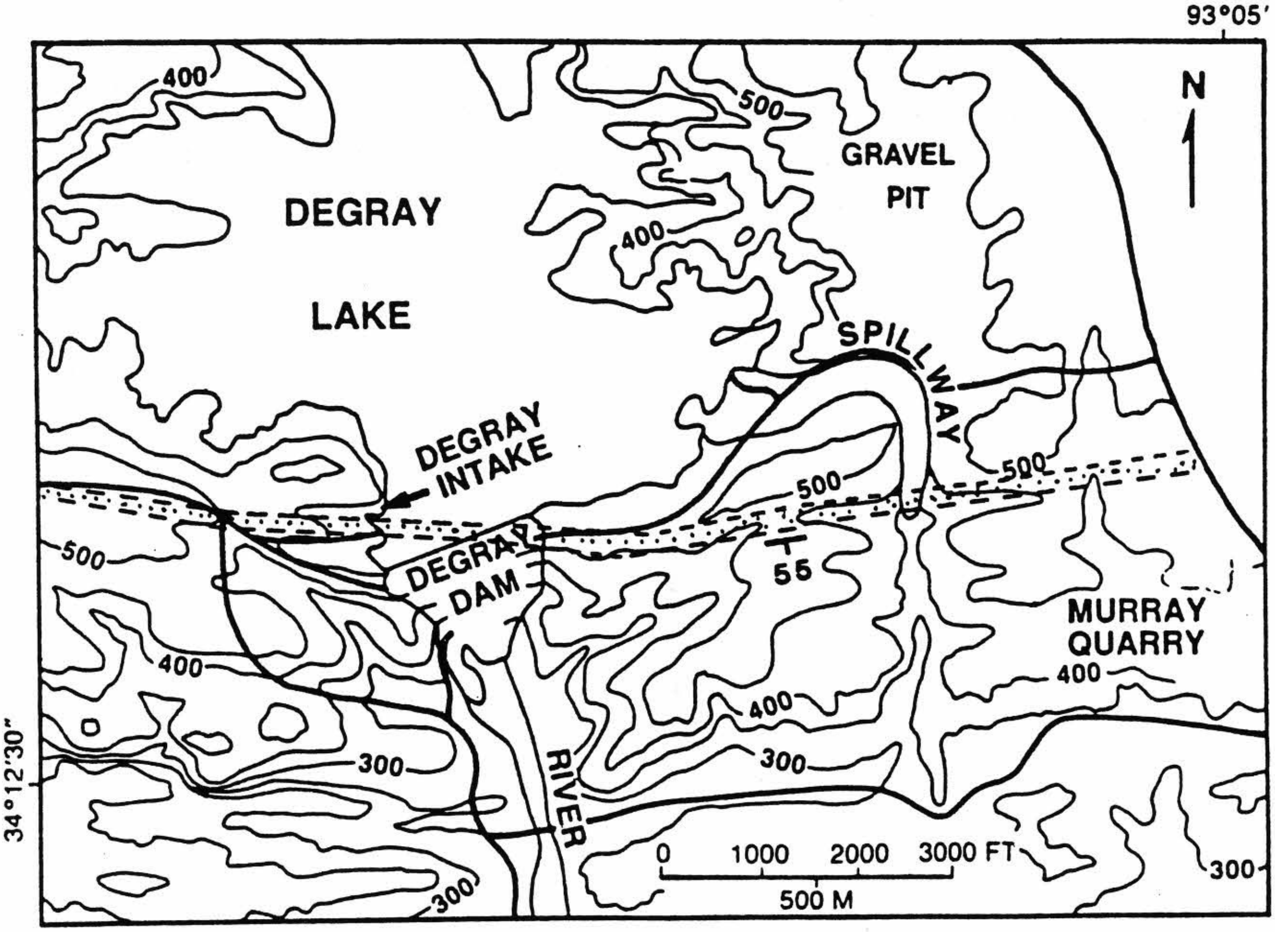


Figure 129. DeGray Dam Spillway outcrops. A) Shale (S), shale slump with large clasts (arrows) sequence, alternating upward thinning small interchannel crevasse splay channels (C) and upward thickening lobes (1) B) Finely laminated shale (80-120') overlying a sandstone filled channel (C) sequence. C) Symmetrical cycle with the thickest bed in the center of the sequence. D) Several stacked channels sequences (dots) which thin and fine upward.



Jordan et al., 1991

Fig. 130 . Location map DeGray Lake area showing spillway, dam and other features. Upper conglomeratic sandstone shown in stippled pattern.

20 inch-thick sedimentation units representing individual flows. The lowest beds are typically blocky and show flat lamination. Higher beds showundulating contacts produced by soft-sediment deformation and foundering of one bed into underlying, liquefied beds. The uppermost beds in a package are usually thinner and show flat bases and tops.

Interbedded distal and proximal turbidites deposited on an abyssal plain (Morris, 1973).
Lower slope- outer fan lobe, lobe fringe and interlobe deposits; massive sandstone/pebbly sandstone interval at the top of the section comprises a major distributary channel (possible middle fan) (Lock and Fisco, 1979).
Prograding sequence of middle fan, submarine fan channel, and outer fan lobe deposits; coarse deposits at the top are middle fan channel fills (Stone and McFarland, 1981; McFarland, 1988).
Middle fan channel and interchannel crevasse splay deposits. Overall prograding system. The uppermost coarser-grained intervals may represent an inner fan transition zone (Moiola and Shanmugan, 1984; Link and Roberts, 1986).
Submarine channel-fill (Shanmugan et. al., 1988).
Middle-fan or suprafan, channel-fill and lobe complex overlain by chaotic and sandy deposits of a major inner-fan channel-fill complex (Breckon, 1988).

Table 4. Published Interpretations of the DeGray Lake Spillway Section.

At DeGray Lake Spillway, there is a continuous spectrum from 3 to 10 foot-thick sandstone packages representing more proximal flow sequences to thin sandstone beds, 1 to 3 feet thick, representing intermediate flow stages; to beds less than 1 inch thick, representing distal flow. These units provide a downslope spectrum representing the various stages in the evolution of mud-poor, fine-grained turbidity currents produced by retrogressive failures.

The Jackfork Formation at the DeGray Lake Spillway section is close to the southeastern limit

of Jackfork outcrops in the Ouachita Mountains. If the bulk of the sediments were derived from the northeast as some have inferred, this section should be located further downfan than that at the Big Rock Quarry. We might expect this to be reflected in a number of ways:

(1) The overall sand/shale ratio might decrease down fan, although submarine slopes and proximal fan areas outside of channels are also characteristically sites of mud deposition. Much of the coarser sand and most of the fine sand bypasses the upper fan and upper mid-fan areas, moving further downslope in turbidity currents and accumulating in lobes or more distal outer fan regions. Perhaps this criterion is less useful than might be expected, especially within predominantly fine-grained systems like the Jackfork.

(2) The coarsest grain sizes should decrease down fan. It is noteworthy that perhaps the coarsest sediment in the Jackfork anywhere in the Ouachitas occurs near the top of the DeGray Lake Spillway section.

(3) The frequency and scale of channeling should decrease down fan. It is difficult to assess the abundance of channels in the DeGray Lake Spillway section compared to that in Big Rock Quarry because lithologic units in the Spillway crop out for only a short distance along strike. Overall, there appears to be less channeling and bed lenticularity and more lateral bed continuity than in the Big Rock Quarry (discussed in detail below). This is verified at the Intake section.
(4) Large scale, slope-related slides and debris flows might decrease down fan, away from the bounding basinal slopes. Two of the largest debris flows that we will see on this trip are located in the upper part of the DeGray Lake Spillway section. One is nearly 30 feet thick. There are, however, fewer small debris flow deposits and nothing like the sand-matrix mudstone-clast breccias seen in the Big Rock Quarry. Evidence for repeated local slides and debris flows is also missing in the spillway section.

It seems that the DeGray Lake Spillway section of the Jackfork probably represents a more basinal setting less directly associated with slopes and slope processes as inferred in Big Rock Quarry or Pinnacle Mountains, although the sediments were probably derived primarily from the south or southeast.

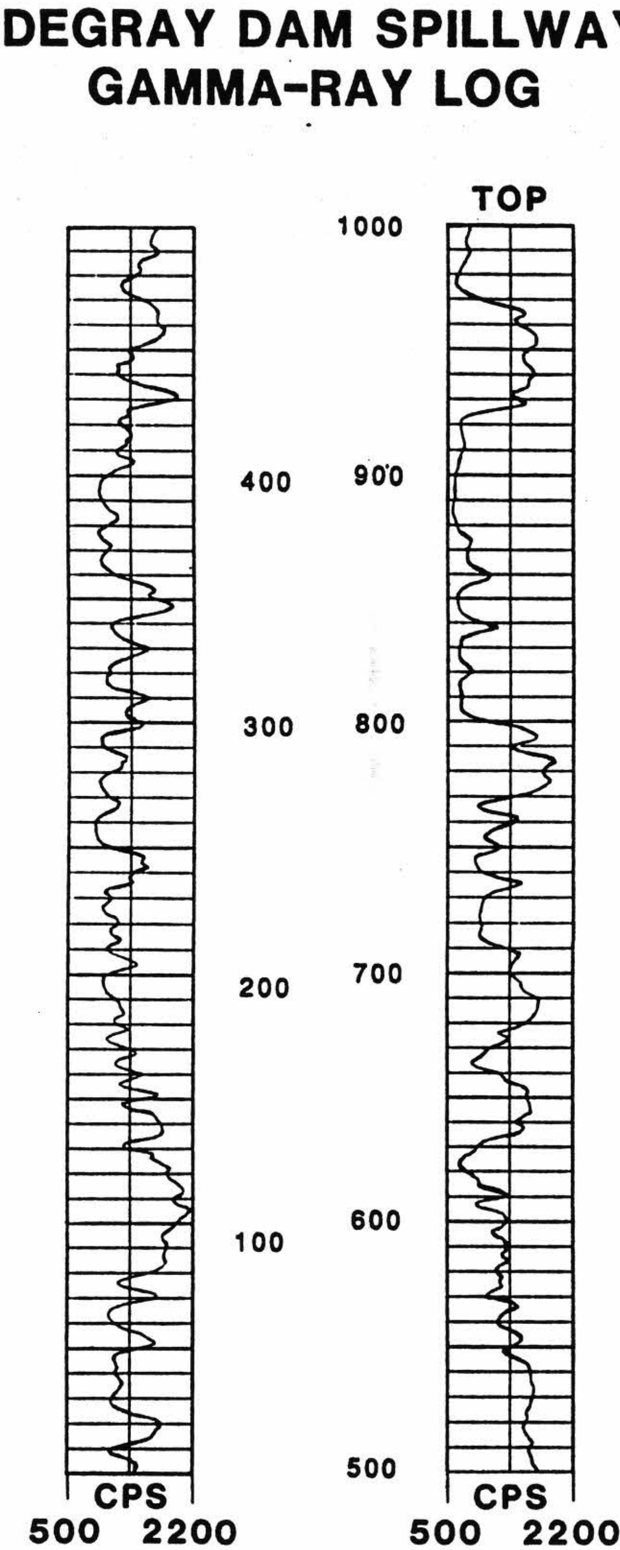


Fig. 131. Gamma ray log produced from measurements by hand-held scintillometer on east wall of DeGray Lake Spillway. Base of section is at lower left. Incremental scale is in 10' intervals. Total section is 1000'.

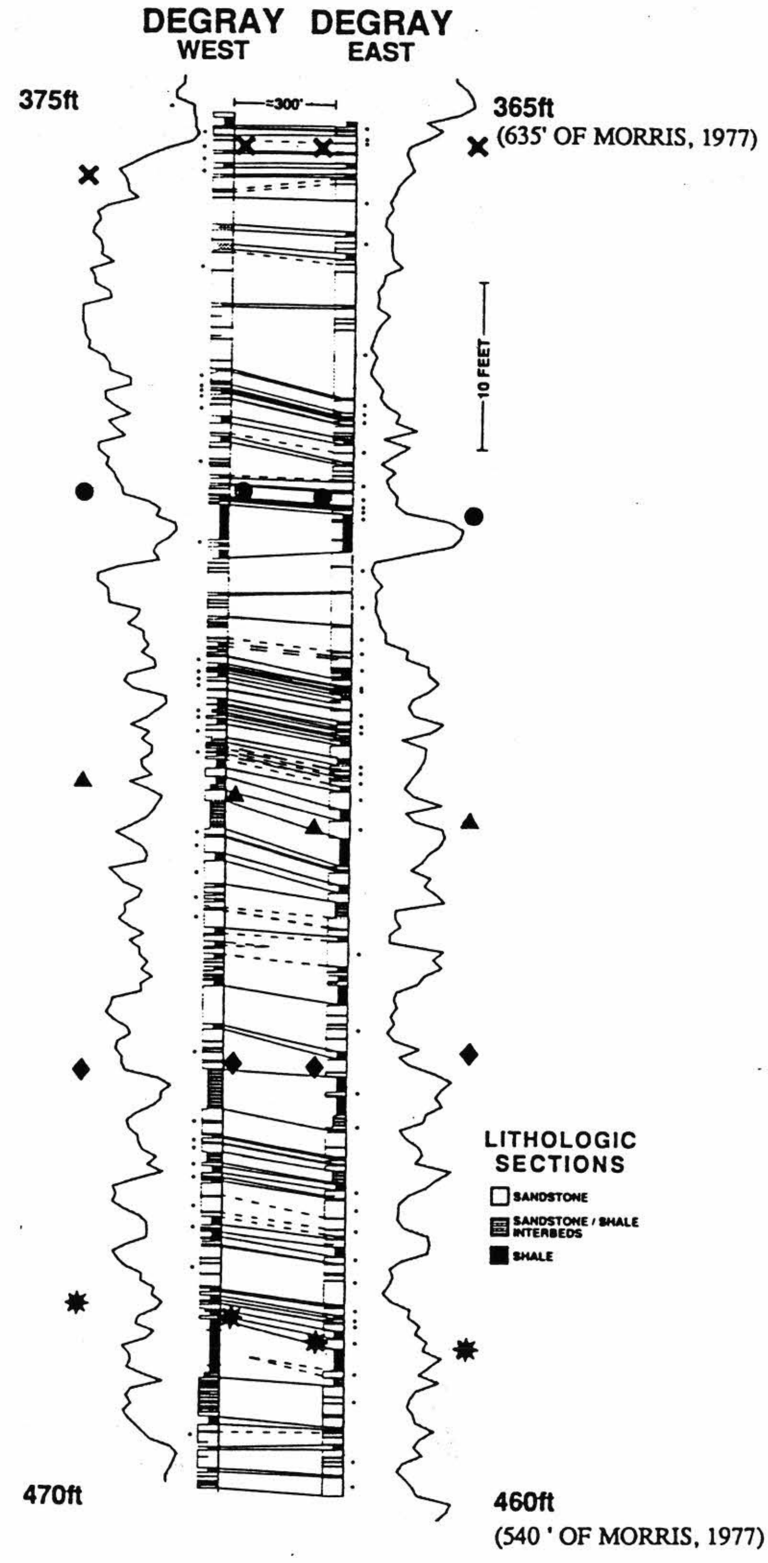


Figure 132. Detailed stratigraphy and half-foot gamma-ray logs for the interval 365-460 and 375-470 feet (log depth) along the eastern and western walls, respectively, of DeGray Lake Spillway.

The large symbols refer to some individual beds correlated between the outcrops and logs. Small dots are located adjacent to the bed that is thickest (relative to the other wall of the spillway).

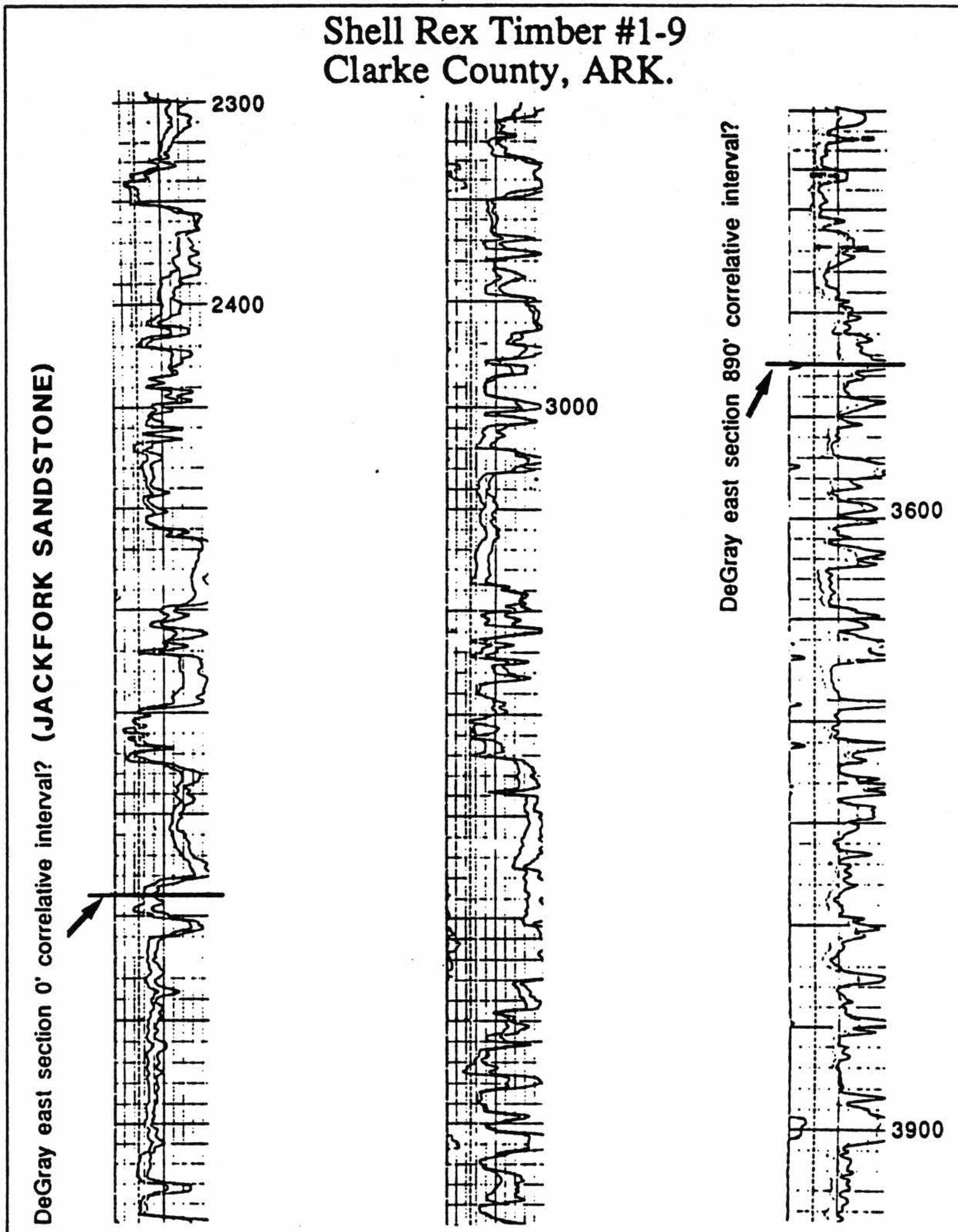


Fig. 133. Subsurface gamma ray log Shell Rex Timber Co. 1-9 located approximately 6 miles SSE of DeGray Dam Spillway. Jordan et al. (1991) interpret the 0-890' section on the east wall of the spillway to be equivalent to the interval from approximately 2700' to 3525' in the well.

(Description by Jordan, et al., 1991;
Breckon, 1988; and others)

STOP 17

Lake DeGray Intake, Bismarck, Arkansas

This stratigraphic section, within view of the DeGray Lake Spillway, provides a good example of changes in long distance (about one mile) lateral continuity of beds as well as laterally continuous conglomeratic beds described by Breckon (1988). The conglomerate beds at the Intake are correlative with those at the south end (uppermost parts of the section) of the Spillway, and the beds can be traced for several miles along strike.

A portion of the stratigraphic section at this locale was measured for bed thickness and gamma-ray log character (at 0.5 and 2-foot sample spacing). This measured section correlates with the detailed section that was measured at DeGray Spillway (at approximately 375-470 feet at the DeGray Lake Spillway west section, log depth, or 550-635 feet on the section by Morris, 1977; compare with Figure 132).

In addition, gamma-ray logs were measured at two additional Jackfork localities that lie near the Dam in a higher stratigraphic interval than the beds at the Intake section.

Much of the interpretation of this section has been interwoven in the comparable section described at STOP 15 (DeGray Lake Spillway). Figure 136 shows the concept of compensation cycles applied to the DeGray Lake Spillway and the DeGray Lake Intake sections. Our gross correlation scheme shows that some intervals become sandier and thicker toward the east while other intervals become sandier and thicker toward the west. As was the case above, it is evident that a knowledge of depositional geometries is essential for a realistic correlation of stratigraphic intervals or depositional cycles at this scale of well spacings (see discussion in previous section STOP 15). This correlation attempt, while not perfect, can be thought of as representative of a typical oil field in which planning of well spacing scenarios for development of hydrocarbon resources from similar deepwater sandstones is critical. By correlating lithogenetic packages instead of "layer-cake" lithologies (or sandstone to sandstone), a better description of the reservoir would result and productivity of the hydrocarbons would be enhanced.

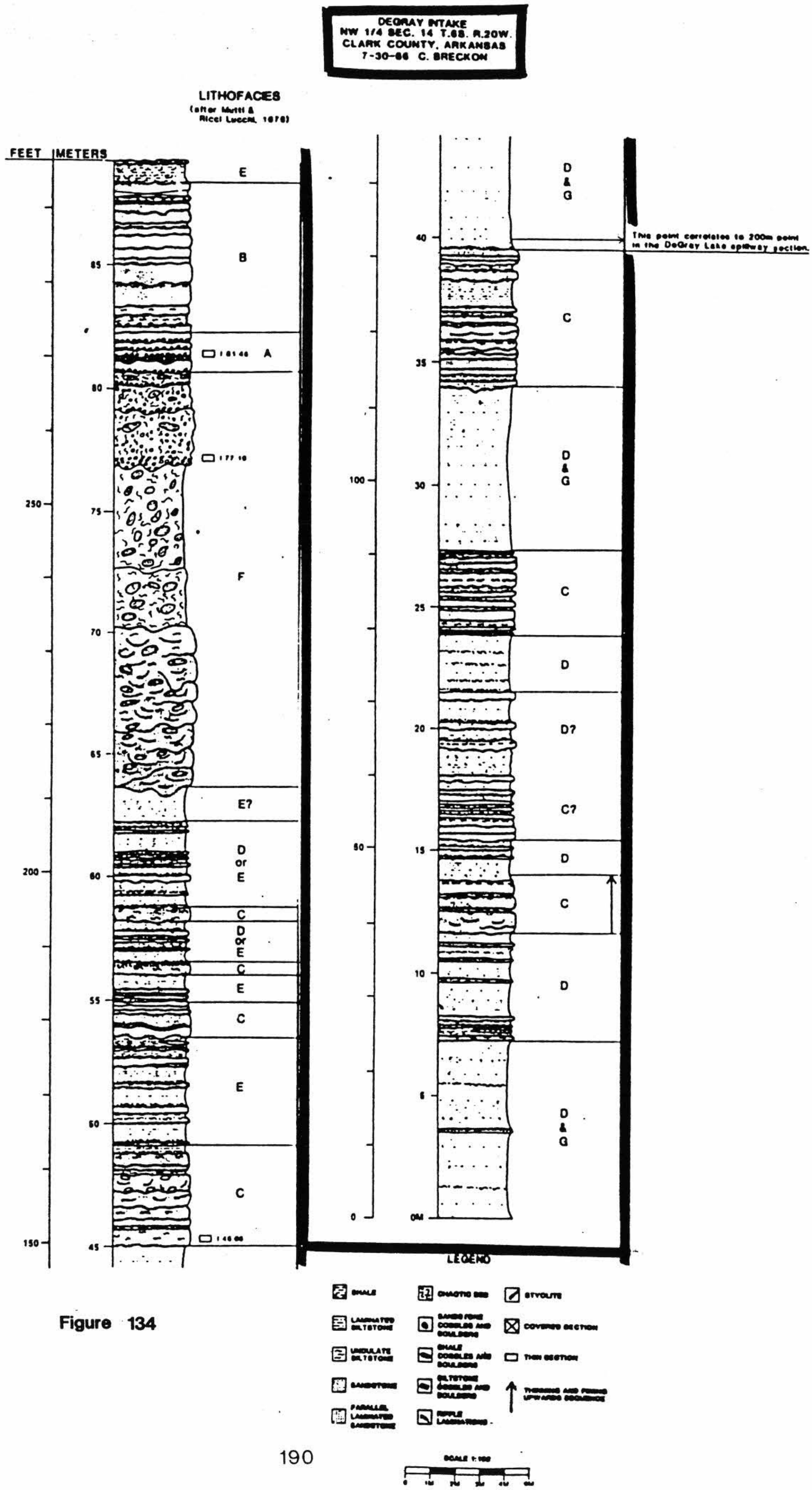

DEGRAY INTAKE
NW 1/4 SEC. 14 T.6S. R.20W.
CLARK COUNTY, ARKANSAS
7-30-86 C. BRECKON
LITHOFACIES
(after Mutti & Ricci Lucchi, 1978)
FEET
METERS
E
B
A
F
E?
D or E
C
D or E
C
E
C
E
C
D & G
This point correlates to 200m point in the DeGray Lake spillway section.
C
D & G
C
D
D?
C?
D
C
D
D & G
LEGEND

Figure 134

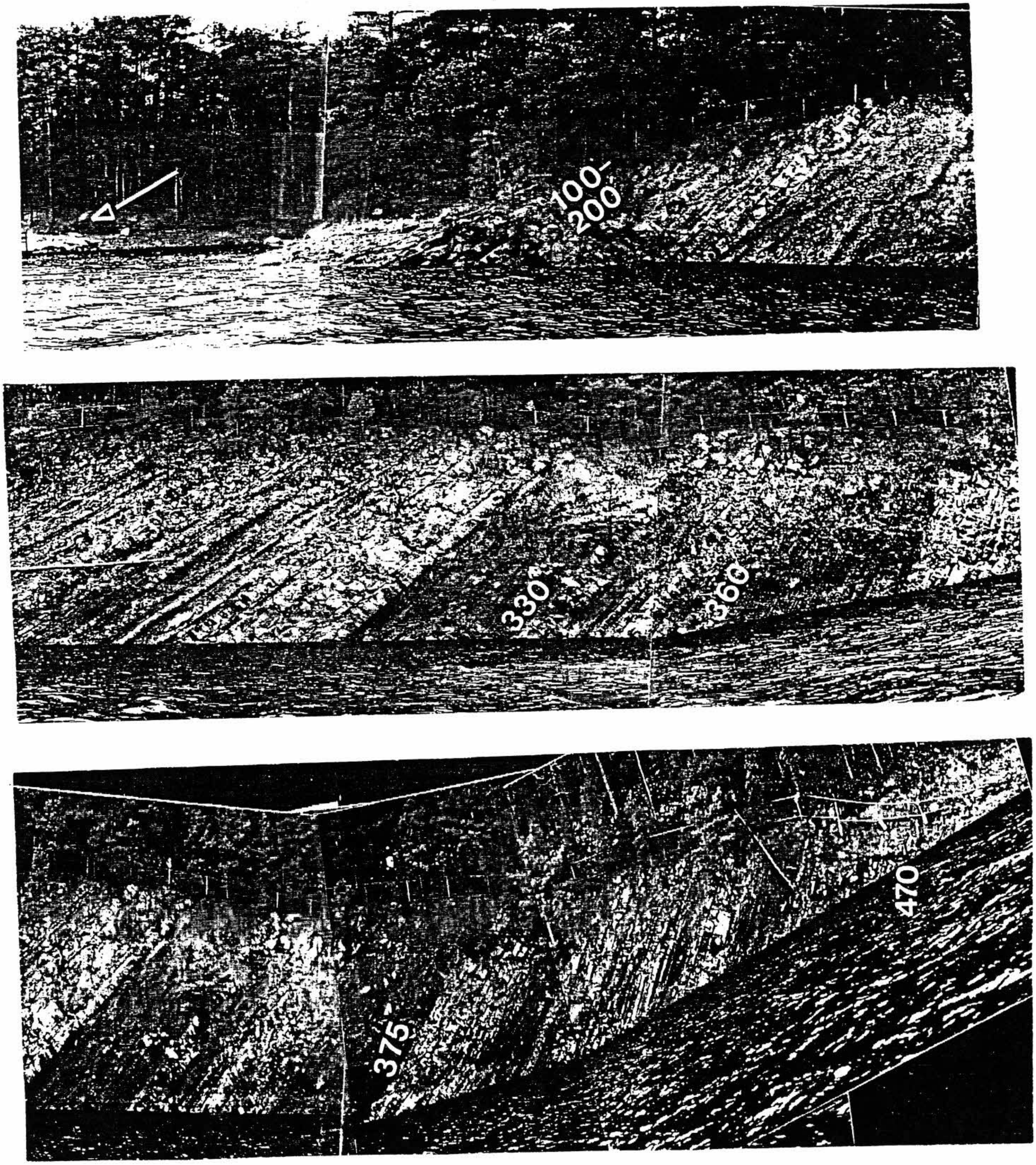


Figure 135 DeGray Intake section. Numbers on photograph refer to correlative interval at DeGray Spillway west section (log depth).
Compare with equivalent interval shown in Figure 132. (A) Upper part of section. Conglomeratic bed shown at arrow. (B) Middle part of section. Section is equivalent to interval at 330-360 feet (log depth) at Spillway west. (C) Lower part of section is equivalent to 375-470 feet (log depth) on the Spillway west section.

CORRELATION OF THE JACKFORK FM. BETWEEN DEGRAY SPILLWAY CUTS AND DEGRAY DAM INLET CUT

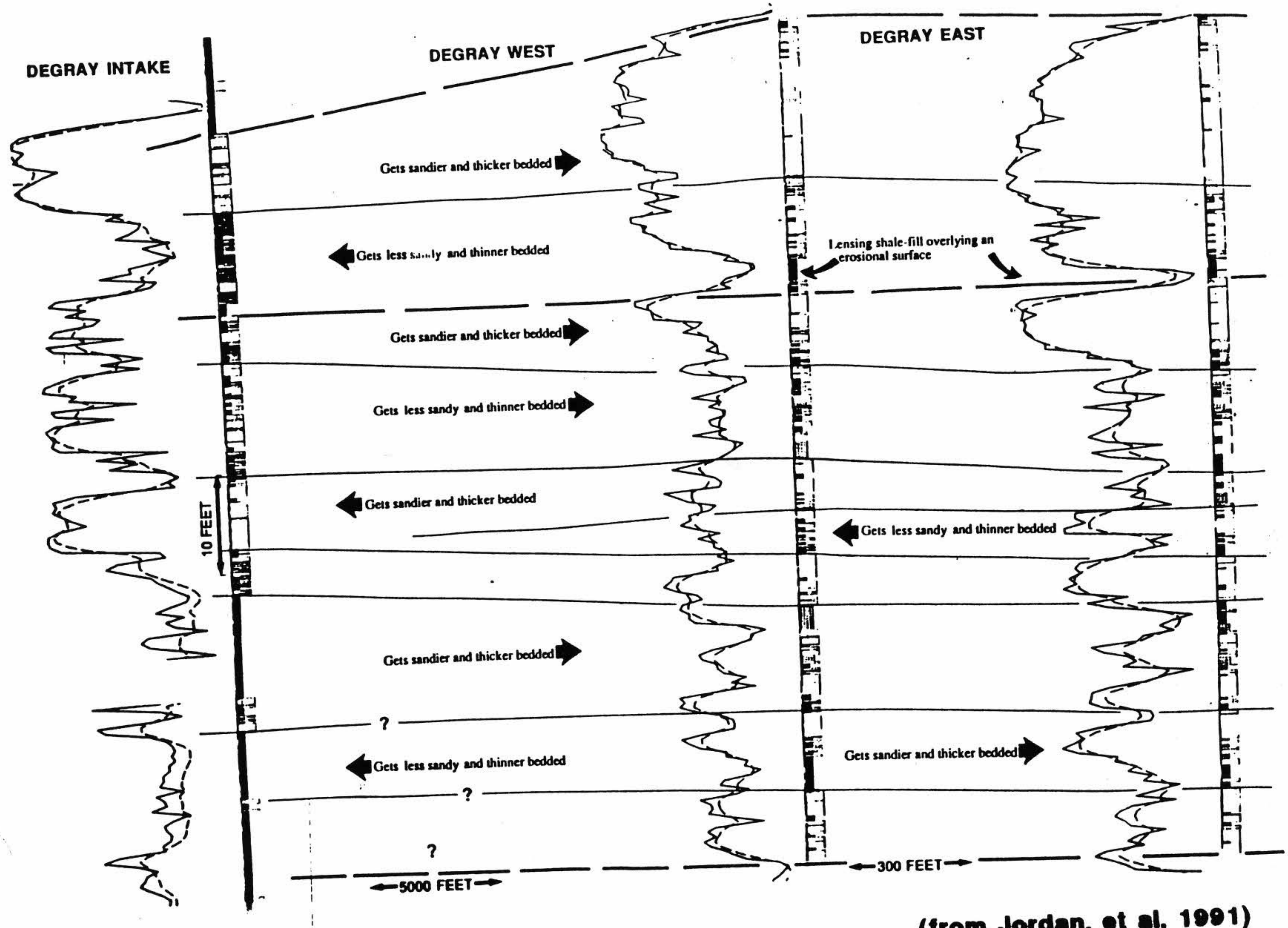
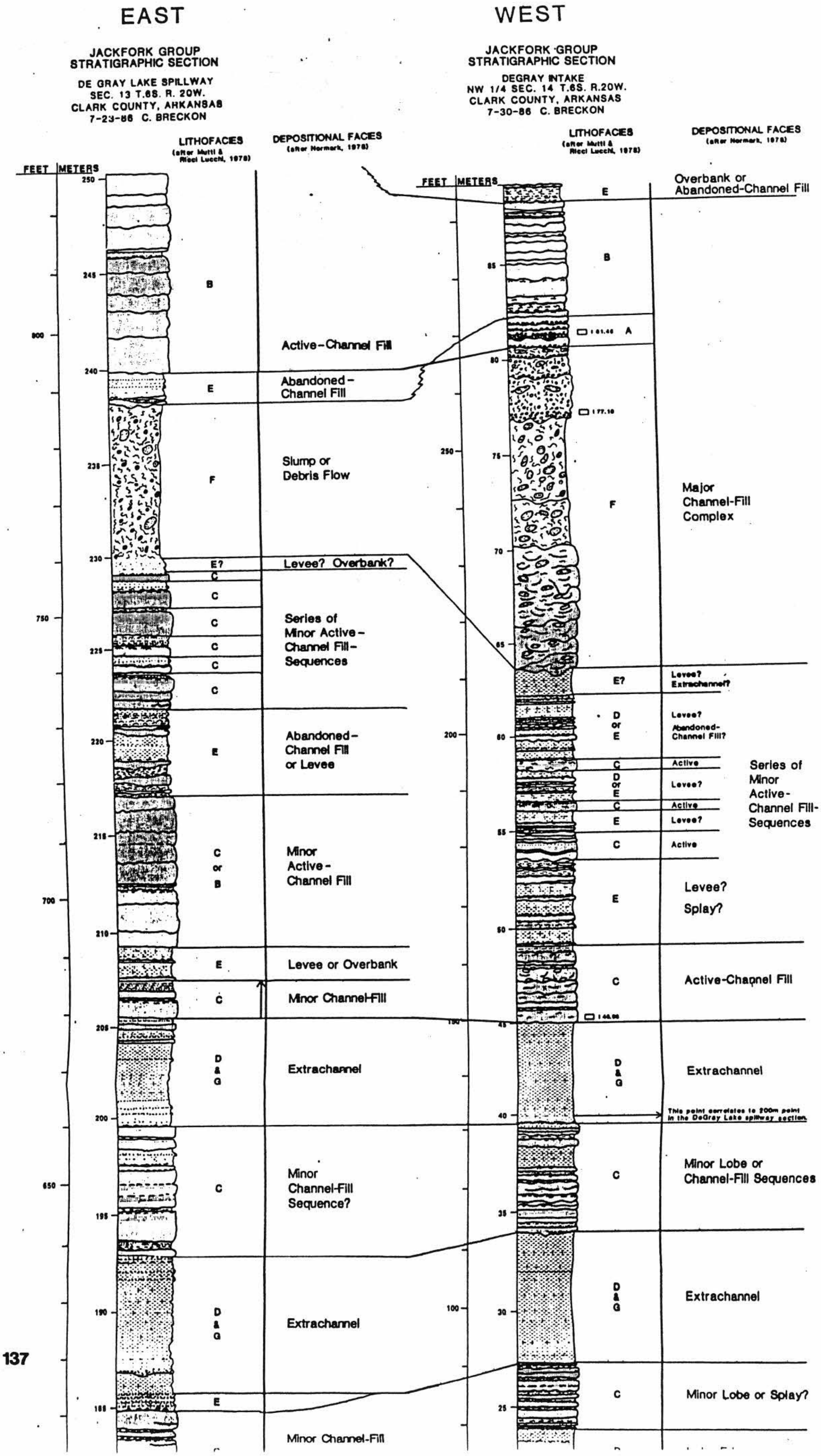


Figure 136

(from Jordan, et al, 1991)



EAST
JACKFORK GROUP
STRATIGRAPHIC SECTION
DE GRAY LAKE SPILLWAY
SEC. 13 T.6S. R. 20W.
CLARK COUNTY, ARKANSAS
7-23-86 C. BRECKON
LITHOFACIES
(after Mutti & Ricci Lucchi, 1978)
DEPOSITIONAL FACIES
(after Normark, 1978)
FEET
METERS
Active-Channel Fill
Abandoned-Channel Fill
Slump or Debris Flow
Levee? Overbank?
Series of Minor Active-Channel Fill-Sequences
Abandoned-Channel Fill or Levee
Minor Active-Channel Fill
Levee or Overbank
Minor Channel-Fill
Extrachannel
Minor Channel-Fill Sequence?
Extrachannel
Minor Channel-Fill
WEST
JACKFORK GROUP
STRATIGRAPHIC SECTION
DEGRAY INTAKE
NW 1/4 SEC. 14 T.6S. R.20W.
CLARK COUNTY, ARKANSAS
7-30-86 C. BRECKON
LITHOFACIES
(after Mutti & Ricci Lucchi, 1978)
DEPOSITIONAL FACIES
(after Normark, 1978)
FEET
METERS
Overbank or Abandoned-Channel Fill
Major Channel-Fill Complex
Levee? Extrachannel?
Levee? Abandoned-Channel Fill?
Active
Levee?
Active
Levee?
Active
Series of Minor Active-Channel Fill-Sequences
Levee? Splay?
Active-Channel Fill
Extrachannel
This point correlates to 200m point in the DeGray Lake spillway section.
Minor Lobe or Channel-Fill Sequences
Extrachannel
Minor Lobe or Splay?

Figure 137

(Description from Coleman & Van Swearingen, 1991; Jordan et al., 1991; Breckon, 1988; and others)

STOP 18

Lake DeGray Dam Powerhouse

The purpose of this stop is to continue our traverse of the Jackfork fan complex, comparing the dominant bed forms and associated processes of the upper Jackfork. This section at the powerhouse access road provides an examination of the upper Jackfork stratigraphically above the Spillway exposure and below the Murray Quarry site.

The strata here are channel-levee complex sandstones and shales, with a probable upper fan - slope component, based on the degree of chaotic bedding and slumping, especially in beds exposed in the raceway outlet.

The stratigraphic top of the Jackfork is not well exposed along the river to the south. However, it is located near the topographic slope change from high ridge to low, flat plain, and is approximately where the small bridge crosses the river. Exposures on the east side of the river, and stratigraphically above the raceway exposure, are poor, but generally suggest a shaly section with thin, minor sandstone beds.

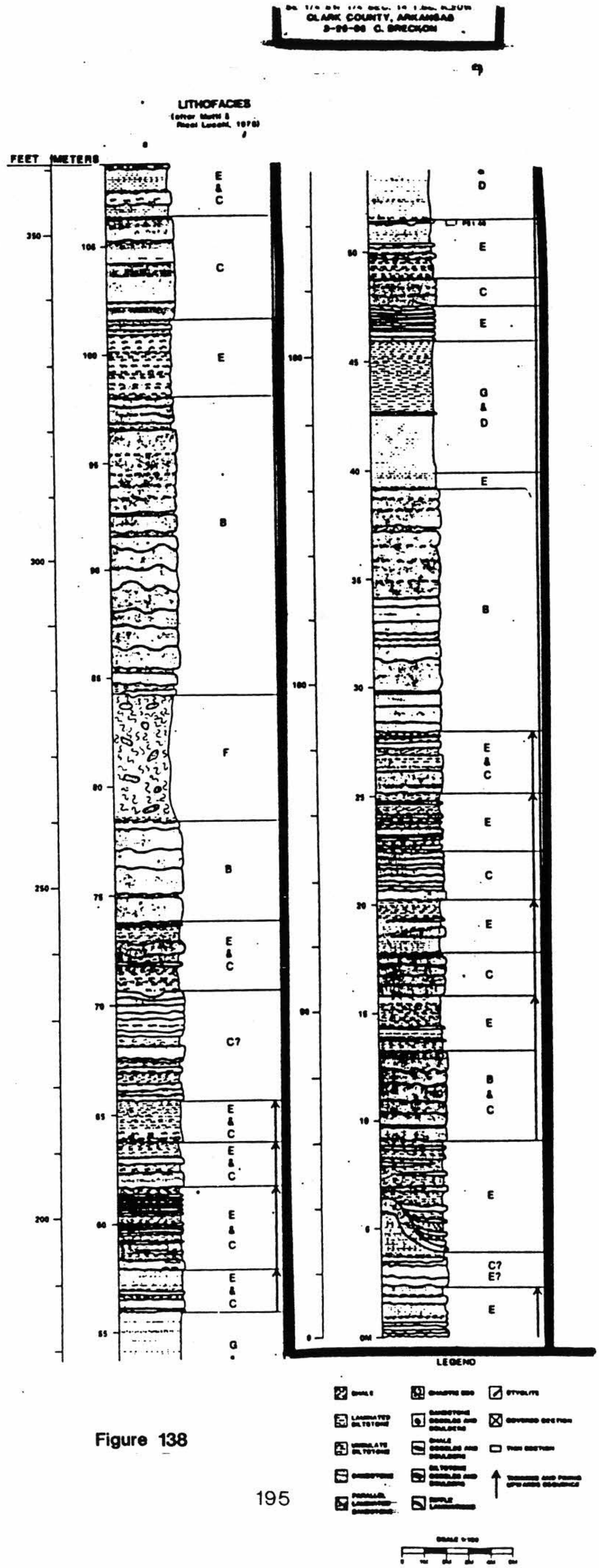


Figure 138

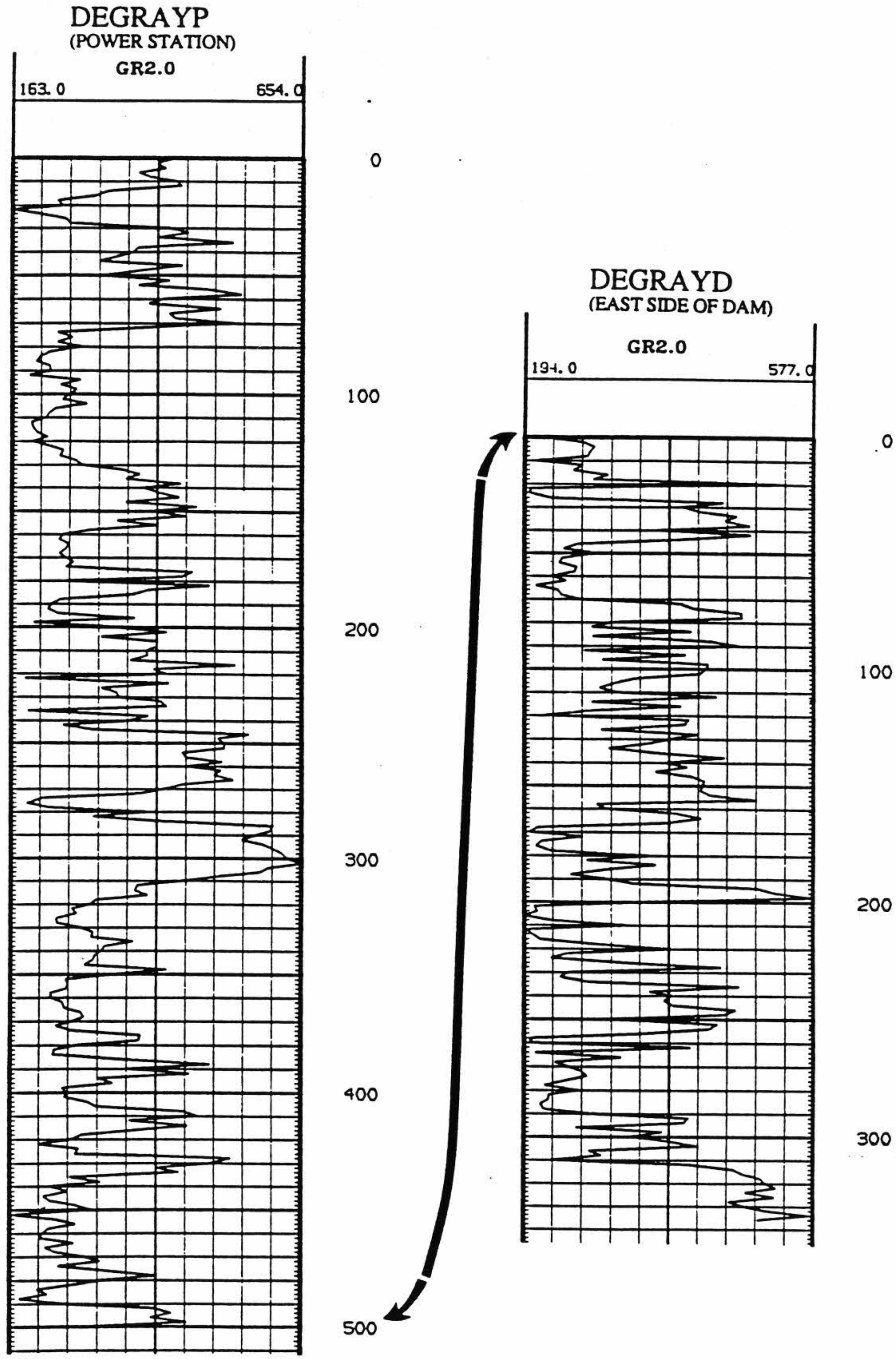


Figure 139.

Gamma-ray logs (2-foot sample spacing) taken at the DeGray Power Station (DEGRAYP) cut (west side of out take) and Dam (DEGRAYD) cut (east side of out take) sites. The section at DEGRAYP is stratigraphically higher.

(Description from Coleman and Van Swearingen, 1991; Stone, 1973; nad Morris 1974)

STOP 19

DeRoche Ridge Roadcut

Roadcut along Arkansas Highway #7 at the south end of large man-made dike on eastern Lake DeGray. We will begin our transact of the lower Jackfork Formation at the base of the unit, where it rests with conformity on the Chickasaw Siliceous Shale member of the upper Stanley Formation. The contact is exposed in the drainage ditch, near the Highway #7 junction with the road to the swimming beach. The Jackfork is approximately 6,200 feet thick in the DeGray area and is 150 miles from the northern shelf, 180 miles from the eastern shelf, and an estimated 100 miles from the southern shelf. For all intents and purposes we are out in the middle of the basin, and at this point, on the basin floor.

This exposure shows alternating massive-to thin-bedded, fine-tomedium-grained (locally some grit layers with crinoid fragments), graded, bottom-marked, convoluted, subgraywacke to quartzitic sandstone; rather soft sandy siltstone with coalified plant fragments; and gray to black shale (containing some small iron carbonate concretions). Bottom marks indicate the paleocurrent was generally from an easterly direction. Is this sequence of beds what you might expect at the bottom of a thick deep sea fan complex? Notice that a pattern of debris flow-->>thick sandstone-->>thin sandstones-->>shale develops at about 10 feet above the base and is repreated, with variations, 6 times up to the fault at about 340 feet onthe section (Fig. 140) (Morris, 1974).

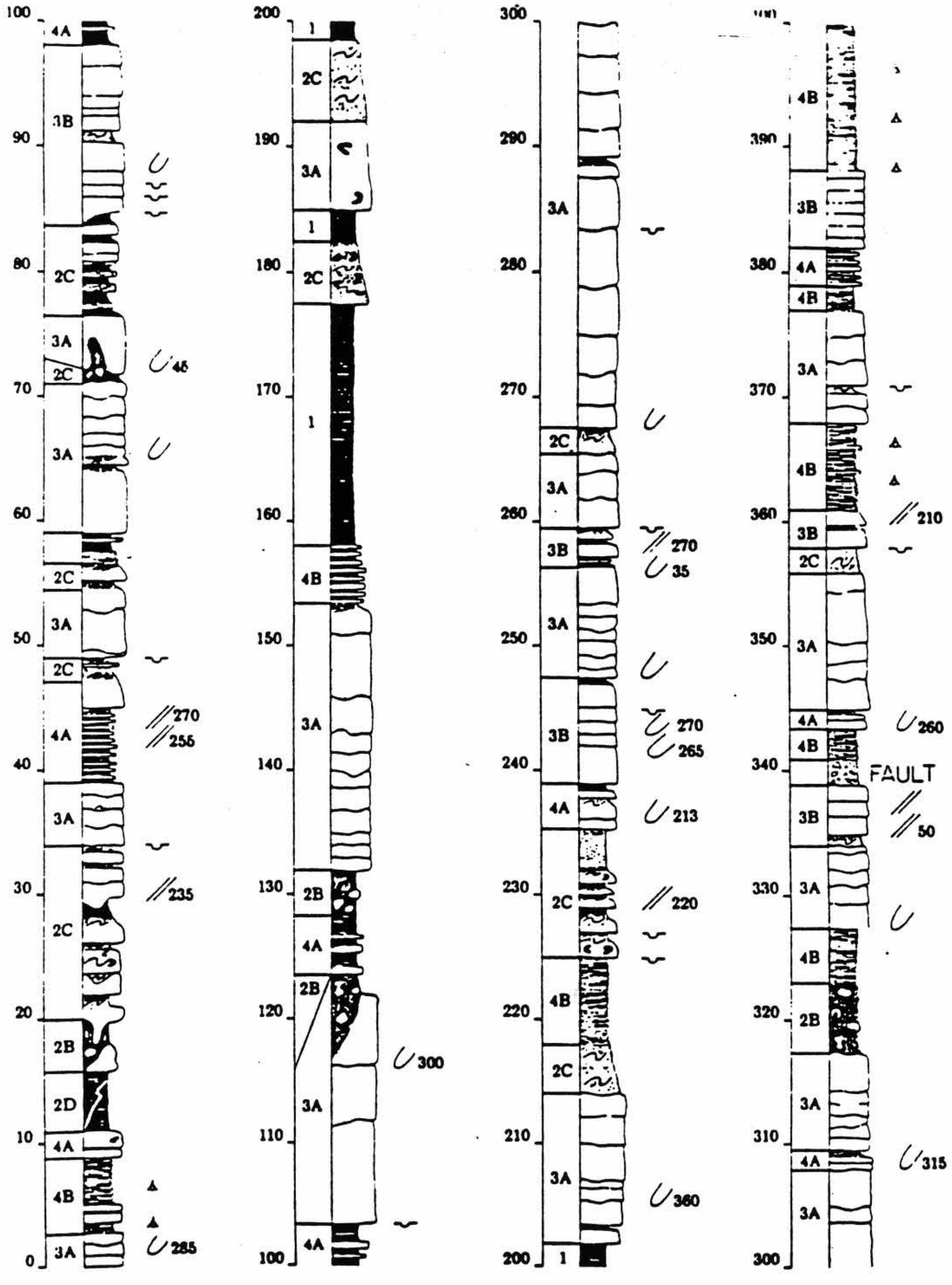


Figure 140. DeRoche Ridge Road Cut. Stratigraphic section of basal Jackfork along Arkansas Highway 7, just south of the dike, sec. 12, T6S R20W. Section measured from base of Jackfork Formation. (from Morris, 1974).

(Description by Jordan, et al., 1991)

STOP 20

Hollywood Quarry, North of Hollywood, Arkansas

The purpose of this stop is to view the sedimentology and bedding characteristics of laterally discontinuous, lenticular and laterally continuous, sheet-like facies, and provide an opportunity to observe and understand a complex three-dimensional "guts of a reservoir" perspective in a deep-water sequence having a structurally complex overprint.

Hollywood Quarry, located about 10 miles west of the DeGray Lake Spillway, offers an excellent opportunity to study a section through a sequence of massive to thickly-bedded upper Jackfork sandstones roughly equivalent in stratigraphic position to units in and between the upper part of the DeGray Lake Spillway and Murray Quarry sections further to the east. The units have been interpreted by Stone and McFarland (1981) as having been deposited in middle submarine fan channels derived from the south-southeast.

This quarry exposes both lenticular and sheet-like turbidite strata; in addition, some faults cut through the quarry. A logging exercise was designed to develop a scaled-down representation of well logs and spacing patterns as might occur in a subsurface oil or gas field. With such a representation, it is possible to demonstrate potential correlation pitfalls which might be encountered in fields with these types of geological complexities.

Bedding strikes uniformly N20-30E and dips 10-15 degrees west. The rocks are cut by a number of minor high-angle faults, most trending east-west to northeast-southwest (Fig. 141). The exposed section totals 140 feet thick (Fig. 142).

The lower 88 feet of section is a lenticular, fining- and thinning-upward channel-fill sandstone sequence (Fig. 143) that rests on shale. The base is a 5-foot-thick layer of quartz-granule conglomerate containing abundant clasts of deformed shale and mudstone ripped up from the underlying unit. Granule conglomerate occurs throughout the lower 33 feet of the channel fill interbedded with poorly-stratified, granule-bearing medium- to coarse-grained sandstone. Large internal channels are locally defined by lenticular conglomerate layers. The sandstones are generally massive, although some show well-developed water-escape structure. Traction structures are absent, suggesting that most of the sediment was deposited by suspended-load fallout from high-density turbidity currents.

The upper 55 feet of the channel-fill sequence consists of well stratified, thick to medium-bedded, predominantly

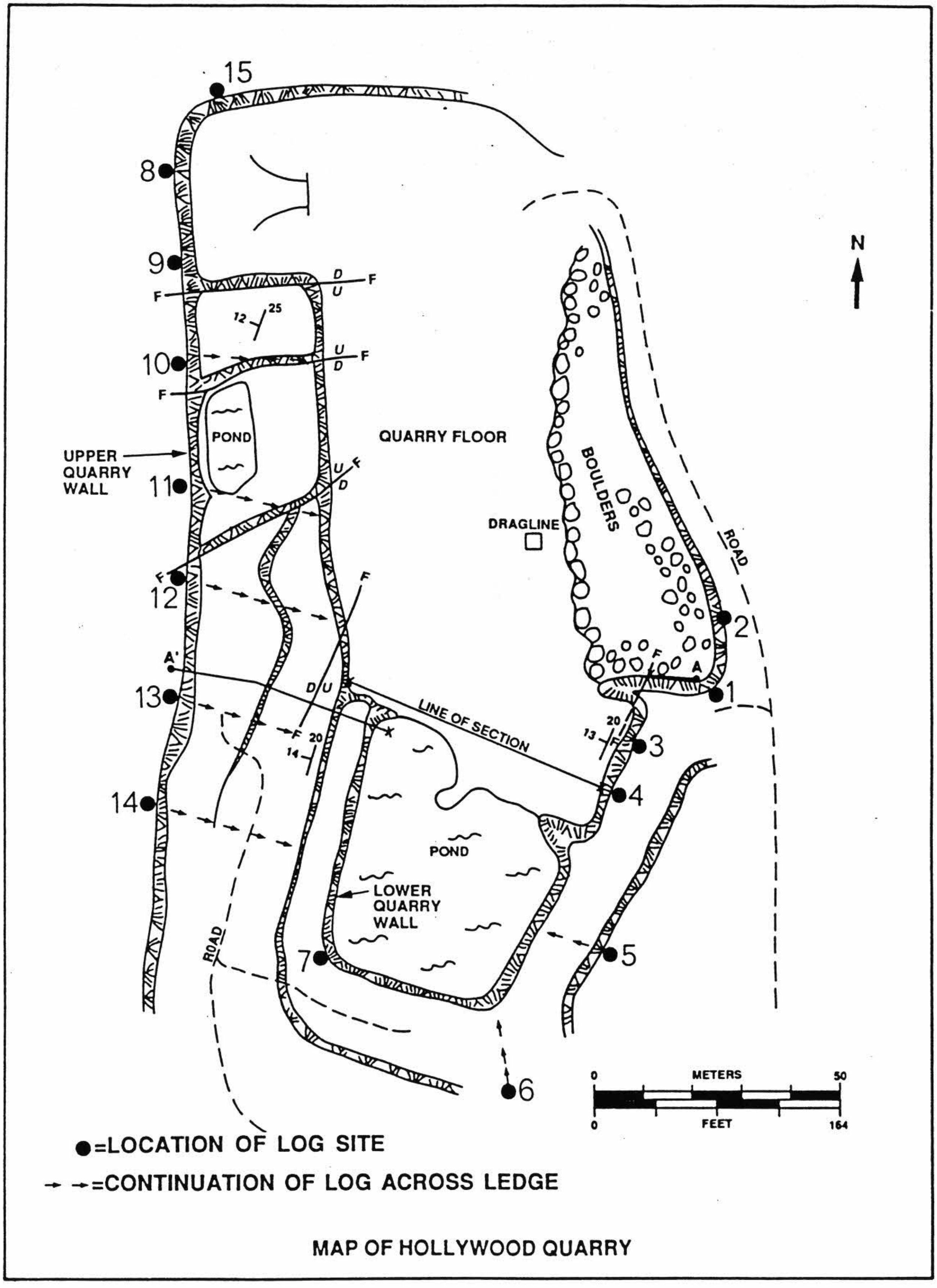


Figure 141. General sketch map of Hollywood Quarry showing line of section (Fig. 142) and principal faults cutting the Jackfork sequence. Location of logging sites shown.

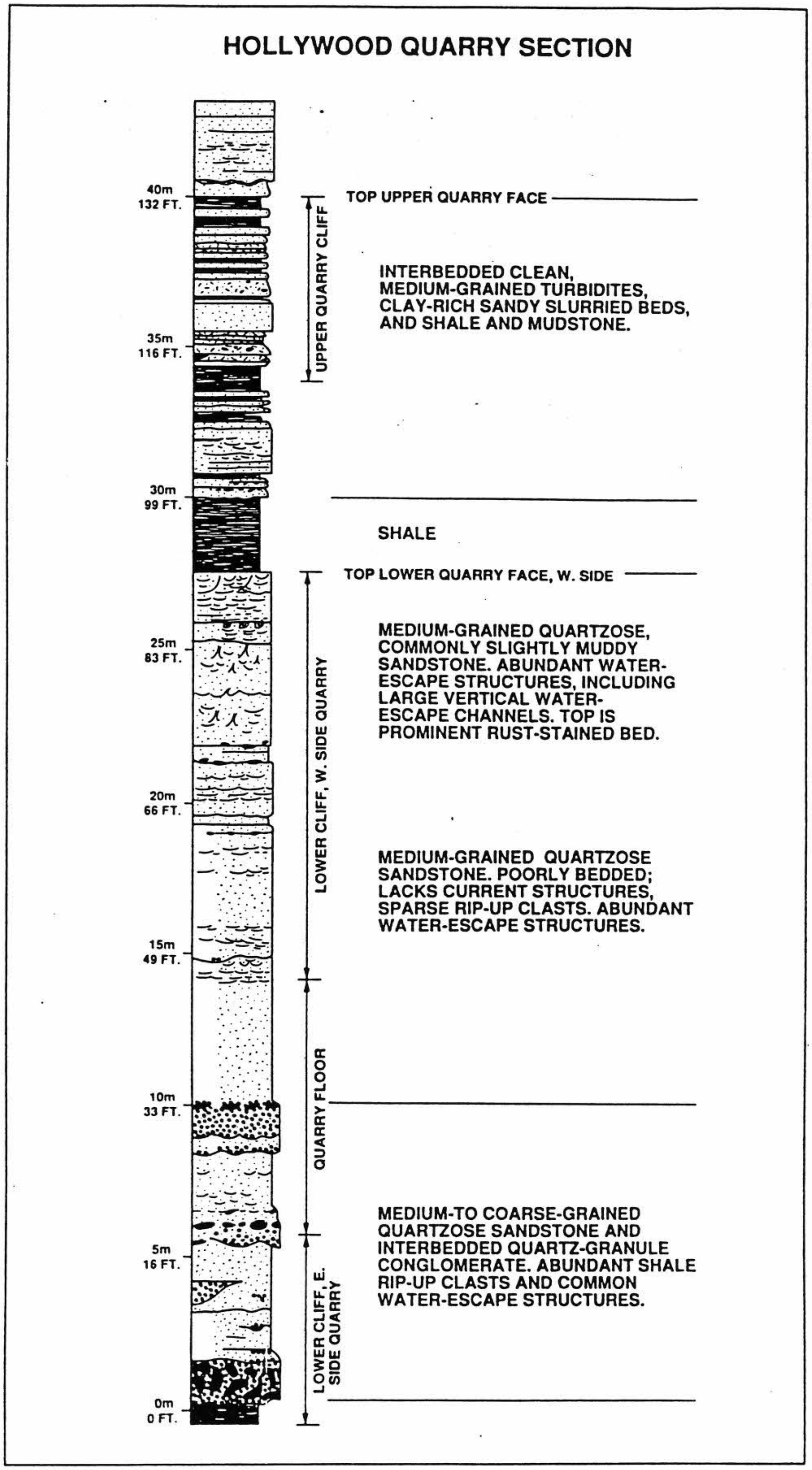


Figure 142. Stratigraphic column of the Jackfork sequence exposed in the Hollywood Quarry, Arkansas. Line of section is located on Figure

HOLLYWOOD QUARRY
SOUTHEAST WALL

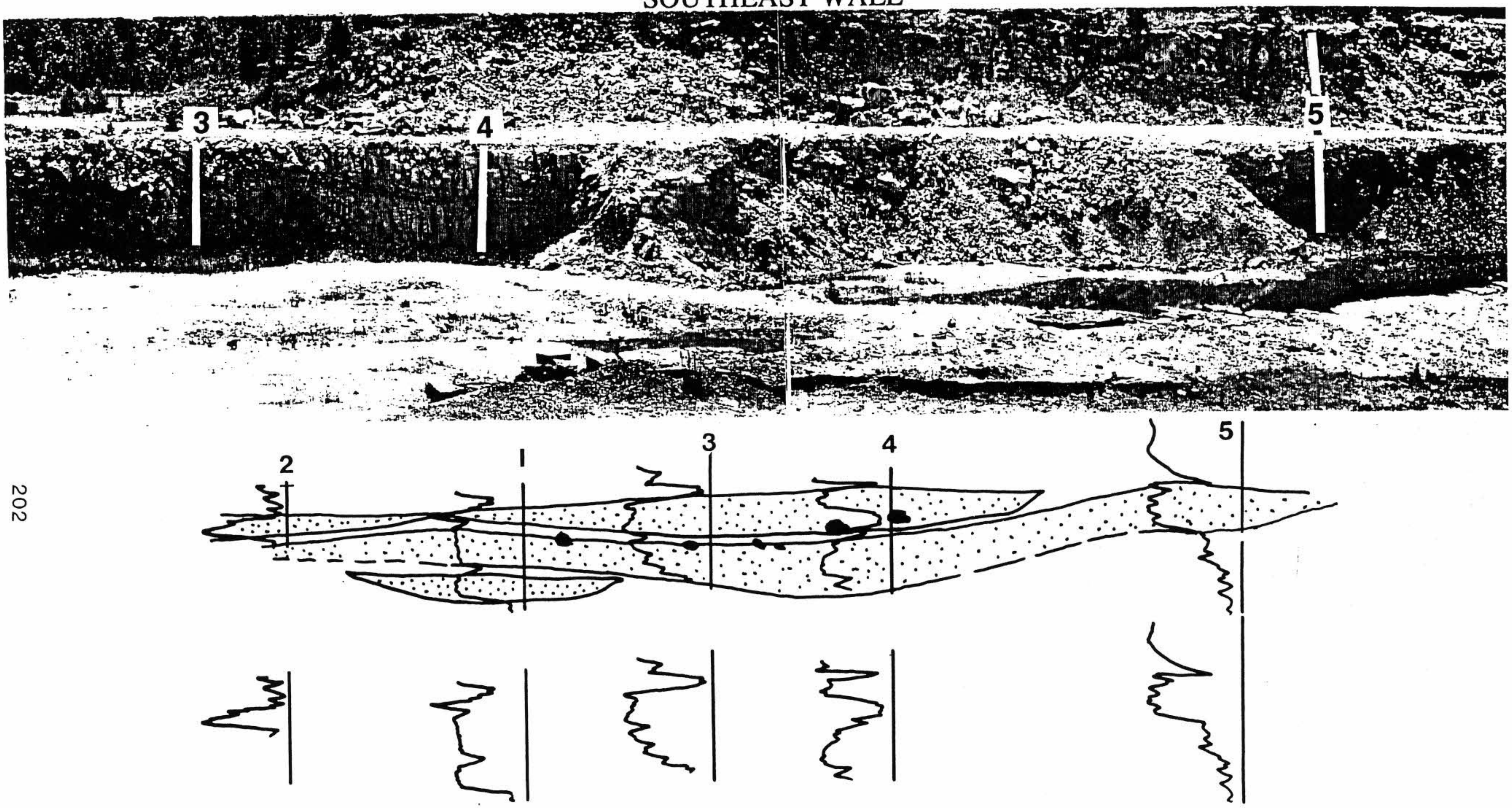
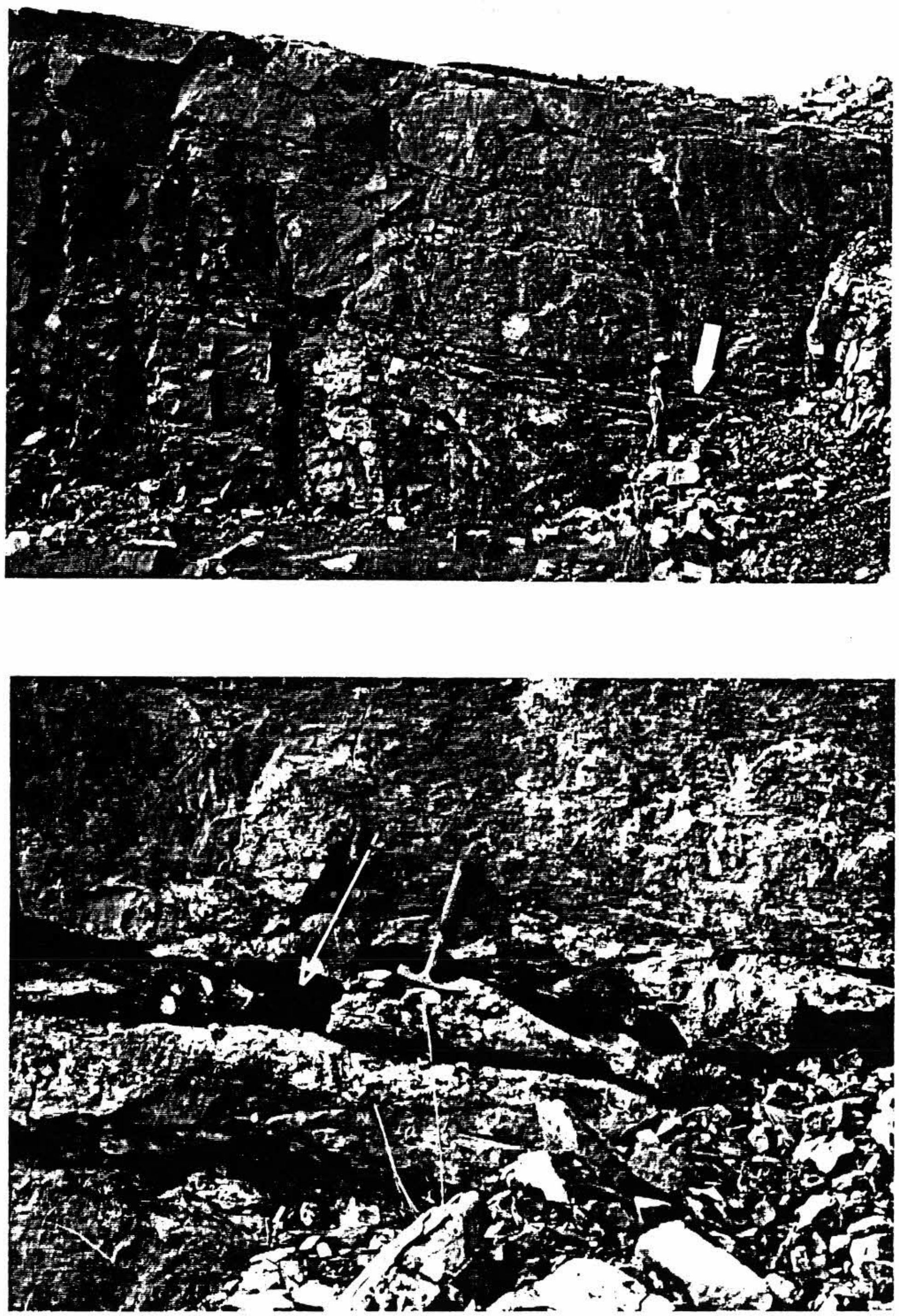


Figure 143. Lower 88 feet of section at Hollywood Quarry consisting of several channels, the lowest one resting on shale. Clasts of deformed shale and mudstone ripped up from the underlying unit are locally present.

medium-grained quartzose sandstone. Sandstone beds average 2 to 4 feet thick (Fig. 144A-B) and show abundant water-escape structures and soft-sediment deformation fea-tures. Near the top, the sandstones become slightly muddy and one or two slurried units are interbedded with the cleaner sandstones. The top of the channel sequence is marked by a prominent rust-stained unit containing large, vertical water-escape channels.

The top 52 feet of the quarry section (Fig. 145A) consists of interbedded units of predominantly medium-grained sandstone and shale. Sandstone units include layers of relatively clean, light gray, quartzose sandstone, from less than an inch to several feet thick, deposited by turbidity currents. Thicker turbidites are dominated by water-escape structures, especially dish structures, and many show ripped-up shale clasts. Only a few of the thinnest turbidites show well-developed traction structures, such as flat lamination and cross-lamination (Fig. 145B).

A number of sandy units in the upper part of the Hollywood Quarry section are composed of very dirty, brown to gray, poorly indurated sandstone. These beds range from 4 to about 24 inches thick and contain a high proportion of mud, commonly 15% to 35%. They lack current and water escape structures but show soft-sediment deformation features and abundant mudstone clasts. These units appear to represent dilute debris flows or slightly cohesive slurries that may have formed as small failures on mud-covered submarine slopes. They are clearly distinct from the cleaner turbidite sands. Figures 146 and 147 show the gamma-ray logs obtained by the logging truck and scintillometer methods (see Figure 141 for locations of logging sites in the quarry). The beds strike about north 25 degrees east and dip 10 degrees to the northwest so that the beds on the east wall of the quarry dip beneath those on the west wall. The upper part of the southern half of the west wall comprises laterally discontinuous strata. The northern half of the west wall contains more laterally continuous, but highly faulted beds. The east wall contains thick, lenticular sandstones and pebbly sandstones which are probably correlative with the pebbly sandstones at the top of DeGray Lake Spillway. In this example, if only logs were available, the correlations shown in Figure 148 most likely would not have been made. Considering that these logs are relatively closely-spaced, this exercise demonstrates some of the uncertainty in well log correlations that might be anticipated in oil and gas fields of complex facies architecture and structure. Coring of development wells for detailed sedimentologic and stratigraphic analysis, utilizing Formation microscanner (FMS) or dipmeter logs for recognizing thin beds, and utilizing 3D seismic surveys for structural and stratigraphic interpretation, in addition to well testing, would reduce this uncertainty in such situations and aid in the understanding of interval heterogeneity.



A

B

Figure 144. Lower cliff, west side of Hollywood quarry. (A) Well-stratified, thick- to medium-bedded, medium-grained quartzose sandstone with sharp basal contact. (B) Close-up of area shown at arrow in A, with large mudstone clasts (arrow).

Figure 145. Upper part of section on the west cliff of the quarry. (A) The top 52 feet of the quarry section consists of interbedded units of predominantly medium-grained sandstone and shale. Sandstone units include relatively clean quartzose sandstone, from less than an inch to several feet thick, deposited by turbidity currents. (B) Well-developed traction structure (cross-lamination) within thin turbidite. Underlying bed is a clay-rich, sandy slurry bed.

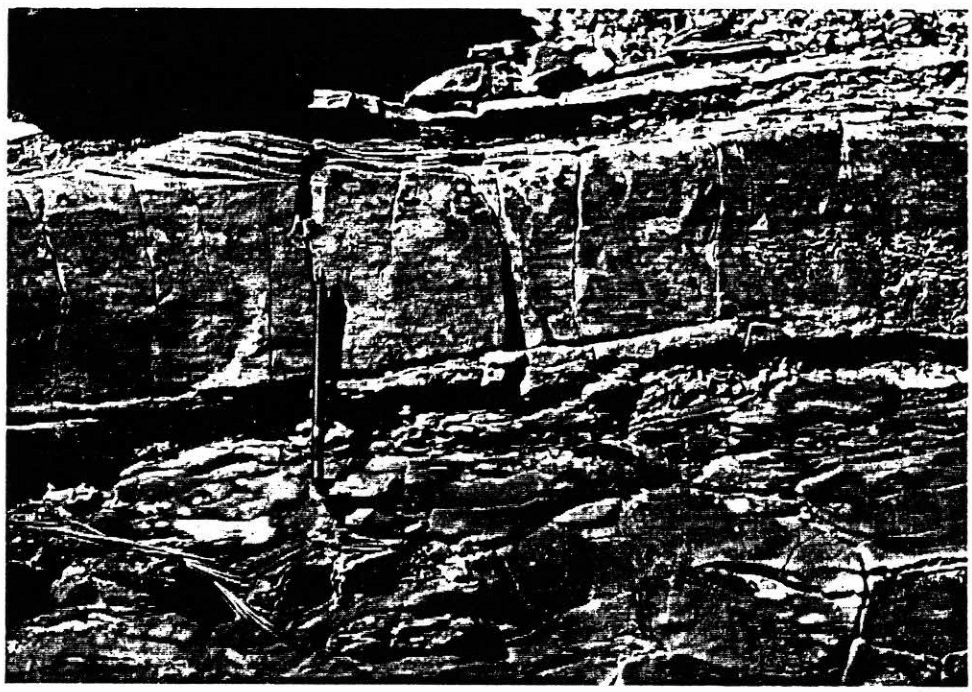


Figure 145B. (caption on previous page).

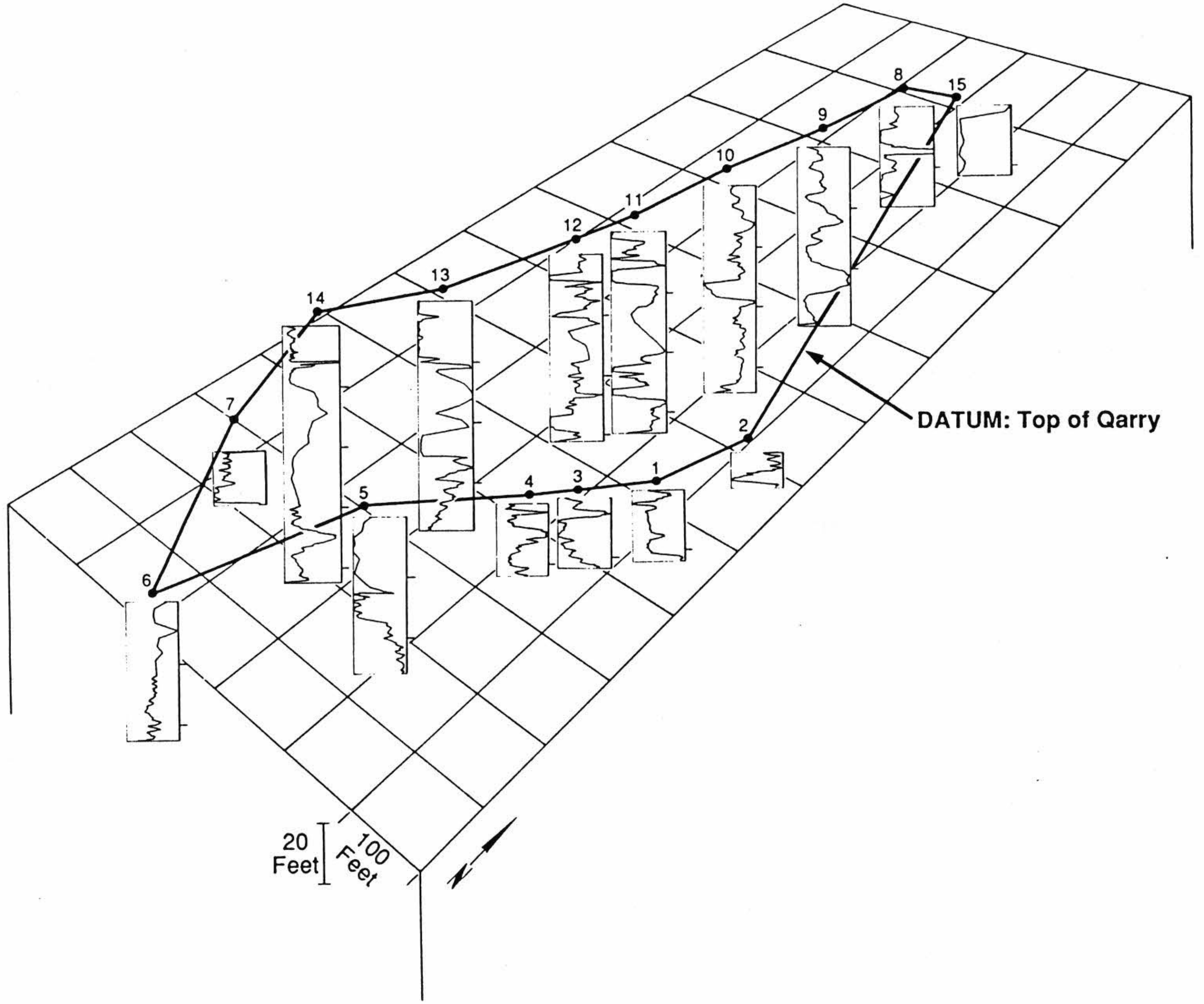


Figure 146. Location of the fifteen gamma-ray logs obtained with the logging truck at Hollywood Quarry. Heavy line connects the top of the cliff wall.

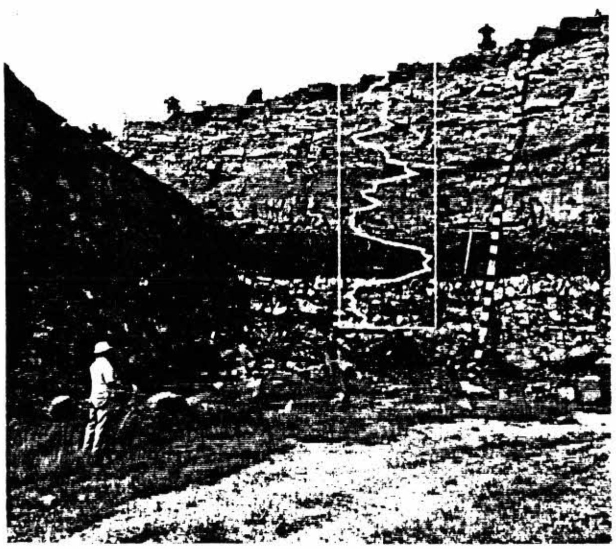


Figure 147. Typical gamma-ray log (log site #9; see Figs. 141 and 146). Gamma-ray sonde (large arrow) being raised from the base of the west wall of Hollywood Quarry by a cable (small arrow) attached to the logging truck (not in view). This logging run produced the gamma-ray log that is superimposed upon the quarry wall in the photograph. Note the scale marked in one foot increments.

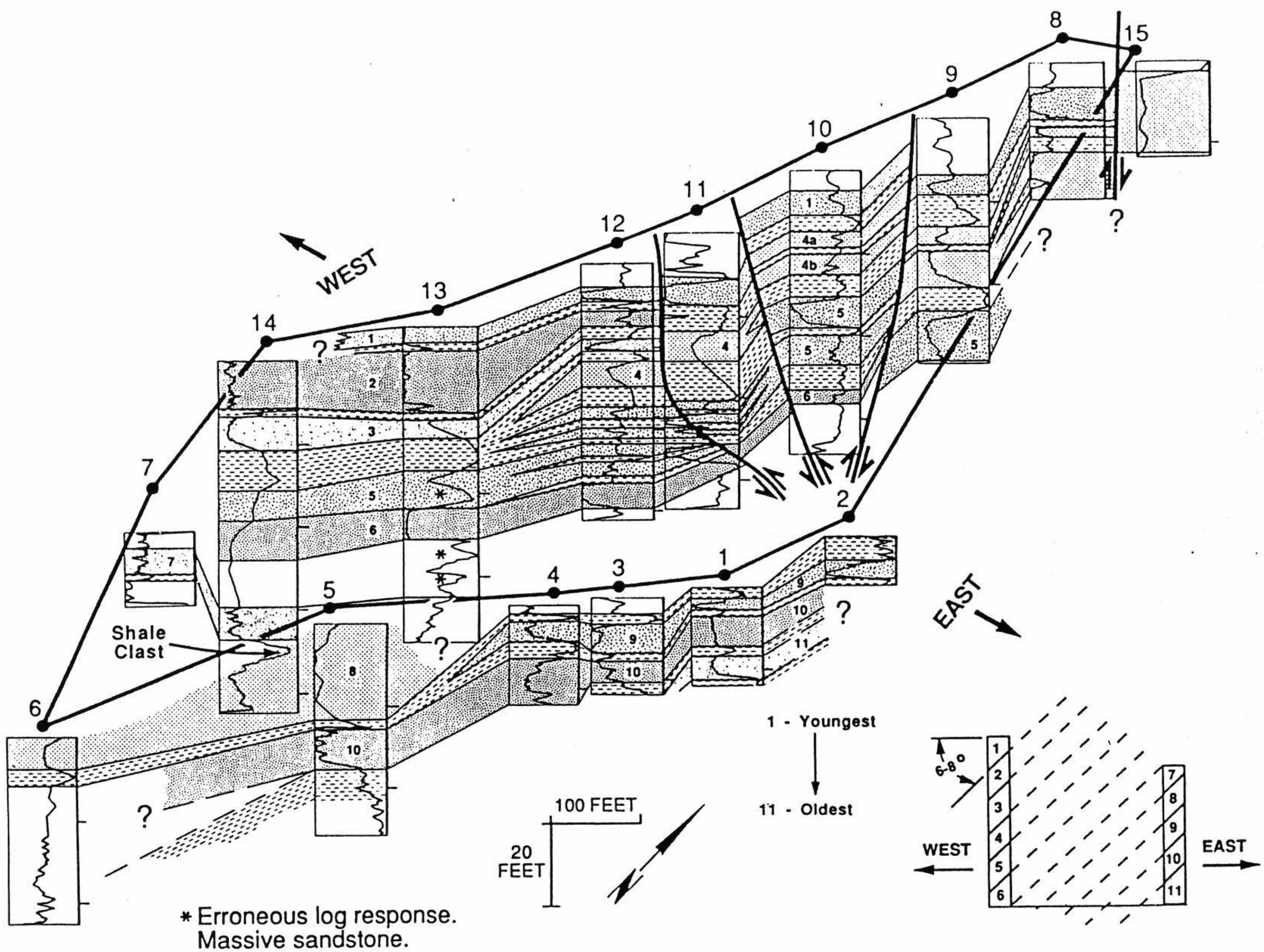


Figure 148. Gamma-ray log correlations at Hollywood Quarry.

FIELD TRIP DESCRIPTION -- DAY THREE

The area in and around Hot Springs National Park offers many fine exposures of Cambrian through lower Pennsylvanian rocks along with some recent formations at the thermal springs. These strata are often steeply inclined to overturned, with many tight folds and numerous thrust faults formed by the late Paleozoic orogeny that uplifted the Ouachita Mountains. These rocks may also contain quartz veins, small Cretaceous-age igneous dikes, and both hot and cold springs. While there are stops in the Ordovician Bigfork Chert and the Devonian-Mississippian Arkansas Novaculite, our primary interest is directed towards the outer fan sequences in the Mississippian Stanley Formation and the slope to mid-fan clastics in the Hot Springs Sandstone Member at the base of the Stanley Formation and in the Ordovician Crystal Mountain and Blakely Formations. It just so happens that one of these stops is located at the largest commercial clear quartz crystal mine in North America (Ron Coleman's).

(Description from Bouma, 1993; and Stone, 1993)

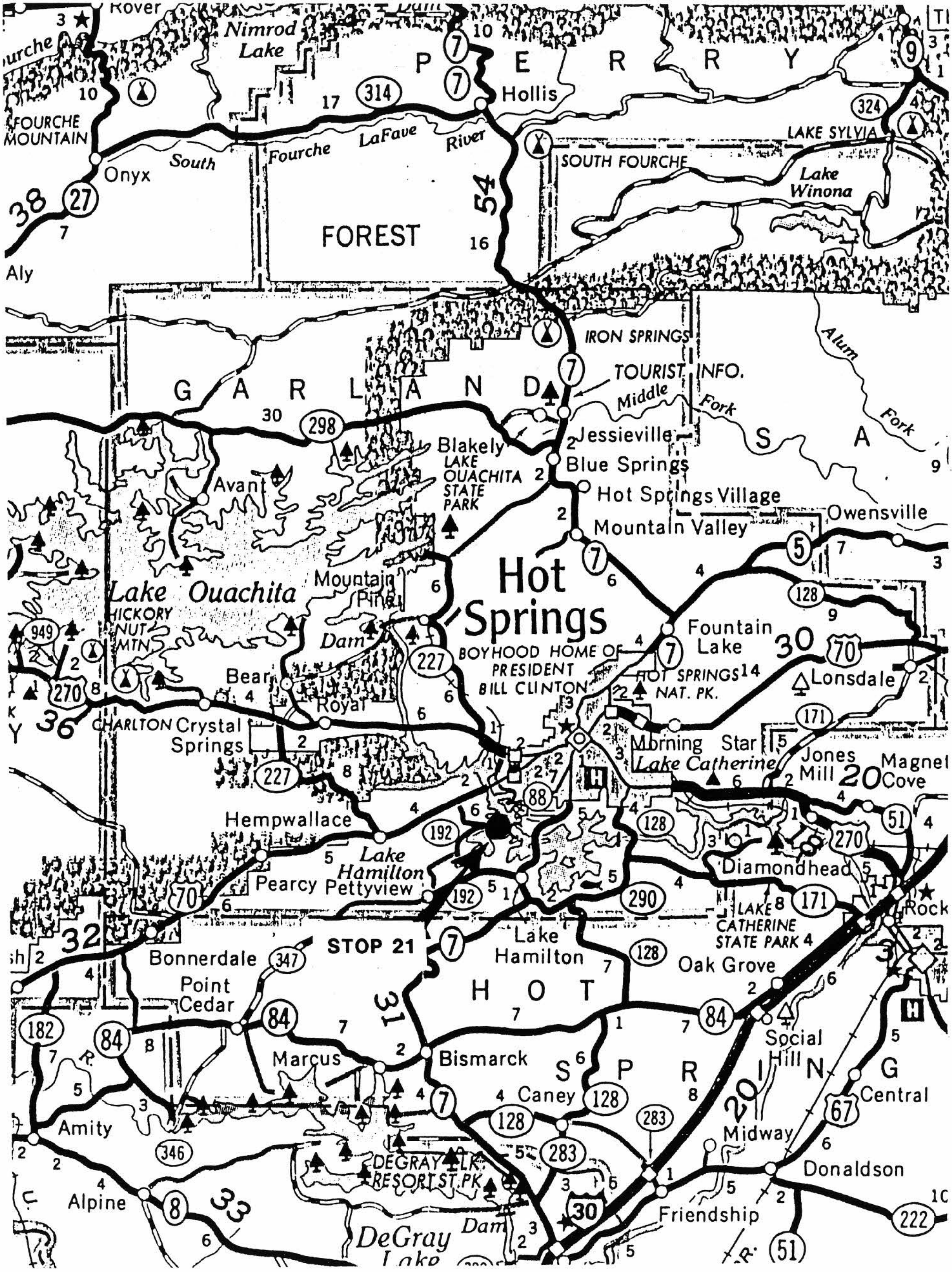
STOP 21

Bypass Around Hot Springs

The new bypass around part of Hot Springs shows some excellent outcrops in the shale-rich Mississippian Stanley Formation which stratigraphically underlies the Jackfork Group. The outcrops are comprised of those intervals of the middle Stanley Formation that have dominant sandstone layers that is typical for this group. The freshness makes it possible to look both at the overall package and at details.

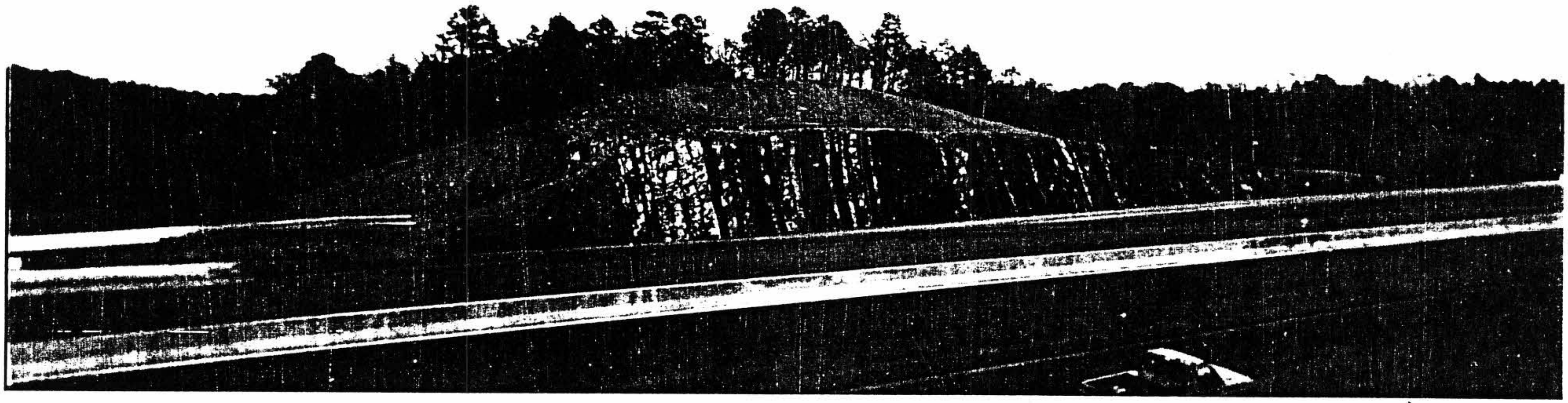
Parallel roadcuts with steeply southward dipping subgraywacke silty sandstones (gray) and shales (black) can be found at several locations. The one at the Lakeshore Drive crossing is by far the best because it contains four parallel outcrops, two on the sides of the bypass and two much closer together along the adjoining access roads.

No studies have yet been conducted on these outcrops and therefore no specifics can be provided. The reason for the stop is that it presents a series with a much lower sand/shale ratio. In spite of that aspect it does not differ noticeably from the other, younger, submarine fan deposits. It is likely that this sequence was derived from an orogenic landmass to the south that spread large submarine fan (lobe and channel) sediments far to the north and northeast in the early forming Ouachita basin. These deposits are presently thought to represent outer fan lobe deposits. Notice the hydrothermal milky quartz veins with greenish chlorite that fill fractures in the sandstones.

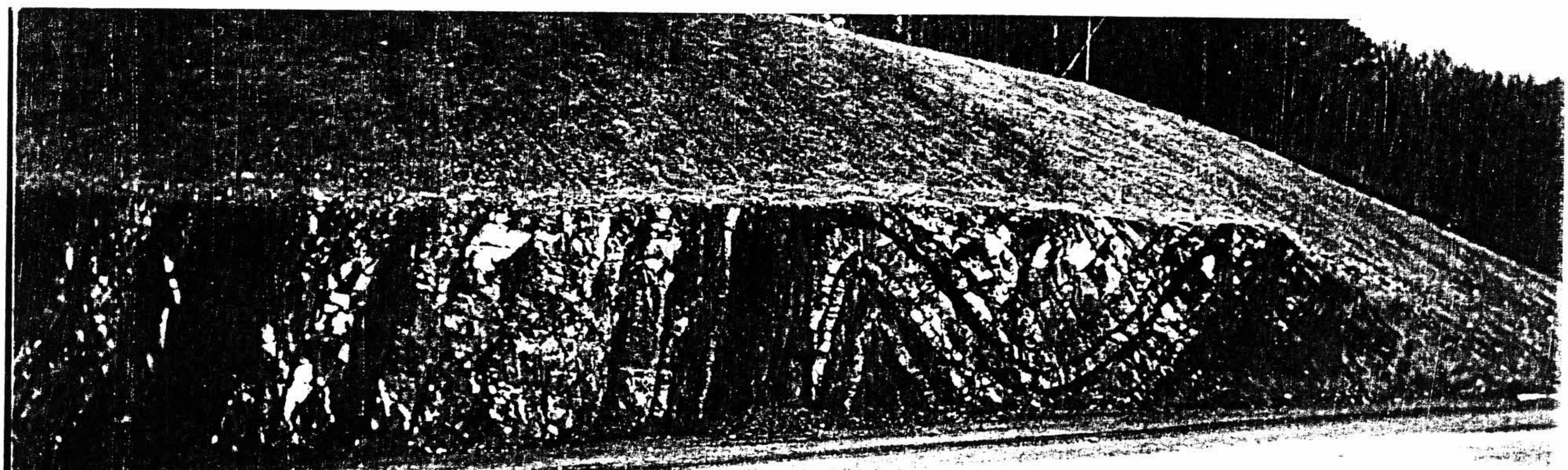


Nimrod Lake
PERRY
Hollis
FOURCHE MOUNTAIN
Onyx
South Fourche LaFave River
FOREST
SOUTH FOURCHE
LAKE SYLVIA
Lake Winona
Aly
IRON SPRINGS
TOURIST INFO.
GARLAND
Middle Fork
Alum Fork
SA
Jessieville
Blakely
LAKE OUACHITA STATE PARK
Blue Springs
Hot Springs Village
Avant
Mountain Valley
Owensville
Lake Ouachita
Mountain Pine
Hot Springs
HICKORY NUT MTN.
Dam
Fountain Lake
BOYHOOD HOME OF PRESIDENT BILL CLINTON
HOT SPRINGS NAT. PK.
Lonsdale
Bear
Royal
CHARLTON
Crystal Springs
Morning Star
Lake Catherine
Jones Mill
Magnet Cove
Hempwallace
Lake Hamilton
Diamondhead
Pearcy
Pettyview
LAKE CATHERINE STATE PARK
Rock
STOP 21
Lake Hamilton
Bonnerdale
Oak Grove
Point Cedar
HOT
Social Hill
Marcus
Bismarck
SPRING
Caney
Central
Amity
Midway
DEGRAY LK. RESORT ST. PK.
Donaldson
Alpine
Friendship
Dam
DeGray Lake

Figure 149



This is an about 1,000' sequence of sandstone and shale in the middle part of the Stanley Group on the new Hot Springs highway bypass.



Complex structures and small thrust faults with midky quartz veins at the northwest end of the new Hot Springs bypass exposure.

Figure 150

(Description from Stone, Howard, and Haley, 1986)

STOP 22

Hot Springs National Park Tour

This is a brief walking tour of the hot springs located off Central Avenue in downtown Hot Springs, Arkansas (Fig. 151). The Grand Promenade Trail provides convenient access to hot springs, and exposures of tufa rock and sandstone. Nearby there is a fine museum at the Hot Springs National Park Headquarters. This was the first Federal Reservation (1832) and became the eighteenth National Park (1921) in the United States.

The U. S. Park Service regulates the use of the water from the world famous hot springs which issue from fractures and joints in the Hot Spring Sandstone along the base and slope of East Mountain.

East Mountain is a westward plunging, faulted, southward overturned anticlinal ridge of the Zigzag Mountain subprovince. Early American Indian tribes, Spanish conquistadors, early settlers and modern man have all exploited the therapeutic properties of the springs. "Tah-ne-co", the place of the hot waters, was the Indian name for the site. About 50 of the original 71 springs produce hot water. According to J. K. Haywood and W. H. Weed (1902) the daily flow aggregates 850,000 gallons with the largest spring yielding a little over 200,000 gallons. The temperature range is from 95.4° to over 147° F. The hot spring water is slightly radioactive, apparently caused by radon gas. A soil-and-vegetation-covered gray calcareous tufa formed by the hot springs covers an area of 20 acres and in places is 6 to 8 feet thick. Measurements of the hot springs' flow to the central collection reservoir have been made periodically since 1970 by the U. S. Geological Survey. These measurements indicate a range in spring flow of 750,000 to 950,000 gpd with an average flow of about 825,000 gpd.

The tritium and carbon 14 analyses of the spring water indicate a mixture of a very small amount of water less than 20 years old and a preponderance of water about 4,400 years old.

Bedinger et al. (1979) state that the geochemical data, flow measurements, and geological structure of the region support the concept that virtually all the hot springs water is of local meteoric origin. Recharge to the hot springs artesian flow system is by infiltration of rainfall in the outcrop areas of the Bigfork Chert and the Arkansas Novaculite. The water moves slowly to depth where it is heated by contact with rocks of high temperature. Highly permeable zones related to jointing or faulting collect the heated water in the aquifer and provide avenues for water to travel rapidly to the surface.

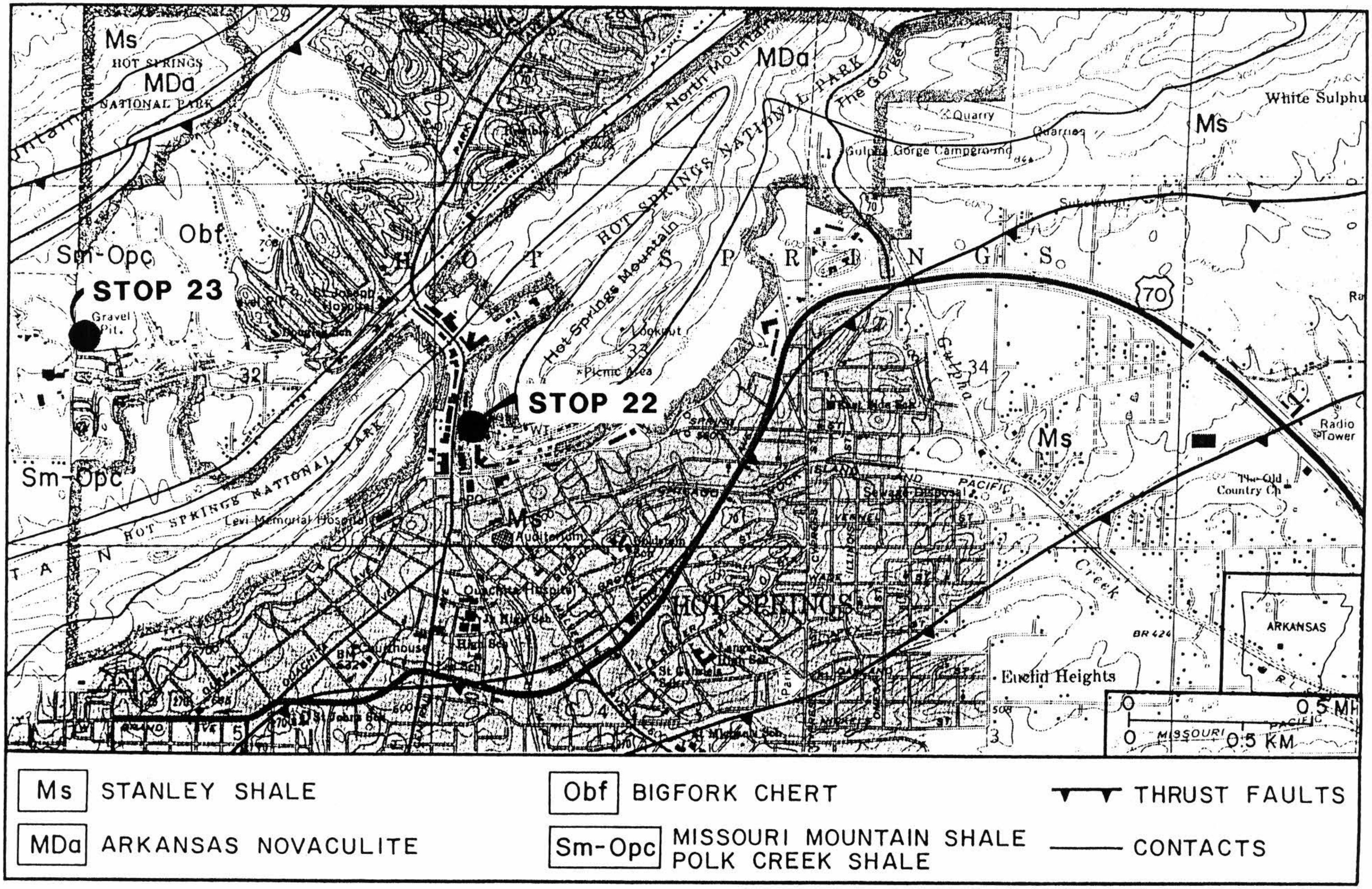


Figure 151. Map showing locations of Stops **22 and 23** in the Hot Springs area.

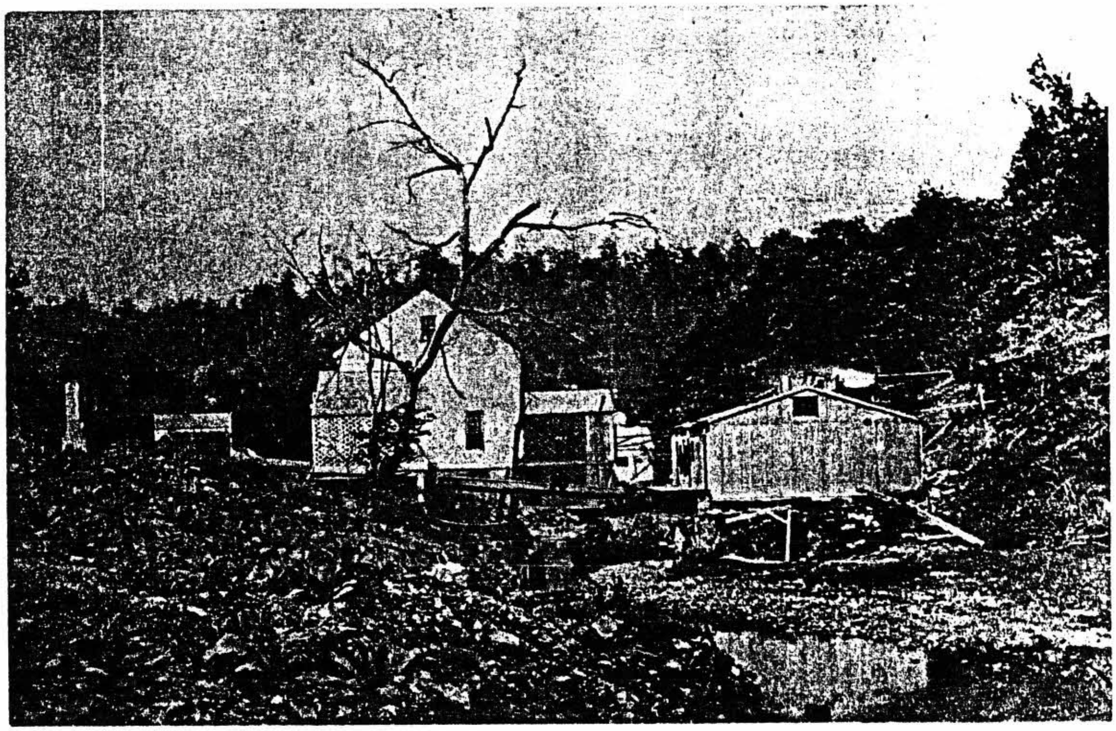


Figure 152. The celebrated old Hale Bath House, east of and adjoining the present site of the Arlington Hotel, Hot Springs.

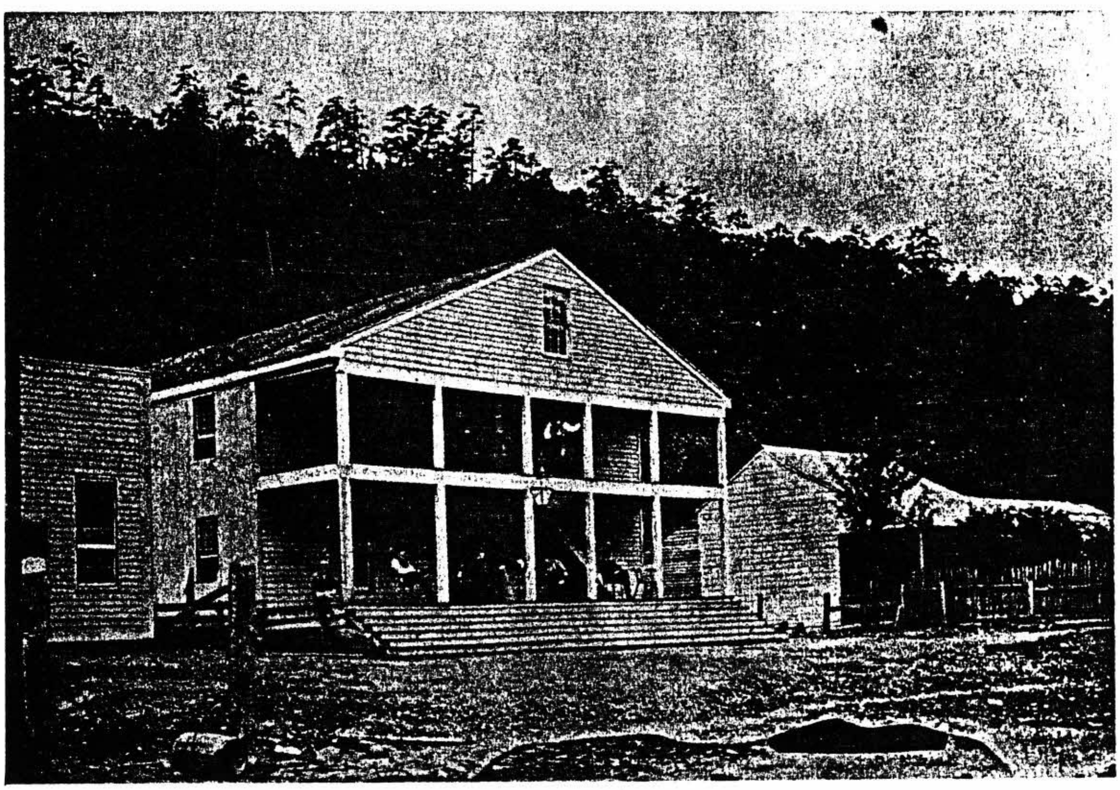


Figure 153. The Earl Bath House at Hot Springs, opposite Hale Bath House.

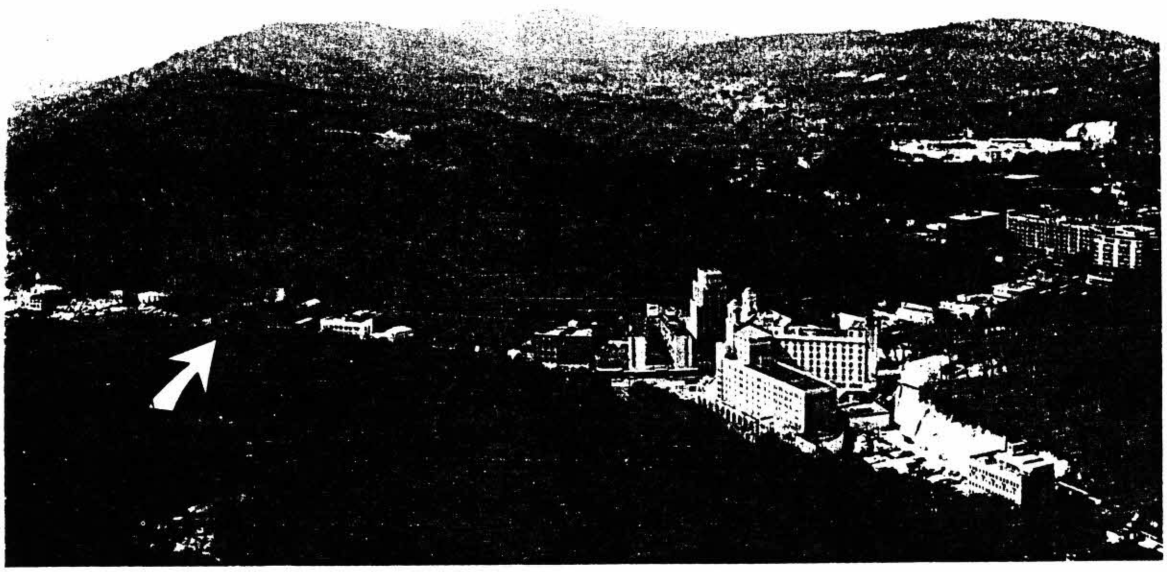


FIGURE 154. Downtown Hot Springs as viewed westward from Hot Springs National Park tower. The hot springs (arrow) issue from fractures in the Hot Springs Sandstone along nose of small southwestward plunging anticline forming hill in nearground. More distant ridges are underlain by the Arkansas Novaculite and outline part of the larger Hot Springs anticline.

(Description from Stone, Howard, and haley, 1986)

STOP 23

City Quarry in Bigfork Chert at Hot Springs

Folded and faulted rocks in the middle part of the Bigfork Chert are exposed in a quarry north of Weyerhaeuser office on Whittington Avenue (Fig. 155). The quarry is near the axis of the south-westward plunging Hot Springs anticline. Massive novaculite of the Lower Division (Devonian) of the Arkansas Novaculite forms the ridges to the north and south.

In the quarry, the Bigfork consists of very thin-bedded and often graded, calcareous (often decalcified) siltstone, gray chert, and minor beds or laminations of siliceous to carbonaceous shale. The basal silty part of each interbedded sequence likely represents minor influxes of fine clastics and bioclastics transported into the Ouachita trough by turbidity and bottom currents and the overlying cherts and siliceous shales represent slowly deposited deep-water pelagic accumulations. Stylolites are often present in the calcareous siltstones and indicate a significant removal and thinning of the section.

The Bigfork is complexly folded in the quarry with a large box fold and associated kink, chevron, and buckle folds inclined both to the south and north. Most of the strata dip to the north, indicating a dominant southward vergence. North-dipping cleavage often refracts across the cherts and calcareous siltstones. Flowage of rock can been seen in some of the tight fold hinges. Several apparently small reverse faults dissect the rocks and they often contain gouge (dickite?) impregnated chert breccia.

The Bigfork Chert is the most reliable aquifer in the Ouachita Mountains. Clear or small chalybeate (iron-rich) springs are often present in the basal outcrops. The water moves along joints, fractures, and bedding planes in the thin-bedded basal sequence. The Bigfork Chert produces some oil and gas from small fields in the northern Ouachita Mountains of Oklahoma.

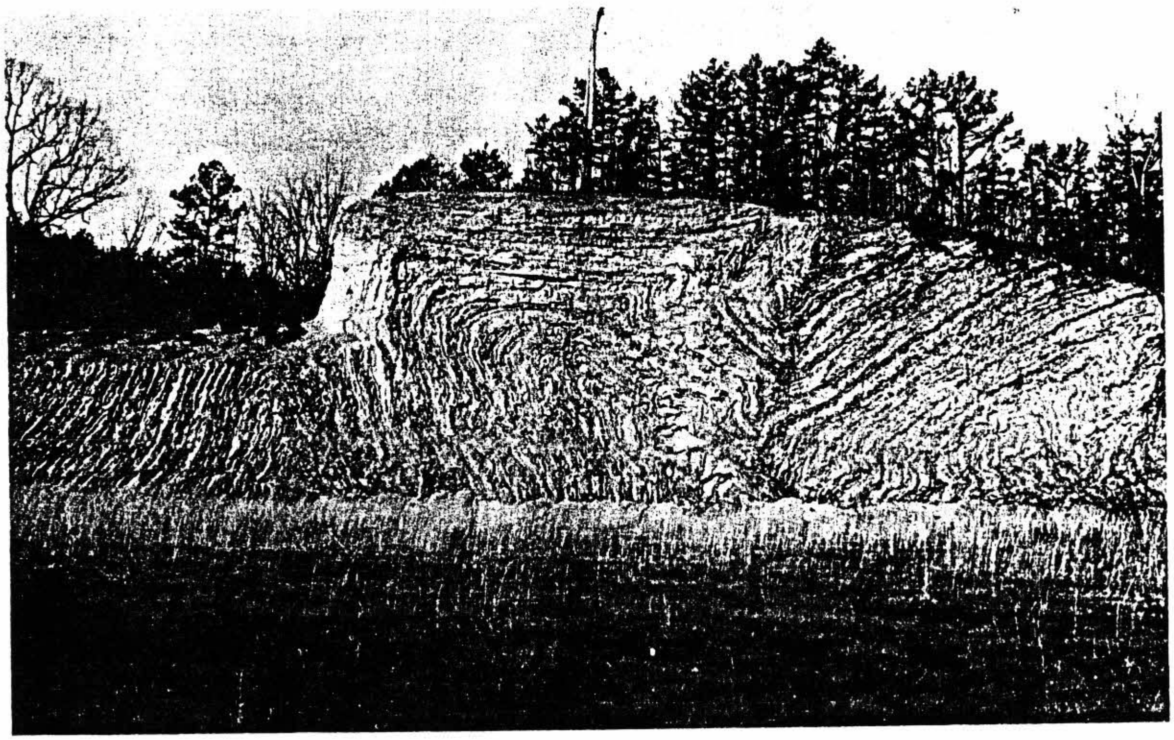


FIGURE 155. Classic box and kink folds in cherts and siltstones of the Bigfork Chert at a quarry north of Whittington Avenue.

(Description mostly from Stone, Haley, and Viele, 1973)

STOP 24

Hot Springs Sandstone at Hot Springs Water Works

We wish to express appreciation to Mr. Alford of the Hot Springs Water Works for granting permission to examine this fine exposure!

The first rock encountered on the eastern edge of the small reservoir (Fig. 156) is an upright section of the Hot Springs Sandstone Member of the basal Stanley Formation. It is about 90 feet in thickness and at the base contains several thick conglomeratic (mostly novaculite clasts) sandstones, overlain by several very clean sandstones which becomes thinner bedded and more silty and shaly towards the top. This unit represents a classic upper(?) submarine fan channel sequence that both thins and fines upward. At a few sites in the area, it contains some early-middle Mississippian shelfal limestone exotics.

Clearly the Hot Springs Sandstone had a source area to the north and northeast (Fig. 157 & 158). It was immediately overlain by orogenic and volcaniclastic facies derived from land masses to the south. The Hot Springs Sandstone Member must represent some dramatic lowering of sea level and regressive deposition in the Ouachita Mountains. Then the overlying very thick sequence of Stanley shales, sandstones, and volcanic tuffs reflect the tremendous basinal subsidence and orogenic spasms that happened immediately afterward. This subsidence and northward progression led to over 45,000 of Carboniferous clastic turbidites in the Ouachita trough prior to eventual emergence in late Pennsylvanian time.

As you proceed along the shoreline of the small reservoir the following sequence of rocks is exposed:

(1) Massive upright, southward-dipping, quartzitic sandstone with some thin conglomerate and shale intervals of the Hot Springs Sandstone Member of the Stanley Shale;

(2) thin black shale at base of the Stanley Shale;

(3) thin beds of tripolitic southward-dipping Upper Division of Arkansas Novaculite;

(4) the shallow northward-dipping shale, siliceous shale and chert of the Middle Division of Arkansas Novaculite which is intensely folded. This is actually part of a large recumbent fold which has been transported to the south. Note shallow northward-dipping axial plane cleavage and flowage of shale from flanks into axes of folds. At the small spillway farther to the north, you can view the Hot Springs Sandstone exposed both above and below the Arkansas Novaculite. A fault truncates the sequence at the upper end of the second small lake.

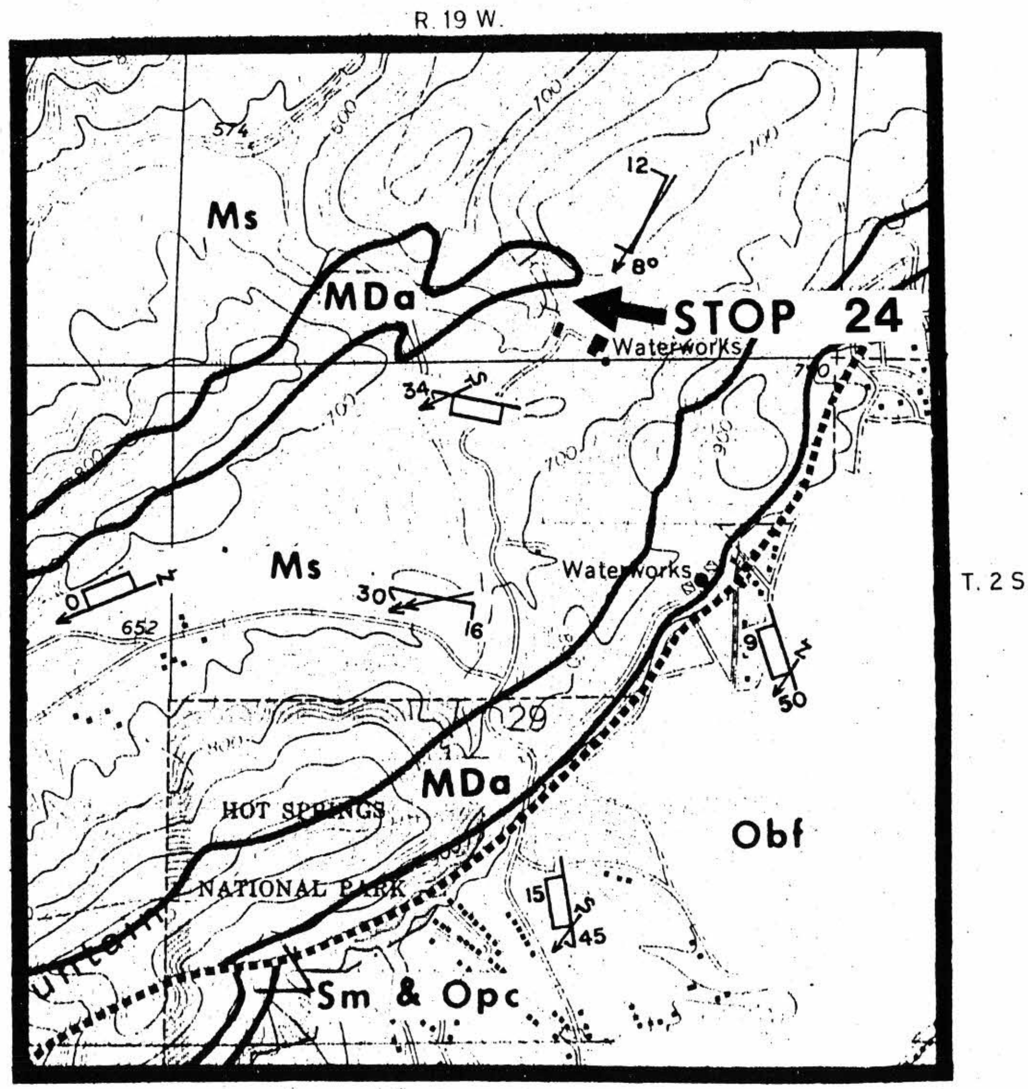
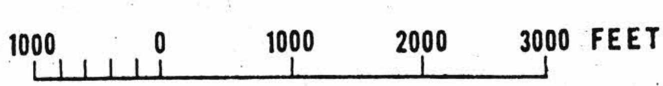


FIGURE 156. GEOLOGIC MAP OF NORTHERN HOT SPRINGS, ARKANSAS-STOP 24.



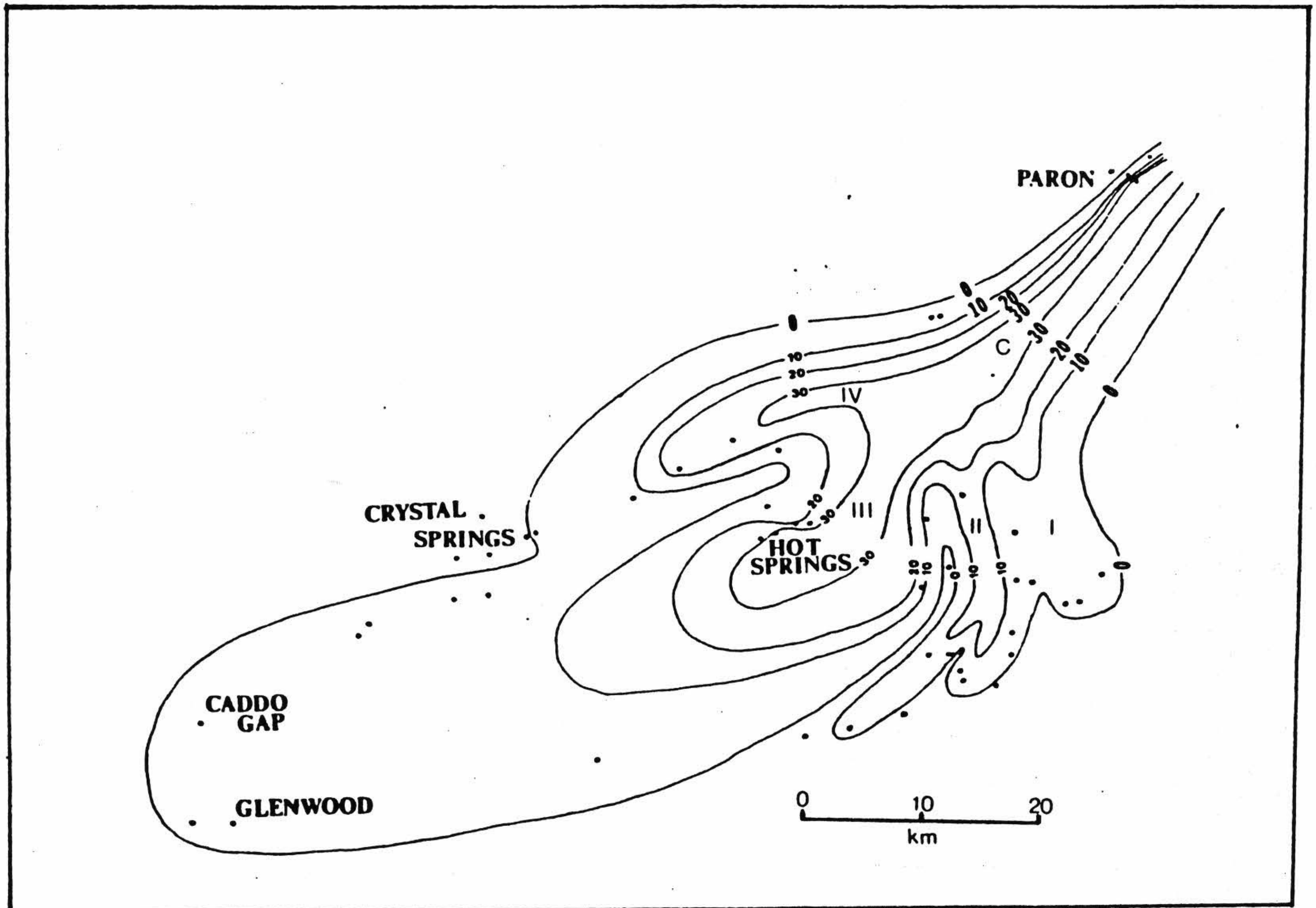


Fig. 157: Net Sandstone isopach of the Hot Springs Sandstone. The contour interval is in meters. (Dickerson, 1986)

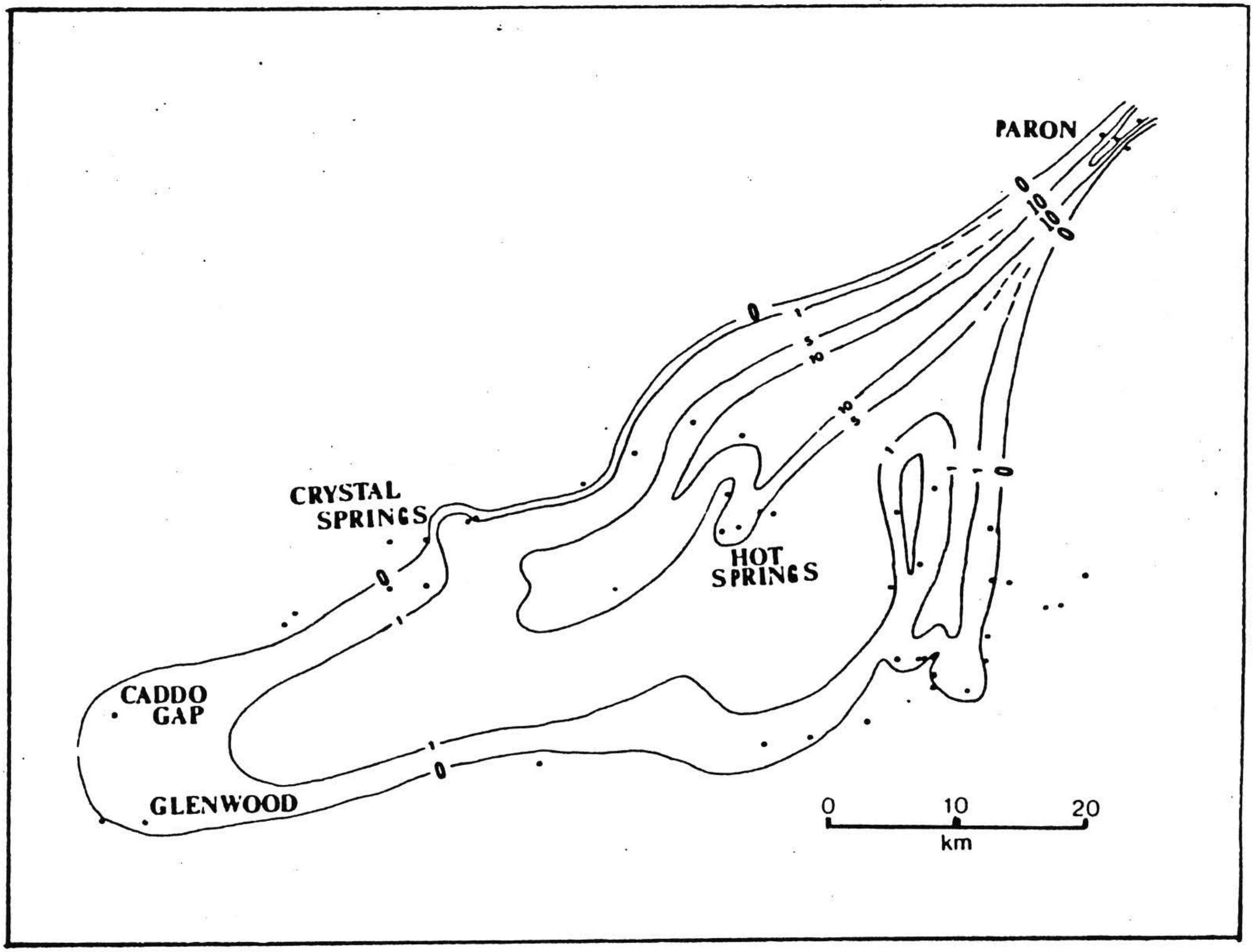


Fig. 158: Net breccia isopach of the Hot Springs Sandstone Note that the contour interval is variable in order to better delineate thickness trends. The contours are in meters. (Dickerson, 1986)

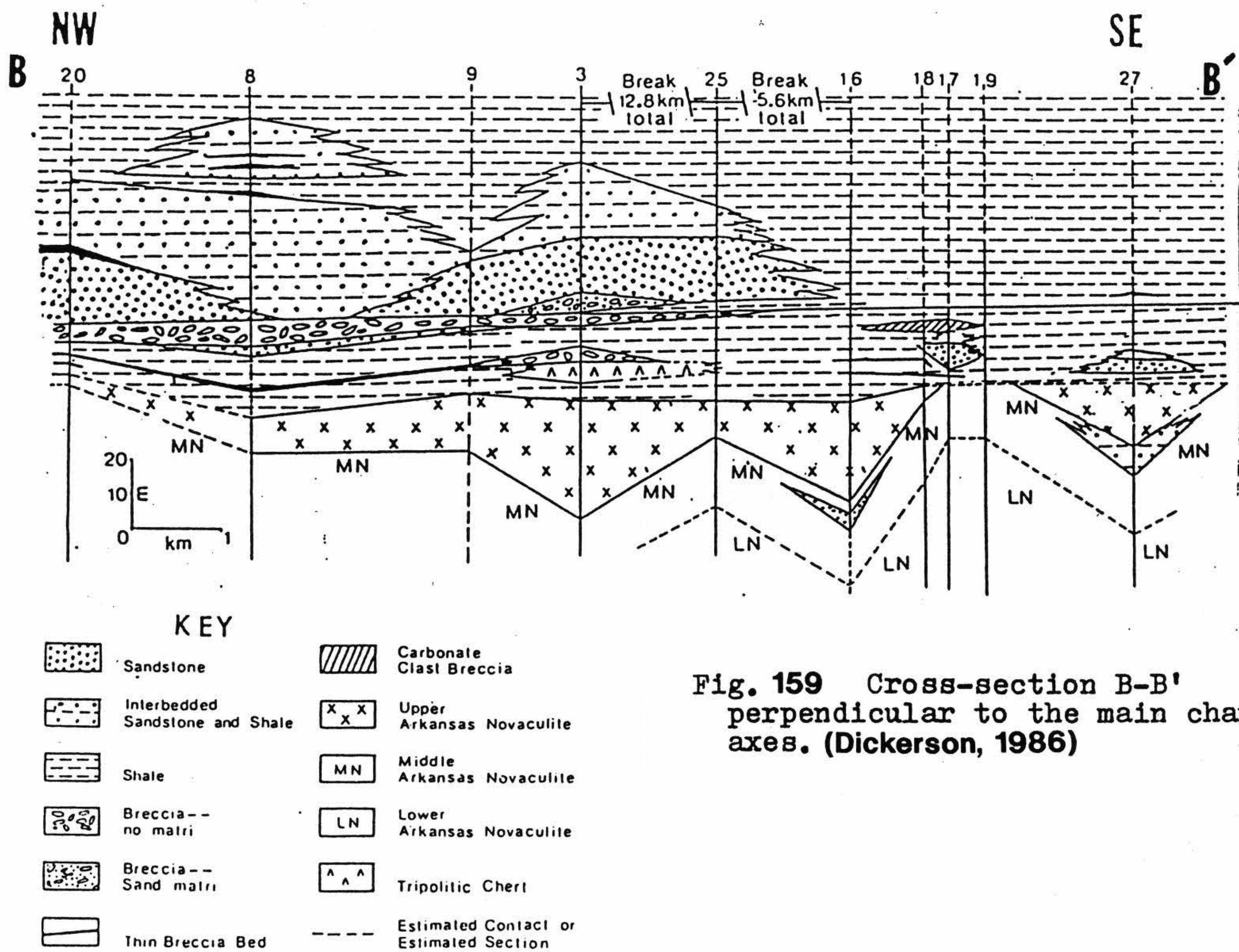


Fig. 159 Cross-section B-B' perpendicular to the main channel axes. (Dickerson, 1986)

(Description by Stone, Howard, and Haley, 198)

STOP 25

Blakely Sandstone at Blakely Mountain Dam

Rocks of the middle and lower Blakely Sandstone can be examined in two fine exposures at Blakely Mountain Dam (Fig. 160). They consist of thin to massive bedded quartzitic sandstone (in part calcareous and conglomeratic), thin-to-thick-bedded siltstone, and gray to black banded shale. Bottom marks, cross-laminations, and graded bedding indicate the tops of the beds are to the southeast, thus these beds are southward overturned. Cleavage in the shale dips northward and refracts across thin sandstone beds especially near the fold hinge lines. This is especially well developed at the western end of both exposures. The sandstone divisions are thinning and fining upward and probably represent upper and middle submarine fan channel deposits derived from submarine scarps and slopes and possible foreland facies to the north and northeast. Quartz veins of late Paleozoic age fill fractures in many of the sandstones. Graptolites are present in some of the shales.

Regional mapping indicates two large channel-lobe like clastic sequences with high sand/shale ratios separated by a thick shale and siltstone interval in the Blakely throughout portions of this area. Does this indicate there were two major low-stands of sea level, or retrogressive slope failures caused for other reasons in Blakely time?

The Hot Springs structural trend (Northeast) as mapped by Purdue and Miser (1923) is the dominant structural feature in this area. Some anomalous directions of fold rotations in outcrops downstream indicate an older period of folding. At least two sets of cleavage are present. Most folds are overturned toward the southeast and hinge lines are near horizontal or rake gently northeast or southwest. Clastic dikes, in part paralleling cleavage, suggest deformation of soft water-saturated sediments. Sedimentary pull-aparts, debris flows, and soft-sediment slump features are present and may be mistaken for tectonic compressional features.

Riprap on the face of the dam is a gray, micritic and conglomeratic limestone quarried from the lower part of the Womble. The quarry is located in the NW¼ SW¼ NW¼ Sec. 34, T. 15 N., R. 22 W. a few miles northeast of the dam. This sequence of limestones represents a large sedimentary slurry deposit containing probable shallow-water pelletoidal and deepwater micritic materials. Repetski and Ethington (1977) made conodont collections from these limestones and indicate a Middle Ordovician age with affinities both to European and North American strata.

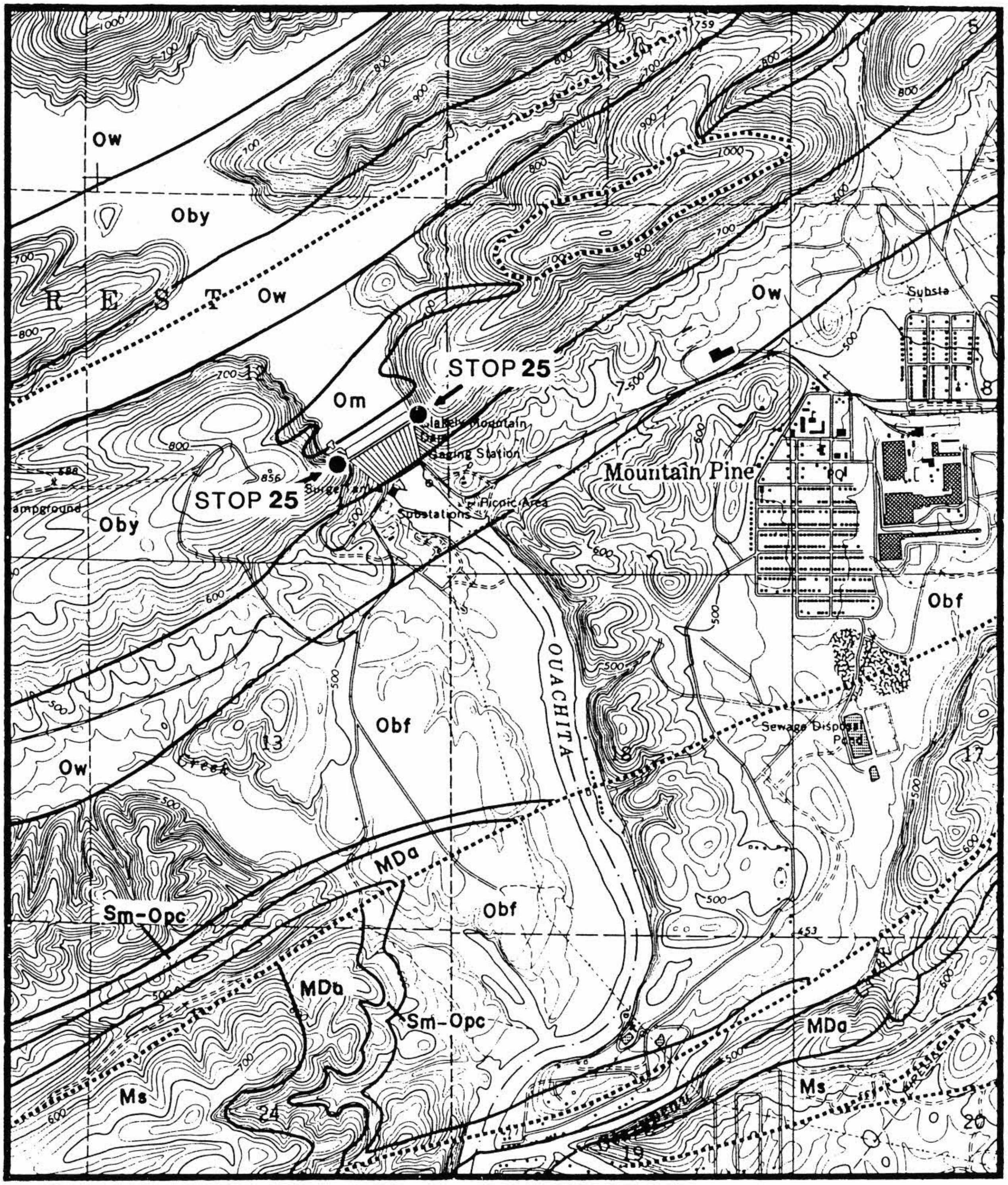
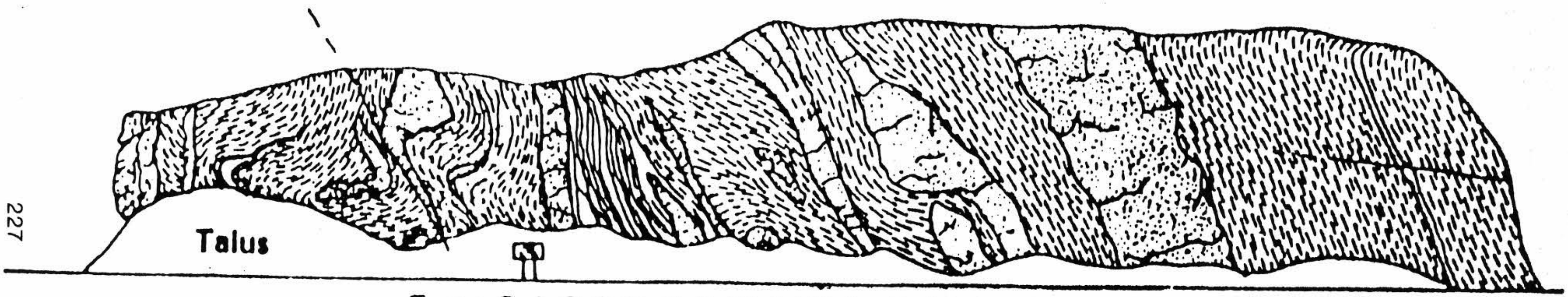


Figure 160. GEOLOGIC MAP OF BLAKELY MOUNTAIN DAM - STOP **25**

1000 0 1000 2000 3000 FEET

Ms	STANLEY SHALE	Oby	BLAKELY SANDSTONE
MDa	ARKANSAS NOVACULITE	Om	MAZARN SHALE
Sm-Opc	MISSOURI MOUNTAIN SHALE POLK CREEK SHALE		
Obf	BIGFORK CHERT	....	THRUST FAULTS
Ow	WOMBLE SHALE	————	CONTACTS

SOUTH END — BLAKELY MOUNTAIN DAM



From S.A.S.G.S Fall Field Trip Guidebook 1978

Figure 161. Geologic sketch of the south end of Blakely Mountain Dam.

Figure 162. Interbedded sandstone and shale forming southward-verging folds in the middle Blakely Sandstone at the south end of Blakely Mountain Dam. Arrow marks a small structure that has been interpreted variously as an earlier superposed fold, a sedimentary pull-apart, and as a tectonically "crowded" interval.

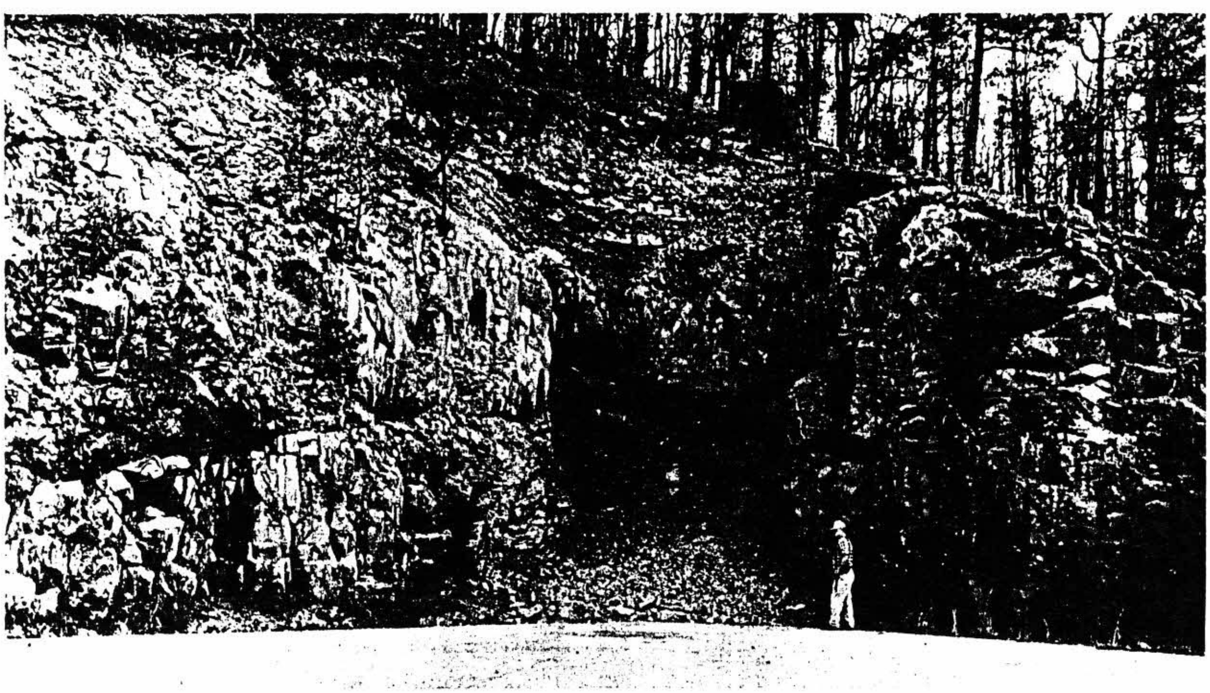


Figure 163. Massive sandstones of the Blakely enclosing a mass of deformed shale ("the terrible corner") at the north end of Blakely Mountain Dam. The shale mass has been interpreted as tectonic boudinage, a faulted block, a clay plug in a submarine fan channel system, and as a sedimentary slump mass.

(Descriptions from Stone & Haley, 1986; and Stone, 1993)

STOP 26

Ron Coleman's Quartz Crystal Mine

Mining of quartz crystals in the Ouachita Mountains of Arkansas has been going on for many years, the first miners probably being the Indian tribes that shaped them into arrowheads. Because of the clarity and perfect shape of many of the individual crystals and crystal clusters, the principal market for the quartz over the years has been as specimens in both individual and institutional mineral collections. During World War II about five tons of clear quartz crystals from Arkansas were used in the manufacture of radio oscillators to supplement the production from Brazil. Currently, the quartz crystals are being used for: manufacturing fusing quartz, which has many chemical, thermal and electrical applications; for seed crystals (lasca) for growing synthetic quartz crystals; and, of course, for mineral specimens. It should be noted that the "Hot Springs Diamonds" for sale in the local rock shops and jewelry stores are cut from Arkansas quartz crystals.

Quartz veins are numerous and are found in a wide belt extending from Little Rock, Arkansas to Broken Bow, Oklahoma, in the central core area of the Ouachita Mountains. These veins, up to sixty feet in width, commonly contain traces of adularia, chlorite, calcite and dickite. In a few places lead, zinc, copper, antimony and mercury minerals are associated with the quartz veins. At relatively few localities, however, within the quartz vein belt individual quartz crystals and crystal clusters attain the size and clarity requisite for mining.

In the Ouachita Mountains, there is a close association of quartz veins with fault zones. It is believed that the quartz veins represent, in part, dewatering processes that took place along the fault zones. The increase in pore fluids may well have contributed to overpressuring and related conditions and enhanced the overall faulting and folding process. The quartz veins with their associated minerals are presumed to be hydrothermal deposits of tectonic origin formed during the closing stages of the late Pennsylvanian-early Permian orogeny in the Ouachita Mountains.

The quartz crystal deposits at the Coleman Mine (Fig. 164 & 165) are also known as the West Chance Area, Dierks No. 4 Mine, and Blocher Lead. The quartz crystals occur in veins in limy sandstone and conglomeratic sandstone beds of the Crystal Mountain Sandstone. Beds of conglomeratic sandstone exposed in the pit contain abundant

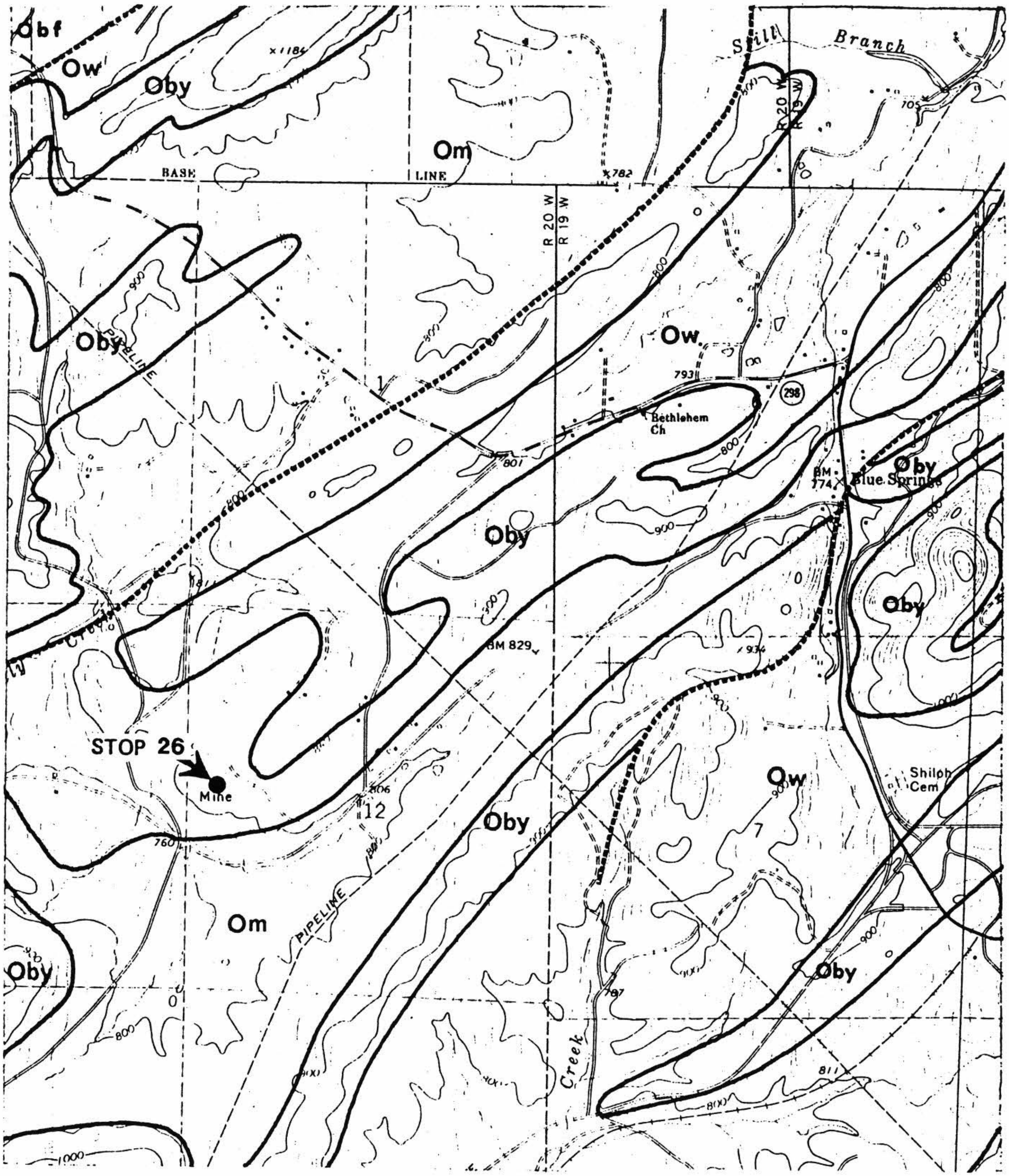
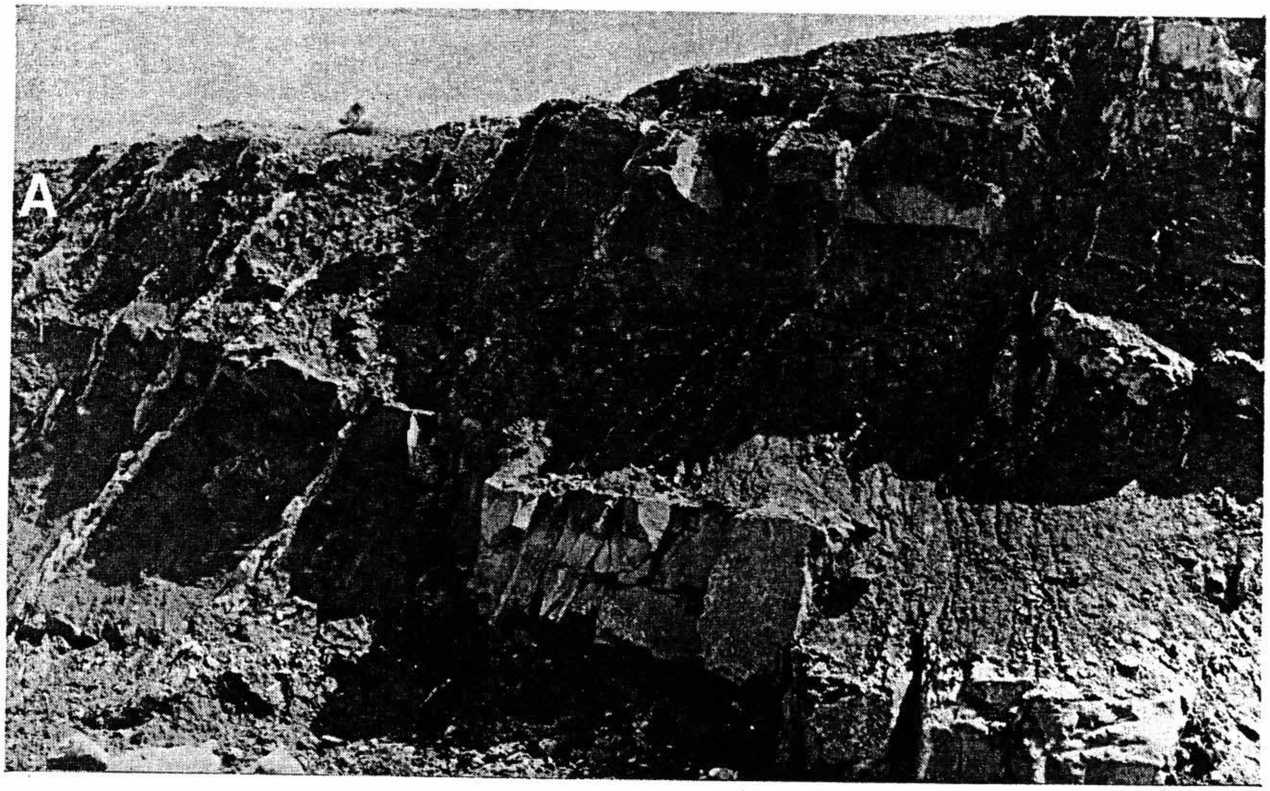


Figure 164. GEOLOGIC MAP OF BLUE SPRINGS AND VICINITY STOP 26

N

1000 0 1000 2000 3000 FEET

Obf	Bigfork Chert	Oby	Blakely Sandstone	········	Thrust Faults
Ow	Womble Shale	Om	Mazarn Shale	———	Contacts



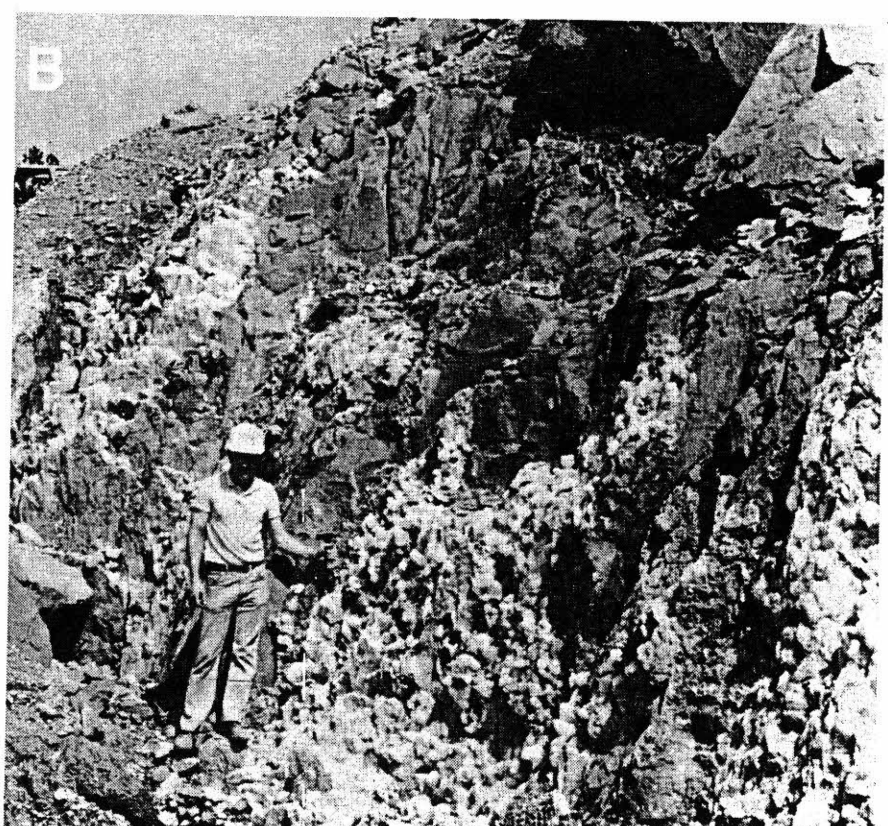


Figure 165

weathered meta-arkose and granitic boulders, cobbles, and pebbles, and some clasts of limestone, chert and shale. It is likely that these sediments were deposited in submarine fan channels and were derived from a granite-rich terrain to the north-northeast. This area is comprised of many thrust-faulted sequences with at least two major periods of folding resulting in differing attitudes in fold hinge lines and axial planes. The mine itself is situated on the nose of a large, complexly-deformed syncline.

The quartz crystal veins are fracture fillings with the larger and more productive cavities being located at the intersection of two veins. Mining operations are relatively simple, consisting initially of removing overburden and loose rock with a bulldozer to expose the crystal-filled cavities, and then removing the quartz crystals with hand tools.

Individual quartz crystals upto five feet in length weighing as much as 400 pounds and clusters 15 feet in length weighing over five tons have been produced from these mines.

It is our present belief that this unit represents a tremendous low-stand of sea level and that vast quantities of clastics from the shelf and slope to the north were deposited along the northern flanks of the early Ouachita trough via major submarine canyon systems. It is likely that the erratic-bearing conglomeratic sandstones and lenticular sandstone masses exposed in this and other nearby pits are classic examples of upper submarine fan deposition. There is also some consideration for a lower submarine canyon depositional site or combinations of both.

(Description by Stone & Bush, 1982)

STOP 27

Tripoli Mine in Arkansas Novaculite

The Malvern Minerals Company of Hot Springs operates this mine and a plant (Fig. 166 & 167) to produce Novacite from tripolitic novaculite of the Arkansas Novaculite. Novacite is a microcrystalline silica (99.6%) used as a filler and extender in paints and plastics and as an abrasive. The Arkansas Novaculite here is overturned to the southeast and the rocks are dipping 32°-45° to the northwest. The tripolitic novaculite is mined from the Upper Division of the Arkansas Novaculite which is about 60 feet thick. It is overlain by the Middle Division (about 20 feet thick) and the Lower Division (about 400 feet thick). The Upper Division is white and friable with an average particle size of 7 microns. The Middle Division consists of a highly siliceous, carbonaceous black shale which the company mines and markets under the trade name Ebony for use as an extender pigment and other purposes. Typical analyses of the Novacite and Ebony are shown in Table 4.

		Novacite	Ebony
Silica	SiO_2	99.49%	60.40%
Carbon	C	0.00%	3.37%
Sulphur	S	0.00%	0.07%
Aluminum oxide	AL_2O_3	0.102%	22.40%
Ferric oxide	Fe_2O_3	0.039%	2.15%
Titanium oxide	TiO_2	0.015%	1.70%
Calcium oxide	CaO	0.014%	2.00%
Magnesium oxide	MgO	0.021%	0.38%
Loss on Ignition		0.190%	9.75%

Table 4

The tripoli in the Upper Division is probably formed by the leaching of calcium carbonate cement form the novaculite. Scanning electron micrographic studies of the novaculite and tripoli by Keller et al. (1977) confirm that the silica has been slightly recrystallized at this locality and that polygonal triple-point texture is present.

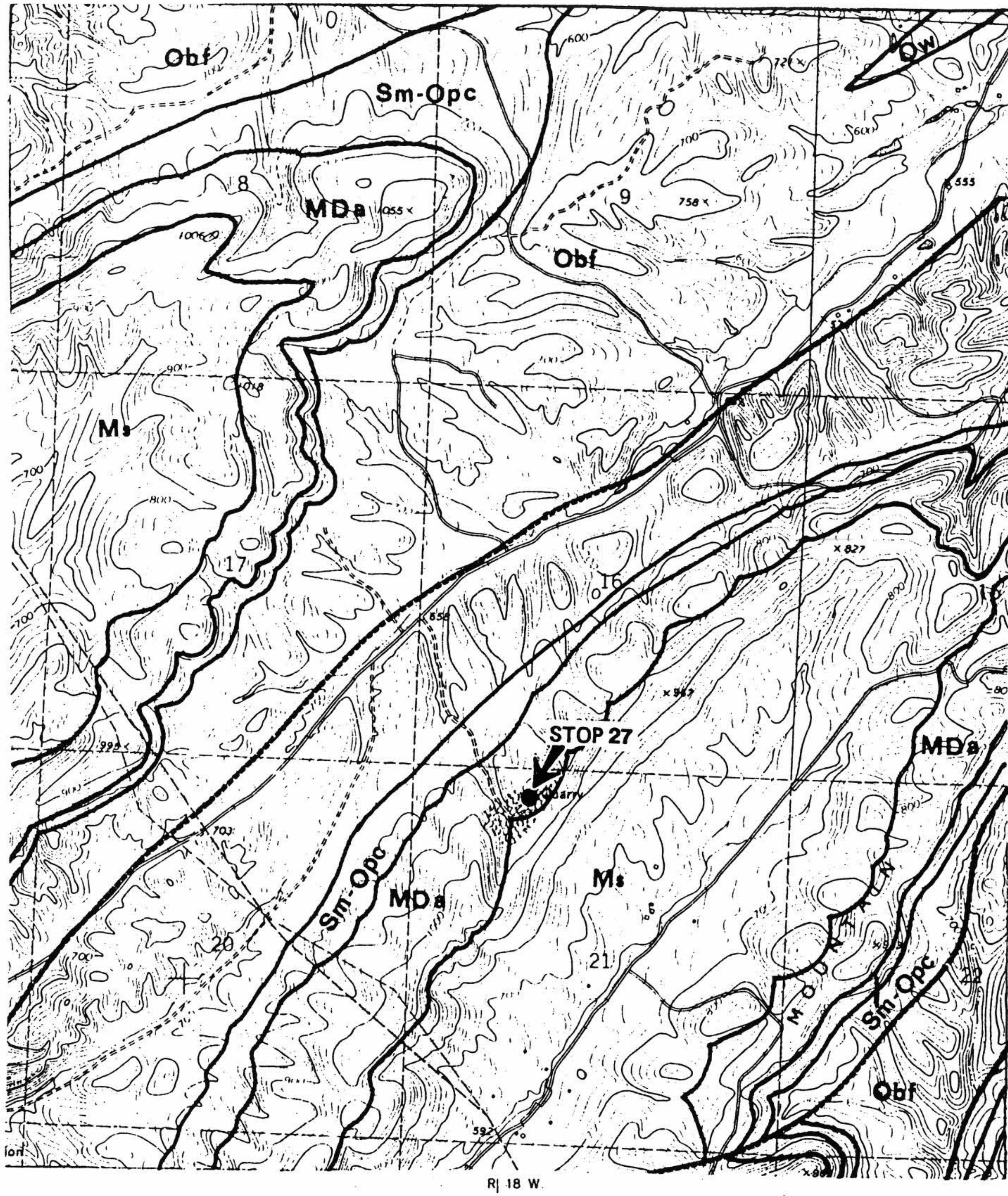
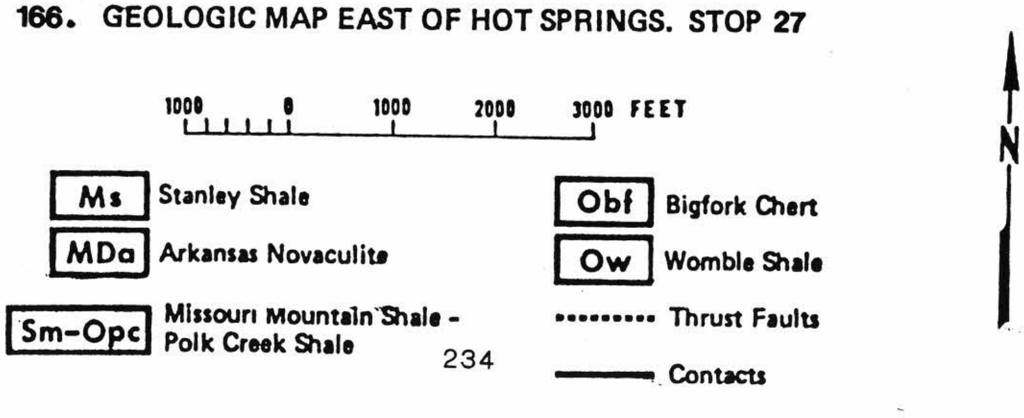


Figure 166. GEOLOGIC MAP EAST OF HOT SPRINGS. STOP 27



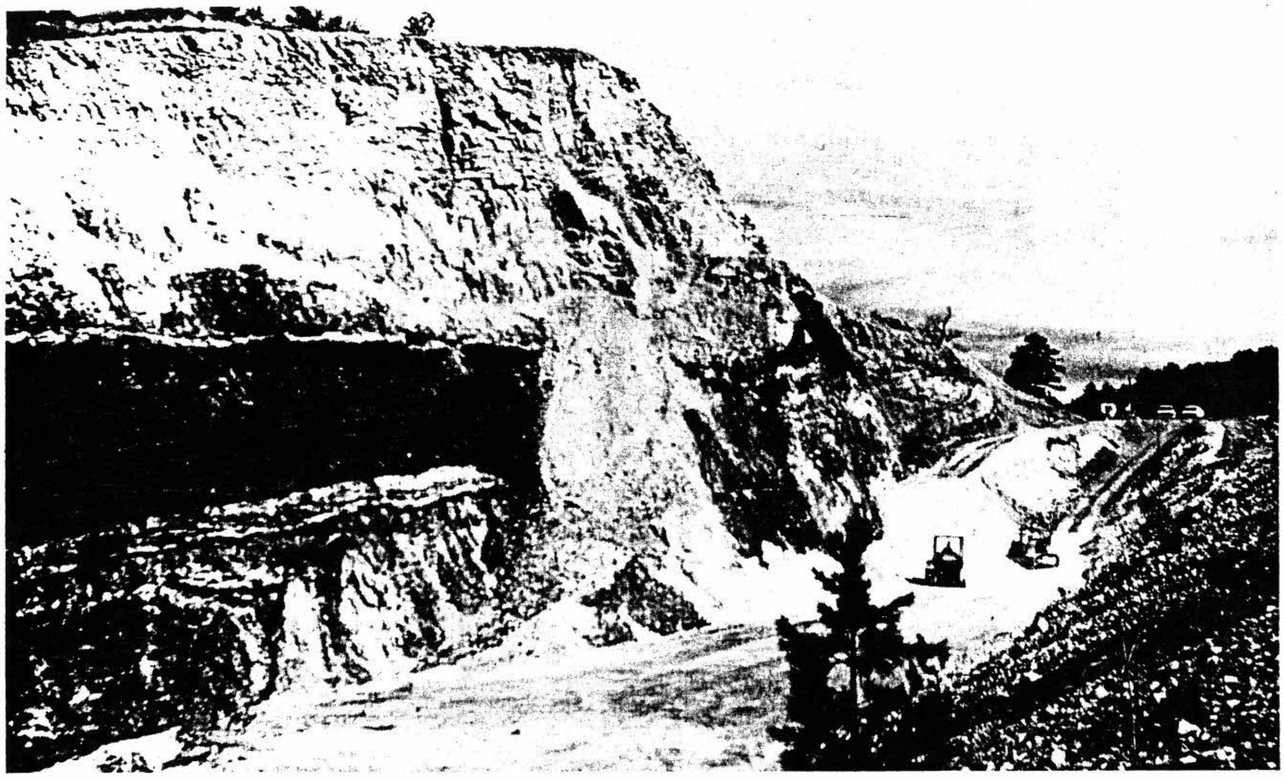


Figure 167. - Open-pit, tripoli ("Novacite®") mine of the Malvern Minerals Company, Garland County, Arkansas. The stratigraphic section is inverted here; the white, tripolitic upper member of the Arkansas Novaculite is at the base of the deposit, the dense novaculite of the lower member is at the top, and the dark shale of the middle member is between them.

At the top of the Lower Division at this locality there is about a 20 foot interval of sedimentary slurried boulder-like novaculite, minor sandstone, and other materials forming a very coarse conglomerate or breccia. Minor granitic fragments are present in similar rocks about ten miles to the north. Honess (1923) reports igneous and volcanic debris in the Arkansas Novaculite in the Broken Bow area of Oklahoma. These deposits likely represent slurries derived from submarine scarps and ridges to the north that, in part, had active extensional faulting and igneous activity in Devonian times. Richard Lane of Amoco Production Company tentatively identified Middle Devonian conodonts from the shales about 18 feet stratigraphically below the boulder-interval.

Other interesting geologic features in or near the quarry are:

1. Secondary oxides, often forming manganese and iron dendrites, coatings and discolorations on fracture surfaces;

2. Unusual development of grooved-like curtains or sheets of tripoli are present on the wall face;

3. Some of the fractures are filled with weathered to fresh igneous intrusives (alkalic dikes, early Late Cretaceous);

4. Novaculite in the Lower Division that is of very high quality for whetstones, including both hard (Arkansas) and soft (Washita) types;

5. Minor thrust and tear faults with slickensides;

6. The ridge south of the quarry is formed by the overlying Hot Springs Sandstone of the lower Stanley Shale. It is about 75 feet thick and is composed of quartzitic sandstone and shale with sandy chert-novaculite conglomerate typically near the base;

7. Thin films of gorceixite (barium phosphate) coating novaculite fractures.

(Description by Stone, Haley and Viele 1973; and Stone, 1993)

STOP 28

Abyssal Plain Deposits Near Crows Arkansas

South of Crows, Arkansas along a rural road and the adjoining Middle Fork of the Saline River (Fig. 168) there is about a 1/4 mile exposure containing very thin layers of calcareous to clean siltstone and gray shale that are intensely crinkled and buckled (Fig. 169) and further cut by pervasive cleavage (in places), and milky quartz veins. This strata occurs in the Lower Ordovician Mazarn Formation and this determination has been recently confirmed by conodont evaluations from a few of the thin gray micritic limestones.

It is commonly thought that these deposits represent waning cycles of turbidity currents on the early Ouachita basinal plain. While there are paleocurrent features (cross-laminations, bottom markings and other features) the source of the sediments is unknown due to the structural complexities. We postulate a sediment transport from the east-northeast. But you are welcome to your own opinions.

Examine this interval and see if you can tell top and bottom, current direction(s), structural repetition, thrust faulting, etc. Also there are some hints of deep-marine trace fossils--that is, if they are not structural boudins, mullions and other phenomena.

What cycles of deposition formed these rather "varve-like" thinly banded, laminated to minutely cross-laminated deep water shales and siltstones? Do these deposits afford a true parameter for the many thousands of episodes of turbidity currents required for their accumulation in this early Ouachita basin? Another related possibility is does each apparent lithic "pair" represent a yearly climatic cycle (winter-summer/hot-cold/wet-dry) with waxing and waning input of fine clastic nutrient. The coarse fractions deposited (wet cycle) distally by weak turbidity or bottom currents and the finer clays (dry cycle) being suspended clouds--hemipelagic materials? You are most welcome to present other hypotheses!

AGAIN THANK YOU FOR YOUR INTEREST AND PLEASE RETURN AGAIN!

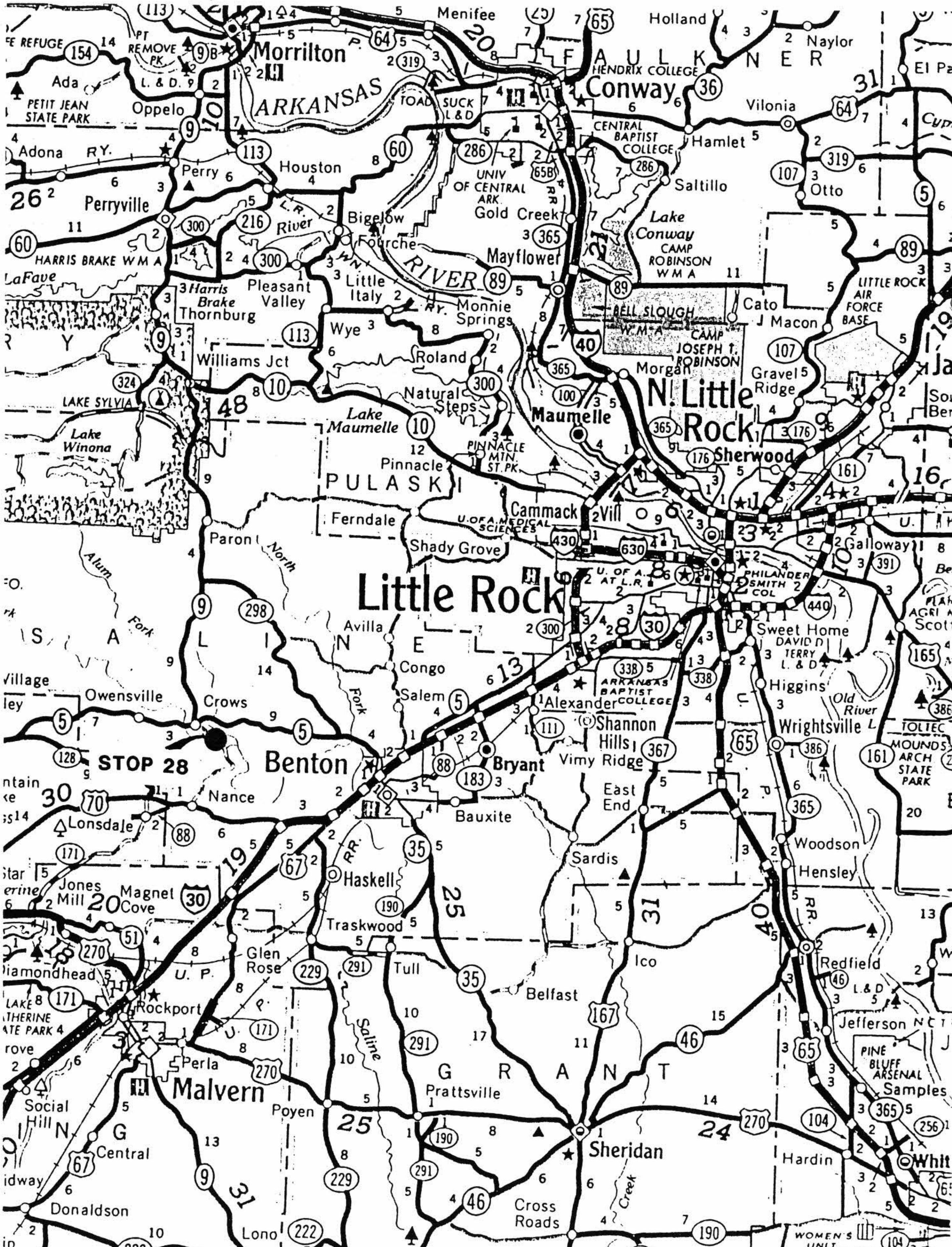

Menifee
Holland
Naylor
Morrilton
PT REMOVE PK.
PETIT JEAN STATE PARK
Ada
Oppelo
ARKANSAS
TOAD SUCK L & D
F A U L K N E R
HENDRIX COLLEGE
Conway
El Pa
Vilonia
Hamlet
CENTRAL BAPTIST COLLEGE
Adona
Perry
Houston
UNIV OF CENTRAL ARK.
Saltillo
Otto
Perryville
Gold Creek
Lake Conway
CAMP ROBINSON W M A
HARRIS BRAKE W M A
Bigelow
Fourche
River
RIVER
Mayflower
LaFave
Harris Brake
Pleasant Valley
Little Italy
Monnie Springs
BELL SLOUGH W M A
Cato
Macon
LITTLE ROCK AIR FORCE BASE
Thornburg
Wye
Roland
CAMP JOSEPH T. ROBINSON
Morgan
Gravel Ridge
Williams Jct
LAKE SYLVIA
Natural Steps
N. Little Rock
Maumelle
Lake Maumelle
Lake Winona
PINNACLE MTN. ST. PK.
Pinnacle
Sherwood
PULASKI
Cammack Vill
Ferndale
U. OF A. MEDICAL SCIENCES
Shady Grove
Galloway
Paron
North
Alum Fork
Little Rock
U. OF A. AT L.R.
PHILANDER SMITH COL
Avilla
Sweet Home
Scott
S A L I N E
Congo
ARKANSAS BAPTIST COLLEGE
Higgins
Village
Owensville
Crows
Salem
Alexander
Old River
Shannon Hills
Wrightsville
TOLTEC MOUNDS ARCH STATE PARK
STOP 28
Benton
Bryant
Vimy Ridge
Nance
East End
Bauxite
Lonsdale
Sardis
Woodson
Hensley
Haskell
Jones Mill
Magnet Cove
Traskwood
Ico
Diamondhead
Glen Rose
Tull
Redfield
Belfast
Rockport
Jefferson
Perla
PINE BLUFF ARSENAL
Malvern
G R A N T
Prattsville
Samples
Social Hill
Poyen
Central
Sheridan
Hardin
Donaldson
Cross Roads
Lono
WOMEN'S UNIT

Figure 168

Figure 169. Thin layers of calcareous siltstone and shale in the Mazarn Formation.

PAPERS ON REGIONAL GEOLOGY, STRATIGRAPHY AND SEDIMENTOLOGY OF OUACHITA MOUNTAINS AND ARKOMA BASIN

The Geology of North America
Vol. A, The Geology of North America—An overview
The Geological Society of America, 1989

Chapter 14

The Ouachita system

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INTRODUCTION

During the Paleozoic, the southern margin of the North American craton underwent one major ocean margin cycle ("Wilson cycle") lasting from the Late Proterozoic into the Permian. Breakup of a preexisting continent may have begun in the Late Proterozoic and lasted to the Middle Cambrian along a series of divergent rifts and transforms, which gave the margin between southern Alabama and northern Mexico an angular appearance. The breakup was accompanied by a system of rift arms and fault systems that reached far into the cratonic interior. The cycle ended with the formation of a collisional orogenic belt known as the Ouachita system.

The Ouachita Mountains of Oklahoma and Arkansas, the Marathon uplift of west Texas, and the Solitario (a small outcrop of Paleozoic rocks west of the Marathon region) are the only outcrop areas of the Ouachita system in the United States (Fig. 1). In northern Mexico, a number of small outcrops of Ouachita-facies rocks appear as inliers of the eastern Cordilleran orogen. The remainder of the frontal elements of the Ouachita system and the entire hinterland are covered by postorogenic Permian, Mesozoic, and Tertiary rocks of the Gulf Coastal Plain. Thus, some 90 percent of this orogenic belt cannot be inspected at the surface and must be mapped from subsurface borehole and geophysical data. Regional tectonic interpretations remain therefore rather tentative, especially in the deep hinterland, where boreholes are sparse.

The foreland of the Ouachita system, on the other hand, abounds in subsurface information, although postorogenic rocks also cover large areas. Because many of the foreland basins are prolific producers of oil and gas, however, thousands of boreholes and kilometers of seismic reflection surveys have yielded an almost unmatched three-dimensional information network.

From Late Cambrian to Early Mississippian time, the continental margin was trailing and passive, receiving stable to moderately subsiding platform sedimentation, while deep-water sediments were laid down off the shelf edge on the slope, rise, and bottom of the adjacent sea. Although this continental margin was the western continuation of the Appalachian margin, the sedimentary sequences, timing, and frequency of orogenic events were in part quite different.

Beginning in middle Mississippian time, deep-water flysch sedimentation with very high sedimentation rates replaced the preceeding slow off-shelf sedimentation. In the Ouachita Mountains, flysch sedimentation ceased in the mid-Pennsylvanian (late Atokan), but in west Texas, some deep-water sedimentation persisted into the Permian. During the Ouachita orogeny, northvergent thrusting transported much of the deep-water facies as an allochthonous thrust belt onto the outer continental shelf.

The tectonostratigraphic history of the Ouachita system has been the subject of many monographs, symposia, and regional summary articles since the early part of the century (e.g., Taff, 1902; Powers, 1928; Miser, 1929, 1934, 1959; van Waterschoot van der Gracht, 1931a, b; King, 1937, 1950, 1975a, b; 1977; Hendricks and others, 1947; Cline and others, 1959; Cline, 1960; Flawn and others, 1961; Keller and Cebull, 1973; Thomas, 1976, 1977a, b; Wickham, 1978; and summary articles in Hatcher and others, 1989). The present chapter serves, therefore, primarily as an update to the earlier publications and as an attempt to point out remaining unresolved problems.

PRE-OROGENIC EARLY AND MID-PALEOZOIC HISTORY OF DEPOSITION AND DEFORMATION

The rifting event that led to the formation of the passive margin of southern North America probably occurred during or sometime before the Middle Cambrian. The onset of rifting cannot be documented in the Ouachita system, but is known to have occurred in late Precambrian time in the southern Appalachians (Rankin, 1975). The overall shape of the margin is generally outlined by the present shape of the compressional thrust belt (Fig. 2) as well as by the distinct gravity signature that identifies the present southern border of the full thickness of continental crust, although the latter may have been somewhat modified by late Paleozoic tectonism and Mesozoic attenuation (Kruger and Keller, 1986). This margin (Fig. 3) can be divided into four, nearly straight segments: (a) a segment trending northwest to southeast between southern Alabama and southeastern Oklahoma; this segment abuts with an almost right angle against the Appalachian margin in southern Alabama; (b) a segment trending north-northeast to south-southwest in central Texas; (c) a south Texas segment trending northwest to southeast; and (d) a

Arbenz, J. K., 1989, The Ouachita system, *in* Bally, A. W., and Palmer, A. R., eds., The Geology of North America—An overview: Boulder, Colorado, Geological Society of America, The Geology of North America, v. A.

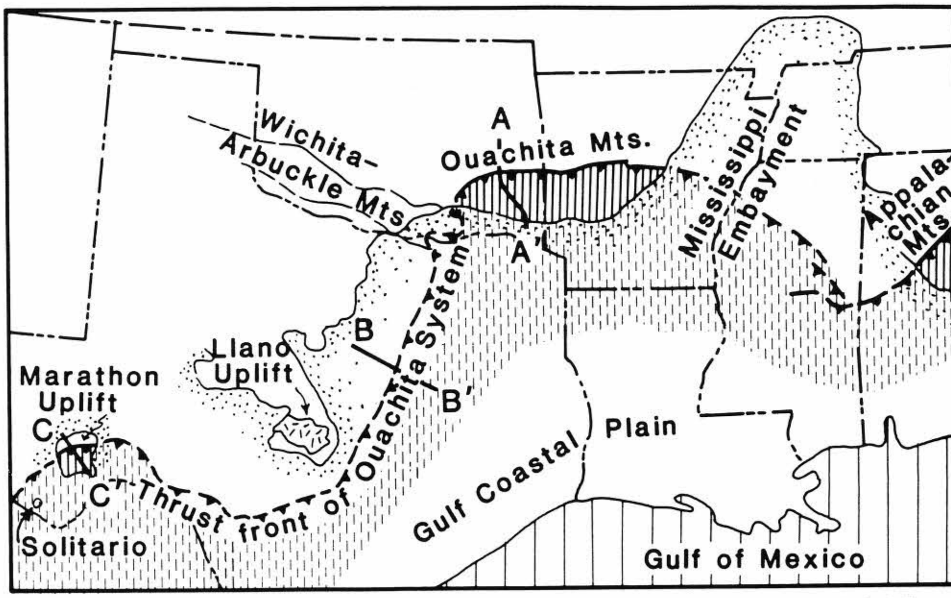


Figure 1. Regions of outcrop and subsurface distribution of the Ouachita system in the United States. For cross sections A–A′, B–B′, and C–C′, see Figure 11. Stipples, northern margin of Cretaceous-Tertiary cover; vertical ruled pattern, outcrop regions of Ouachita system; vertical dashes, subcrop region of Ouachita system.

Marathon segment trending northeast to southwest. Thomas (1976), following Keller and Cebull (1973) and Cebull and others (1976), suggested that this jagged margin was the result of alternating divergent rift and transform sections, and later (Thomas, 1977a) named the two southward-jutting prongs the Alabama and Texas promontories, respectively, and the two northerly reentrants the Ouachita and Marathon Embayments, respectively. During the late Paleozoic orogeny these promontories and embayments became recesses and salients in the thrust front of the Ouachita system.

From the continental margin, several rift and rift-transform basins extended into the interior of the continent (Fig. 3) from the newly opening ocean (Hoffman and others, 1974; Walper, 1977). From east to west these are (a) the extensive Reelfoot–Rough Creek–Rome trough rift and its more easterly subsidiary, the Birmingham basement fault; (b) the southern Oklahoma aulacogen or "leaky transform" (Thomas, 1985); and (c) the poorly documented Delaware aulacogen (Walper, 1977). The Reelfoot–Rome trough rift was the site of a thick, primarily clastic suite of exclusively sedimentary rocks of Early to Middle Cambrian age (Houseknecht and Weaverling, 1983; Ammerman and Keller, 1979). The Southern Oklahoma aulacogen, in contrast, had extensive Cambrian igneous activity (Ham and others, 1964) consisting of an earlier basaltic-gabbroic suite intruded and deposited upon older metasedimentary rocks, and followed by widespread rhyolitic extrusions and granitic epizonal intrusions. These rocks were uplifted and eroded during late Paleozoic orogeny and are well exposed in the Wichita and Arbuckle Mountains of southern Oklahoma. The Delaware rift or aulacogen has not been documented by borehole data and is inferred largely from gravity anomalies (Keller and others, 1989). With the exception of minor rejuvenations of some faults, the rifting stage had ended by Middle Cambrian time, and the southern margin of North America became a passive, trailing margin.

Foreland facies

Relative tectonic stability and proximity to equatorial latitude in the Cambrian (Van der Voo, 1988) were responsible for establishing a huge carbonate platform extending from Canada to west Texas. In the foreland of the Ouachita system this platform displayed major, broad undulations of pericratonic basins and arches. Basinal sags became established above and across the above-mentioned rift arms (Thomas, 1976), while the inter-rift areas remained tectonically positive (Fig. 3). From east to west these are the Nashville dome, the Reelfoot–Mississippi Valley basin, the Ozark uplift, the Oklahoma basin, the Concho arch, the Tobosa basin, and the ancestral Pedernal-Diablo platform (Galley, 1958; Nicholas and Rozendal, 1975; Thomas, 1976, 1977a, b, 1989). These basins and arches retained their tectonic character from the Late Cambrian until the end of the Devonian, the basins receiving more continuous and thicker suites of sediments, the arches being covered by thin, intermittent sections replete with multiple disconformities (Sloss, 1963).

The lithology of the platform sections (Figs. 4, 5) in the foreland of the Ouachita system can be crudely divided into three parts, (1) a basal transgressive part including arkosic, quartzose, and generally glauconitic sandstones and shales (e.g., Lamotte, Timbered Hills, Riley, Bliss); (2) a Late Cambrian to Middle Ordovician shallow-marine to coastal-carbonate section (e.g.,

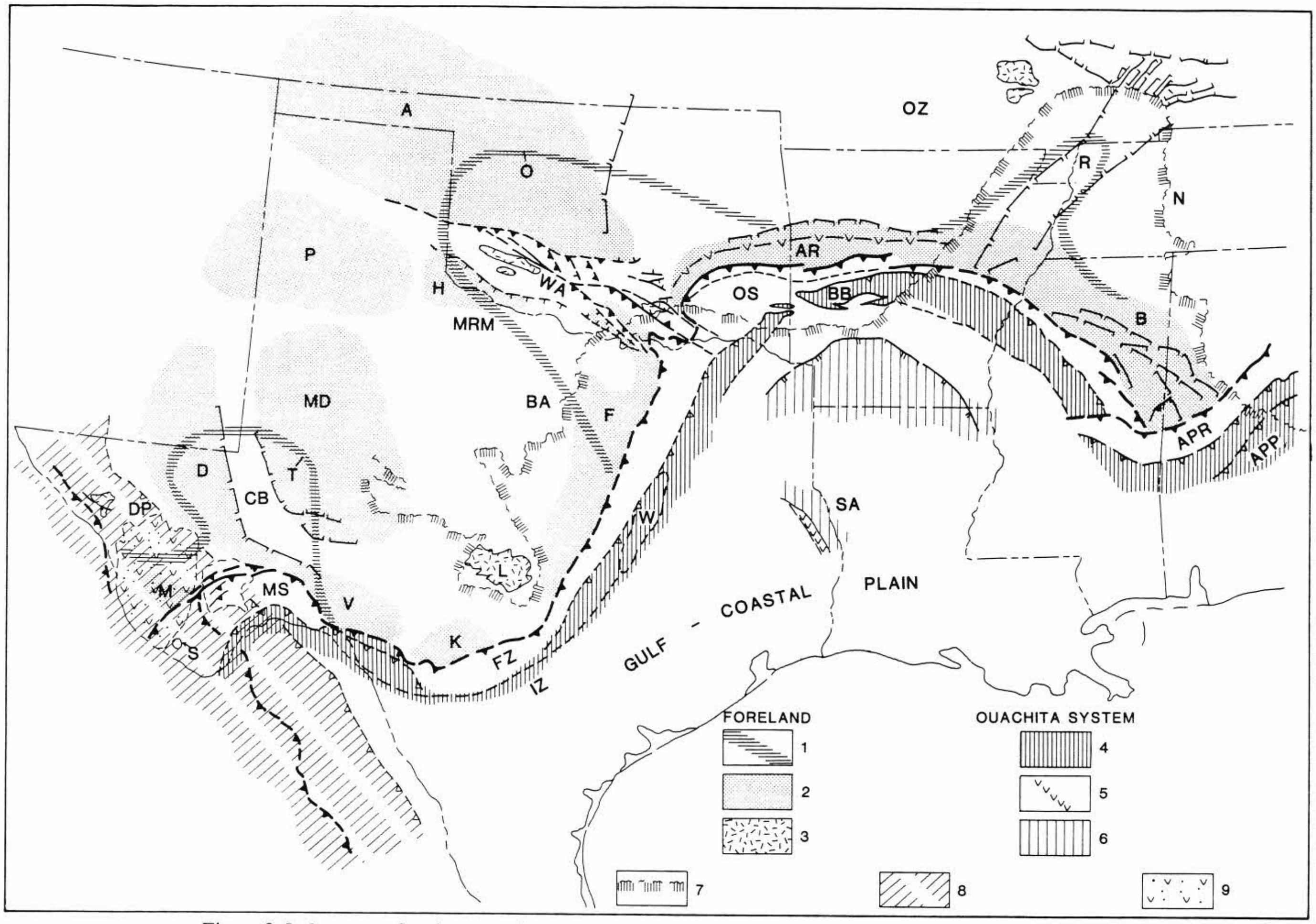


Figure 2. Index map of major tectonic elements of the Ouachita system and its foreland (subsurface after Thomas, 1989, Nicholas and Waddell, 1989). Legend patterns: 1, early Paleozoic basins; 2, late Paleozoic basins; 3, basement outcrops (Precambrian cratonic and Cambrian igneous complex of Wichita Mountains); 4, central uplifts and Interior zone (metamorphosed); 5, late Paleozoic volcanics (Sabine uplift); 6, Desmoinesian and younger successor basin; 7, northern margin of Gulf Coastal Plain; 8, mild overprint by Cordilleran orogen; 9, Davis Mountains volcanics (mid-Tertiary). **Basins on cratonic platform** (from W to E): **Early Paleozoic:** T = Tobosa, O = Oklahoma, R = Reelfoot–Mississippi Valley. **Late Paleozoic synorogenic:** M = Marfa, D = Delaware, V = Val Verde, K = Kerr, MD = Midland, F = Fort Worth (Strawn), P = Palo Duro, H = Hardeman, A = Anadarko, AR = Arkoma, B = Black Warrior. **Uplifts on cratonic platform** (from W to E): DP = Diablo platform, CB = Central Basin platform, L = Llano, BA = Bend arch, MRM = Matador–Red River–Muenster arch, WA = Wichita-Arbuckle, OZ = Ozark, N = Nashville. **Ouachita system** (W to E): S = Solitario, MS = Marathon salient, R = Red River, FZ = Frontal zone, IZ = Interior zone, W = Waco uplift, OS = Oklahoma salient, BB = Broken Bow–Benton uplift, SA = Sabine uplift (postorogenic). **Appalachian system:** APR = Appalachian fold and thrust belt (Ridge and Valley province), APP = Talladega slate belt and Piedmont.

Arbuckle, Wiberns-Ellenburger, El Paso); and (3) a more varied Late Ordovician to Late Devonian suite. During the latter period, craton-derived quartzose sandstones (e.g., St. Peter, Simpson) and transgression-related marine shales (Sylvan, Cason) found their way onto and across the platform of carbonate (Viola, Montoya, Hunton) and cherty (Penters) sediments. The widespread Chattanooga-Woodford transgression brought the preorogenic sequence to a close in much of the Ouachita foreland, although platform conditions continued to exist in some areas that were not affected by the orogeny (e.g., Ozark uplift, Diablo platform).

Ouachita facies

Ever since the early decades of this century it has been recognized that the early to middle Paleozoic rocks exposed in the Ouachita Mountains of Oklahoma and Arkansas, the Marathon uplift, and the adjacent Solitario intrusive uplift of west Texas are drastically different from their foreland sequences (Figs. 4, 5); they were probably emplaced as allochthonous thrust masses and represented some sort of basinal, "geosynclinal" facies (e.g., Miser, 1929) commonly referred to as "Ouachita facies." Particu-

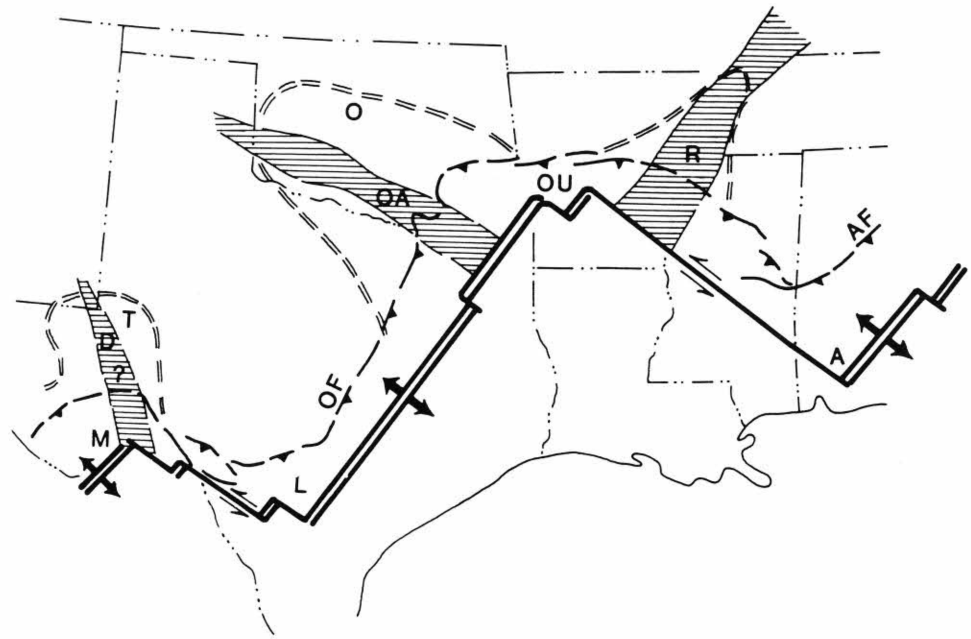


Figure 3. Late Precambrian to Middle Cambrian rift-transform breakup and failed rift arms of the southern continental margin of Paleozoic North America (for possible outboard configuration, see Fig. 14). A = Alabama promontory, AF = late Paleozoic Appalachian thrust front, D = Delaware aulacogen(?), L = Llano (or Texas) promontory, M = Marathon Embayment, O = Oklahoma basin (early Paleozoic), OA = Oklahoma aulacogen, OF = late Paleozoic Ouachita thrust front, OU = Ouachita Embayment, R = Reelfoot rift (with early Paleozoic Mississippi Valley basin), T = Tobosa basin (early Paleozoic). After Cebull and others, (1976), and Thomas (1977a).

larly since the 1950s it has become evident that a good part—if not all—of this sequence must have been deposited in deep water.

Furthermore, it was noted early that the older Paleozoic rocks of the Benton and Broken Bow uplifts of the Ouachita Mountains had undergone mild metamorphism (greenschist facies), are cut by numerous quartz veins, and are mildly to severely tectonized (Honess, 1923; Purdue and Miser, 1923; Miser, 1929; Miser and Purdue, 1929), while the older Paleozoic outcrops of the Potato Hills and the Black Knob Ridge (Oklahoma Ouachitas) and the Marathons (King, 1937) were not metamorphosed. In outcrop, fossil evidence has identified ages from Late Cambrian to Early Mississippian for the preorogenic rocks in these sequences. Although fossils are sparse and the record incomplete, the sections of both the Ouachitas and the Marathons appear to be conformable. In addition, many of the taxa have distinct North American affinities (Ethington and others, 1989).

In modern terms these sequences of dark gray to black shales, silty shales, siliciclastic and carbonate sandstones, detrital limestones, cherts and cherty limestones, conglomerates, and boulder beds are thought to be off-shelf, deep-water (slope to abyssal) deposits laid down by pelagic fallout, turbidity currents, and other mass-flow mechanisms. The following brief remarks are summarized from Lowe (1989) and McBride (1989).

Dark gray and black shales, with minor amounts of greenish gray shales and thin-bedded siltstones, make up nearly 60 percent of the lower and middle Paleozoic of the Ouachita Mountains (Collier, Mazarn, Womble, Polk Creek, and Missouri Mountain Shales). Because they are intensely deformed and cleaved, accurate thickness measurements are difficult to obtain (Bush and others, 1977; Stone and others, 1981). In the nonmetamorphic outcrops of the western Ouachitas of Oklahoma, the black shales of the Womble and Polk Creek display a high organic content typical of rich hydrocarbon source beds (Curiale and Harrison, 1981; Curiale, 1983). In the section of the Marathon uplift, shales are more subordinate (Woods Hollow, Alsate), but shale interbeds in cherts, limestones, and sandstones account for as much as 30 percent of the stratigraphic section (McBride, 1989).

Siliciclastic sandstones are present as major formations in thicknesses of several hundred meters (Ouachitas: Crystal Mountain, Blakely, Blaylock; Marathons: Dagger Flat) and as thin unnamed members in the shales. The petrography of these sandstones can vary considerably, from highly quartzose (Crystal Mountain, Blakely) to lithic (Blaylock, sandstones in Womble) and arkosic (Dagger Flat). They display a variety of bedding and internal sedimentary structures typical of turbidite-fan complexes. In the Broken Bow uplift, Lowe (1989) recognizes two classes of facies and provenance. (a) The first is a quartzitic to slightly arkosic facies associated with clay shale and limestone fragments and with rare paleocurrent indications toward the south and southwest (Crystal Mountain Sandstone and Blakely Sandstone). This facies is inferred to be derived from the craton and to be petrologically related (but not age equivalent) to well-known quartzose sandstones on the shelf (St. Peter–Simpson). The Dagger Flat Sandstone of the Marathons also appears to belong to this facies. (b) The second class is a lithic, overall finer facies (as much as 50/50 quartz to lithic components) characterizing the sandstones and siltstones in the Mazarn and Womble Shales and the Silurian Blaylock Sandstone. Based on paleocurrent directions, this facies seems to be derived from an easterly source. Likewise, a phosphatic sandstone and conglomerate facies in the upper Womble Shale near Little Rock seems to have an easterly source. The Blaylock Sandstone is present only in the Broken Bow uplift and in the southern part of the Benton uplift. The origin of this facies is less distinctly cratonic than facies (a) and—because it is nearly devoid of volcanic debris—suggests as a source the uplift of a passive margin sequence to the northeast, east, or southeast (Lowe, 1989).

Evenly bedded limestones, as micrites, calcisiltites, calcarenites, and conglomerates, generally rhythmically interbedded with shale partings, form a major part (~45 percent) of the Ordovician section in the Marathon uplift (Marathon Limestone, Fort Pena Formation, Maravillas Formation). Most of the limestones are fragmental, and many fragments are clearly derived from shallow-marine environments, such as ooids, stromatolitic structures, whole fossils, and fossil fragments, whereas many shale interbeds contain graptolites. These observations suggest emplacement of the limestone beds into deep water by turbidity currents. In the section of the Ouachita Mountains, limestone plays a subordinate role. Some mainly lens-shaped limestone units up to 15 m thick appear commonly near the top of the Collier Shale (others occur in the middle Mazarn Shale and in the upper Womble Shale). The limestones in the Collier are platy and contain fine- to medium-grained allochems composed of ooids and pellets. Because of the intense deformation it is often difficult to decide whether these lenses are mechanically disrupted, formerly continuous units, or olistoliths. Late Cambrian trilobites with North American affinities have been recovered from such limestone lenses in the Benton uplift near Hot Springs, Arkansas (Hart and others, 1987), indicating that the age span of pre-Mississippian rocks exposed in the Ouachitas and Marathons is the same.

Cherts and novaculites become important and dominant fa-

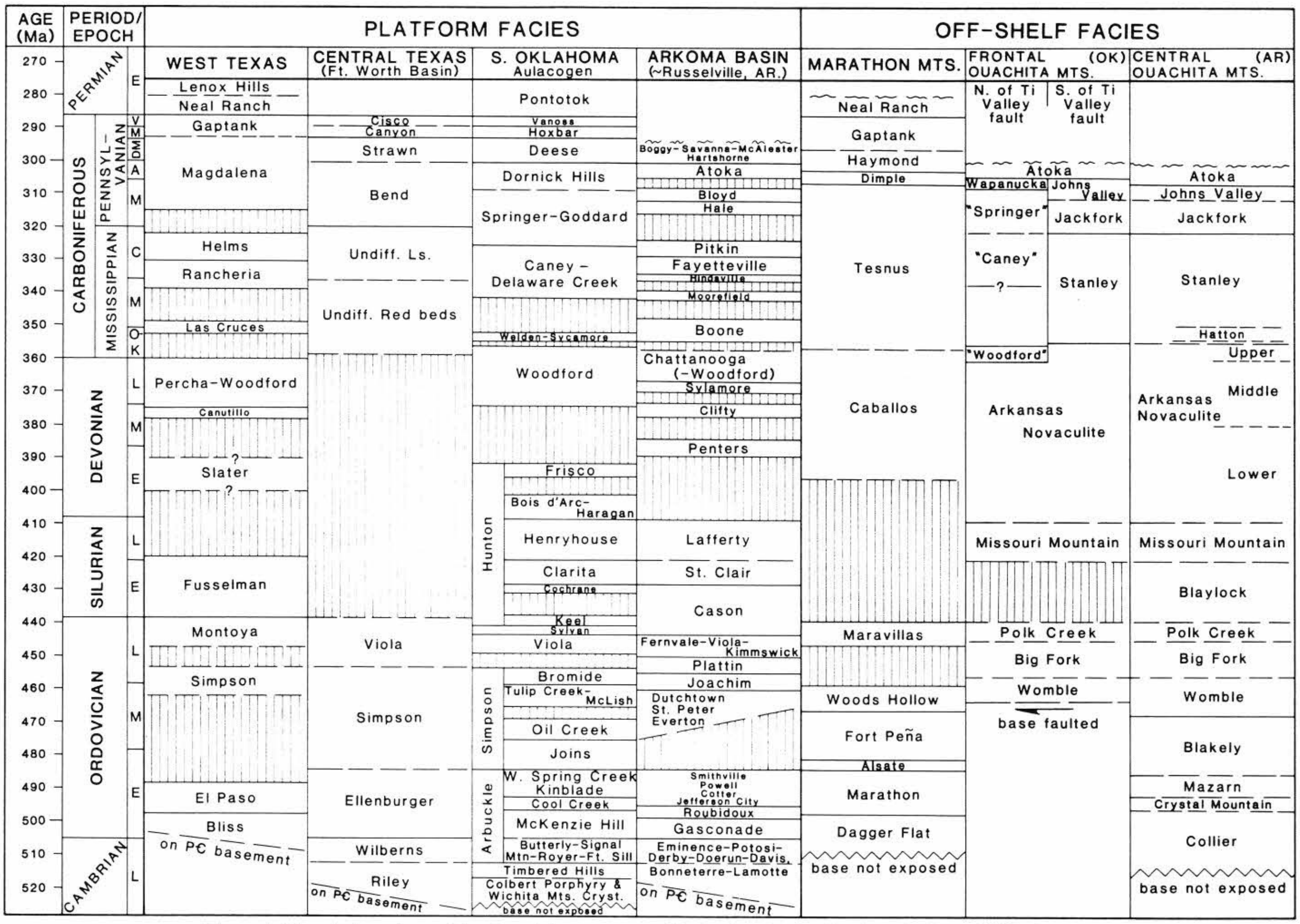


Figure 4. Correlation chart of the Ouachita system and its foreland (after Bush and others, 1977; Gordon and Stone, 1977; Cook and others, 1975; Ethington and others, 1989; Mankin, 1987; Ross, 1986; Sutherland and Manger, 1979). Abbreviations of Carboniferous series (in ascending order): K = Kinderhookian, O = Osagian, M = Meramecian, C = Chesterian; M = Morrowan, A = Atokan, DM = Desmoinesian, M = Missourian, V = Virgilian.

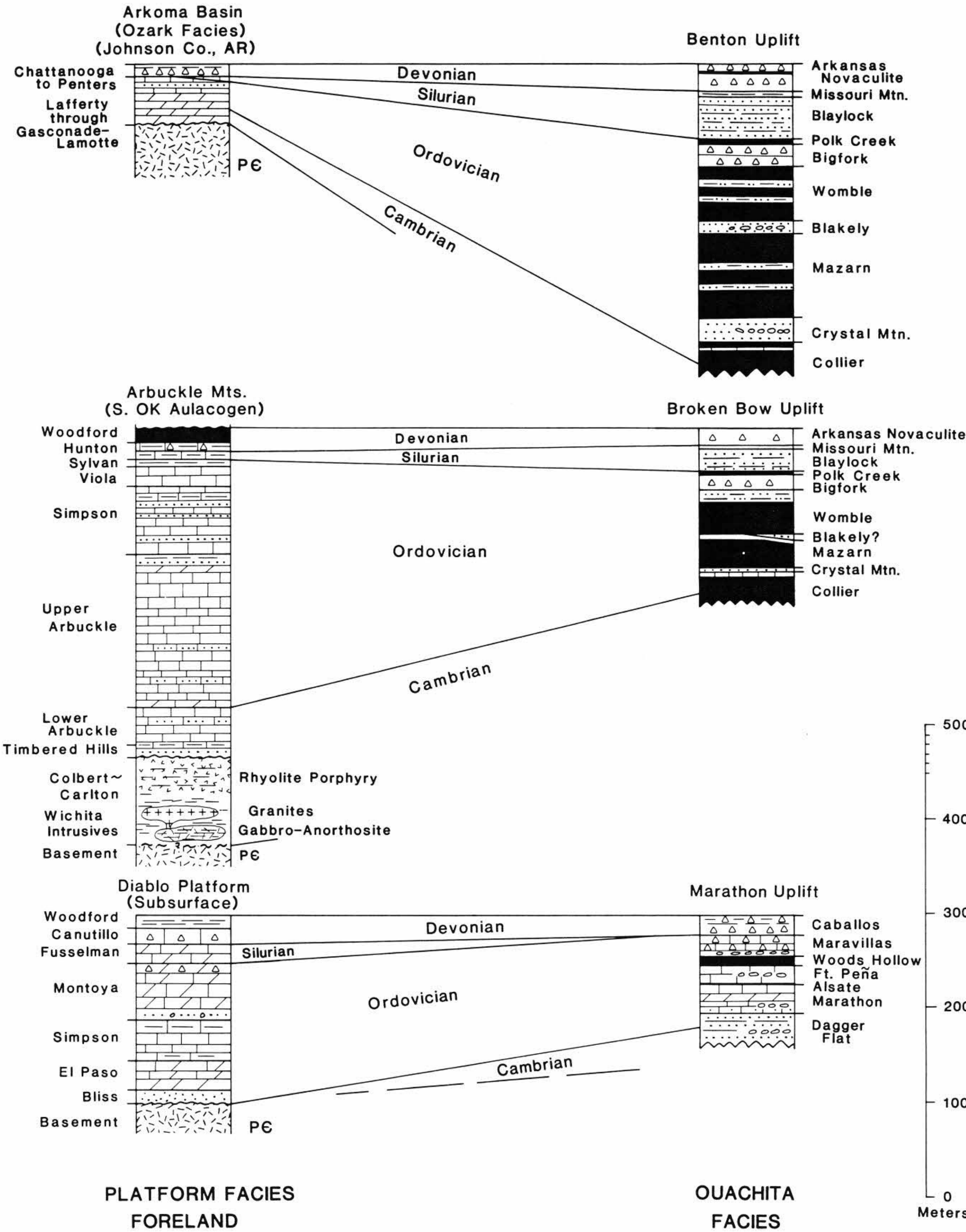


Figure 5. Generalized lithologic columns of major tectonic provinces of the pre-Carboniferous (preorogenic) platform and basin facies (after Bush and others, 1977; Cook and others, 1975; Ham and others, 1964; Lowe, 1989; McBride, 1989; Ross, 1986).

cies near the end of the preorogenic sequence in both the Ouachitas and the Marathons from Late Ordovician to the Early Mississippian. The Ordovician Bigfork (as much as 250 m) and Maravillas (as much as 150 m) Formations (Ouachitas and Marathons respectively) contain abundant beds, 1 to 60 cm thick, of uniformly bedded, organic-rich, black to gray microcrystalline chert, separated by black shale and interbedded with dark, mostly fetid, fragmental limestones and occasional conglomerates. Many pure cherts display ghosts of radiolarians and sponge spicules, but other cherts have textures of chertified calcarenites and calcisiltites. Deposition seems to have occurred in an often anoxic marine basin under hemipelagic and turbiditic conditions.

The Devonian and Early Mississippian Arkansas Novaculite of the Ouachita Mountains and its equivalent in the Marathon uplift, the Caballos Formation, are significant chert-bearing formations comprising one or two massive, milky white novaculite members up to 140 m thick, which pass downward and upward into thin-bedded units of multicolored cherts with some admixtures of quartz sand and silt, limestone, and chert conglomerate, and with shale partings of varied thickness and color. The cherts frequently contain radiolarians, sponge spicules, and spores. Lithologic units in the Marathons are quite persistent; in the Ouachitas they are regionally lenticular, indicating regional disconformities, areas of nondeposition or removal, and lateral gradation into shales. Most of the massive white chert members in the Benton and Broken Bow uplifts of the Ouachita Mountains have undergone a late Paleozoic thermal recrystallization from a cryptocrystalline chert to a polygonal microquartzite (grains up to 100 μm in diameter), which is known and mined as one of the best quality whetstones (Keller and others, 1985). Because of their resistance to erosion, the two chert formations are regional ridge formers that have contributed greatly to the mappability and structural understanding of the Ouachita and Marathon thrust belts.

The origin of some of the chert units remains controversial. Regional pinchout of members of the Arkansas Novaculite and the Caballos Formation, local in-situ brecciation and current-transported conglomerates with quartz detritus, birds-eye–like structures, apparent soil horizons with iron and manganese cements and crusts are some features that characterize these units. Several authors (e.g., Folk, *in* Folk and McBride, 1978; Lowe, 1977, 1989) have concluded from this evidence that during some periods of Devonian sedimentation, shallow-marine and even subaerial conditions may have existed. However, other workers (e.g., McBride and Thomson, 1964; Sholes, 1977; McBride, 1989; Viele and Thomas, 1989) defend a deep-water origin for both chert-bearing formations.

Many of the formations of preorogenic Ouachita facies (Fig. 5) contain conglomerates and boulder beds (Stone and Haley, 1977; Lowe, 1989; McBride, 1989) including in-situ brecciated units, current-transported conglomerates with angular to rounded fragments, and polymodal boulder sands and shales of debris flow and/or gravity slump origin. The fragments consist of a variety of lithologies, some clearly intraclastic (both soft and indurated), some identifiable as debris from shallow-marine environments of the same or older age, and some of truly exotic lithology (igneous, metamorphic, and sedimentary), the provenance of which is not known.

Uncertainties about water depths and about provenance of many units of the preorogenic Ouachita facies, together with an almost total ignorance regarding rock units and terranes in the deep hinterland, lead to even greater uncertainties about the tectonostratigraphic and paleobathymetric framework within which these sequences were deposited. Some possible alternatives will be touched upon at the end of this chapter.

CARBONIFEROUS HISTORY OF DEPOSITION AND DEFORMATION

Flysch facies

The long period of low sedimentation rates of the Silurian and Devonian came to an abrupt end in the Mississippian (Meramecian or earliest Chesterian) with the onset of the late Paleozoic orogeny (Figs. 6, 7) that accompanied the closing of the Paleozoic proto-Atlantic and ended in the collision of Gondwana with North America, and with the assembly of Pangaea. During this phase, enormous amounts of terrigenous clastic sediment poured into the Ouachita trough and came to rest as turbidite fan complexes in an expanse of turbiditic and hemipelagic muds and silts.

In the Ouachita basin proper (i.e., overlying deep-water, preorogenic Ouachita facies) about 6,000 m of such flysch assemblages were deposited in the Mississippian and Morrowan alone (Figs. 6, 7; Morris, 1974, 1989). Paleocurrent indications in the fan complexes are mainly axial from east to west with a dominant sediment supply from northeast and southeast, although there is good evidence of sediment input from a source to the south—possibly an approaching subduction-generated orogenic thrust front (Owen, 1984). A southern source is also indicated for a group of 7- to 40-m-thick rhyodacitic tuffs, which have been mapped near the base of the Mississippian Stanley Group along the southern margin of the outcrop belt of the Ouachita Mountains and pinch out to the north (Niem, 1977). Thicknesses of the Mississippian and Morrowan flysch remain fairly uniform regionally, with the exception of some noticeable thinning of the Morrowan to the northwest near Atoka, Oklahoma, at the western end of the Ouachita Mountains (Hendricks and others, 1947).

During the Atokan, deep-water flysch sedimentation migrated northward onto the continental margin along the southern portions of the present Arkoma basin, where a series of mainly down-to-the-south, Atoka-age, extensional growth faults created a deep-water basin above shallow-marine pre-Mississippian rocks (Buchanan and Johnson, 1968; Berry and Trumbly, 1968; Houseknecht, 1986). These growth faults possibly developed as a response to thrust loading by the approaching orogenic thrust front in the distant south. In the southern Arkoma basin of Ar-

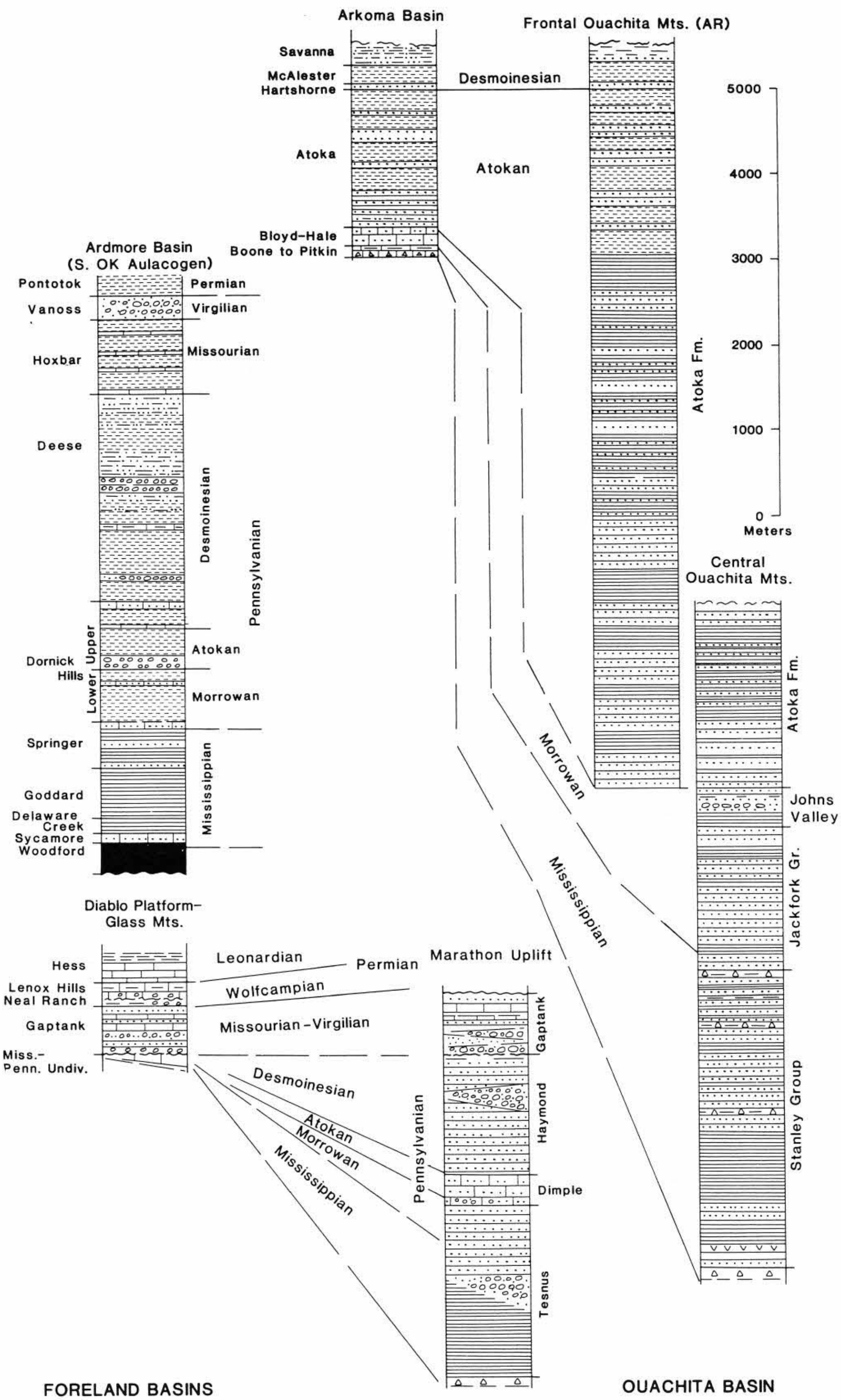


Figure 6. Generalized lithologic columns of major tectonic provinces of the Carboniferous (orogenic) platform and basin facies (after Bush and others, 1977; Cline, 1960; McBride, 1989; Morris, 1989; Sutherland and Manger, 1979; Tomlinson and McBee, 1959).

kansas, some 9,000 m of Atokan flysch have been recognized. Westward axial transport is dominant in the Atokan, but some eastward dispersal existed near the northern basin margin in Oklahoma (Ferguson and Suneson, 1988), and several local, northern source, turbidite fans have been recognized (Morris, 1974, 1989; Houseknecht, 1986). In the central and southern Ouachita Mountains, as much as 2,000 m of Atokan flysch are preserved as erosional, conformable outliers in synclines (Fig. 8). By late Atokan time the flysch basin had filled up, and sedimentation became shallow marine and ultimately fluvial in the Desmoinesian of the Arkoma basin.

As in most flysch assemblages, fossil control is extremely sparse, and many major lithosomes appear to be mildly diachronous. However, throughout the flysch basin of Oklahoma and western Arkansas, thin units of widespread siliceous shale and chert serve as time markers and have been selected as group, formation, and member boundaries (Harlton, 1938; Hendricks and others, 1947; Cline, 1960).

In the Marathon region, the entire flysch sequence is thinner (McBride, 1989) and wedge-shaped, thinning from a maximum of about 4,200 m in the southeast to some 1,500 m in the northwest. Substantial thinning also occurs along strike from northeast to southwest (Muehlberger and others, 1984). Paleocurrent directions are less uniform, indicating a northwesterly dispersal in the Mississippian (Tesnus Formation); a southeasterly, off-shelf direction in the Atokan (Dimple Limestone, a predominantly calcareous turbidite-fan package); and axial, northeast to southwest, in the Desmoinesian (Haymond Formation). Deep-water sedimentation lasted longer than in the Ouachitas in this northwestward migrating and narrowing basin, and ceased only in the earliest Permian (Ross, 1986).

On the basis of bedding sequences and sedimentary structures, a wide range of turbidite-fan subenvironments can be interpreted in the flysch deposits of the Ouachita Mountains and Marathon uplift. These are usually classified according to the criteria developed by Mutti and Ricci Lucchi (1975).

Besides the dominating sand-silt-shale assemblages of the flysch sequences in the Ouachita system, there are also numerous units of pebble-to-boulder conglomerates, olistostromes with scattered boulders, blocks, and slabs as wide as several hundred meters of penecontemporaneously deformed and disrupted slump horizons, and mélange-like chaotic intervals. The boulder beds contain intraclasts (i.e., clasts of lithified and unlithified intraformational material), as well as a vast variety of angular to rounded boulders of older rocks of both known provenance and of unknown origin (Shideler, 1970). Today most workers agree that these conglomeratic units occupy stratigraphic horizons of local to regional dimensions and are the products of a variety of deep-water transport regimes ranging from grain flow to gravity slump, and originating both on the continental margin of North America and the southern orogenic flanks of the basin.

Because many conglomerates are encased in shales or em-

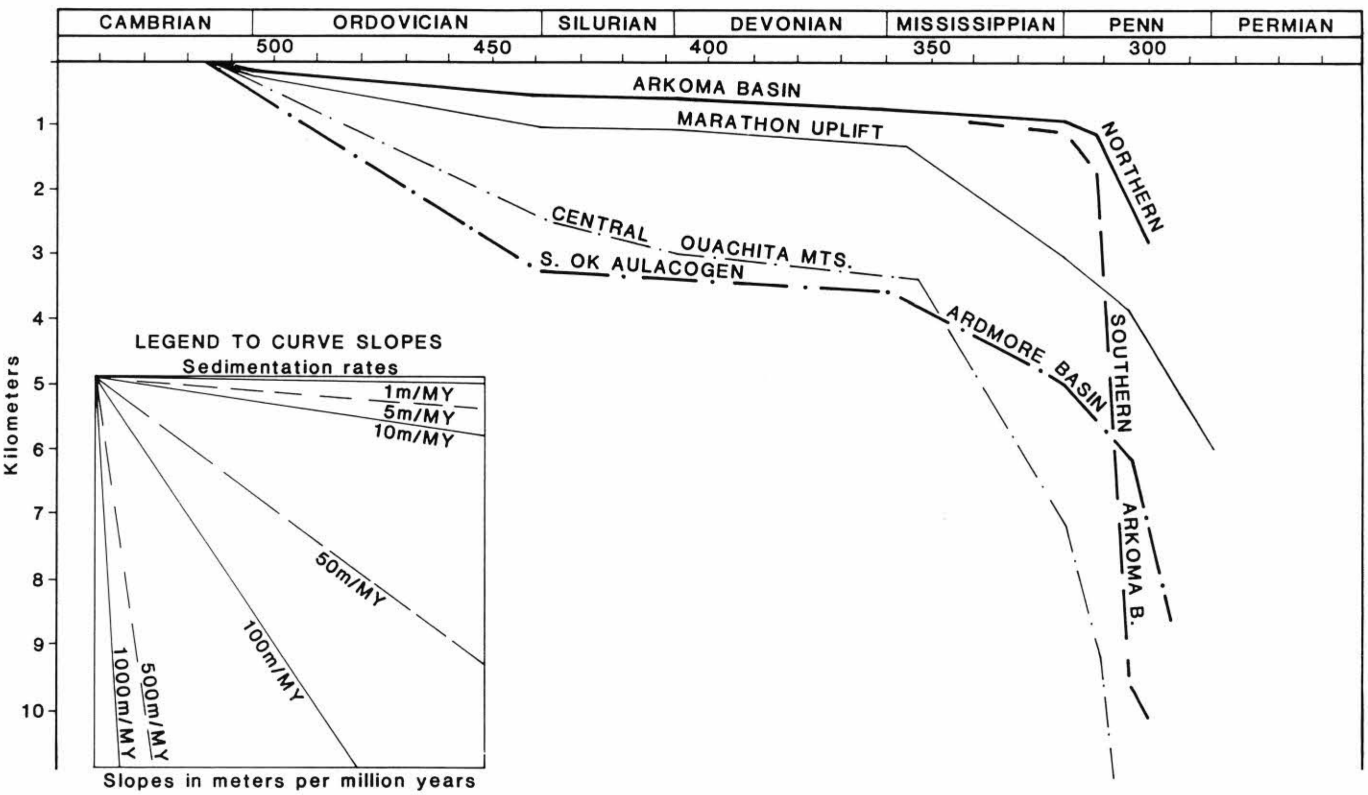


Figure 7. Sedimentation rates in meters per million years for several regions of the Ouachita system and its cratonic foreland.

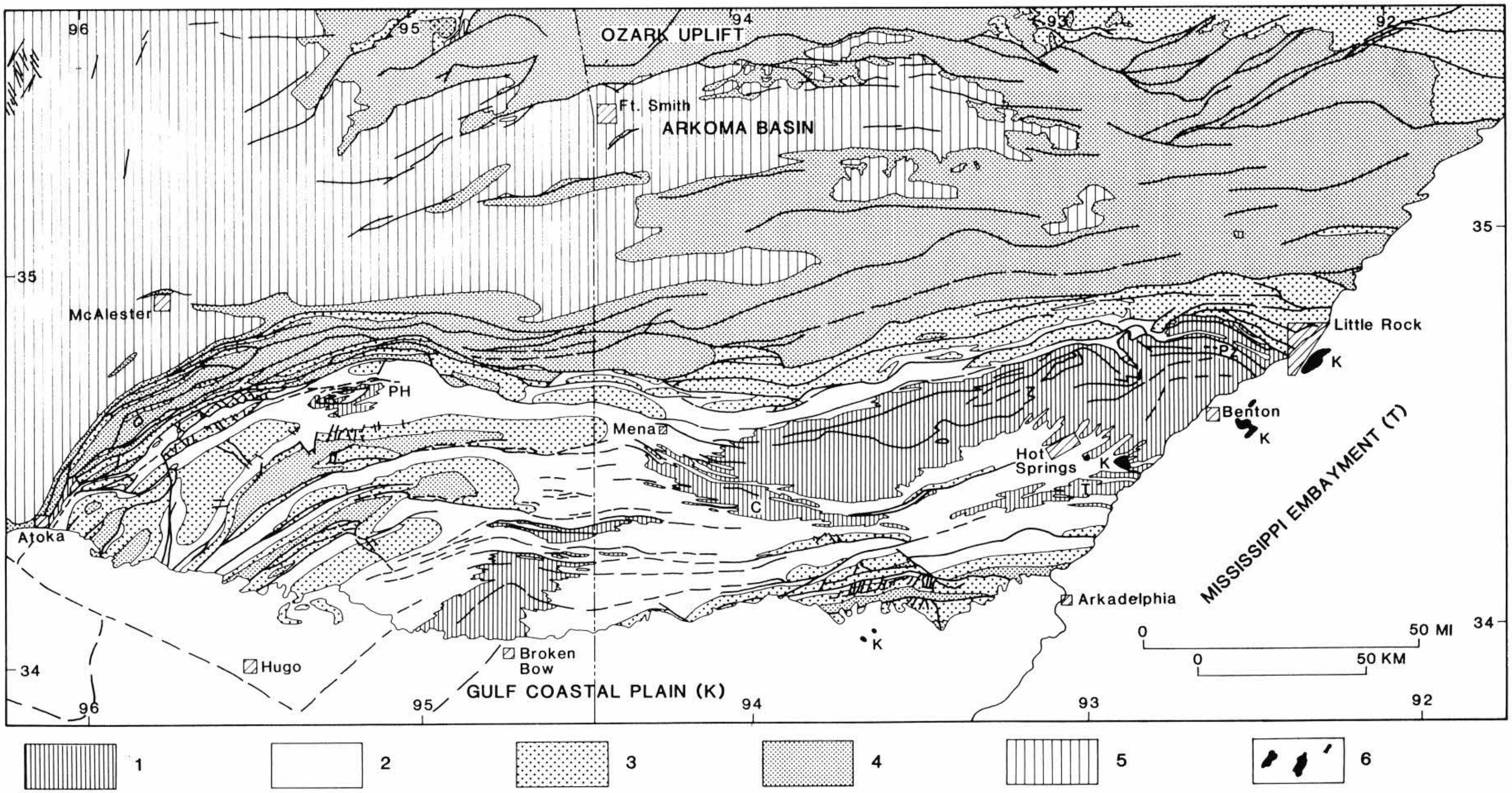


Figure 8. Geologic map of the Ouachita Mountains and the Arkoma basin (simplified after Plate 8, Hatcher and others, 1989). 1. Early and middle Paleozoic (Cambrian through Early Mississippian). 2. Middle and Late Mississippian (Stanley Group of Ouachita facies). 3. Morrowan (Jackfork Group and Johns Valley Formations of Ouachita facies). 4. Atokan. 5. Desmoinesian. 6. Intrusives and/or olistoliths of Paleozoic (Pz) and intrusives of Cretaceous (K) age. PH = Potato Hills, C = Cossatot Mountains, T = Trap Mountains.

bedded in a shaly matrix, they were more readily and more intensely tectonized, especially when these shales became the site of bedding plane faults. Furthermore, it is quite probable that deformation occurred before the flysch was completely dewatered, because the thrust front proceeded in a marine environment. Thus, the deformational fabrics and textures take on many characteristics of mélange-wedges of accretional subduction prisms. This similarity to certain accretionary phenomena of modern deep-sea trench walls has led Viele and Thomas (1989) to propose a model that makes tectonic units out of many of these disrupted, conglomeratic elements. Only very detailed stratigraphic and paleontologic subdivision and mapping will verify or contradict these concepts, which would require that the active, accretionary thrust front had advanced northward into the vicinity of the frontal Ouachitas already in early Morrowan time.

In the flysch series of the Marathon uplift, boulder beds have been known for many decades (Sellards, 1931) and yield some important paleotectonic information. They contain a richly varied assortment of shallow-marine boulders from the North American shelf, but also a multitude of exotic igneous and metamorphic boulders. Of the latter, Denison and others (1969) have determined Middle Devonian to Late Silurian ages, indicating that these boulders could not have come from the North American side of the basin but must have originated from some orogenically uplifted foreign terrane to the south. Recently, unmetamorphosed limestone cobbles of Middle Cambrian age were discovered in a conglomerate band within the Haymond Formation (Palmer and others, 1984); this is an age that is older than any sedimentary rocks known today from the adjacent platform region (e.g., Llano uplift and surrounding subsurface) or from the underlying basinal facies. These boulders indicate (a) that a carbonate platform had been established in the Middle Cambrian on the passive margin, and (b) that they were evidently fed into the basin from a more seaward position of provenance, which must have become buried by the allochthonous thrust sheets.

STYLES OF DEFORMATION

Surface maps from the two major outcrop regions (Figs. 8, 9, 10), amplified by subsurface reflection seismic and borehole information, are the principal sources from which to draw conclusions on the structural styles and tectonic units of the Ouachita system. The system is bordered along its front by a north- and northwest-vergent thrust and fold belt containing multiple and, at some places, folded thrust sheets (Fig. 11; Flawn and others, 1961; King, 1937; Hendricks and others, 1947; Blythe and others, 1988; Bush and others, 1977). North of the Ouachita Mountains, a mildly compressed thin-skinned fold belt, occupying the southern Arkoma basin, lies in front of the highly telescoped frontal imbricated thrust belt. North vergence prevails throughout most of the Ouachita Mountains, with the exception of the central uplifts or core areas (Broken Bow and Benton uplift) where a complex strain history has resulted in south-vergent folds. South of the Benton uplift, north vergence is again present and appears to continue beneath the Mesozoic cover as far as north Texas and Louisiana, where seismic control becomes inconclusive. Northwest vergence also persists throughout the Marathon thrust belt and disappears under the Mesozoic of the Gulf Coastal Plain. Thus, the true width of the thrust belt is unknown.

The vertical distribution of competent, load-bearing units and ductile, shaly, and silty intervals in the outcrop regions is characterized by an abundance of ductile formations in the older deeper parts of the section, and a dominance of very thick, massive, and mostly competent rocks in the Morrowan and younger flysch sequence (Figs. 5, 6, 11). It is evident that this general ductility/competence distribution leads to a structural style that differs considerably from the well-known and much-described style of miogeoclinal thrust belts, which have their main load-bearing members deep in the section and their more ductile units in the younger rocks.

Inspection of surface geologic maps (Figs. 8, 10) and cross sections (Fig. 11) shows that the youngest structural level (upper Mississippian and younger) forms broad synclines and sharp, usually faulted anticlines involving the 4,000+ m of Late Mississippian and younger turbidite fan complexes. In the Ouachitas this competent section comprises the upper Stanley Group, the Jackfork Group, the Johns Valley Formation, and the Atoka Formation. In the Marathons a similar style is produced by the combined competent rocks of the upper Tesnus, Dimple, and Haymond Formations. In plunge view this fold train has the shape of cuspate-lobate folds (Ramsay and Huber, 1987) with the lobes facing the more ductile, less viscous underlying units.

In the older, shale-dominated section, the more competent units are quite thin and tend to seek out their own preferred dominant wave lengths, which are much smaller. Because of the well-bedded nature of the competent rocks (e.g., cherts, well-bedded limestones, and sandstones), chevron folds are common. On plunging map configurations and in mesoscale exposures, several competing wavelengths can often be observed to form complex polyharmonic fold trains (Currie and others, 1962; Ramsay and Huber, 1987). In the Ouachita Mountains, the leading competent units in this overall ductile package are the Crystal Mountain Sandstone, the Blakely Sandstone, and the combined Bigfork-Blaylock-Arkansas Novaculite section. In the Marathons, the Caballos stands out as the dominant competent unit that controls the size and shape of folds.

The intervening ductile shale and limestone sections, which separate differently folded units, must accommodate major and minor disharmonic detachments. Because of the poor exposure of these units, the mechanical details of the decoupling are usually not resolvable. Some authors show these detachment zones as bed-parallel faults or "domain boundaries" (e.g., Fig. 10; Haley and others, 1976; Muehlberger and Tauvers, 1989), while others might interpret them as submarine unconformities, especially where the older rock units seem to be more tightly folded (Welland and others, 1985).

Numerous major thrust faults are recognized (Figs. 9, 11) where contacts of competent marker beds are displaced. Poor

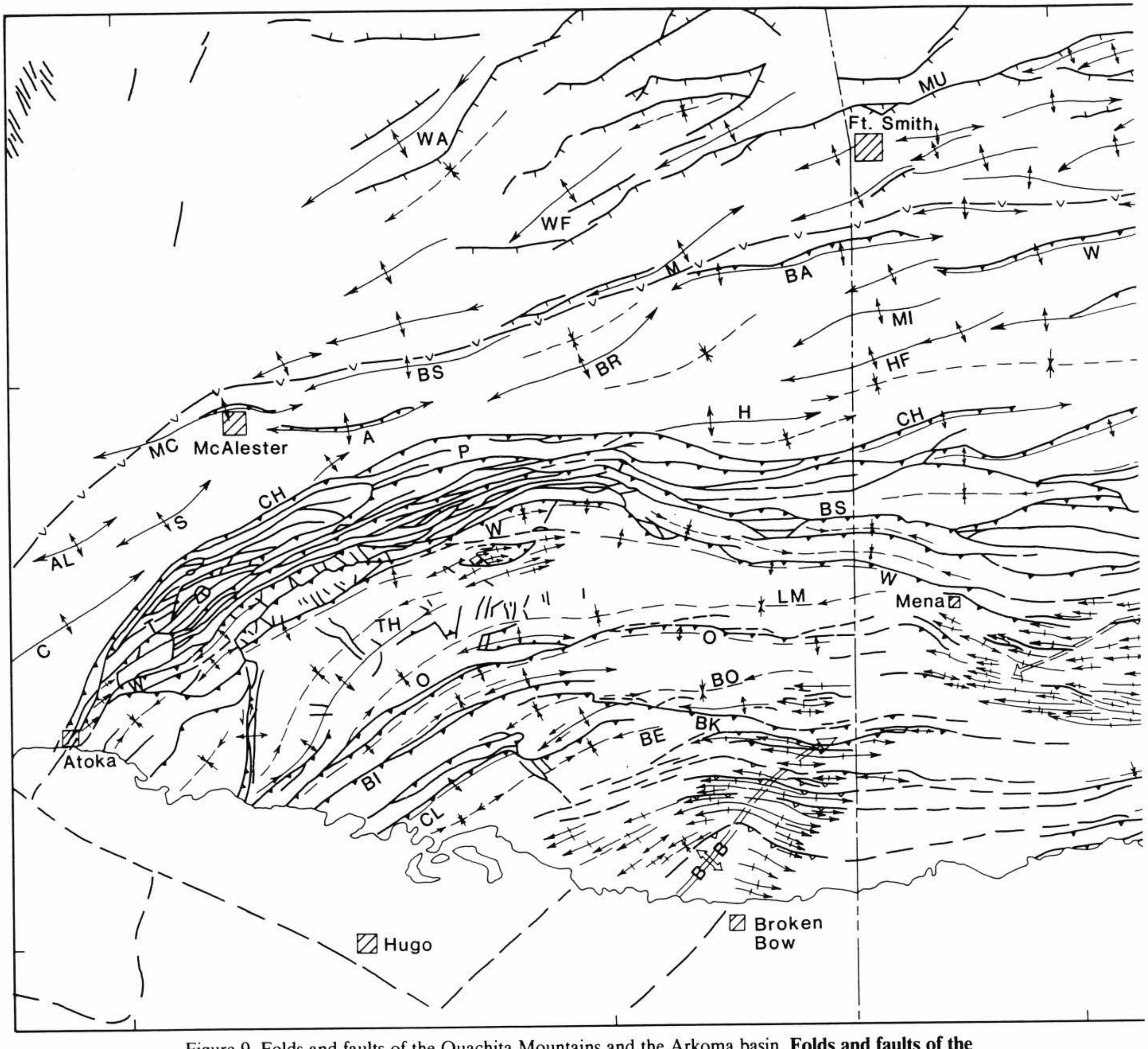
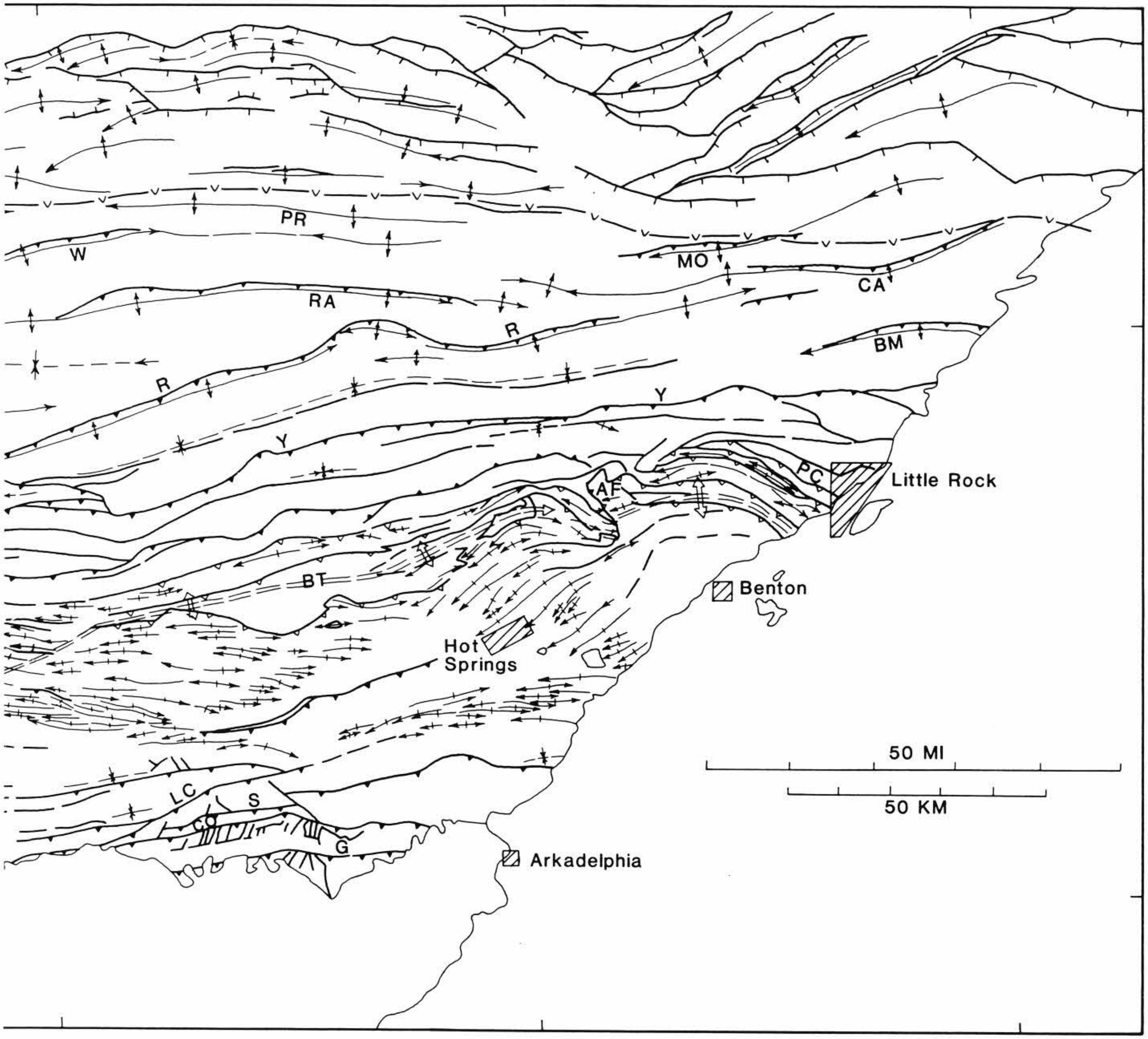


Figure 9. Folds and faults of the Ouachita Mountains and the Arkoma basin. **Folds and faults of the Ouachita Mountains:** AF = Alum Fork thrust, B = Briery fault, BB = Broken Bow uplift, BE = Bethel syncline, BI = Big One fault, BK = Boktukola fault, BO = Boktukola syncline, BT = Benton uplift, CH = Choctaw fault, CL = Cloudy fault, CO = Cowhide fault, G = Graysonia fault, LC = Little Creek fault, LM = Lynn Mountain syncline, O = Octavia fault, P = Pine Mountain fault, PC = Panther Creek fault, S = Skelton fault, T = Ti Valley fault, TH = Tuskahoma syncline, W = Windingstair fault, Y = Y City fault. **Folds and faults of the Arkoma basin and the southern Ozark uplift:** A = Adamson anticline with Carbon fault, AL = Ashland anticline, BS = Burning Springs anticline, BA = Backbone anticline, BM = Bayou Meto anticline, BR = Brazil anticline, C = Coalgate anticline, CA = Cadron anticline, H = Heavener anticline, HF = Hartford anticline, M = Milton anticline, MC = McAlester anticline and Penitentiary fault, MI = Midland anticline, MO = Morrilton anticline, MU = Mulberry fault, PR = Pine Ridge anticline, R = Ross Creek fault, RA = Ranger anticline, S = Savanna anticline, W = Washburn anticline, WA = Warner horst, WF = Whitefield uplift.


PR
W
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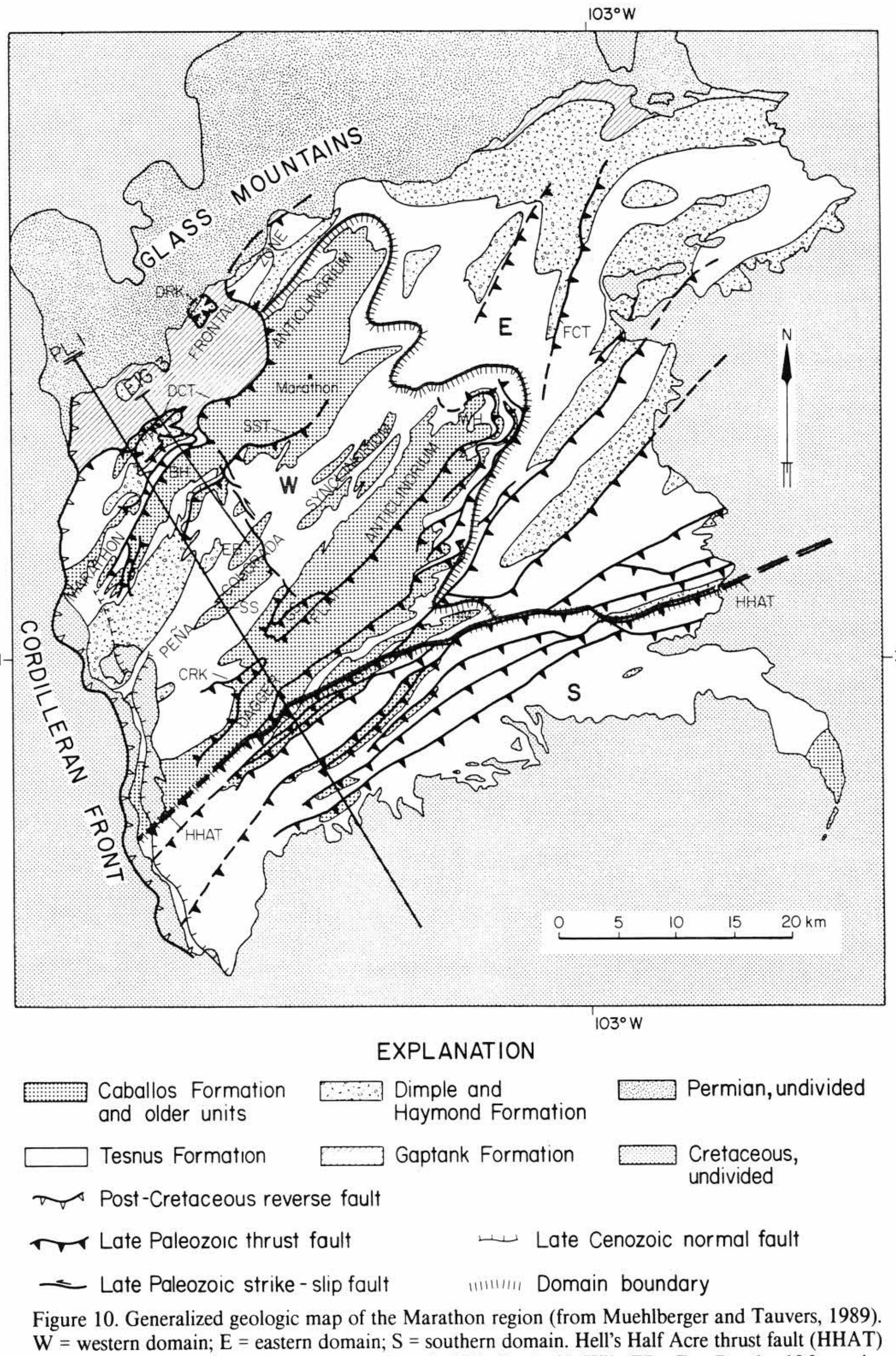

Figure 10. Generalized geologic map of the Marathon region (from Muehlberger and Tauvers, 1989). W = western domain; E = eastern domain; S = southern domain. Hell's Half Acre thrust fault (HHAT) marks the northern border of the southern domain. BH = Beckwith Hills; EB = East Bourland Mountain; GS = Garden Springs area; HM = Horse Mountain; PH = Payne Hills; SS = Simpson Springs Mountain; WH = Warwick Hills; CRK = Combs Ranch klippe; DRK = Decie Ranch klippe; DCT = Dugout Creek overthrust; FCT = Frog Creek thrust.

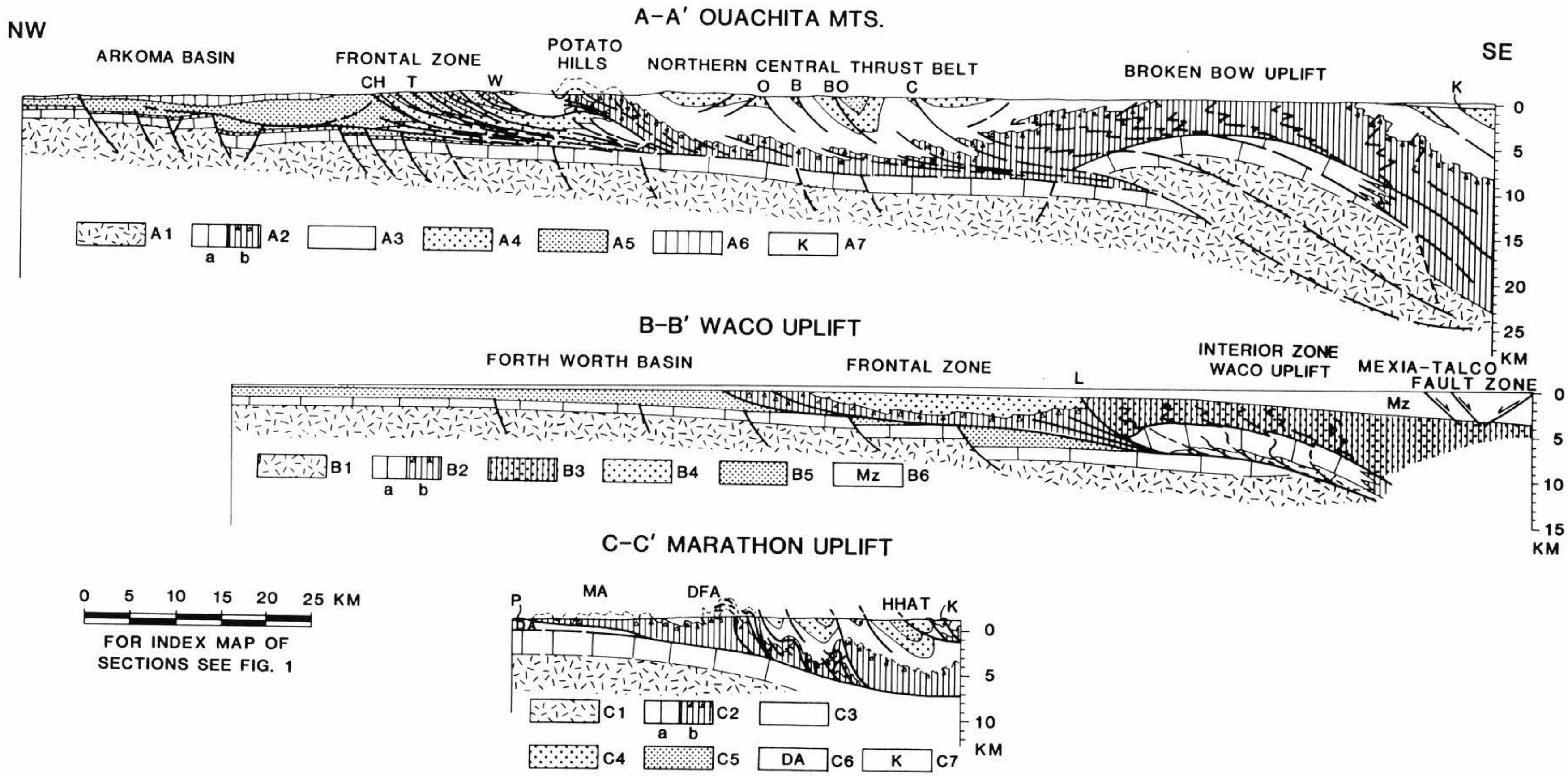

Figure 11. Generalized cross sections across the frontal Ouachita system in Oklahoma and Texas (simplified from Plate 11, Hatcher and others, 1989). Explanation: **Section A–A′:** A1, Precambrian basement; A2a, Cambrian through Mississippian platform rocks; A2b, Cambrian through Early Mississippian deep water rocks (pre-orogenic), triangles represent Ordovician to Devonian cherts; A3, Mississippian flysch; A4, Morrowan flysch and platform rocks; A5, Atoka Formation; A6, Desmoinesian; A7, Cretaceous. Abbreviations: CH = Choctaw fault, T = Ti Valley fault, W = Windingstair fault, O = Octavia fault, B = Boktukola fault, BO = Big One fault, C = Cloudy fault. **Section B–B′:** B1, Precambrian basement; B2a, Cambrian and Ordovician platform rocks; B2b, Cambrian through Early Mississippian deep-water rocks (preorogenic); B3, Cambrian through Carboniferous deep-water rocks; undifferentiated; B4, Morrowan and Atokan flysch; B5, Atoka Formation; B6, Mesozoic undifferentiated. Abbreviations: L = Luling front. **Section C–C′:** C1, Precambrian basement; C2a, Cambrian through Mississippian platform rocks; C2b, Cambrian through Early Mississippian deep-water rocks (preorogenic); C3, Mississippian flysch; C4, Morrowan and Atokan flysch; C5, Haymond Formation; C6, Dugout allochthon, Late Pennsylvanian (DA); C7, Cretaceous. Abbreviations: P = Permian, MA = Marathon anticlinorium, DFA = Dagger Flat anticlinorium, HHAT = Hell's Half Acre thrust.

mappability in the shales usually prevents us from determining whether these major faults actually cut through the entire sequences, including the thick ductile units, or whether they disappear in a multitude of minor faults in the ductile, cataclastically flowing shales. The major thrust faults, where mappable by seismic or borehole control, appear to be listric with little evidence of angular ramps and flats. At the front of the Oklahoma salient of the Ouachita Mountains a great amount of shortening is accomplished by numerous, tightly imbricated thrust slices (Hendricks and others, 1947; Arbenz, 1989).

Late-stage footwall thrusting (duplexing) has resulted in folding of thrust faults. This is particularly well displayed in the Marathon region, where King (1937) demonstrated by down-plunge projection the folding of multiple thrust sheets. In the Ouachita Mountains, similar map configurations indicated folded faults have been recognized, for instance in the Potato Hills (Miser, 1929; Arbenz, 1968).

A very large duplex structure of some 45 km width has been interpreted to underlie the central Benton and Broken Bow uplifts (Fig. 11), where COCORP (Consortium for Continental Reflection Profiling) and proprietary reflection seismic surveys have detected a large antiformal feature beneath the surface anticlinoria (Nelson and others, 1982; Lillie and others, 1983; Arbenz, 1984; Blythe and others, 1988). The special characteristics of the surface geology of these central uplifts had been noticed and described early in this century (Honess, 1923; Purdue and Miser, 1923; Miser and Purdue, 1929) and are summarized by Nielsen and others (1989). The most noteworthy features of these uplifts are summarized below:

a. They represent the major regional outcrop areas of pre-Mississippian rocks in the Ouachita Mountains. The axis of the uplifts trends obliquely across the regional grain of the fold-and-thrust patterns (Fig. 9).

b. Folds are mostly south overturned to recumbent, but show transition to north overturning along the northern and southern margins of the uplifts (Zimmerman, 1984, 1986).

c. Most major faults show down-to-the-north displacements. Honess (1923) shows the faults on his cross sections to have "normal" fault geometry. At two significant places on the Benton uplift, down-plunge views demonstrate major north-vergent thrusting of several kilometers with an overprint of south-vergent folding and minor faulting, for example, the Alum Fork fault (Haley and others, 1976; Nielsen and others, 1989) and the major unnamed fault east of Mount Ida (Soustek, 1979), which itself became refolded. This relationship implies an initial north-vergent thrusting, followed by a complex event of south overturning of both thrust plates and fault surfaces. Nielsen (in Nielsen and others, 1989) believes that the inverse sequence may apply in the Broken Bow uplift. Wherever south overturning is present, it is accompanied by a pervasive north-northwest–dipping cleavage.

d. A large area of both central uplifts shows a low-grade metamorphism (Fig. 12), accompanied by elevated vitrinite reflectance (Desborough and others, 1985; Guthrie and others, 1986; Houseknecht and Matthews, 1985; C. G. Stone, personal communication, 1988), quartz novaculitization (Keller and others, 1985), and by a pervasive invasion by quartz veins (Miser, 1959). Denison and others (1977) reported late Paleozoic radiometric ages for this metamorphism, in places overprinting an earlier Devonian event.

e. Leander and Legg (1988) described the results of a borehole near the top of the Broken Bow uplift, which penetrated 3,570 m of metamorphosed Ouachita facies and then entered and bottomed in foreland facies carbonates at a total depth of 5,784 m. This facies relationship involving the above-mentioned antiform confirms very similar conditions encountered near Waco, Texas, in a borehole described by Vernon (1971) and Nicholas and Rozendal (1975) drilled on an antiformal subsurface structure named the Waco uplift (Figs. 2, 11), where marbleized foreland(?) carbonates and underlying basement rocks were also encountered beneath Ouachita facies rocks. These authors proposed a model of a duplex structure of foreland basement and its cover below a refolded and faulted Ouachita allochthon, a model that was later modified by Nelson and others (1982) in their interpretation of the COCORP seismic survey in the western Benton uplift.

The central uplifts of the Ouachita Mountains and the Waco uplift are parts of a regional positive gravity anomaly or a major gravity gradient (Nicholas and Rozendal, 1975). This gravity feature has received much attention since and is now generally recognized as the southern outboard limit of the North American Paleozoic continental crust (Lillie and others, 1983; Kruger and Keller, 1986; Keller and others, 1989). The crust south and southeast of this gravity anomaly has been modeled to be attenuated transitional or possibly even oceanic crust on which Ouachita facies rocks were deposited. This implies further that the thrust sheets that carry the northernmost occurrence of this facies had to originate south of the central uplifts (i.e., the thrust sheets overlying the Ti Valley, Briery, and Y-City faults of the Ouachitas).

SUBSURFACE CONNECTIONS

The vast area of the Ouachita system that lies beneath late Paleozoic, Mesozoic, and Cenozoic rocks of the Gulf Coastal Plain is still poorly known, and a concise picture of the hinterland and the Pangean suture zone cannot yet be made. The frontal elements, especially in the segment west and south of the Ouachita Mountains, have been penetrated by hundreds of wells as the search for hydrocarbons in the adjacent foreland proceeds beneath the thrust front (Flawn and others, 1961). Also, because small amounts of hydrocarbons have been discovered in Ouachita-facies rocks (Chenoweth, 1989) since the mid-1970s, geophysical exploration and drilling have intensified, but few results have been published (Petroleum Information, 1985). Another problem in interpreting subsurface data is the virtual impossibility of differentiating from drill cuttings the many formations of the flysch series, or of separating them from middle and older Paleozoic siltstones and shales; this leaves only the cherts as somewhat diagnostic lithologies. Around 70 km south and east of the thrust front, usually just southeast of the above-mentioned gravity anomaly, the base of the postorogenic rocks descends to great depths, and borehole information remains very sparse. Nicholas and Waddell (1989) and Thomas (1989) have summarized all existing information and concepts, updating the fundamental framework laid three decades ago by Flawn and others (1961).

Western segment (south and west of the Ouachita Mountains)

On the basis of the visionary groundwork by Powers (1928) and van Waterschoot van der Gracht (1931a, 1931b), and a careful inspection of all existing well data, Flawn and others (1961) established the existence of a nonmetamorphic thrust zone (their "frontal zone") passing southward and eastward into a low-grade metamorphic zone (their "interior zone"). The outcrops of the Marathon belt belong to the frontal zone, and so would that part of the Ouachita Mountains lying north and west of the central Broken Bow and Benton uplifts. Nicholas and Rozendal (1975) and Nicholas and Waddell (1989) pointed out that the northwestern border of the above-mentioned gravity anomaly also correlates with the northwestern limit of the "interior zone," and that the "interior zone" may therefore be a reflection of the band of basement duplex structures recognized from the Waco uplift to the north. In the outcrop south of the Benton uplift and in many bore holes in southern Arkansas and adjacent northern Louisiana and Texas, it is evident that the metamorphism of the Benton uplift drops off to the south (Fig. 12). In south-central and western Texas the well control south of the gravity anomaly is insufficient to document a similar cooling on the seaward side of the interior zone, but the overall gravity interpretation indicates a much-attenuated crustal configuration (Kruger and Keller, 1986; Keller and others, 1989).

A somewhat special feature along the postulated margin of the North American craton in the subsurface of west Texas is the Devils River uplift (Fig. 2), a thrust-faulted basement uplift with a mildly metamorphosed foreland-facies cover (Nicholas, 1983), but without a Ouachita facies allochthon overlying the uplift. It occupies the same regional gravity anomaly, and its metamorphism places the Devils River uplift into the Ouachita interior zone (Keller and others, 1989). The lack of a Ouachita-facies allochthon can be ascribed to pre-Mesozoic uplift and erosion probably during the final stages of the continental collision.

Another tectonostratigraphically significant feature of the Ouachita hinterland is the existence of a large, downbent, postorogenic successor basin (or "episutural" basin, Bally, 1975), in the region of southern Arkansas and northeast Texas (Fig. 2), the base of which rests with angular unconformity on nonmetamorphic flysch. The northern margin of this basin is fairly well delin-

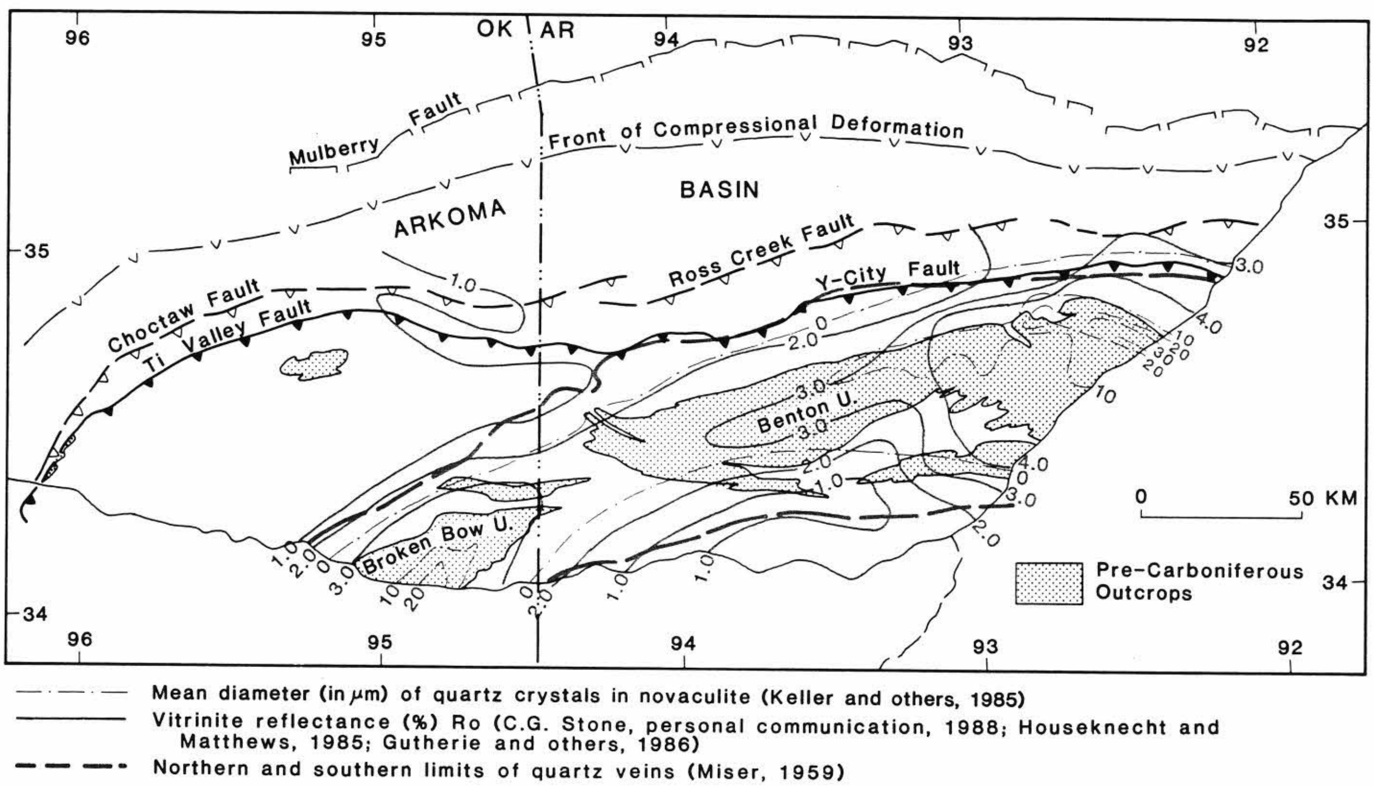

Figure 12. Thermal anomaly of the central uplifts of the Ouachita Mountains manifested by quartz recrystallization, vitrinite reflectance, and quartz veins (from Desborough and others, 1985; Guthrie and others, 1986; Houseknecht and Matthews, 1985; Keller and others, 1985; Miser, 1959).

eated by boreholes and by exploration seismic surveys, and swings in an arc from northeastern Texas to northeastern Louisiana. The eastern, southern, and western limits of this basin are not defined as yet. The youngest synorogenic flysch rocks, truncated by the successor basin sequence, are quite probably Atokan (judging from the outcrops to the north); the oldest rocks of the successor basin consist of a shallow-marine, fossiliferous Desmoinesian carbonate sequence, which grades upward into Late Pennsylvanian and Permian marine carbonates and clastic rocks. This basin was described by Paine and Meyerhoff (1970), Vernon (1971), and Woods and Addington (1973), and was given the somewhat inappropriate name "Texarkana platform." The most significant aspect of this basin is the fact that it dates the latest orogeny in this part of the hinterland with great accuracy as late Atokan. Rocks of the same age and facies have been penetrated by several wells on the Sabine uplift, a long-known domal structure straddling the state line between Texas and Louisiana. There the Desmoinesian carbonates rest on late Paleozoic rhyolite porphyry and tuff, which in turn lie unconformably on Carboniferous flysch (Fig. 2; Nicholas and Waddell, 1989).

In the states of Chihuahua, Coahuila, Nuevo Leon, and Tamaulipas of northern Mexico, rocks from all zones of the Ouachita system and its foreland are known from a number of small outcrop inliers and some exploratory wells in the frontal portions of the Sierra Madre Oriental (Flawn and others, 1961; Handschy and others, 1987). Because of their small size and wide scatter, these occurrences can give only some general constraints to the still-speculative southwestern continuation of the Ouachita system.

Eastern segment

Subsurface control in the Ouachita system between the eastern Ouachita Mountains and the juncture with the Appalachian system remains limited (Thomas, 1976, 1977a, b, 1985, 1989). Subsurface information on the adjacent foreland has delineated the structure and stratigraphy of the Black Warrior Basin as a homoclinal, extensionally step-faulted south flank of the southeastern midcontinent. Much less has been published (Thomas, 1989) about the Reelfoot–Mississippi Valley basin, mainly because commercial accumulations of hydrocarbons have yet to be discovered, and the number of deep well penetrations has consequently stayed low.

The front of the thrust belt east of the Ouachita Mountains is fairly well defined to eastern Mississippi, although identification of formational units is still tentative. Farther east the thrust front is less distinct and continues as a series of right-stepping, en echelon folds of thick Carboniferous rocks, indicating a probable right-lateral strike-slip component along this segment of the Ouachita thrust front. A more interior slate belt, probably a continuation of the central uplift of the Ouachitas, can be traced into central Mississippi (Thomas, 1989).

Because of its much more distinct lithofacies and excellent resolution on proprietary seismic sections, the southwestern and western continuation of the subsurface Appalachians can be traced successfully into central Mississippi, where the information is finally lost beneath the Mesozoic Mississippi salt basin. From these data it is evident that the subsurface thrust system of the Appalachians cuts off and overrides the southeastern extension of the Black Warrior Basin as well as the Ouachita fold and thrust belt farther to the west (Thomas, 1989).

TIMING OF OROGENIC DEFORMATION

The earliest indirect evidence of a constriction of the Ouachita depositional basin and an orogenic uplift of one or more regions of provenance comes in the Meramecian or earliest Chesterian (Mississippian), with the rather sudden influx of large amounts of terrigenous clastic sediments (Fig. 7). The majority of these sediments are being funnelled down the axis of a deep-water trough that parallels the continental margin. The basal contact of the flysch is gradational and regionally conformable. The nearly simultaneous appearance of volcanic tuffs in the southern outcrop of the Ouachita Mountains can be used as confirming evidence for subductional events to the south, while the occurrence of sedimentary, bedded barite deposits at the very base of the flysch sequence in a band along the southern border of the Benton uplift (Hanor and Baria, 1977) may still be related to extensional pre-flysch phenomena rather than to orogenic ones.

Neither in the Ouachita Mountains nor in the Marathon region can one detect any unequivocal angular unconformity between the base of the flysch section and the late Atokan in tectonically undisturbed areas. Some regional wedging and pinchout of mappable units may be attributed to the depositional shape of turbidite fans. This observation of regional conformity is based on extensive mapping and stratigraphic measurements by numerous authors (e.g., Cline, 1960; Haley and others, 1976; Harlton, 1938; Hendricks and others, 1947; Honess, 1923; King, 1937; Miser, 1929; Miser and Purdue, 1929; Morris, 1974; Walthall, 1967), and implies that structural deformation had not proceeded into the region of the present outcrops. It is evident that this conclusion conflicts with Viele's (in Viele and Thomas, 1989) proposed origin of the Maumelle chaotic zone as an active accretionary prism, which would have affected the entire Ouachita flysch basin since Early Pennsylvanian.

It was pointed out earlier that some of the tectonic disharmonic detachment zones (e.g., the one in the upper Mississippian flysch) could be interpreted as unconformities. However, unpublished mapping by the author shows that the Mississippian structural disharmony never displays any significant stratigraphic cutouts, and that the disharmony does not stay at the same level of the stratigraphic section, indicating that this is indeed a tectonic feature and not an angular unconformity representing an orogenic pulse.

It was also mentioned before that the continental platform under the southern Arkoma basin subsided strongly along several basement growth faults (Fig. 11) during the Atokan (and probably before). This faulting was probably related to thrust loading

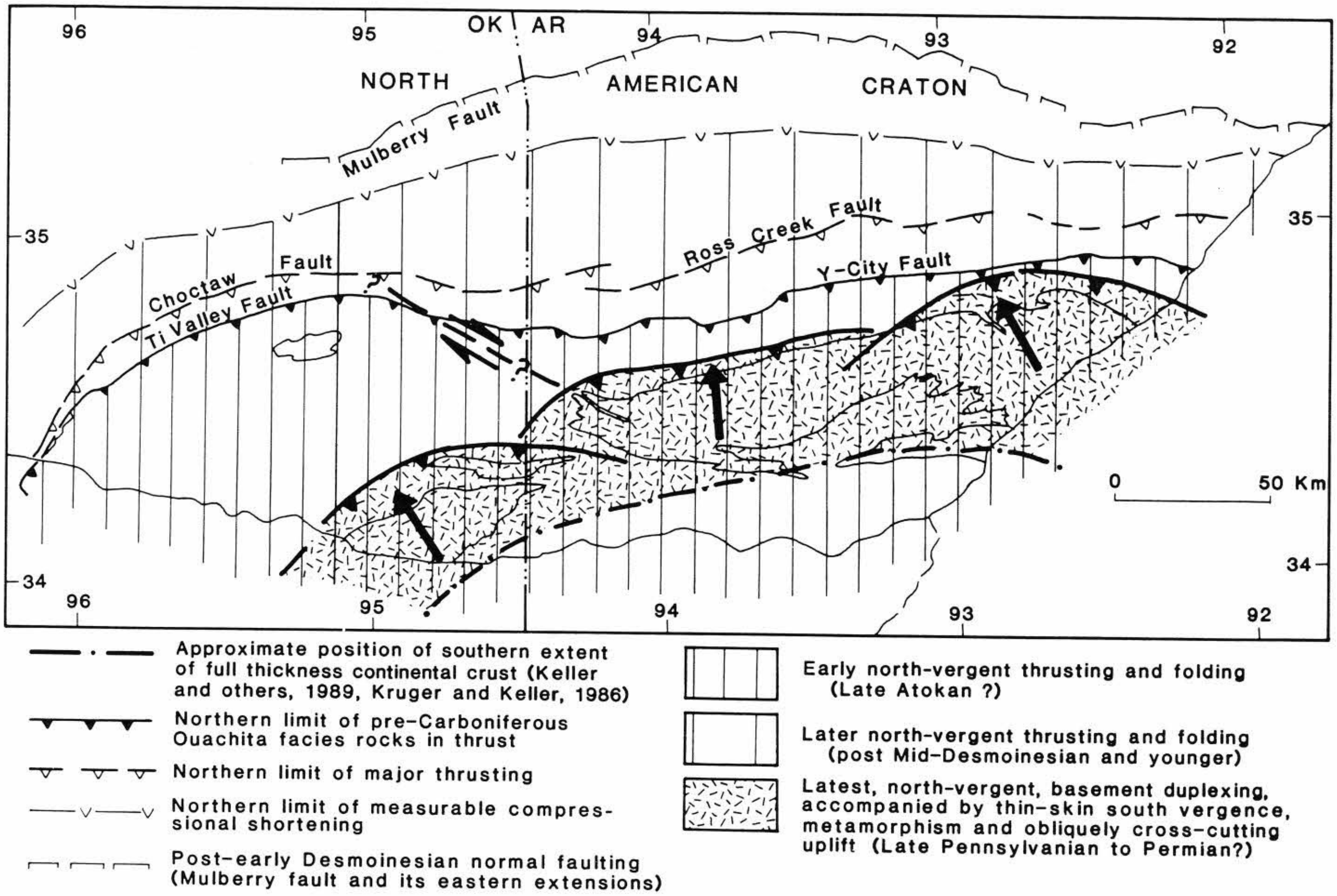

Figure 13. Tentative timing of emplacement and deformation of the major tectonic elements in the Ouachita Mountains.

during the Morrowan, which must have occurred in the hinterland, south of the present surface and subsurface control. The existence of the Desmoinesian and younger, postorogenic successor basin sequence in the subsurface of southern Arkansas and Texas gives clear evidence that Atokan-age thrusting, folding, and subsequent peneplanation had to be terminated by early Desmoinesian. The main compressional event in that part of the orogen had to happen in late Atokan, because all previous flysch units appear to be conformable.

Yet in the Arkoma basin, north of the thrust front of the Ouachita Mountains, beds as young as middle Desmoinesian are still conformably infolded in the synclines of the fold belt. An unconformity separating the Desmoinesian Hartshorne Sandstone from the underlying Atoka Formation, described by Hendricks and Parks (1950) and Haley (1966, 1968), appears still to be more closely related to the extensional growth faulting than to the compressional folding. Thus, it is evident that along the northern margin of the orogen, deformation did not start until middle or late Desmoinesian. Indirect evidence of a major tectonic event at that time comes also from the Ardmore basin, a foreland basin west of the Ouachita thrust front, where the earliest preflysch Ouachita-facies detritus appears in the eastward-coarsening Desmoinesian Devils Kitchen Conglomerate (Tomlinson and McBee, 1959; Denison, 1989).

In the Marathon region, direct evidence for the termination of the orogeny is documented by an angular unconformity between rocks of Wolfcampian and Leonardian (Early Permian) age (Ross, 1986). A Permian age for the close of the orogeny is further suggested by numerous Permian isotopic ages from west Texas to southeastern Oklahoma (Denison and others, 1977; Denison, 1989). On the basis of well-known north-south–trending zones of en echelon, right-stepping faults in north-central Oklahoma (suggesting left-lateral, minor strike slip), Melton (1930) proposed the presence of an effect of Ouachita tectonism in the foreland outside the Ouachitas in Early Permian.

Based on present information, the events of the Ouachita orogeny can be tentatively summarized as follows.

a. Earliest orogenic basin constriction and source activation for flysch sedimentation occurred in the deep hinterland in Mississippian (Meramecian to early Chesterian).

b. Orogenic sediment influx continued through Late Mississippian and Early Pennsylvanian (Morrowan and Atokan), but deformation front did not reach the presently exposed flysch basin until late Atokan.

c. Thrusting and folding deformed the central and southern Ouachita Mountains region in late Atokan, probably obducting and emplacing a Ouachita-facies allochthon on the North American continental margin at that time (Fig. 13).

d. The present deformation front was reached in mid-Desmoinesian, while in the hinterland an unconformable successor basin was established during the same period.

e. A collisional duplex structure was formed along much of the former continental margin west of the Mississippi River in Late Pennsylvanian to Permian, resulting in a band of antiformal ramp structures, an accentuation of the platform margin gravity anomaly, and a thermal event of mild metamorphism.

A more complete record of late Paleozoic orogenic pulses is preserved in the outcrops and the heavily drilled subsurface of the southern Oklahoma aulacogen, the Arbuckle Mountains, the Ardmore basin, the Criner Hills–Wichita Mountains, and the southern margin of the Anadarko basin. Repeated orogenic uplift, followed by truncation and dispersal of coarse clastic sediment, has been documented in Late Mississippian, late Morrowan, Desmoinesian, and late Missourian, with mild rejuvenation of positive structures in the Early Permian (Tomlinson and McBee, 1959; Reedy and Sykes, 1959; Latham, 1970; Brewer and others, 1983; Denison, 1982, 1989). How exactly this record of deformation relates to the Ouachita orogeny is still unresolved.

TECTONIC SUMMARY

The frontal length of the Ouachita system is some 2,100 km, of which only about 430 km are exposed in the Ouachita Mountains and in the Marathon uplift. The entire orogenic hinterland is deeply buried beneath the Gulf Coastal Plain and is known to have undergone varying amounts of extension during the opening of the Gulf of Mexico. Constraining data, especially from the hinterland, are therefore fragmentary, and the regional tectonic history remains the subject of much debate. Many authors have proposed models of the tectonic evolution for parts or all of the Ouachita system, particularly since the emergence of plate tectonics in the 1960s, but space is too limited in this summary to discuss adequately the merits of these proposals. The interested reader is referred to the following list of publications, all of which concern some geotectonic aspects of the Ouachita system: Cebull and others, 1976; Cebull and Shurbet, 1980; Graham and others, 1975; Flawn and others, 1961; Ham and Wilson, 1967; Handschy and others, 1987; Keller and Cebull, 1973; Keller and others, 1989; King, 1959, 1975a, b, 1977; Kruger and Keller, 1986; Lillie, 1984; Lillie and others, 1983; Link and Roberts, 1986; Lowe, 1985, 1989; Meyerhoff, 1973; Nelson and others, 1982; Nicholas and Rozendal, 1975; Paine and Meyerhoff, 1970; Pindell and Dewey, 1982; Rippee,1985; Ross, 1986; Thomas, 1972, 1976, 1977a, b, 1985, 1989; Viele, 1973, 1979; Viele and Thomas, 1989; Walper, 1977; Wickham and others, 1976; Zimmerman and others, 1982.

Most geologists and geophysicists tend to agree today that the southern margin of an intact and nearly autochthonous North

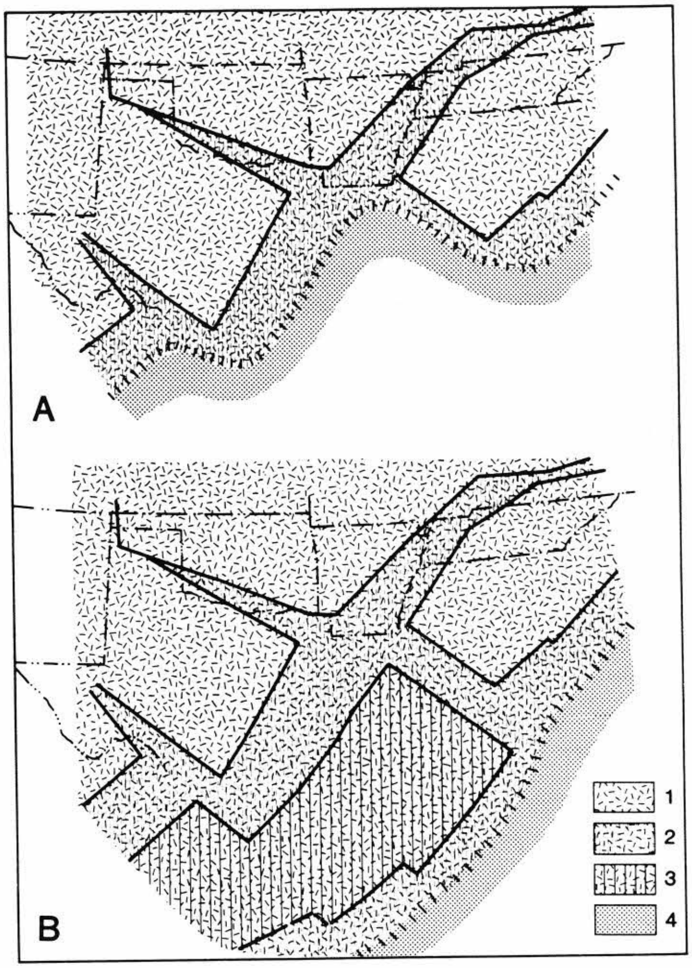

Figure 14. Two possible configurations of the southern continental margin of North America and the tectonic setting of the deep-water Ouachita basin during the early Paleozoic: A. Trailing rift-transform margin facing the open proto-Atlantic Ocean (Iapetus) with three rift basins extending into the craton (after Cebull and others, 1976; Thomas, 1977a; Viele and Thomas, 1989). B. Ouachita trough as a rift basin separated from the open proto-Atlantic by a rifted microcontinent or oceanic plateau that served as an intermittent sediment source and topographic restricting agent (after Lowe, 1985, 1989). Definition of symbols: 1, Continental crust with platform facies cover; 2, Attenuated continental, transitional, and possibly oceanic crust; 3, Attenuated continental crust (less attenuated than 2); 4, Oceanic crust.

American continental crust lay close to the present southern limb of the Benton–Broken Bow uplift in the Ouachita Mountains, and regionally just south and east of the above-mentioned gravity anomaly that extends from west Texas to the Mississippi Embayment. The deep-water oceanic basin, in which the typical preorogenic and early orogenic Ouachita facies rocks were deposited, was probably fault-bounded along its northern rim, judging from the presence of olistostromes and the abrupt appearance of the deep-water facies in the allochthonous thrust belt. That the early Paleozoic continental margin was the result of an angular breakup geometry of alternating rift and transform segments

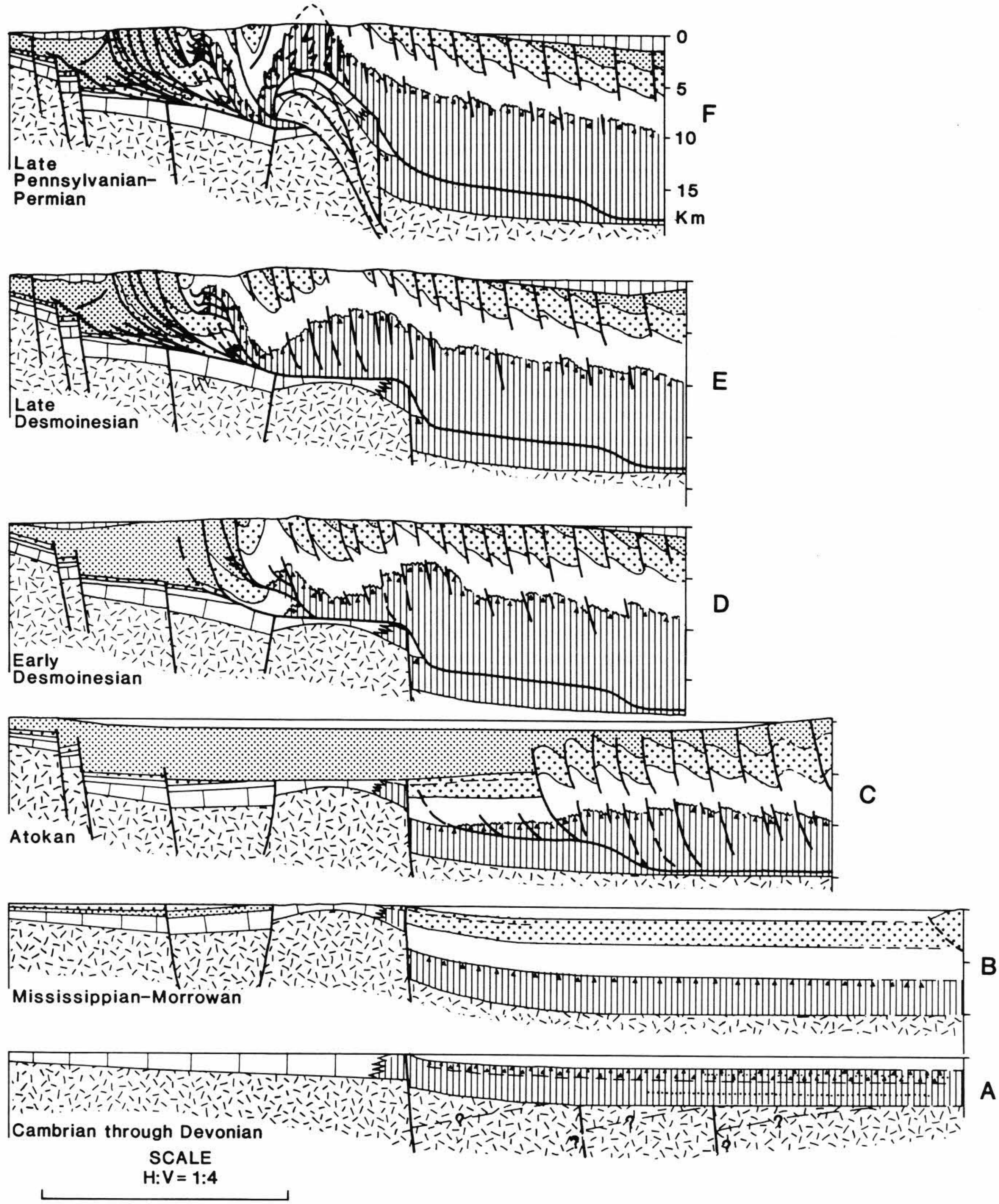

Figure 15. Diagrammatic evolution of depositional and deformational stages of the Ouachita system continental margin (along Section A–A′, Ouachita Mountains, Fig. 11). For discussion see text.

(Fig. 4) is now accepted by many authors (Cebull and others, 1976; Thomas, 1976, 1977b, 1989; Viele and Thomas, 1989). The beginning of rifting is poorly constrained, but by Middle Cambrian a continental carbonate platform had been established on the Llano promontory (Palmer and others, 1984), while igneous activity continued in the south Oklahoma aulacogen.

Less unanimity exists about the depth and size of the ocean basin, in which the Ouachita facies rocks were deposited, and therefore about the nature of the crust underlying the basin. One model (Fig. 14A), originally presented by Cebull and others (1976) and since favored by several authors (Thomas, 1976, 1977b, 1989; Kruger and Keller, 1986; Viele and Thomas, 1989), proposes that the rift-transform margin, mentioned above, became truly the southern continental margin of North America in Cambrian time, facing an open ocean floored by a Paleozoic oceanic crust. Thus, the Ouachita facies rocks of early and middle Paleozoic age represent slope, rise, and/or abyssal sediments, which became obducted and emplaced as an allochthonous mass during the closing phases of that ocean.

The other model (Fig. 14B), suggested by Lowe (1985) and favored by this author, is a plate-tectonic application to ideas dating back to Schuchert (1923), claiming that the Ouachita basin remained a wide rift basin flanked on the outboard side by one or more attenuated microcontinents or ocean plateaus, which could act intermittently as a source of clastic sediments and olistoliths. These outboard barriers could also be the cause for repeated restriction and anoxic conditions, and for basin-center turbidite fans in the rift. The water depth of such a rift could be bathyal but not necessarily abyssal, and the underlying crust could be an attenuated continental (i.e., "transitional") crust, which could exist within a few kilometers of the continental margin. Geographic alignment would associate this rift with the Reelfoot–Rough Creek–Rome trough rift system of the eastern North American craton.

Most authors agree that by the end of the Early Mississippian the Ouachita basin became a narrowing trough receiving vast amounts of clastics and some southern-source volcanics (Fig. 15B). Depending on one's preferred geotectonic model, the south flank of this basin probably was a subductional accretionary thrust front of (a) approaching Gondwana, (b) a mid-oceanic volcanic arc, (c) an unknown foreign terrane, or (d) a returning piece of North America separated from the craton by a Cambrian rifting event. Olistostromes were derived from the cratonic margin (Ouachitas and Marathons) and from the southern tectonic front (Marathons). Much of the continental platform edge from Alabama to west Texas underwent extensional faulting (Fig. 15C)—possibly triggered by continued thrust loading in the hinterland—from Morrowan into mid-Pennsylvanian time, creating deep-water conditions in several foreland basins (e.g., Arkoma and Val Verde basins).

The orogeny reached the Ouachita outcrop region in late Atokan, creating a topographic divide, which separated the depositional environments of the postorogenic Desmoinesian marine successor basin in the south from the continuing foredeep in the Arkoma basin (Fig. 15D). By late Desmoinesian the Atokan and Desmoinesian deposits of the remaining foredeep became themselves mildly compressed (Fig. 15E).

In the Marathon basin, deep-water sedimentation continued in an ever-narrowing channel (Ross, 1986) into the earliest Permian, while an encroaching allochthon transformed the basin into a thrust belt from the close of the Desmoinesian to the late Wolfcampian. Into this Late Pennsylvanian to Early Permian time bracket fall apparently all of the latest observable events affecting the continental margin west of the Mississippi River. The duplexlike basement thrusts (Fig. 15F)—which produced the central uplifts of the Ouachita Mountains (the Broken Bow–Benton uplift), the Waco uplift, the Devils River uplift, and the Sierra del Carmen (Big Bend region), with their attendant metamorphism (Denison and others, 1977; Denison, 1989) and gravity signatures (Keller and others, 1989)—all have isotopic ages indicating latest thermal imprints in the Permian. On the Benton–Broken Bow uplift this duplexing seems to be also the most plausible and opportune time for the formation of the south-vergent overturning and the formation of the pervasive north-dipping cleavage.

On a continental scale these Permian events signal the final closing of the proto-Atlantic Ocean and the assembly of Pangaea. After a period of relative stability in the Late Permian, renewed rifting in the Ouachita-Appalachian hinterland initiated the opening of the Gulf of Mexico and the North Atlantic.

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Pennsylvanian paleogeography for the Ozarks, Arkoma, and Ouachita basins in east-central Arkansas

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ABSTRACT

In early Pennsylvanian time, east-central Arkansas was part of a foreland basin developing over a downwarped Cambrian to Mississippian carbonate shelf. The deep-marine Ouachita basin lay to the south and was tectonically closing by northward advancing thrust sheets. Over 30,000 feet of clastic rocks were deposited in the Ouachita foredeep in the early Pennsylvanian.

In Morrowan time, 200 to 300 feet of carbonate and shale shelf facies accumulated in northeastern Oklahoma. Eastward in north and central Arkansas, the Morrowan facies are mainly fluvial, deltaic, and shallow-marine, quartzose clastic rocks up to 2,000 feet thick. In the Ouachita basin to the south, 7,500 feet of Morrowan quartzose sandstones and shales were deposited as submarine fan, basin plain, and slope facies. Morrowan clastics were derived from the northeast via fluvial channels that cut across the Illinois basin and from the rising Appalachians to the east and southeast. In the Ouachita basin, sediment transport directions were dominantly to the west. In Atokan time, 600 to 6,000 feet of north-derived fluvial, fluvial-deltaic, and shallow-marine clastic rocks were deposited in the Ozarks and northern Arkoma basin areas. To the south, over 20,000 feet of shallow- to deep-marine clastic rocks accumulated in the center of the closing foreland basin. South of the Atokan shelf, over 10,000 feet of north- and east-derived turbidite sandstones and shales were deposited as submarine fans, basin plain and slope facies in early Atokan time. The foreland basin shallowed through Atokan time with the upper Atokan clastics being solely of fluvial, deltaic, and shoreline facies.

The foreland basin gradually closed due to continental collision and was filled primarily from east to west. A few hundred feet of Desmoinesian fluvial and deltaic facies, originally deposited by westward flowing streams, are the last deposits preserved in central Arkansas. Afterwards, continental collision culminated in the middle Pennsylvanian to early Permian and uplift and denudation of much of the basin followed.

INTRODUCTION

Mississippian and Pennsylvanian strata form a nearly continuous belt of rocks deposited in a late Paleozoic foreland or foredeep basin(s) that extend from the southern Appalachians to the Mexican border in southwest Texas (Flawn et al., 1961). Intersecting this foreland belt are the Anadarko-Ardmore aulocogen in Oklahoma and Texas, and the Reelfoot rift in Arkansas. This foreland belt is approximately 1,240 miles long and is largely buried except for outcrops in the Appalachian Valley and Ridge province, Oua-

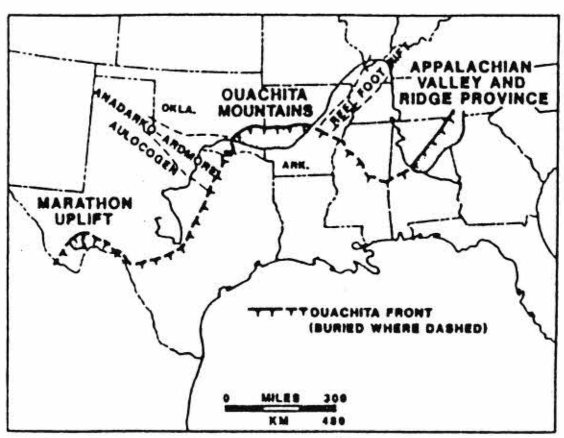

Figure 1. Map of southeastern United States showing outcrop distribution of late Paleozoic foreland basin deposits, trace of Ouachita front, and location of related aulocogens. Modified from Moiola and Shanmugam (1984).

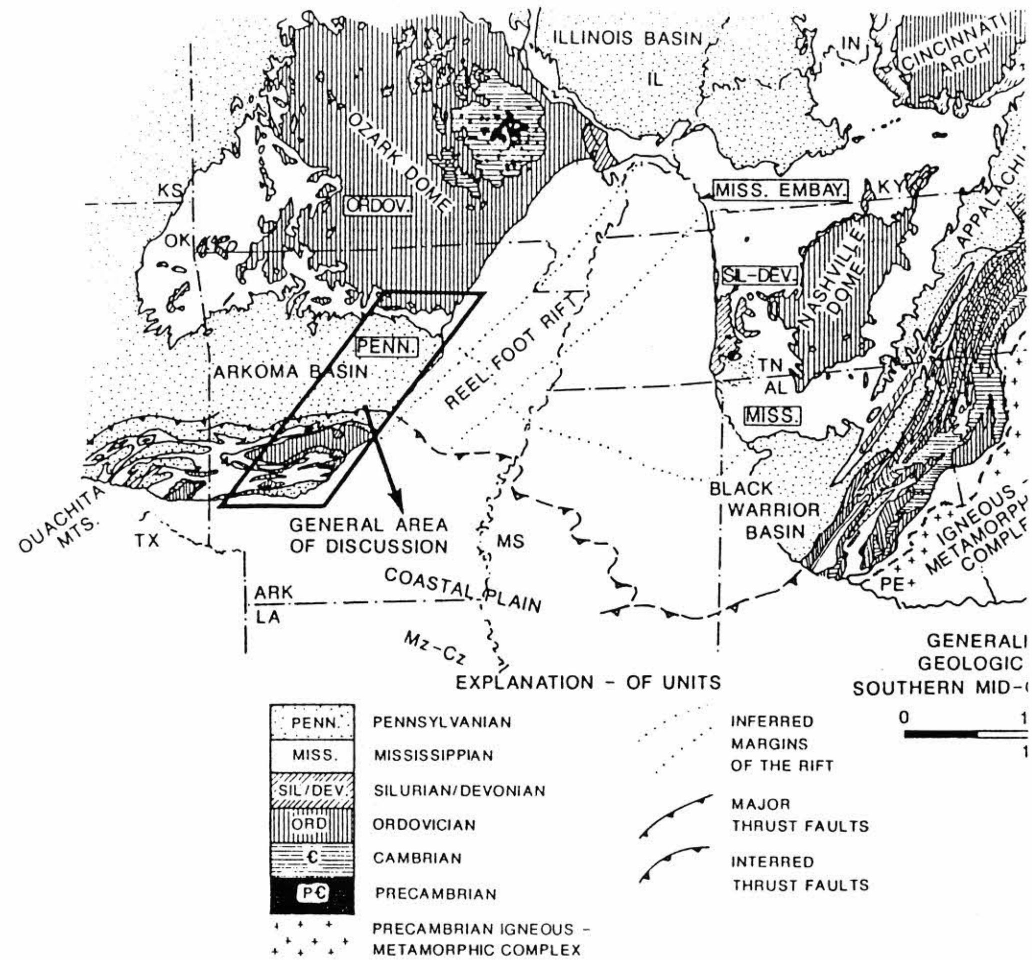

Figure 2. Generalized geologic map of part of the southeastern United States show units, major structural features, and area discussed in this paper.

chita Mountains/Arkoma basin/Ozark region, and the Marathon uplift (Fig. 1). The rocks of the Arkansas portion of the foreland basin are locally over 45,000 feet thick and conformably overlie Cambrian to early Mississippian rocks (Fig. 2). These older rocks were deposited along an Atlantic-type margin during which a period of reduced sedimentation (a starved basin facies) occurred from the latest Ordovician to earliest Mississippian. Sedimentation rates increased in the late Mississippian and very rapid rates of sedimentation occurred in the early Pennsylvanian (Atokan time). During the Atokan, large volumes of sediment derived from the east and north filled the subsiding basin. The high rate of subsidence was due to thrust-sheet loading during the closing of the Ouachita basin from south to north. At the close of the Paleozoic, contine culminated with the emplaceme ment duplex under the core of t (Broken Bow-Benton uplift). The f was in part uplifted and its uplands denudation (Fig. 3). Discussions tectonics and geologic history of t be found in Viele, 1973; Graham Wickham et al., 1976; Briggs and I and Arbenz, 1984.

The purposes of this paper are: 1) scribe the Pennsylvanian stratigra tonic relationships in east-central A 2) to present some interpretative Pennsylvanian paleogeography f The area of discussion extends ship, Hot Spring County, to Batesv

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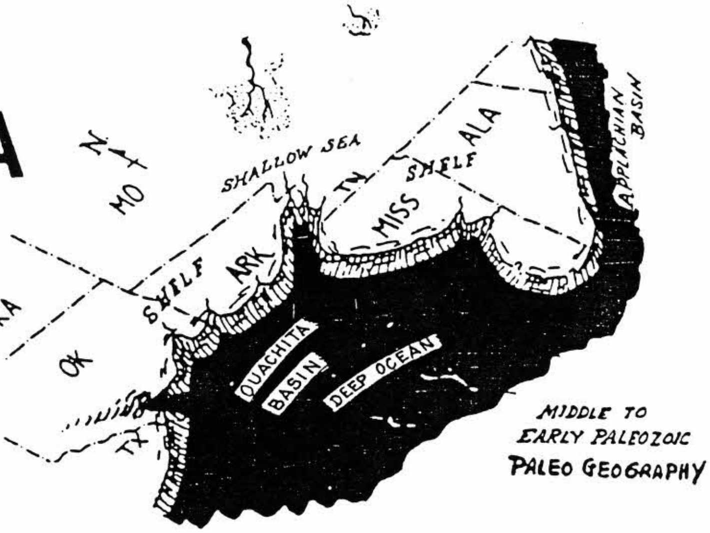
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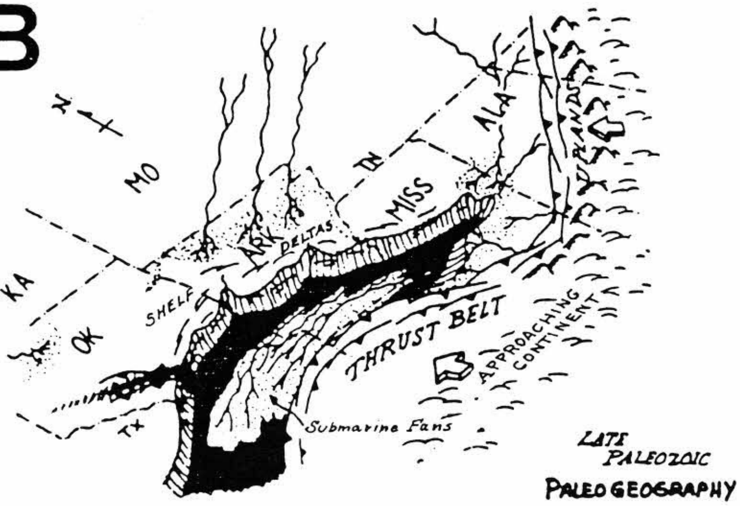
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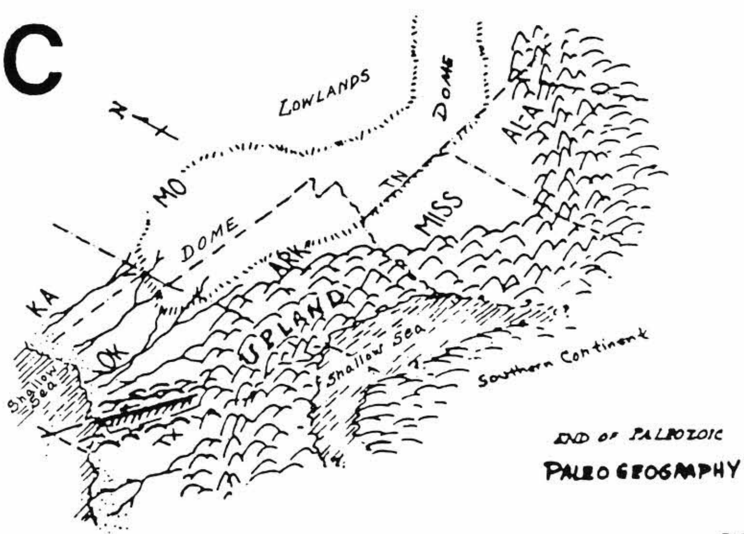
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PALEO GEOGRAPHY

Figure 3. Paleogeographic maps in the Appalachian and Ouachita basins for the early to late Paleozoic. A) Cambrian to Mississippian interval representing a dominantly passive Atlantic-type continental margin. B) Mississippian to early Pennsylvanian interval, showing the closure and uplift of the southern Appalachian basin and narrowing of the Ouachita basin by northward advancing thrust sheets. C) Middle Pennsylvanian to Permian interval depicting the ultimate closing of the Ouachita basin, reactivation of the Ozark dome, and uplift and erosion of the foreland basin. Figure A modified from Chamberlain, 1978.

dence County, Arkansas, and includes the easternmost exposures of Pennsylvanian rocks in the Ouachita Mountains, Arkoma basin, and the Ozarks (Fig. 2). Pennsylvanian deep-water facies have been described in some detail in this area by Cline, 1966, 1970; Briggs and Cline, 1967; Morris, 1971, 1974; Moiola and McBride, 1978; Thomson and LeBlanc, 1975; Stone and McFarland, 1981; Stone and Bush, 1984; and Moiola and Shanmugam, 1984, among others. The nonmarine and shallow-marine facies in this area, in contrast, have not received as much attention. Equivalent rocks to the west in western and northern Arkansas and eastern Oklahoma have been described in some detail by Stone, 1968; Sutherland and Manger, 1977, 1979; Zachry, 1975, 1977; Zachry and Haley, 1975; and Stone and McFarland, 1981.

GEOLOGIC SETTING

Between 1,000 and 30,000 feet of Pennsylvanian strata accumulated in parts of the Ozarks, Arkoma basin, and Ouachita basin in east-central Arkansas (Fig. 4). In the Ozarks and Arkoma basin, these rocks overlie about 5,000 feet of shelfal lower Cambrian to Mississippian rocks that in turn rest on Precambrian granitic basement. The Pennsylvanian section is 1,000 feet thick in the northern Arkoma basin and ranges from Morrowan to Desmoinesian (lower to middle Pennsylvanian) in age. The equivalent section in the Ouachita basin to the south (Ouachita facies) is over 30,000 feet thick and overlies about 20,000 feet of Ordovician-Mississippian basinal strata. There are marked differences in the lithologic character, thickness, and depositional patterns of age-equivalent units between these two areas.

The Ozarks, Arkoma basin, and Ouachita Mountains in Arkansas are, respectively, a genetically related late Paleozoic cratonic dome (Ozarks), a foreland (foredeep) basin (Arkoma basin), and a thrust-belt system (Ouachita Mountains) (Fig. 3). In the earlier Paleozoic, the dome and foreland basin areas were a broad continental shelf separated by a narrow slope from a deep basin to the south (Ouachita trough). A thrust-belt sequence developed late in Mississippian (?) time to the south in the basinal area, and in Pennsylvanian time the basinal facies were thrust over the shelf and slope facies (Fig. 3B). In the early stages of basin closure (by Atokan time), a successor foreland basin had developed on the older shelf and slope of the tectonically foundered margin. This area was the site of deep-water sedimentation until mid-Atokan time. The foreland basinal axis trended east-west and followed the older Paleozoic slope and continental rise. Paleocurrents were to the west in the deep-water facies (Briggs and Cline, 1967; Morris, 1974, among others). The axis of the basin shifted northward through the Pennsylvanian as the basin closed.

The nomenclature for the upper Mississippian and lower Pennsylvanian strata in Oklahoma and Arkansas is complex (Fig. 5). It is related to the dramatic facies changes from shelf to basin deposits and the thickening of shelfal units toward eastern Arkansas (Mississippi embayment) and toward the south (basinward). Facies changes include changes in lithologies from thin carbonates in the west to thick clastic sequences in the east and south in the Morrowan. The upper Mississippian (Chesterian) in Arkansas consists of, in ascending order, Hindsville (Batesville), Fayetteville, Pitkin, and Imo formations on the Ozark shelf, and most of the Stanley Group in the basin to the south. The shelf units are dominantly carbonates with shale interbeds (Handford, 1986), and the deep-water rocks are mainly shale-rich clastics with some interbedded volcanic tuffs and tuffaceous sandstones (Niem, 1976).

The youngest Paleozoic strata that crop out in Arkansas are early to middle Pennsylvanian (Morrowan, Atokan, and Desmoinesian) in age. On the Ozark shelf in northeast Oklahoma, the

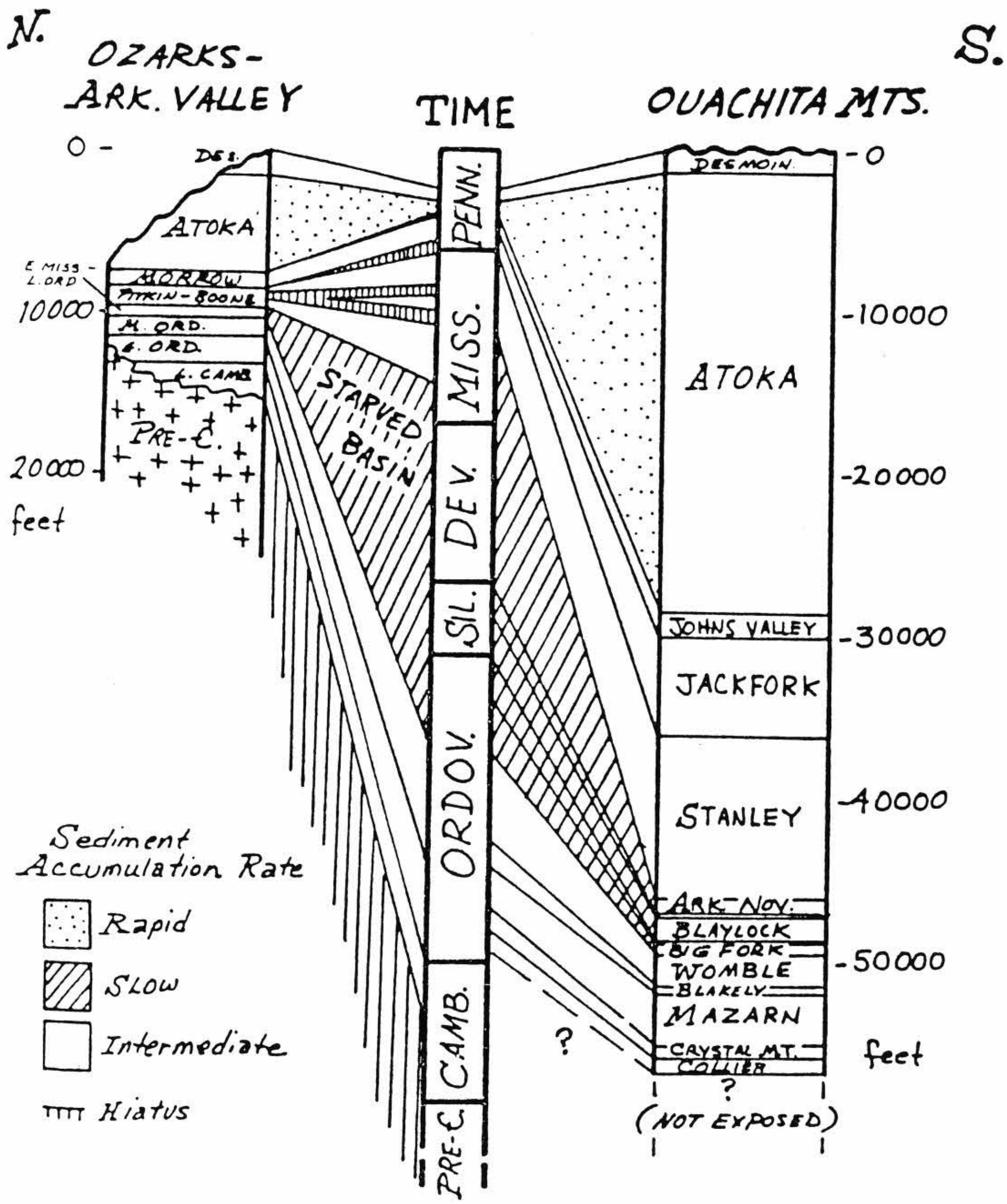


Figure 4. Generalized north to south stratigraphic sections between the Ozarks/Arkansas Valley (the shelf) and the Ouachita Mountains (the basin). Relative thickness and sediment accumulation rate are shown to depict (1) a period of starved basin sedimentation from the Late Ordovician to Early Mississippian, and (2) a period of rapid sedimentation in the Pennsylvanian.

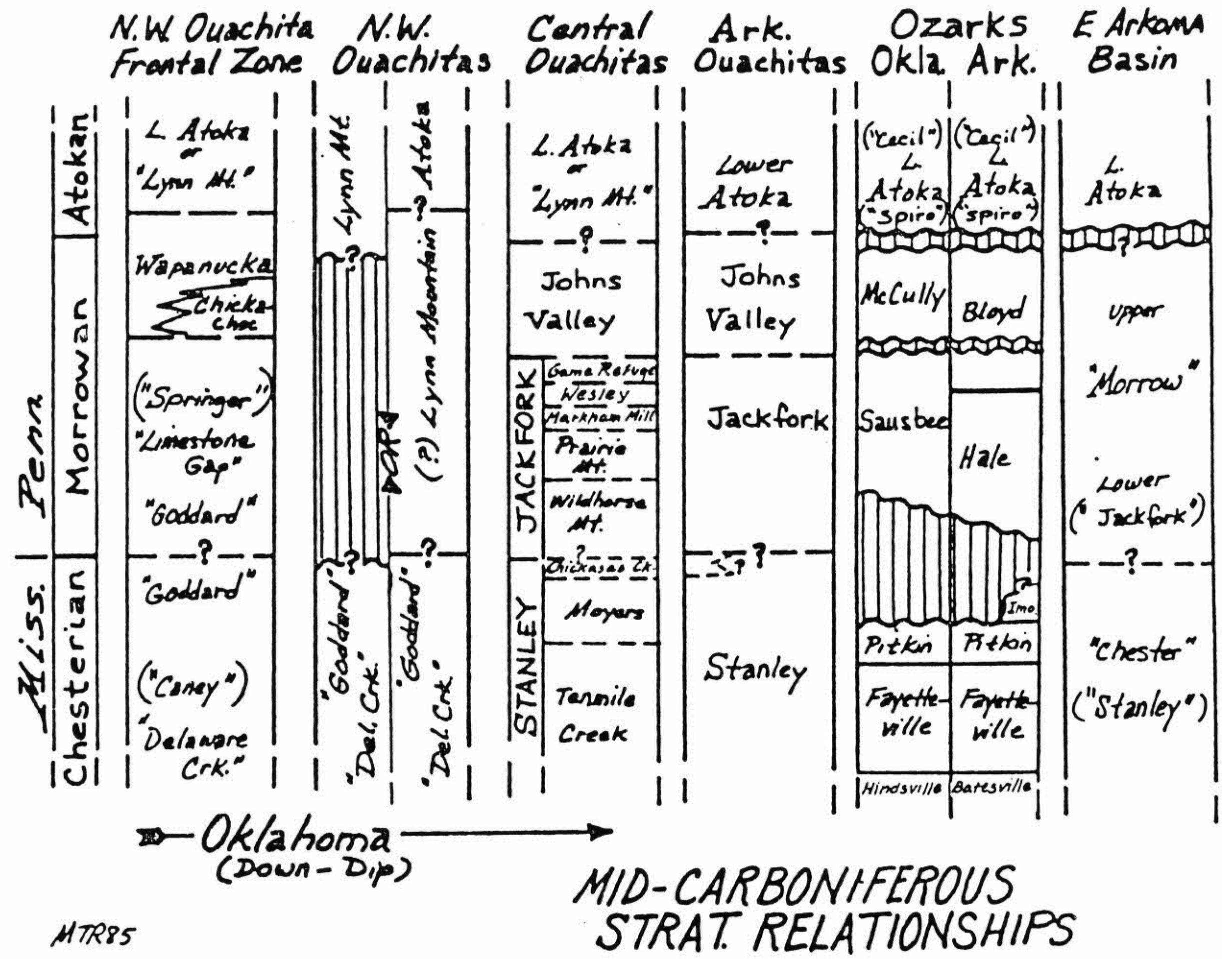


Figure 5. Stratigraphic nomenclature for mid-Carboniferous strata in the Ozarks, Arkoma basin, and Ouachita Mountains. Vertical line pattern denotes erosional gap or depositional hiatus, taken in part from Sutherland and Manger, 1979.

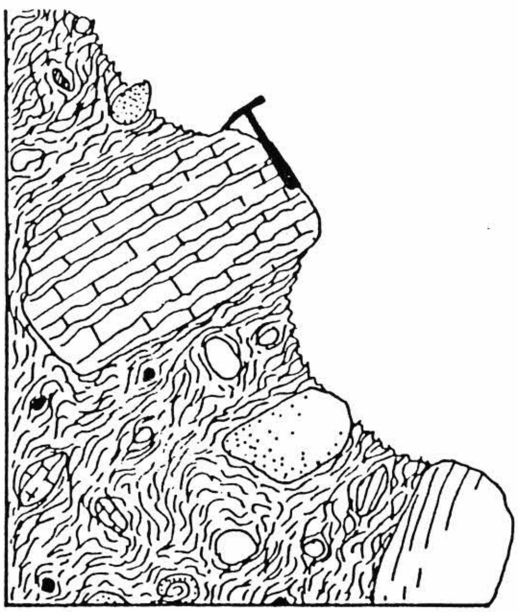
Morrowan units are thin, dominantly carbonate or mixed carbonate/clastic sequences, and are known as the Sausbee and McCully formations. They are unconformably overlain by fluvial-deltaic clastics of the Atoka Formation. The Oklahoma Morrowan units grade laterally eastward into the Hale and Bloyd formations in northwest Arkansas. Morrowan facies thicken toward east-central Arkansas and to the south, where correlation of upper and lower Morrowan units with the Hale and Bloyd subdivisions or units becomes difficult. The Morrowan strata on the northern shelf range in thickness from less than 200 feet in northeastern Oklahoma to over 2,000 feet in east-central Arkansas. Basinward to the south, these units are known as the Jackfork Formation, Sandstone, or Group and the Johns Valley Shale. The Jackfork is an alternating quartzose sandstone and shale sequence about 6,000 feet thick, which has been informally subdivided into five units in the central Ouachitas (Wildhorse Mountain, Prairie Mountain, Markham Mill, Wesley, and Game Refuge, in ascending stratigraphic order) (modified after Briggs and Roeder, 1978). The Johns Valley Shale is a shale sequence up to 1,500 feet thick that contains exotic clasts from many older Paleozoic units representing three major facies (Fig. 6, 7). Some sandstone interbeds are also present. It is found only in the basinal facies overlying the Jackfork. In southern Oklahoma, Jackfork-equivalent shelf and slope units are the upper Goddard(?), Limestone Gap, or "Springer" sequences, which are dominantly shales. Johns Valley Shale is equivalent to the Wapanucka Formation. The

	SERIES	ARBUCKLE FACIES	OUACHITA FACIES	OZARK FACIES
PENNSYLVANIAN	Atokan	Lake Murray - Atoka Fms	Atoka Fm	Atoka Fm
PENNSYLVANIAN	Morrowan	Golf Course - Wapanucka Fms ● Springer Fm	Johns Valley Sh Jackfork Gp	Bloyd Fm ● Hale Fm ●
MISSISSIPPIAN	Chesterian	Caney Sh ●	Stanley Gp	Pitkin Ls Fayetteville Sh Hindsville Ls
MISSISSIPPIAN	Meramecian			Moorefield Fm ●
MISSISSIPPIAN	Osagean	Sycamore Ls ●	Arkansas Novaculite ●	Boone Fm
MISSISSIPPIAN	Kinderhookian	Woodford Fm ●	Pinetop Chert ●	Chattanooga Fm
DEVONIAN	Upper & Middle			
DEVONIAN	Ulsterian	Hunton Gp: Frisco Ls Bois d Arc Ls. ● Haragan Ls	absent ?	Sallisaw Fm ● Frisco Ls
SILURIAN	Niagaran	Hunton Gp: Henryhouse Ls ●	Missouri Mountain Fm	St. Clair Ls.
SILURIAN	Alexandrian	Hunton Gp: Chimneyhill Ls ●	Blaylock Ss	absent ?
ORDOVICIAN	Cincinnatian	Sylvan Sh Fernvale Ls ●	Polk Creek Sh	Sylvan Sh Fernvale Ls.
ORDOVICIAN	Trentonian	Viola Ls ●	Bigfork Chert ●	Fite Ls ●
ORDOVICIAN	Blackriverian	Simpson Gp: Bromide Fm ● Tulip Creek Fm	Womble Sh	Tyner Fm ●
ORDOVICIAN	Chazyan	Simpson Gp: McLish Fm ● Oil Creek Fm ● Joins Fm ●		Jasper Ls ● Burgen Ss
ORDOVICIAN	Canadian	Arbuckle Gp: West Spring Creek Fm ● Kindblade Fm ● Cool Creek Fm McKenzie Hill Fm	Blakely Ss Mazarn Sh Crystal Mountain Ss Collier Fm	Cotter Dol ● Jefferson City Dol Roubidoux Fm Gasconade Dol
CAMBRIAN	Croixan	Arbuckle Gp: Butterly Dol Signal Mountain Fm Royer Dol Fort Sill Fm ● Honey Creek Fm Reagan Ss	not exposed	Eminence Dol Bonneterre Dol Lamotte Ss
PRECAMBRIAN		Granite & Rhyolite		Spavinaw Granite

Figure 6. **Stratigraphic chart showing formations from which exotic clasts (solid circles) have been found in the Johns Valley Shale. From Stone et al. (1979) after Shideler (1970).**

Wapanucka is a shelf-edge limestone deposit and its Chickachoc chert member is more of a shaly slope sequence.

The lower Atoka Formation overlies Morrowan units in Oklahoma and northwest Arkansas. Lower Atoka units are variously called the Spiro Sand, Lynn Mountain Formation, and Trace Creek Shale; in most places, the Atoka is subdivided into lower, middle, and upper members (Stone, 1968). The lower Atoka consists of thin fluvial-deltaic facies in the north and thick turbidite facies in the south. By upper Atoka time, fluvial-deltaic deposition dominated all areas of Atoka accumulation. In Oklahoma and Arkansas, the subsurface nomenclature for the various units of the Atoka Formation is very complex and changes from one township to another (see Branan, 1968, for example). The Atoka ranges from a few hundred to over 20,000



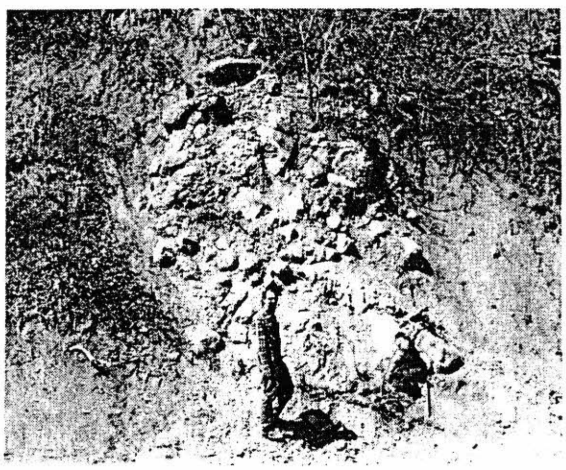


Figure 7. Sketch and photograph of two outcrops of the Johns Valley Shale near Stapp, Oklahoma. Disorganized shale fabric and matrix-supported exotic clasts are typical of this olistostromal facies.

feet in thickness. Desmoinesian strata overlie the Atoka in Oklahoma and in the western part of Arkansas, and include from older to younger, the Hartshorne, McAlester, Savanna, and Boggy formations. These are coal-bearing fluvial, coastal plain, and marginal marine deposits totaling about 5,000 feet in thickness.

PENNSYLVANIAN PALEOGEOGRAPHY

The Pennsylvanian paleogeographic reconstructions for east-central Arkansas presented here involve mainly three time slices: the Morrowan, Atokan, and Desmoinesian intervals. The Desmoinesian section, for the most part, is poorly preserved here. It is better preserved in synclines in parts of western Arkansas and eastern Oklahoma. The Morrowan and Atokan intervals are similar in their facies associations and both show the evolution of the gradual closing of a foreland basin, culminating in the uplift of this area in late or post-Desmoinesian time.

Morrowan

During the Morrowan on the Ozark shelf, shallow-marine carbonates were deposited in eastern Oklahoma and quartzose fluvial-deltaic clastic facies were deposited to the east in northern Arkansas. Coeval slope and deep-water quartzose clastic facies accumulated in the Ouachita basin to the south (Fig. 8). The Morrowan sediments expanded from a few hundred to thousands of feet in thickness from north to south and northwest to southeast(?), aided by growth faults along the shelf margin and slope. Seismic reflection data, as interpreted by the authors, show that these growth faults were localized over older basement-involved normal fault blocks and flatten above the early Mississippian or older platform units. Submarine landslides and related debris flows occurred on the slope and the resultant olistostrome deposits accumulated in the basin and the lower part of the slope to form parts of the Jackfork and Johns Valley formations.

Figures 9A and 9B illustrate the relatively thin (under 200 feet thick) Morrowan carbonate and mixed carbonate/clastic sequences (McCully and Sausbee formations) from the northeast Oklahoma shelf. Time-equivalent clastic units (Bloyd and Hale formations in northwest Arkansas) interfinger with the carbonates to the west. The carbonate rocks represent a shallow-marine, perhaps Bahama-like carbonate shelf facies, and their distribution and facies relationships are discussed by Sutherland and Henry (1977) and Sutherland and Manger (1979). To the south and southwest, the Oklahoma equivalents are the "Springer" slope facies and overlying Wapanucka shelfal limestone facies, which crop out in the Arbuckle Mountains and frontal Ouachita thrust-belt.

MORROWAN

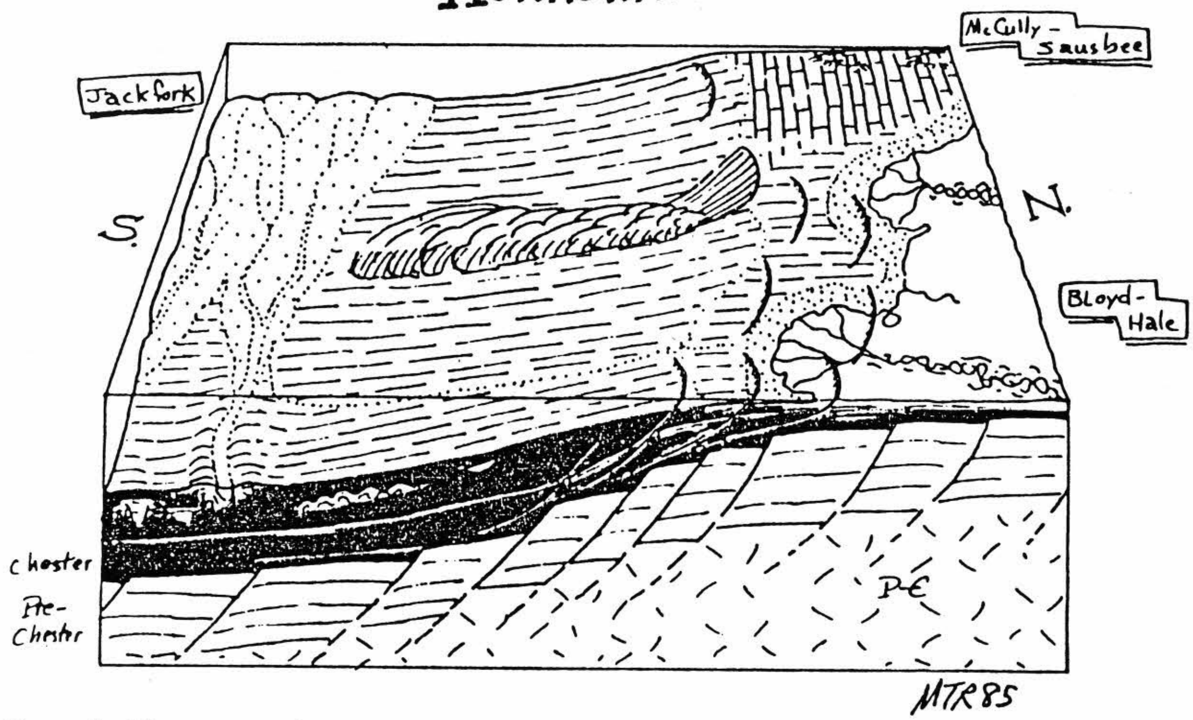


Figure 8. Morrowan paleogeographic sketch showing fluvial, deltaic, shallow-marine, and shelf facies to the north and slope, submarine fan, and basin facies to the south. Note the position of growth or listric faults localized at the outer shelf over pre-existing rift features. On the shelf, the western carbonate units in Oklahoma are the McCully and Sausbee formations in Arkansas, which are equivalent in age to the eastern clastic facies of the Bloyd and Hale formations. In the basin, the coeval units are the Jackfork and Johns Valley formations (not shown).

Eastward, in Arkansas, the dominantly clastic Bloyd and Hale formations constitute the Morrowan (Figs. 9C-F). They expand in thickness from less than 200 feet up to 2,000 feet (Fig. 10) from west to east. The Bloyd Formation in northwestern Arkansas is interpreted to consist of paralic coastal plain, fluvial-deltaic, and transgressive marine units (Zachry, 1977). Paleocurrent measurements from the cross-bedded fluvial units suggest rivers flowing to the south-southwest. In east-central Arkansas, the Bloyd and Hale sections are much thicker and have not been differentiated completely. The facies are similar to those of western Arkansas but are thicker and contain better developed deltaic, delta plain, and sandy fluvial deposits (Figs. 9C and 9D). Major braided stream deposits are transitional southwestward into smaller and more numerous meandering streams, and eventually into distributary channels that fed prograding deltas. Major braided systems are recognized near Gaither Mountain and in the Concord-Almond area (Fig. 9C, E) in northern Arkansas. The deltas are shelf-depth, relatively thin (ca. 40 feet), and widespread in their distribution (Figs. 9D, 9F). For the most part, they are fluvial- to wave-dominated systems and are characterized by highly bioturbated prodelta facies and thin to poorly preserved foreshore facies, locally capped or cut into by river-mouth bar and/or distributary channel facies. Morrowan deltaic deposits are well exposed in quarries in White County, especially near Bald Knob and Judsonia (Fig. 9D).

The Morrowan slope facies, for the most part, is not well exposed in the area, although the Jackfork units of the frontal thrusts in the Ouachitas may contain lower slope facies (for example, near Lake Maumelle). Possible slope facies are poorly exposed in the core of Bayou Meto anticline (C. G. Stone, 1985, personal commun.). The deep-marine Jackfork Formation and Johns Valley Shale have been more

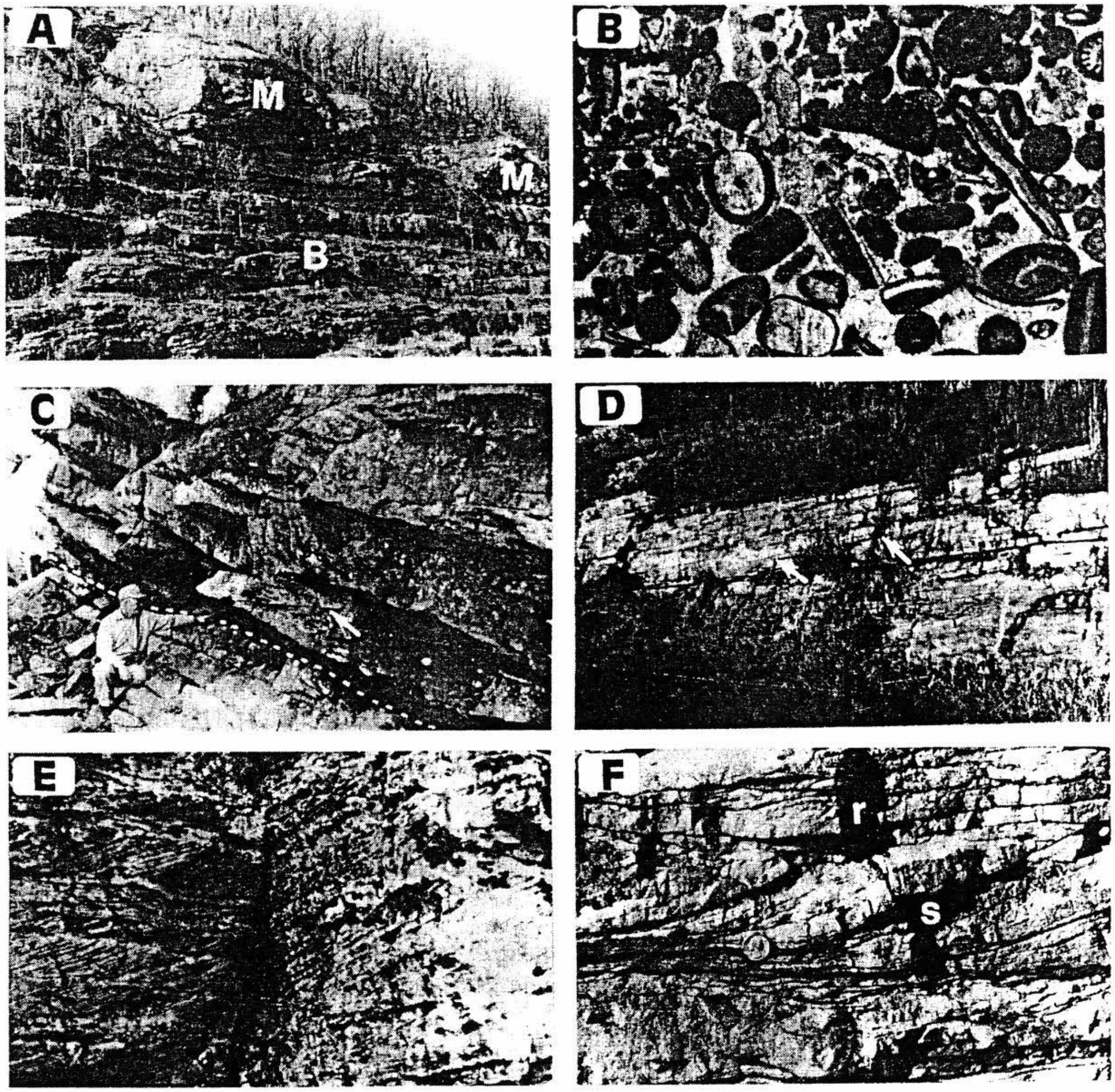


Figure 9. Photographs of Morrowan shelfal carbonates, fluvial and shallow-marine clastic rocks. A) Sausbee Formation at Webber Falls, Oklahoma; B--Briggs, and M--Brewer's Bend members. Note the large carbonate mounds (M) in the Brewer's Bend member. B) Photomicrograph of a carbonate-cemented oolite grainstone with abundant bryozoan and crinoid fragments; Sausbee Formation at Bragg's Mountain. C) Morrowan meandering fluvial deposits incised into marine units at Salado Creek, Arkansas. Note laterally accreting sigmoidal cross-bedded units (arrows) at the base of the channel fill. D) Morrowan foreshore (beach?) accretionary bedding (arrows) along Highway 67 near Judsonia, Arkansas. E) Unimodal, directional, planar-laminated cross-bedding in Morrowan braided stream deposit (view about 5 feet high), near Almond, Arkansas. F) Shallow-marine, tidal, sigmoidal (s) bedforms overlain by ripple-laminated beds (r); along Highway 67 near Judsonia.

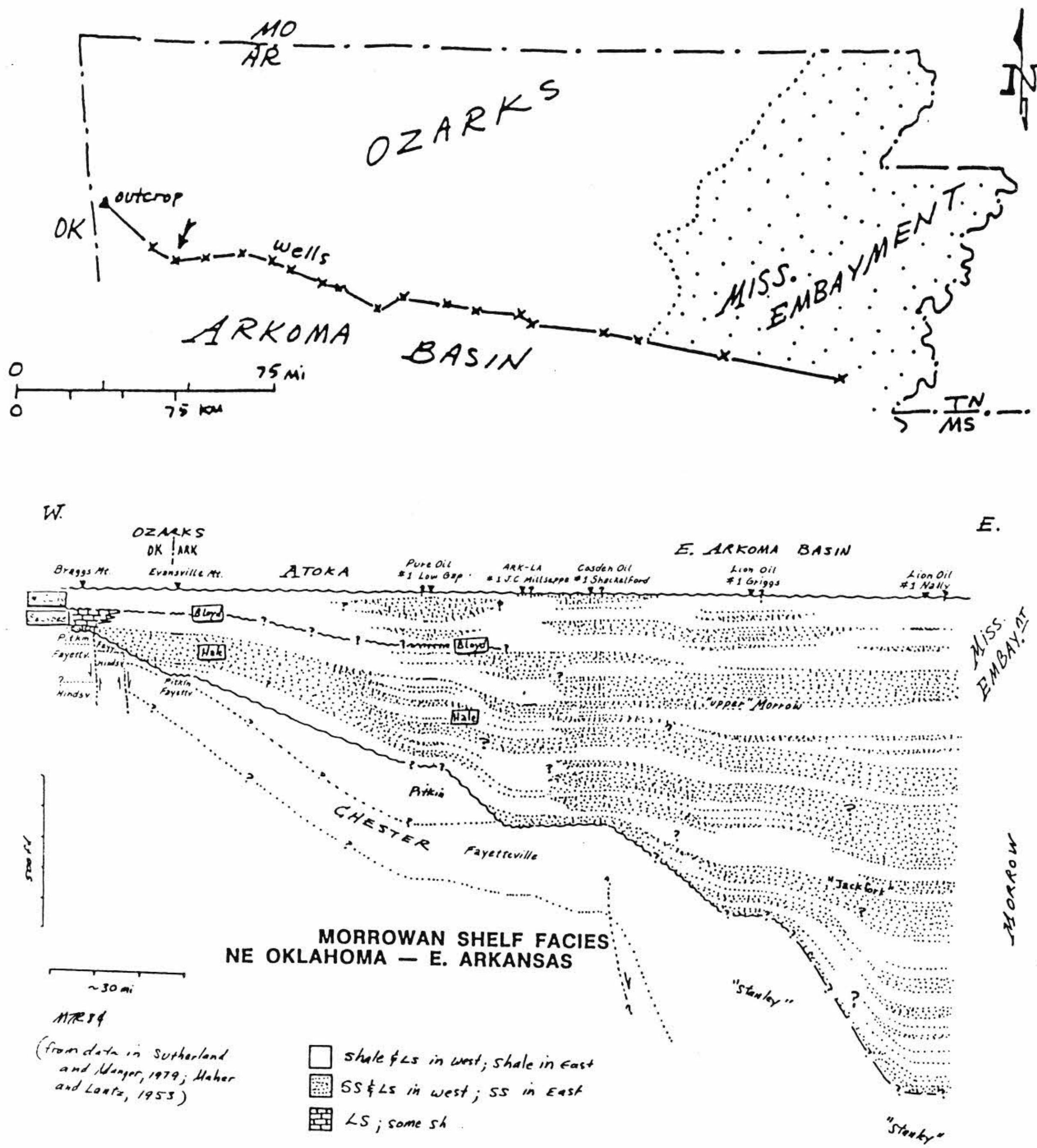
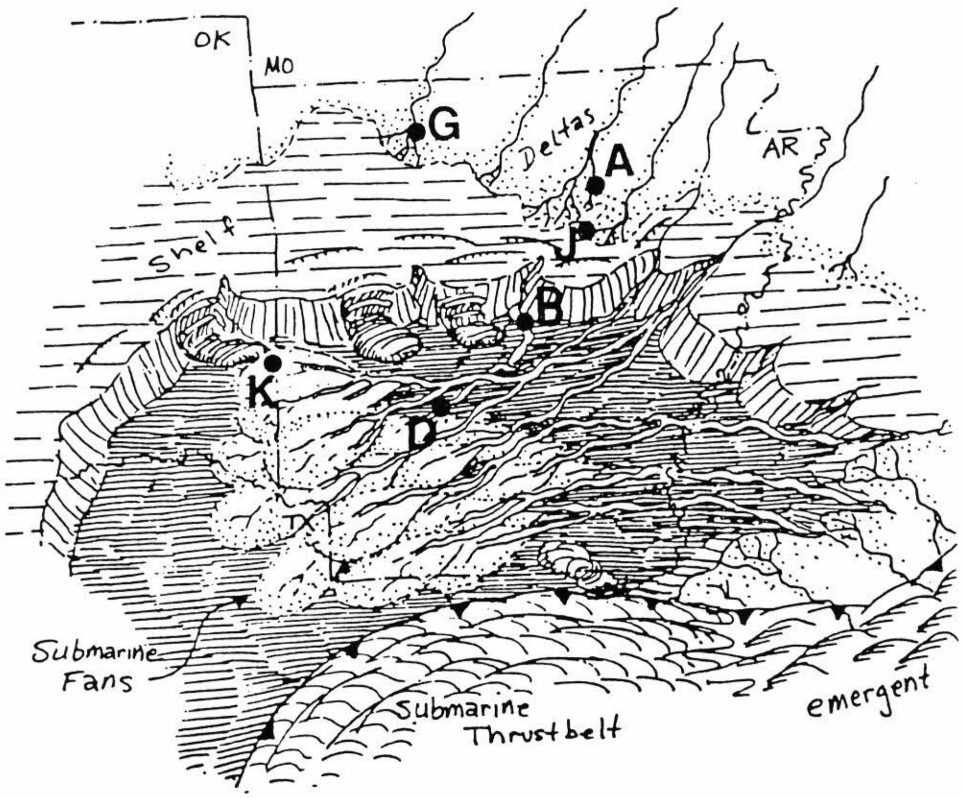


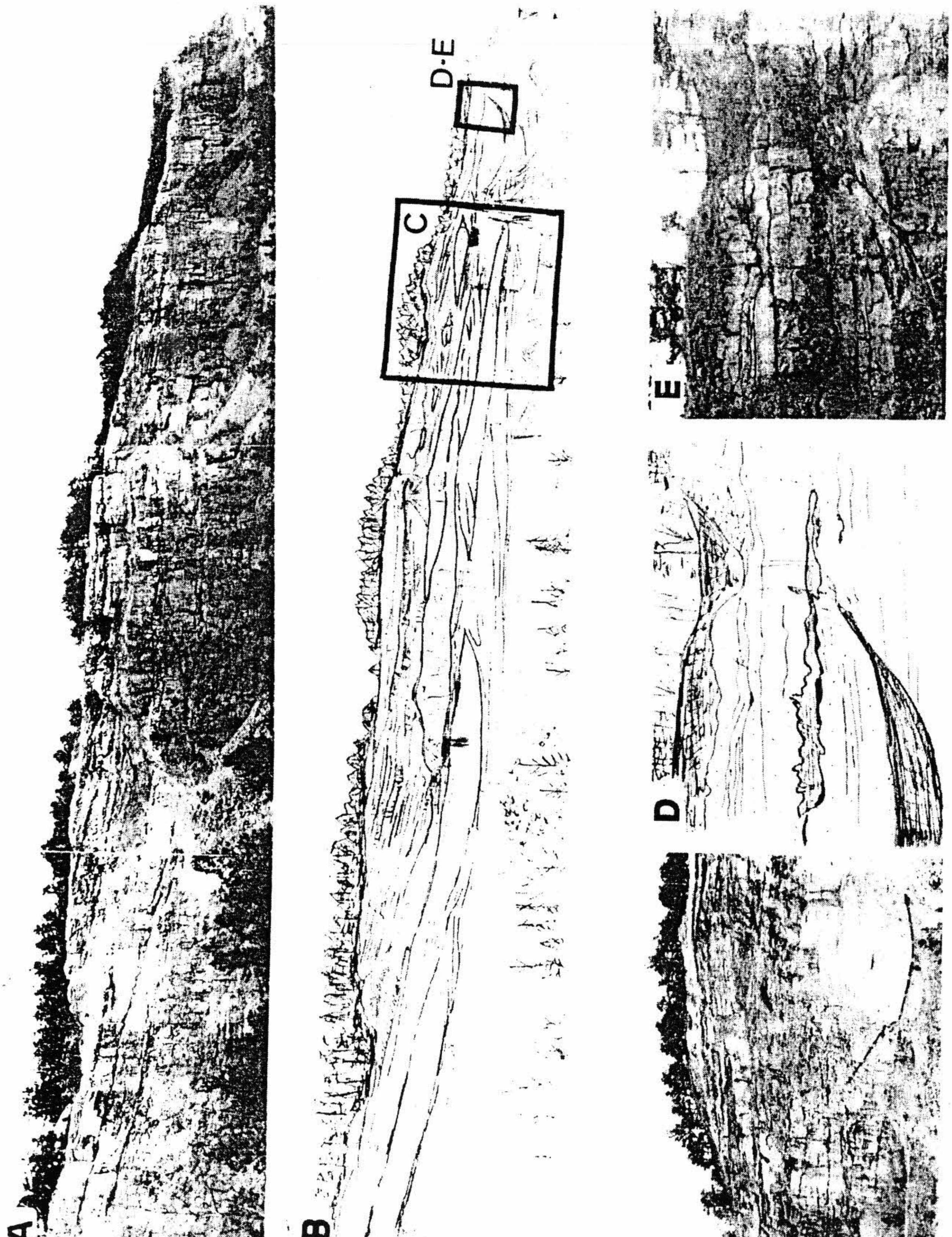
Figure 10. An east-west section showing the stratigraphic relations for the Morrowan. Note the facies change from carbonates in the west to thick clastic sequences to the east. Constructed from data in Sutherland and Manger (1979), and Maher and Lantz (1953). Line of section shown on sketch map above.



DIAGRAMMATIC PALEOGEOGRAPHY—JACKFORK

Figure 11. Diagrammatic map showing paleogeography during deposition of the Jackfork Formation. Abbreviations: B--Big Rock Quarry in North Little Rock; D--DeGray Dam/Friendship area; G--Gather Mountain; J--Judsonia; and A--Almond. Note that the Jackfork fan system consists of submarine canyons and inner fan channels to the northeast near Little Rock, middle fan-channel complexes in the DeGray Dam-Friendship area, and outer fan lobes at Kiamichi Mountain in Oklahoma.

Figure 12. Photographs and sketches of submarine canyon or inner fan-channel complex in the Jackfork Formation, Big Rock Quarry, North Little Rock. A) North to south view of the east face of the Big Rock Quarry. B) Sketch of Figure 12A, noting the lenticular, laterally aggrading channel-fill sandstone deposits and the more shaly upper part. Boxes denote position of Figures 10C-E. C) Photograph showing some of the major angular discordances between channel and the channel-margin deposits where levees are inclined away from the channels. D) Sketch of Figure 12E showing the side of a channel-fill where strata are onlapping the side of channel. E) Photograph of Figure 12D showing the side of the channel outlined.


A
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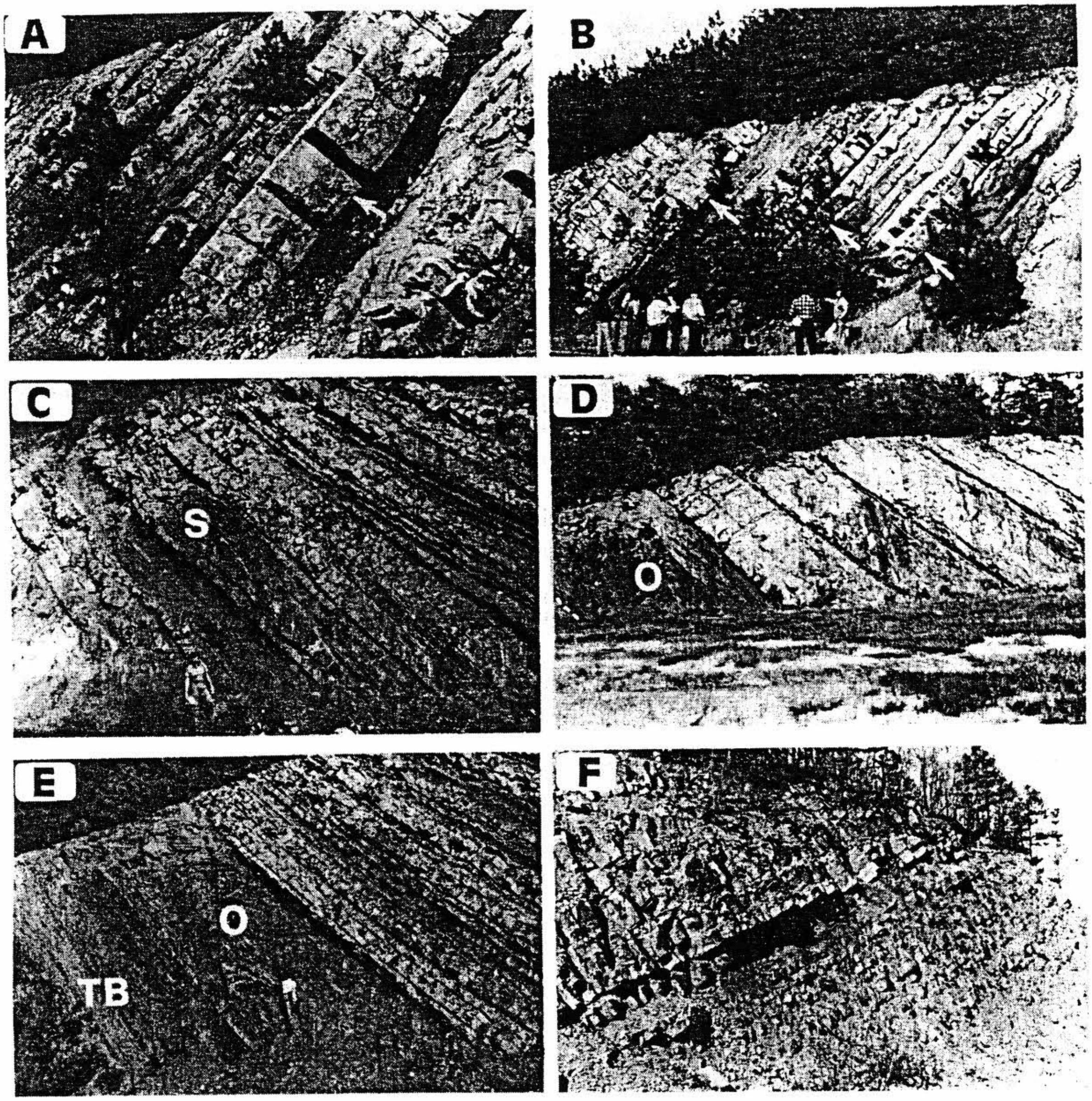


Figure 13. Photographs of submarine fan facies in the Morrowan Jackfork Formation. A) and B) Channel-fill or channel-lobe transition deposits which thin and fine upward to left; note the erosive basal contacts (arrows) of successive packages; Friendship area along U.S. Interstate Highway 30. C) Channel-fill deposits (C) overlain by a slumped unit (S), which is in turn onlapped and overlain by laterally continuous cravasse splay lobe(?) deposits; DeGray Dam. D) Uppermost major channel-fill deposit at DeGray Dam. Note the olistostrome (O) unit at the base (--) of the channel whose top is to the right. E) Thin-bedded (TB) turbidites overlain by an olistostrome (O) unit which is in turn overlain by alternating lobe(?) or crevasse splay beds; DeGray Dam. F) Depositional lobe which thickens and coarsens upward; Kiamichi Mountain along Highway 259.

extensively studied. The Jackfork Formation reaches over 6,000 feet in thickness and consists of alternating sandstone and shale packages interpreted to be submarine fan, basin plain, and slope facies (Fig. 11). The Jackfork contains about 30 to 60 percent sandstone with individual sand-rich packages being up to 1,000 feet thick and composed of 60 to 80 percent sandstone. Slope, submarine canyon, and inner fan deposits are recognized in the northeasternmost Jackfork outcrops in the Little Rock area. An excellent exposure of a submarine canyon fill or an inner fan-valley system occurs in Big Rock Quarry in North Little Rock (Fig. 12). Middle fan-channel complexes occur in central Arkansas in the Friendship-DeGray and Dierks Dam areas, and middle to outer fan lobes occur in western Arkansas and eastern Oklahoma, from near Mena (Rich Mountain) to Kiamichi Mountain (Thomson and LeBlanc, 1975; Moiola and Shanmugam, 1984) (Fig. 13). Paleocurrents are mainly to the west in these rocks. We envision a large submarine fan system with elongate lobes, a complex channel/lobe transition zone, and major submarine canyon/inner fan/slope. The sediments were deposited from east to west for the Jackfork Formation (Fig. 14). Similar elongate, deep-water, foreland basin assemblages have been described by Mutti (1985) in Spain.

The Johns Valley Shale is up to 1,500 feet thick and overlies the Jackfork Formation. It consists of shale with exotic clasts up to boulder size (olistoliths) and large slump masses that range from Cambrian to Morrowan in age (Figs. 6, 7) (Ulrich, 1927; Shideler, 1970; Stone et al., 1973; Stone and McFarland, 1981). Turbidite sandstone beds also occur in the Johns Valley Shale. The Johns Valley has been interpreted by many authors to represent submarine slump and landslide deposits derived mainly from northern and northwestern sources in Arbuckle and Ozark shelf facies. The source terrane for these clasts and slide masses, some of which are thousands of feet across, may have been large normal fault scarps along the northern and western basin boundaries produced during tectonic foundering of the basin margin by thrust-sheet loading to the south (Fig. 15). Seismic data in Oklahoma show huge normal fault scarps in the shelf facies, overidden by Ouachita facies thrusts. The facts that the coeval shelf-carbonate Wapanucka facies was deposited near sea level and the the Johns Valley Shale contains exotic clasts from the Wapanucka through the Cambrian section, strongly suggest that the Johns Valley was deposited near the base of a fault scarp or series of scarps with at least a mile of stratigraphic throw. Water depths are estimated to be over

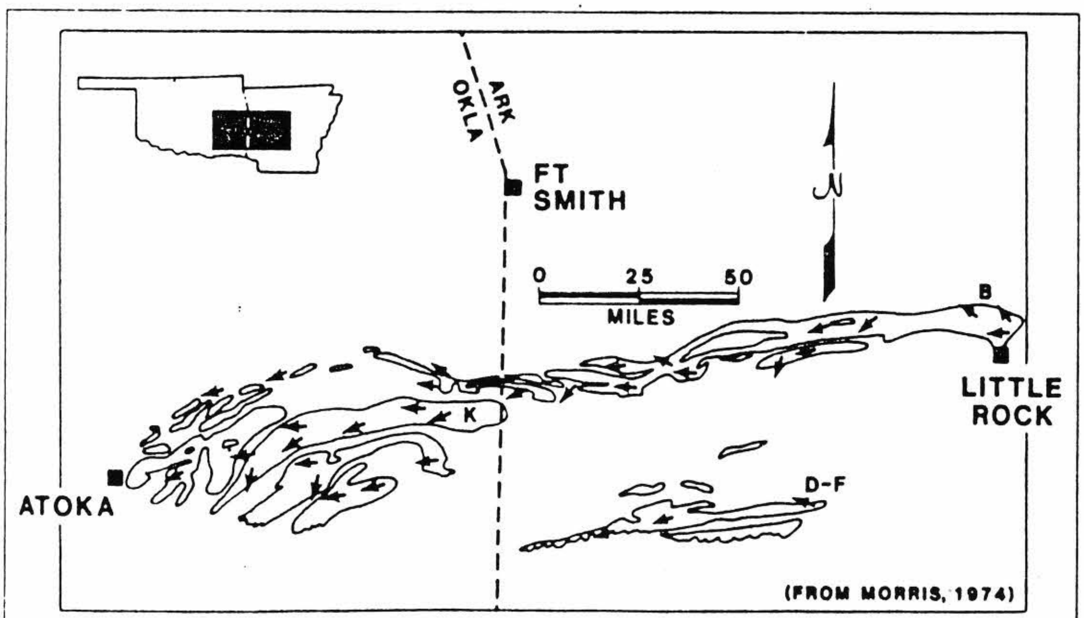


Figure 14. Map of paleocurrent directions in the Jackfork Formation. Note the pronounced westward trend.

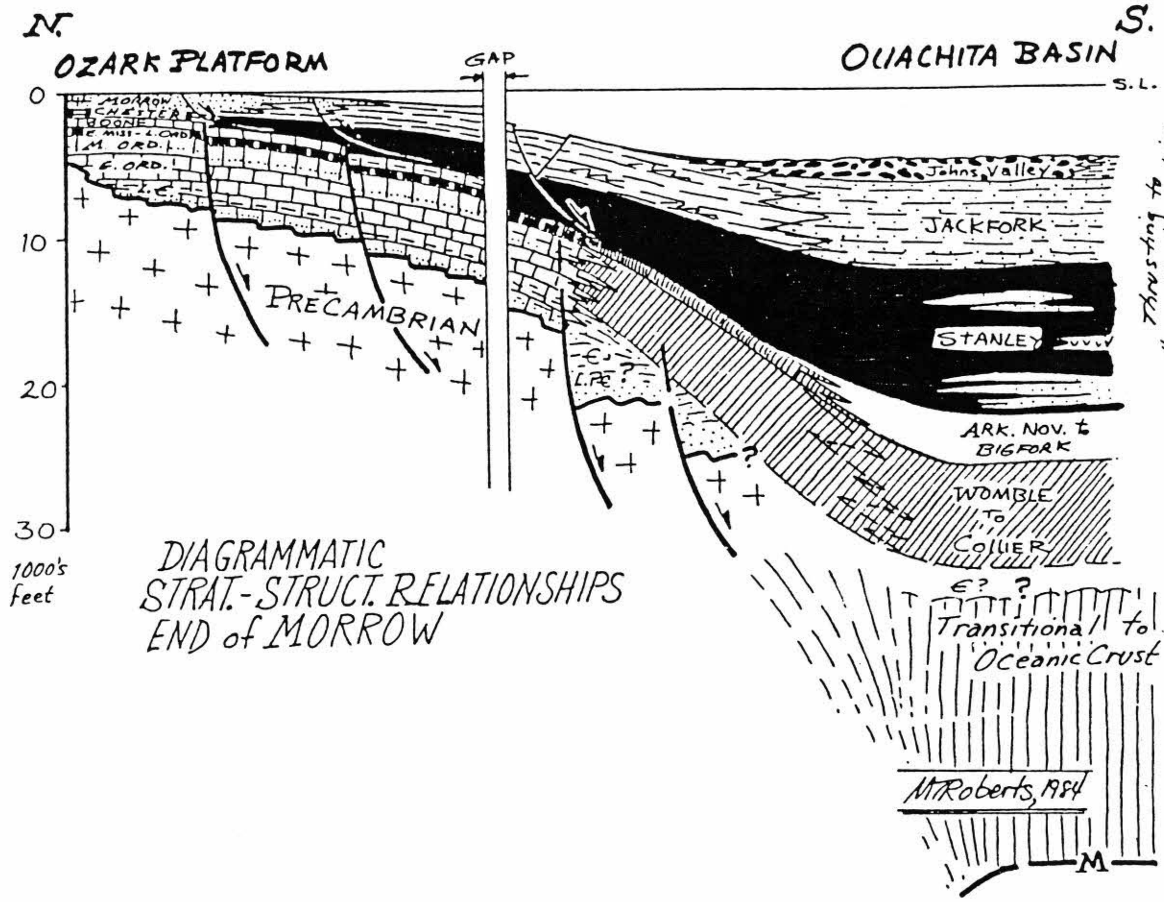


Figure 15. Diagrammatic north-south stratigraphic and structural cross section at the end of the Morrowan. Note the positions of the Ozark platform in relation to the Ouachita basin and the facies changes that occur in the shelf/slope/basin transition. The growth faults are localized over older rift structures and the Morrowan sections thicken on the down-thrown sides of the listric faults. The Cambrian to Mississippian shelfal units must have been exposed locally to erosion, perhaps in the gap indicated, because lower Paleozoic carbonates are found in the Johns Valley Shale.

5,000 feet, perhaps 7,000 feet, during Johns Valley Shale deposition.

Atokan

During the Atokan interval, the foreland basin narrowed and the axis migrated northward markedly due to the advancing thrust sheets from the south (Fig. 16). The basin shoaled through Atokan time due to sedimentary infilling. Fluvial-deltaic and shallow-marine facies were deposited on the shelf to the north, and slope and deep-water facies accumulated in the basin to the south in early to middle Atokan time. By late Atokan time, only shallow-marine and fluvial-deltaic facies were deposited in the basin. Relatively high angle growth faults, which become flatter above the Cambrian-Mississippian platform rocks, are common near the shelf-slope break. In post-Atokan time, some of these listric faults were reutilized as thrusts during the culmination of the Ouachita orogeny.

The Atoka Formation is about 27,000 feet thick and consists of shale, sandstone, siltstone, and some coal. The Atoka is informally subdivided into three mappable (lower, middle and upper) members (Stone, 1968). On the shelf in the northern Arkoma basin, Atokan fluvial-deltaic and shallow-marine facies include wave-, river-, and tidally-influenced deltas, distributary channel, and river-mouth bar deposits, transgressive

ATOKAN

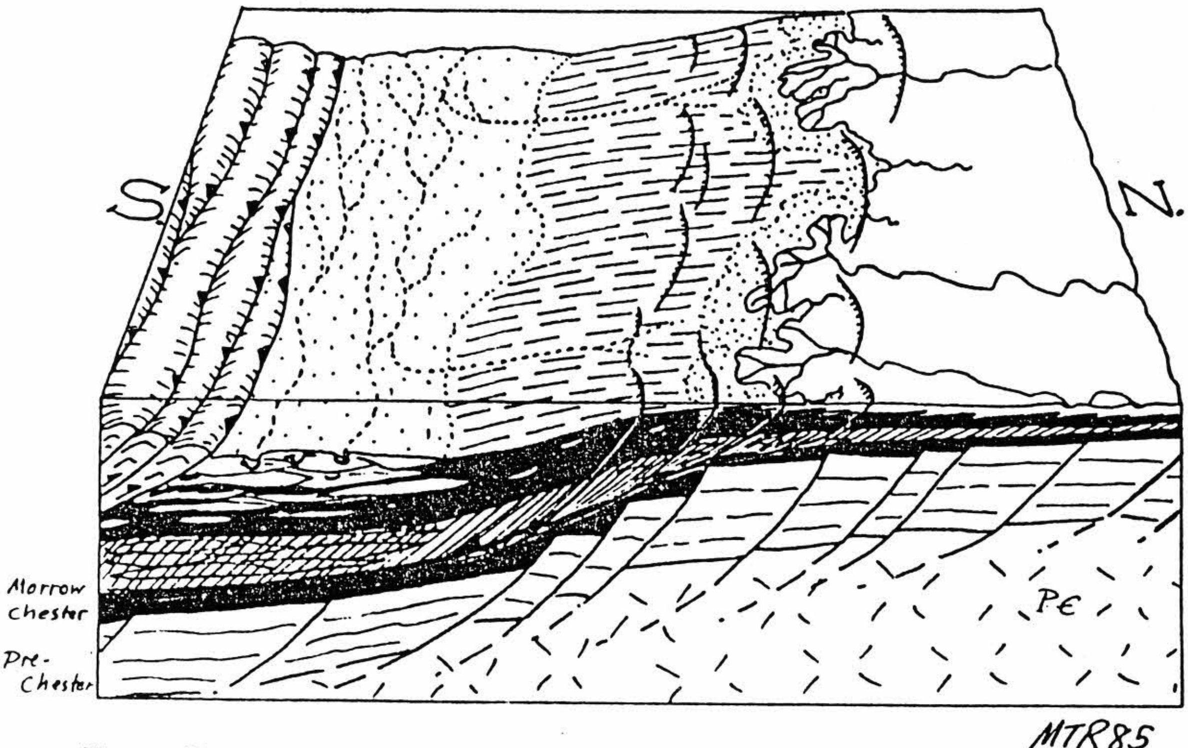


Figure 16. Paleogeographic sketch for the Atokan showing fluvial, deltaic, shallow-marine, and shelf facies to the north, and slope, submarine fan, and basin facies to the south. The advancing thrust sheets from the south can be seen restricting and narrowing the elongate basin. Growth faults are depicted at the outer shelf edge with local development of shelf edge deltas. Sediment transport is to the south/southwest on the shelf and to the west in the basin.

marine shelfal units, deltaic coastal-plain deposits, coal, and meandering stream sequences (Stone, 1968; O'Donnell, 1983) (Figs. 17A-17F). The coal and transgressive marine units (Fig. 17A) are relatively thin and locally widespread. The tidal influence on the deltas apparently became more pronounced as the basin constricted in late Atokan time, as tidal bedforms are more abundant in these deposits (Fig. 17E).

To the south (basinward), the lower Atoka and part of the middle Atoka are composed of sandstone-rich, deep-water deposits, similar to the Jackfork facies, but more than twice as thick. The deep-water facies were confined largely to an elongate east-trending trough that formed the southern part of the Arkoma basin. Paleocurrents were mainly to the west (Morris, 1974) in this basin. Deposition occurred during active closing of the basin by thrusts advancing from the south.

The Atokan interval represents a period of very rapid sedimentation in the foreland basin (Fig. 18). This high rate of sediment accumulation is interpreted to be due to the closing and narrowing of the basin that confined the sediment and increased the volumes of clastics contributed from rising source areas to the east in the Appalachian region.

Desmoinesian

In Desmoinesian time, the northward advancing thrust sheets and increasing influx of sediments had nearly closed the foreland basin, shifting the depo-center to western Arkansas and Oklahoma. The Ozark dome was reactivated as a positive lowlands feature to the north (Fig. 19).

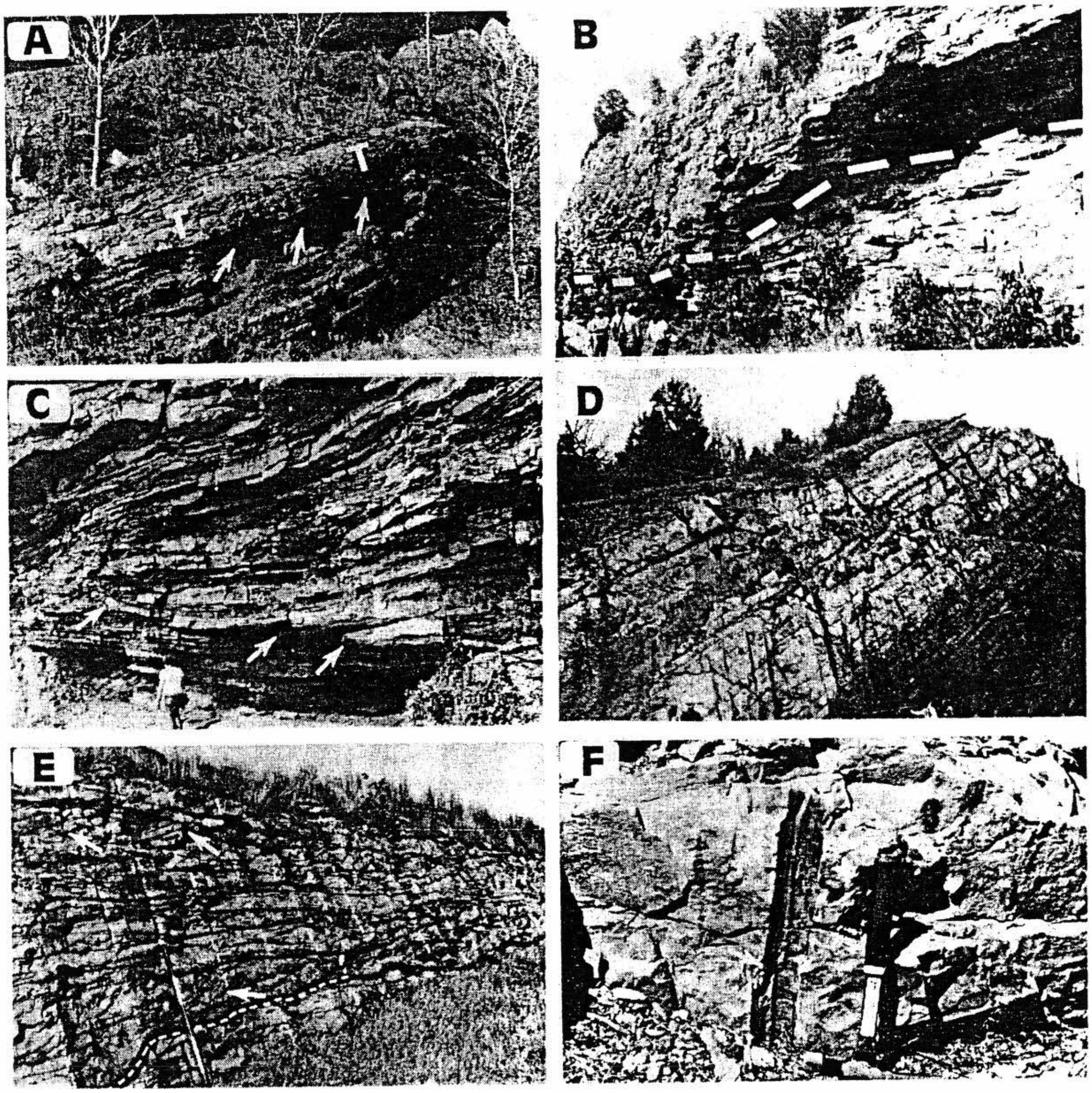
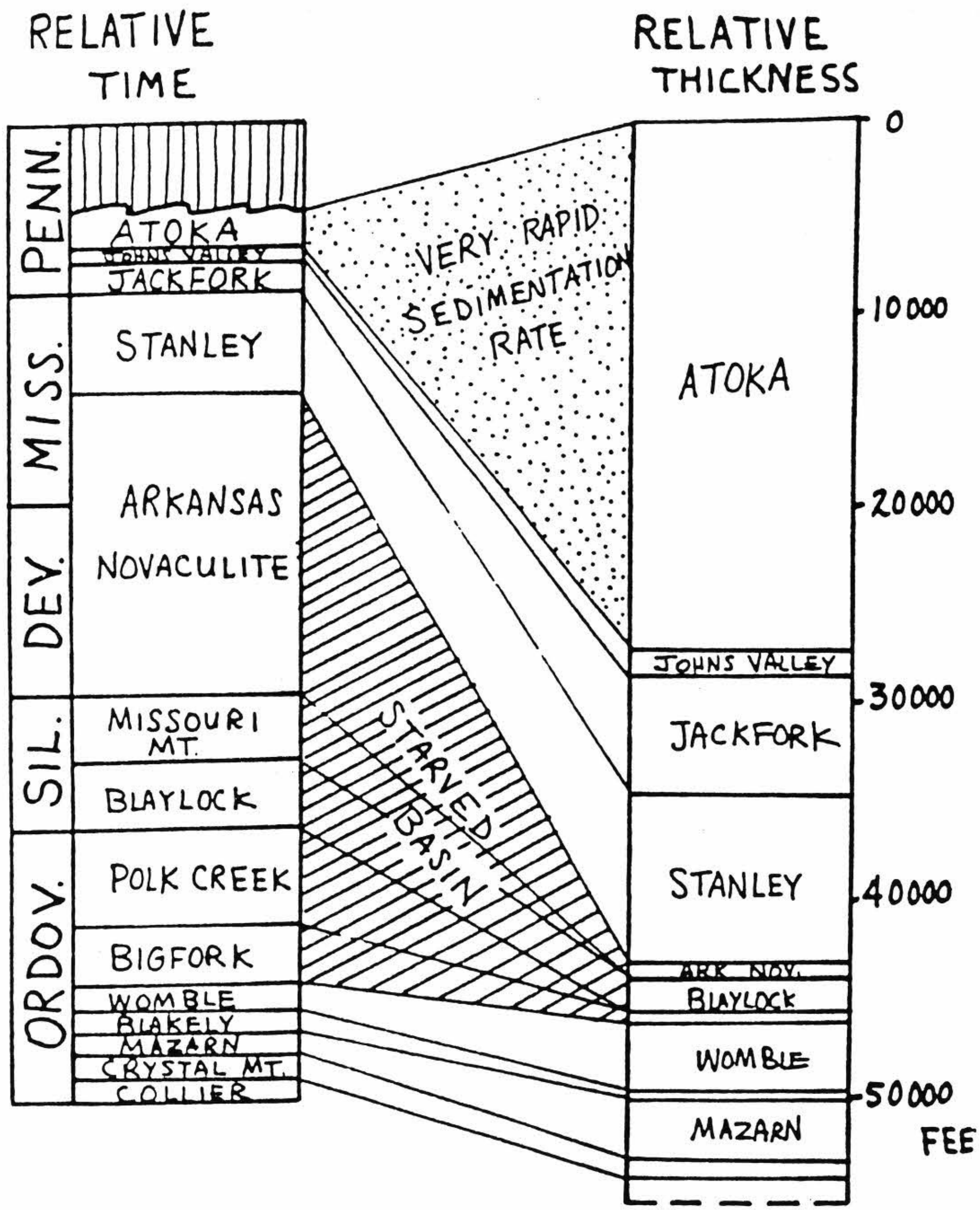


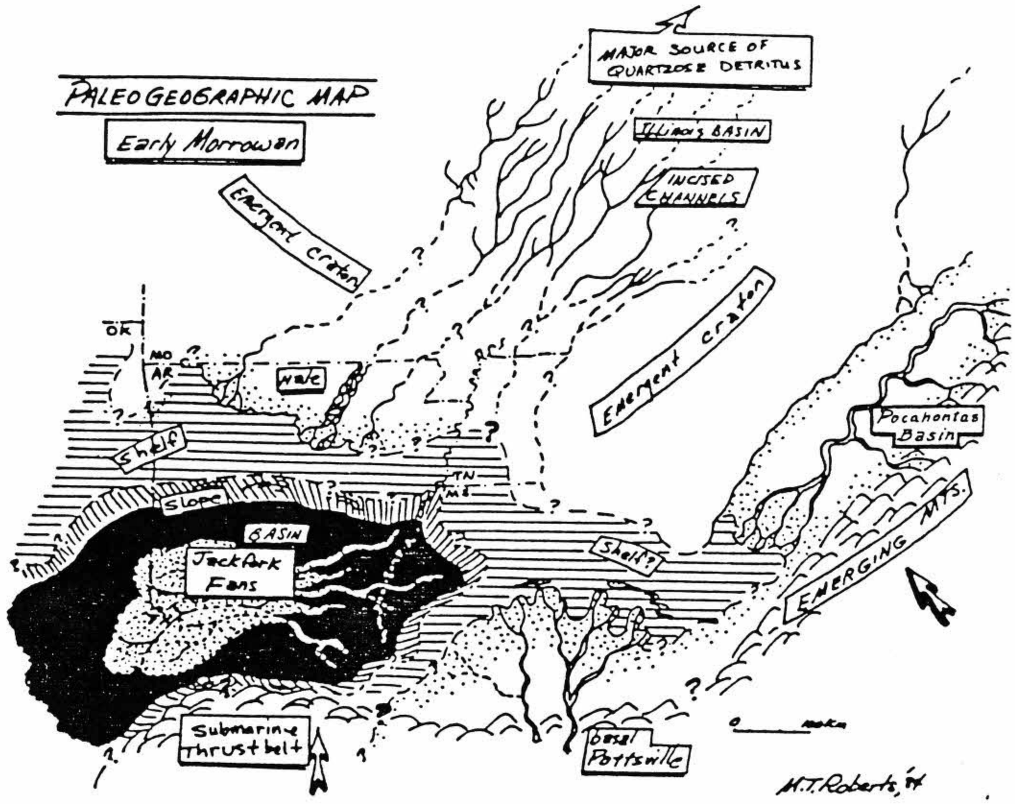
Figure 17. Photographs of Atokan and Desmoinesian outcrops. A) Atokan basal transgressive marine unit (T) overlain by pro-delta shales and underlain by well-developed karst topography (arrows) on Morrowan carbonates; Webber Falls, Oklahoma. B) and C) Large, delta distributary channel complex in the upper Atoka at Ozark, Arkansas. Note the erosional downcutting (outline) in Figure 17B and laterally accreting point bar surfaces (arrows) in Figure 17C. D) Thickening- and coarsening-upward deltaic sequence in the middle Atoka along Highway 9 near Morrilton, Arkansas. Note shale plug (abandoned shale-filled channel) at the top of the sequences (highlighted area). E) Tidally-influenced channel deposit along Highway 35 near Ozark, from the upper Atoka (scale = 1.5m). Note the scour base (highlighted) and sigmoidal bed forms (arrows). F) Mold of upright tree (Calamites) (arrows) in the Desmoinesian McAlester Formation along Highway 59 near Heavener, Arkansas. Several well-developed coal seams are present at this delta-plain outcrop.



PALEOZOIC UNITS
OUACHITA MTS., ARKANSAS

(MTR 80 from data in Stone et al., 1973; Haley, 1976)

Figure 18. Paleozoic stratigraphic section showing the relative thickness of units i the basinal part of the Ouachita basin as related to time. Note the starved-basi (reduced sedimentation) phase for the Ordovician Bigfork and Polk Creek, Siluria Blaylock and Missouri Mountain, and Devonian-Mississippian Arkansas Novaculit formations. Note also the period of very rapid sedimentation in Atoka time.


PALEOGEOGRAPHIC MAP
Early Morrowan
MAJOR SOURCE OF QUARTZOSE DETRITUS
ILLINOIS BASIN
INCISED CHANNELS
Emergent craton
Emergent craton
OK
MO
AR
Hale
Shelf
Slope
BASIN
Jackfork Fans
TN
MS
Shelf?
Pocahontas Basin
EMERGING MTS.
Submarine Thrustbelt
basal Pottsville
M.T. Roberts, '84

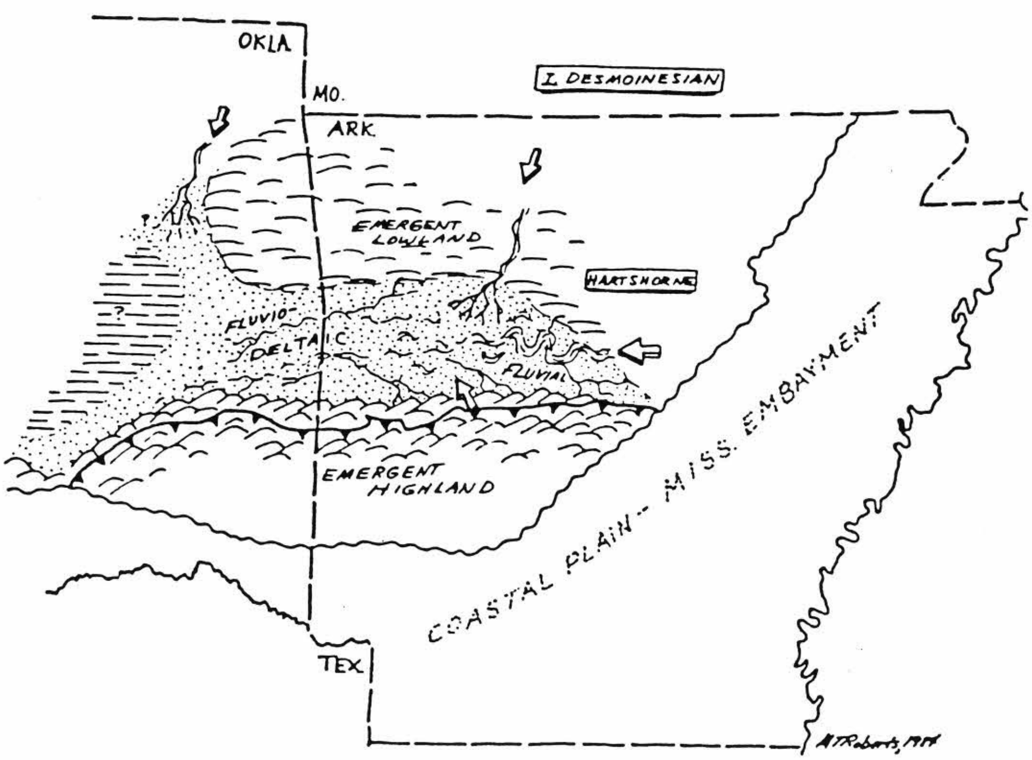

OKLA
MO.
ARK.
I. DESMOINESIAN
EMERGENT LOWLAND
HARTSHORNE
FLUVIO-DELTAIC
FLUVIAL
EMERGENT HIGHLAND
COASTAL PLAIN - MISS. EMBAYMENT
TEX
M.T.Roberts, 1984

Figure 19. Paleogeographic maps showing probable source areas for the early Morrowan and late Desmoinesian rocks. The major source for quartzose sediment is inferred to have been to the northeast and derived from the Canadian shield and interior drainage basin of the Appalachian orogenic belt. In the Morrowan, sediments for the most part by-passed the Illinois basin and Reelfoot rift(?) and were deposited to the south in the Ouachita basin. Sources to the south (Llanoria) and east (Black Warrior basin) probably also contributed sediment to the basin. The strongest evidence for a southern source is in the volcanic-bearing sediments for the Stanley Formation. In late Desmoinesian, the Ouachita basin gradually closed. Uplands to the south and lowlands to the north caused the basin to be restricted and sediments to be funneled to the west.

The foreland basin closed diachronously, starting in the southern Appalachians and proceeding westward into Oklahoma. By the end of Desmoinesian time, the eastern part of the central Arkansas foredeep basin had been filled. The sea retreated westward into Oklahoma and west-flowing rivers across Arkansas fed deltas near the Oklahoma border (Houseknecht and Kacena, 1983). These rivers were confined by the Ozark dome (north) and the frontal Ouachitas (south). Desmoinesian sediments were folded during the middle Pennsylvanian-early Permian culmination of the Ouachita orogeny. The Permian record is partly preserved to the west in Oklahoma.

The preserved Desmoinesian (Hartshorne) section in central Arkansas consists mainly of fluvial deposits. It is transitional into coal-bearing fluvial-deltaic and shallow-marine units in western Arkansas and eastern Oklahoma (Houseknecht and Kacena, 1984; Houseknecht and Matthews, 1985) (Fig. 17F). Middle Pennsylvanian to Permian rocks are not preserved in east-central Arkansas and, presumably, were never deposited or were eroded away during the orogenic mountain-building events caused by continental collision. These same strata are locally well preserved in Oklahoma and Kansas, where these areas contain cratonal, shallow subsiding basins that are influenced by wrench fault structures such as the Wichita-Arbuckle, Criner and Nemaha trends.

Provenance

The foreland basin developed over a downwarped shelf or platform sequence of Cambrian to Mississippian rocks dominated by carbonate deposition. In contrast, the clastic sediments of the foreland basin are mainly fine- to medium-grained quartzose (>90% quartz) sandstone and shale derived externally. Paleocurre petrographic data show that during the l history, quartzose sediment was derive the north, northeast, and east across wh now the Illinois basin, Ozark dome, Reelf Nashville dome, and Black Warrior basir The major source areas for this sedimen the Canadian Shield and the eastern U.S rior drainage basins of the Appalachia genic belt (Fig. 19). Clastic sediments l bypassed the interior cratonal areas, nc shallow seas, during low sea level stanc were deposited in the foreland basin frc southern Appalachian Mountain area. Ozark and Nashville domes rose durin stages of the orogeny, isolating the basin from the Arkoma foreland basin.

With continental collision, the various fc basins that flanked the late Paleozoic "A and "Gulf Coast" margins began to clos the east and south. Beginning in late l sippian (Stanley time), the Ouachita bas gan to close south of east-central Arkans some south-derived clastic and volcanic rial occur in the Stanley Shale (Niem, Some of the Pennsylvanian sediments also likely derived from the east from the southern Appalachians via the Black \ basin and perhaps from the southeast fr postulated land mass of Llanoria (Grahan 1975; Mack et al., 1981; Mack et al., Owen, 1984; Owen and Carozzi, 1986).

Starting in the middle Pennsylvanian t Permian, the southern margin of the fc basin was folded, uplifted, and subjected nudation. Reworked older Paleozoic mainly chert conglomerates and lithic clasts, were shed to the north and nor (Houseknecht and Kacena, 1983; Sutherl preparation) into the gradually closing

The Ouachita thrust-belt segment of central Arkansas and southeastern Oklahoma was probably submarine until the late stages of the orogeny.

SUMMARY AND CONCLUSIONS

1. The Pennsylvanian strata in east-central Arkansas are part of a late Paleozoic foreland basin system that originally extended from the southern Appalachians to west Texas.

2. Early Pennsylvanian Morrowan and Atokan rocks of east-central Arkansas are composed of quartzose sandstone and shale derived mainly from the northeast from the North American craton and Appalachian orogenic belt. These clastics buried the Cambrian to Mississippian carbonate-dominated platform rocks, the last remnant of which were the Morrowan carbonate rocks of northeast Oklahoma.

3. The clastic sedimentary facies that make up the Morrowan and Atokan intervals are similar in type and paleogeographic distribution. Fluvial, deltaic, shallow-marine, and shelf facies occur on the northern flank of the basin, while slope, submarine fan, and basin plain facies are present to the south. Sediment transport on the shelf was mainly to the southwest, and in the basinal area it was to the west. Thicknesses increase drastically from north to south, aided by listric growth faults along the shelf edge and slope.

4. In Atokan through early Desmoinesian times, the foreland basin gradually closed by advancing thrust sheets from the south. Sediments accumulated in the basin at extremely high rates. By late Atokan time, fluvial, deltaic, shallow-marine, paralic, and transgressive units were deposited throughout this shoaling basin. Ouachita uplands to the south and the Ozark dome to the north restricted sedimentation in the narrowing basin by Desmoinesian time and funneled sediments to the west.

5. Continental collision culminated in the post-Desmoinesian (middle Pennsylvanian to early Permian) uplift of the area, followed by erosion and/or nondeposition.

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