Sedimentary Response to Intrabasinal Salt Tectonism in the Upper Triassic Chinle Formation, Paradox Basin, Utah

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Sedimentary Response to Intrabasinal Salt Tectonism in the Upper Triassic Chinle Formation, Paradox Basin, Utah

By Joseph E. Hazel, Jr.

EVOLUTION OF SEDIMENTARY BASINS—PARADOX BASIN

A.C. Huffman, Jr., Project Coordinator

U.S. GEOLOGICAL SURVEY BULLETIN 2000–F

A multidisciplinary approach to research studies of sedimentary rocks and their constituents and the evolution of sedimentary basins, both ancient and modern

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Sedimentary Response to Intrabasinal Salt Tectonism in the Upper Triassic Chinle Formation, Paradox Basin, Utah

By Joseph E. Hazel, Jr.

ABSTRACT

The Upper Triassic Chinle Formation in the Salt Anticline region of the Paradox Basin, Utah, was deposited in fluvial and fluvial-lacustrine depositional environments that developed and evolved in response to salt diapirism. In the vicinity of the Cane Creek anticline, the Chinle consists of four narrow to broad sheet sandstone bodies. The superposition of these sheet sandstone bodies is a major architectural feature that affords an example of nonmarine sedimentary responses to intrabasinal tectonism. The sand bodies, the informal Kane Springs strata and Black Ledge of the Chinle Formation, are characterized by a distinctive suite of calcirudite, calcarenite, and quartzarenite. Stratigraphic mapping using bounding-surface hierarchy documents the localized cyclic reoccurrence of sheet geometry. Sheet sandstones are traceable for tens of kilometers and are separated by fifth- and sixth-order erosional surfaces.

Salt tectonic influences in the Chinle are indicated by (1) intraformational and interformational unconformities; (2) cyclic reoccurrence of sheet geometry and abrupt changes in fluvial style; (3) localization of axial drainage; (4) variation in sandstone story thickness as a result of differential subsidence in intervening synclinal areas; (5) evidence for cannibalization and redistribution of floodplain soil horizons (as calcirudite); and (6) evidence for topographically triggered avulsion via tectonic tilting.

Although allocyclic controls were important in Chinle depositional systems, salt tectonism probably was a catalyst on both extrinsic and intrinsic controls responsible for syndepositional patterns and systematic changes in fluvial style. Episodic rainfall resulted in sedimentation by floods on the Chinle alluvial plain. Sheet geometries were produced during periods of negligible regional subsidence due to salt anticline uplift. Ribbon geometries formed when the salt diapirs were inactive and the area was subsiding at moderate regional rates.

INTRODUCTION

The character of sedimentary deposits in nonmarine depositional environments is determined by partly interdependent variables including tectonism, climate, sediment supply, and base level. Differences of opinion exist as to the origin of cyclicity (autocyclic versus allocyclic processes) in sedimentary deposits and the relation between variable change and resultant sedimentary response. The relationship between changes in fluvial style and intrabasinal tectonics is also poorly understood. Field studies of alluvial architecture and internal stratigraphic variations ascribable to tectonosedimentary interactions are few in number (Alexander and Leeder, 1987).

In this paper, I present a case study in which one control, intrabasinal tectonism, can be confidently linked to the sedimentary product. Architectural facies variability in the Upper Triassic Chinle Formation, proximal to the Cane Creek salt anticline, is attributed to localized salt tectonic interactions within the Paradox Basin of southeastern Utah (fig. 1). In the Salt Anticline region (Cater, 1970) of the Paradox Basin, the Chinle was interpreted by Blakey and Gubitosa (1983, 1984) to represent distinct depositional phases resulting in two fining-upward cycles. Blakey and Gubitosa suggested that the two cycles contain several smaller cycles, all of which are related to salt tectonics. Drawing on these studies, I examined Chinle facies architecture across the Cane Creek anticline and used stratigraphic mapping of bounding-surface hierarchy and independent evidence from other studies to examine the controlling factors in generating the different alluvial stratigraphies (Hazel, 1991). Specific outcrop sequences were integrated and compared with the known regional depositional setting (Stewart and others, 1972; Blakey and Gubitosa, 1983, 1984; Dubiel, 1987, 1989; Hazel and Blakey, 1992) (fig. 2).

The Chinle alluvial plain extended far (>100 km) from source areas and was dominated by seasonal precipitation and low-gradient deposition. In this paper, I attempt (1) to show how a part of this alluvial system was affected by

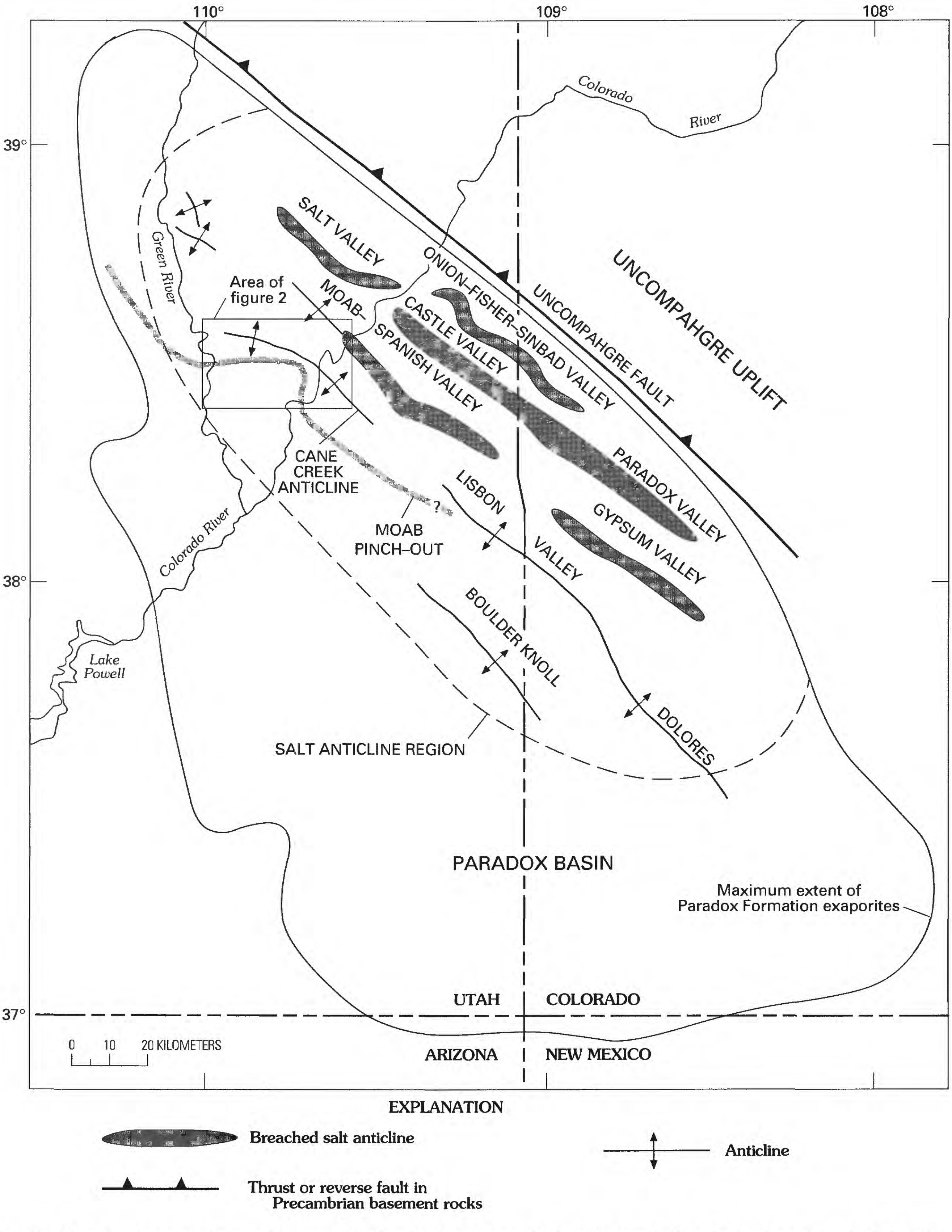

Figure 1. Map showing major structural features in the Paradox Basin region, Utah and Colorado. Breached salt anticlines, the Salt Anticline region (Cater, 1970), Moab pinch-out (Finch, 1959), and maximum extent of Paradox Formation exaporites (Hite, 1960) are also shown. Modified from Doelling (1985).

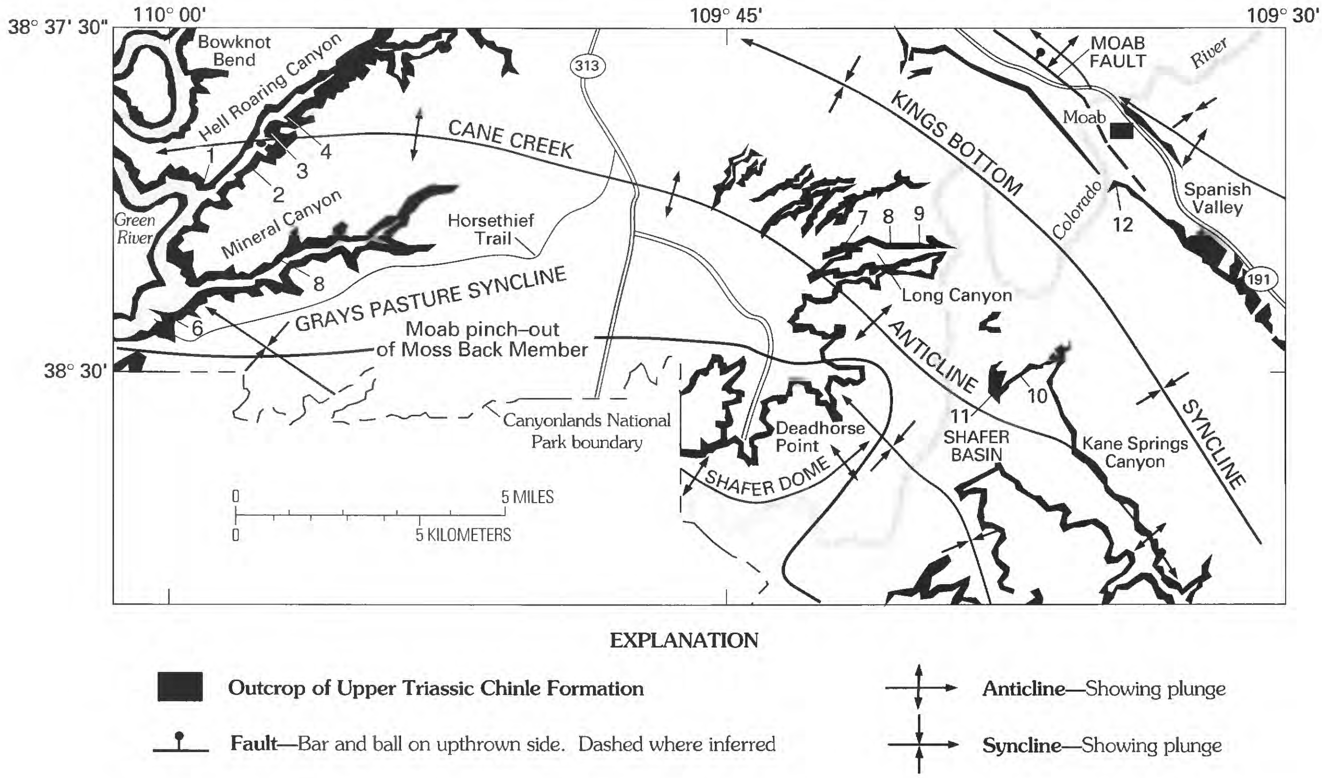

Figure 2. Map of the study area in the Paradox Basin, Utah, showing structure and outcrops of Chinle Formation. Numbers refer to the following locations: 1, location of figure 6; 2, location of figures 10, 16, and 24; 3, location of figure 11; 4, location of figure 17; 5, location of figures 19 and 28; 6, location of figure 22; 7, location of figures 23; and 8, location of figure 27. Area of map is shown in figure 1.

deformational events within the basin and (2) to differentiate autocyclic mechanisms from allocyclic controls on the depositional systems. Autocyclic processes arise from energy distribution within the depositional basin (Beerbower, 1964). Intrabasinal tectonism attributed in part to salt tectonics controlled syndepositional patterns and the systematic changes seen in fluvial style.

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REGIONAL SETTING

TECTONIC SETTING

The Upper Triassic Chinle Formation was deposited during a major global reorganization of plate regimes that included the initial breakup of Pangea, the opening of the Atlantic Ocean, and the consequent westward motion of the North American plate (Coney, 1978). An arc-continent collision and accretion of terranes along the Cordilleran margin terminated Early Triassic marine sedimentation (Moenkopi Formation) that represented a continuation of typical miogeoclinal Paleozoic depositional patterns (Speed, 1978). This post-Antler orogeny event, the Sonoma orogeny, initiated an Andean-type continental margin tectonic style that persisted throughout much of the Mesozoic and Cenozoic.

Middle to Late Triassic subduction and associated continental arc magmatism created a nonmarine back-arc basin (Coney, 1978; Dickinson, 1981). The Chinle Formation was deposited on the southwestern edge of this basin in an area centered about the Four Corners region of Colorado, Utah, Arizona, and New Mexico (Stewart and others, 1972; Blakey and Gubitosa, 1983; Dubiel, 1987, 1989), approximately 5°–15° north of the paleoequator (Van der Voo and others, 1976; Habicht, 1979; Ziegler, and others, 1983; Parrish and Peterson, 1988; Bazard and Butler, 1991) (fig. 3). Correlation of Chinle continental deposits with marine strata of the same age in central and northern Nevada suggests that an outlet for Chinle fluvial systems was present on the Late Triassic marine shelf (Lupe and Silberling, 1985; Elison and Speed, 1988). Convergent plate margin tectonism to the southwest, however, may have at least partly isolated the Chinle basin from the Pacific Ocean (Blakey and Gubitosa, 1983).

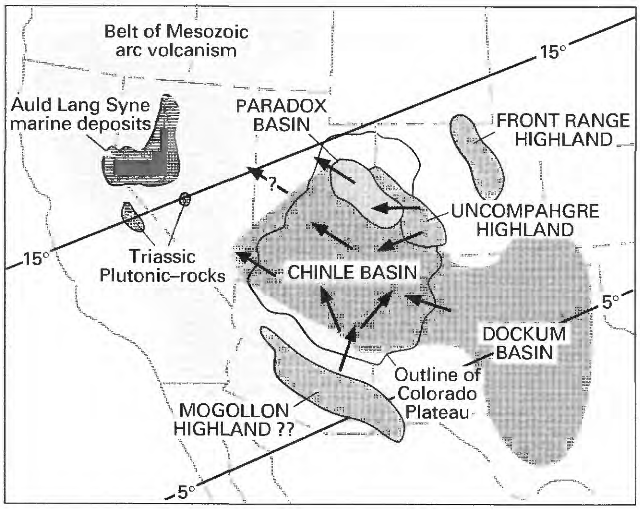


Figure 3. Map showing regional tectonic elements in the southwestern United States in Middle to Late Triassic time. Arrows indicate general paleocurrent trends within the Chinle depositional basin. Modified from Cater (1970), Speed (1978), and Dickinson (1981).

Several tectonic features bordered the Chinle depositional basin and were sources of detritus. Sediment was supplied from the south by Precambrian igneous and metamorphic rocks and Paleozoic sedimentary rocks associated with uplift of the Mogollon highlands (Cooley and Davidson, 1963; Stewart and others, 1972; Dickinson, 1981) (fig. 3). This region was also the source for abundant volcanic detritus (Stewart and others, 1972). The exact location and origin of these highlands is controversial (Bilodeau, 1986; Stewart and others, 1986; Wyman, 1987). Volcanic ash was supplied to the basin from the inferred volcanic arc off the west coast of the Cordilleran margin (Stewart and others, 1972; Blakey and Gubitosa, 1983). In the Salt Anticline region, remnants of the Uncompahgre highlands, the westernmost part of the Ancestral Rocky Mountains to the east, were a sediment source, as were penecontemporaneous Chinle deposits (Stewart and others, 1972; Blakey and Gubitosa, 1983; Dubiel, 1987, 1989) (fig. 3). Coeval sedimentary basins were situated to the southeast, north, and northwest of the Chinle depositional basin (Blakey and Gubitosa, 1984).

Fossil flora and fauna indicate perennial flow in Chinle streams (Ash, 1972; 1975; Gottesfeld, 1972; Colbert, 1985; Dubiel and others, 1989); however, paleoclimatic interpretations vary considerably, ranging from arid and semiarid to humid tropical (Dubiel, 1987). Robinson (1973) described the climate along the western edge of Pangea, the location of the Chinle basin, as monsoonal, and in a summary of sedimentologic and paleontologic data from the Chinle, Dubiel and others (1991) provided evidence to support a Triassic climate that was dominated by monsoonal circulation and alternating wet and dry seasons. Evidence for increasing aridity toward the end of deposition of the Church Rock Member of the Chinle was documented by Stewart and others (1972), Blakey and Gubitosa (1983, 1984), Dubiel (1987), Blodgett (1988), and Blakey and others (1988). The increase in aridity may reflect the continued northward movement of Pangea away from monsoon-dominated, tropical latitudes (Parrish and others, 1986; Parrish and Peterson, 1988).

STRUCTURAL SETTING

The northern part of the Chinle depositional basin inherited the Paradox Basin, a late Paleozoic feature of the Colorado Plateau. The Paradox Basin covers an area of approximately 27,000 km^2 and contains thick cyclical salt beds (Pennsylvanian Paradox Formation) (figs. 1, 3). The boundaries of this elongate, northwest-trending basin are defined by the maximum geographic extent of salt deposition (Hite, 1960).

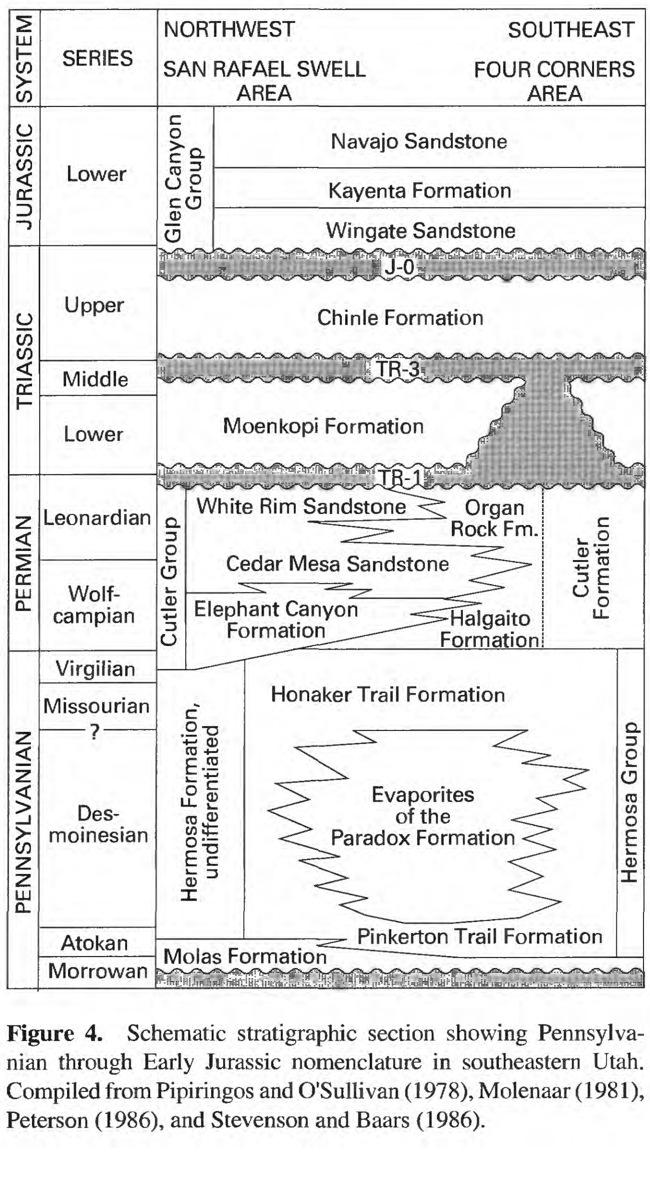


Figure 4. Schematic stratigraphic section showing Pennsylvanian through Early Jurassic nomenclature in southeastern Utah. Compiled from Pipiringos and O'Sullivan (1978), Molenaar (1981), Peterson (1986), and Stevenson and Baars (1986).

The present-day surface manifestation of the area in which salt deposition was greatest has been termed the "Paradox fold and fault belt" (Kelley, 1955). Also known as the Salt Anticline region (Cater, 1970), the belt is adjacent to the southern flank of the Uncompahgre uplift and parallel with the axis of the basin (fig. 1). It is approximately 90 km wide and 200 km long and consists of northwest-trending elongate salt diapirs that overlie the deeper structural depression of the Paradox Basin where evaporites of the Middle Pennsylvanian Paradox Formation attained a depositional thickness in excess of 3,000 m (Jones, 1959). The diapirs formed as a result of loading by overlying strata and were elongated into anticlinal features by salt flowage against and over deep-seated, northwest-trending basement faults (Baars and Stevenson, 1981) (fig. 4). Synclines developed in adjacent areas as salt flowed into the diapiric structures. The most active period of salt movement began after deposition of the first salt bed in Pennsylvanian time and lasted about 75 million years. Diapirism was locally active throughout the Mesozoic as evidenced by stratigraphic thinning or absence of post-Pennsylvanian strata near the salt anticlines (Doelling, 1988). Although most structural growth had ended by Cretaceous time, Colman (1983) documented movement on the Onion Creek diapir throughout the Quaternary, as recently as 250,000 B.P.

CANE CREEK SALT ANTICLINE

The Cane Creek anticline, similiar to larger salt structures in the Paradox Basin, strikes northwestward, parallel with the axis of the basin (figs. 1, 2). The Kings Bottom syncline separates the Cane Creek anticline from the collapsed Moab-Spanish Valley salt anticline to the east. The dip of strata between the two anticlinal axes is as much as 11°. Gravity lows over the Moab-Spanish Valley and Salt Valley anticlines are related to thickened salt cores (Joesting and others, 1966). The residual low at Moab Valley indicates a salt core rising about 2,400 m above the salt on either side (fig. 5). Angular intraformational unconformities within the Chinle Formation attest to salt activity during the Late Triassic (fig. 6). Although the gravity anomaly contrast is not as dramatic as the low over Moab Valley, the total thickness of salt at the Cane Creek anticline, on the basis of well data, is about 1,500 m (Joesting and others, 1966). By analogy, the Cane Creek anticline probably has an origin similiar to that of the Moab-Spanish Valley anticline but has not undergone collapse, breaching, and dissolution. Canyons and side canyons of the Colorado and Green Rivers bisect both limbs of the anticline (fig. 2) and have created accessible, laterally extensive exposures that permit evaluation of the alluvial architecture. Thus, the Cane Creek anticline is an excellent location in which to study progradation of a fluvial system over an active intrabasinal feature.

STRATIGRAPHY

In southeastern Utah, the Chinle Formation unconformably overlies Lower and Middle (?) Triassic coastal-plain deposits of the Moenkopi Formation (fig. 4).

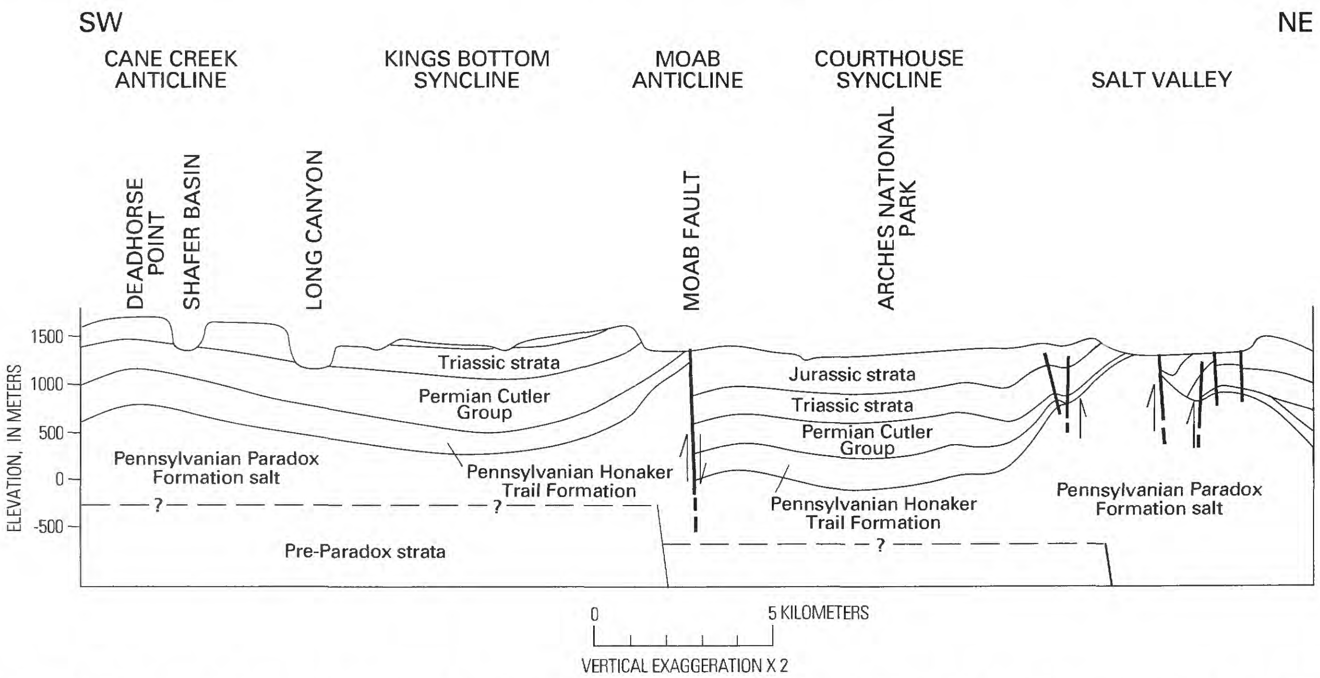


Figure 5. Schematic structural cross section, Paradox Basin, Utah, drawn normal to tectonic strike. Subsurface interpretations from Joesting and Case (1960) and Stevenson and Baars (1986).

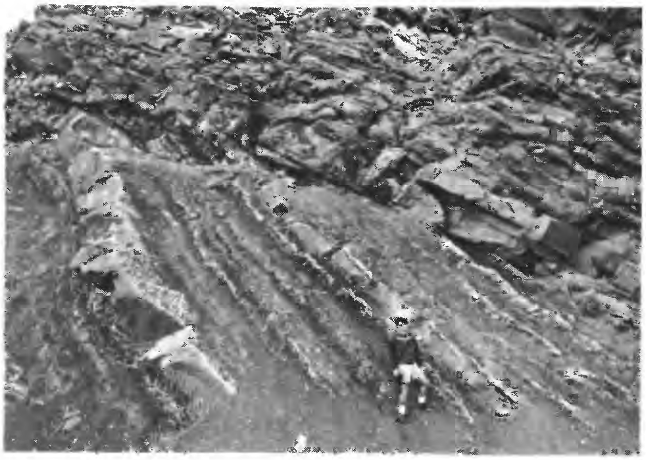


Figure 6. Intraformational unconformity within the Chinle Formation at the Portal of the Colorado River along the west side of the Moab-Spanish Valley salt anticline. Lower Chinle strata were pushed by diapirism above levels of deposition and consequently exposed to erosion before the remainder of the formation was deposited. Location is same location as figure 43 of Doelling (1988); location is shown in figure 2.

Structural discordance of beds at the contact between the Moenkopi and Chinle is slight except locally near the salt cores (Shoemaker and others, 1958). Locally on the flanks of some of the salt structures, the Moenkopi pinches out entirely and the Chinle unconformably overlies upturned strata of the Cutler and Hermosa Groups. On the Uncompahgre uplift, the Chinle rests directly on Precambrian basement (Molenaar, 1981). A widespread erosional event marked by the J–0 unconformity (Pipiringos and O'Sullivan, 1978) beveled the Chinle to successively older units to the west and separates the Chinle from the overlying Lower Jurassic Wingate Sandstone of the Glen Canyon Group (fig. 4). Biostratigraphic studies indicate an age of late Carnian to the later middle Norian for the Chinle (Ash, 1972, 1975; Breed, 1972; Colbert, 1972, 1985; Gottesfield, 1972; Litwin, 1986; Conrad, 1988).

Lithologic heterogeneities in the Chinle caused by abrupt changes in depositional environment make it difficult to divide the Chinle Formation in the Salt Anticline region into the members that are defined elsewhere on the Colorado Plateau. Stratigraphic subdivisions that are useful in the study area include, in ascending order, mottled strata, Moss Back Member, Kane Springs strata, Black Ledge, and Church Rock Member (fig. 7). The Chinle in the Salt Anticline region ranges from 0 to more than 300 m in thickness (Doelling, 1988).

Stewart and Wilson (1960) identified a northwest-trending belt of coarse-grained facies extending from southwest Colorado across the Salt Anticline region to central Utah. Finch (1959), in a uranium reconnaissance, noticed this belt of increased channel density and high sandbody interconnectedness and postulated that the system had been held static and localized along what he termed the "Moab-pinchout" (fig. 1), an irregular east-west line along which the Moss Back Member to the south pinches out against topographic highs along parts of the Cane Creek and Shafer anticlines (figs. 2, 8). The high sandstone to mudstone ratio on the north side of the Moab-pinchout represents a stacking of fluvial axes that is not present south of the Moab-pinchout or toward the northeast. The sand bodies deposited by the localized northwest-flowing fluvial systems are the focus of this study.

The basal contact of the Chinle with the Moenkopi is commonly marked by a thin, discontinuous white conglomeratic sandstone referred to as the "basal sandstone unit" or "mottled strata" (Stewart and others, 1972) (fig. 7). This unit is lithologically distinct from the overlying Moss Back Member and rarely is more than 6 m thick (Stewart and Wilson, 1960). Shoemaker and Newman (1959) noticed that this unit was not correlative with the type Shinarump Member of the Chinle to the south and suggested that it might be a basal unit of the Chinle, as believed by Dane (1935), or that it might represent an interval of Middle Triassic time that elsewhere on the Plateau was marked by a regional unconformity between the Moenkopi and Chinle Formations (fig. 4).

In the area of the Moab-pinchout, Gubitosa (1981) studied a zone of facies change in the Moss Back Member that Blakey and Gubitosa (1983) informally named the Kane Springs strata. The Kane Springs strata differ from the type section of the Moss Back Member of Stewart (1957) in that they are mostly intraclastic conglomerate rather than dominantly siliceous. The Kane Springs strata are characterized by fluvial, fluvial-lacustrine, and lacustrine facies that are stratigraphically equivalent to parts of the Moss Back, Petrified Forest, and Owl Rock Members of the Chinle to the south and west (fig. 8). The Kane Springs strata are about 50 m thick and consist of a lower unit of intraclastic conglomerate, calcarenite, and quartzarenite, a middle unit of bentonitic and calcareous mudstone, and a upper, discontinuous unit of intraclastic conglomerate, quartzarenite, micritic limestone, and calcareous mudstone (Blakey and Gubitosa, 1983; Hazel, 1991) (fig. 7).

Strata above the Kane Springs are assigned to the Church Rock Member of the Chinle. The Church Rock Member was defined by Stewart (1957) as strata above the Moss Back and Owl Rock Members in southeastern Utah. At or near the base of the Church Rock is a sandstone unit, informally called the Black Ledge (Stewart and others, 1959). The Black Ledge erosively overlies, intertongues, or is gradational with Kane Springs strata (fig. 7). It ranges from 10 to 12 m in thickness and consists of horizontal and cross-stratified sandstone that grades upward through horizontally and ripple-laminated siltstone into structureless or thin- to thick-bedded siltstone typical of the Church Rock Member (Stewart and others, 1972).

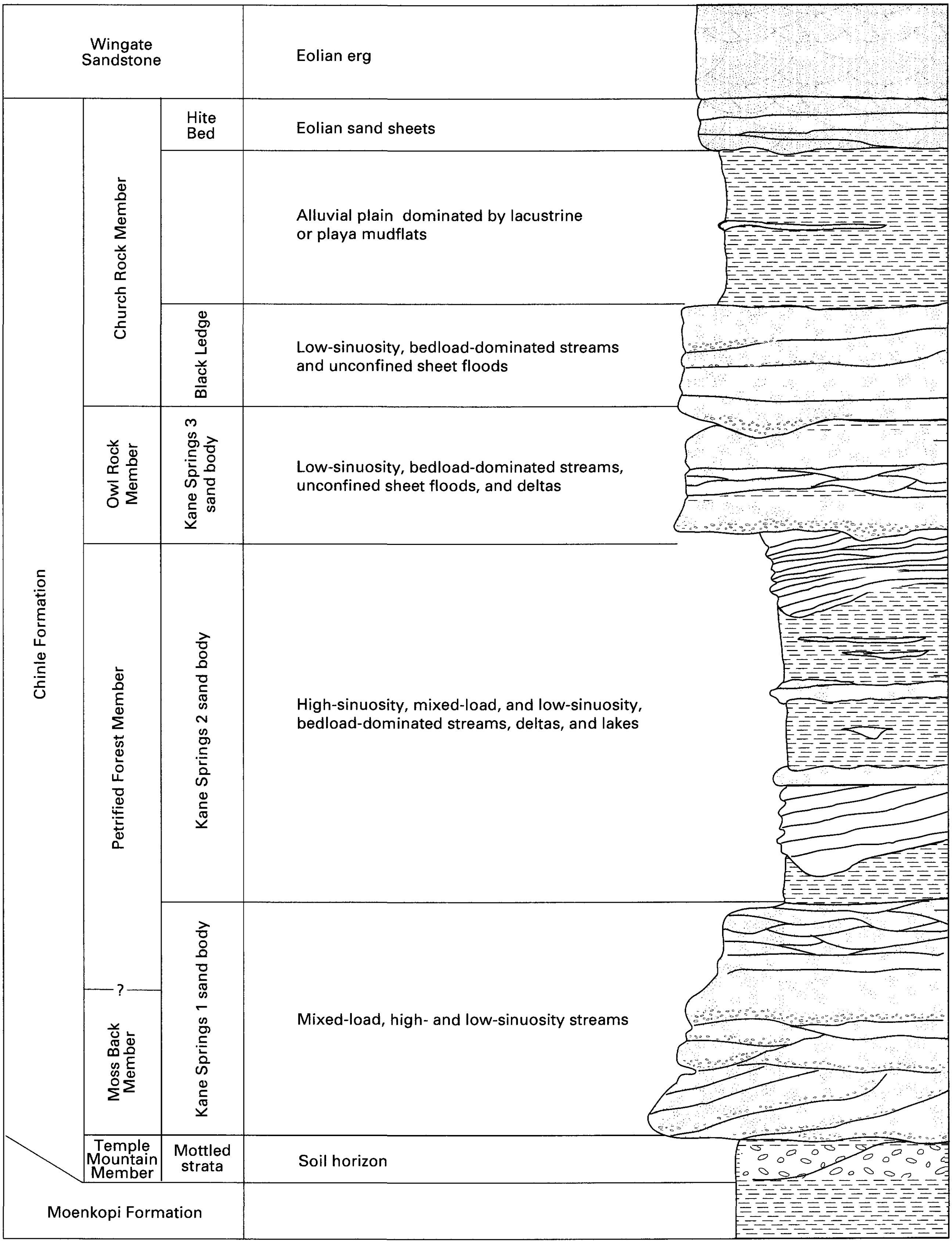


Figure 7. Formal and informal stratigraphic terminology, depositional systems, and lithofacies of the Upper Triassic Chinle Formation generalized from measured sections in the study area, Paradox Basin, Utah. Thickness is in meters.

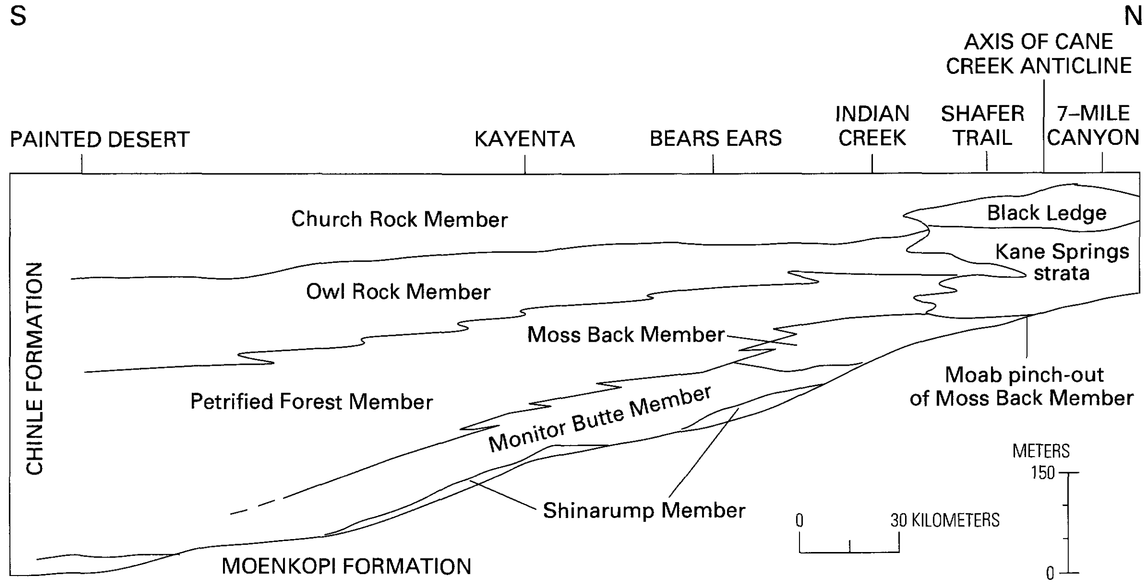


Figure 8. Generalized stratigraphic relationships within the Chinle Formation, southeast Utah and northeast Arizona. Modified from Blakey and Gubitosa (1983).

FACIES TERMINOLOGY AND METHODOLOGY

Lateral- and vertical-facies studies were conducted throughout the area to examine the spatial variability of sandstone units in the area of the Cane Creek anticline. Lithofacies were described, interpreted (table 1), and subdivided into architectural elements based on internal organization (lithofacies associations), bounding relationships, and external geometry (table 2). Three-dimensional facies architecture was described using two-dimensional vertical profiles. The profiles were developed by mapping bounding surfaces and lithofacies in the field on photomosaics. The approach used here is similar to the methodology of Allen (1983) and Miall (1985, 1988), but the hierarchy of bounding surfaces was redefined akin to an adaptation by Soegaard (1990). The hierarchy of facies and corresponding bounding surfaces that separate individual architectural elements is explained below.

The architectural elements (table 2) are defined by a sixfold hierarchy of bedding contacts (Miall, 1985, 1988). First-order contacts bound or envelop individual facies and record boundaries within microform and mesoform deposits in which little or no internal erosion is apparent. Deposits bounded by these surfaces represent the continuous sedimentation of a train of bedforms of similar type at a given point in time (Miall, 1985). Second-order contacts outline cosets of genetically related facies (facies sequence) and define groups of microform and mesoform deposits. Mesoforms include larger scale bedforms such as dunes and antidunes (Jackson, 1975). Second-order surfaces separate facies sequences (Soegaard, 1990) and indicate changes in flow conditions or flow direction but no significant break in time. Lithofacies above and below a second-order surface can be different, and only minor scour is associated with the contacts. Third- and fourth-order surfaces bound genetically related cosets and define larger scale depositional elements that constitute facies associations. Third-order surfaces outline single depositional architectural elements and indicate changes in fluvial stage or bedform orientation but no significant change in sedimentary style. For example, third-order contacts bound laminated sand sheets, gravel bars, sandy bedforms, smaller channels, and individual lateral-accretion units in a lateral-accretion macroform.

In the scheme of Miall (1985), fourth-order contacts represent the upper bounding surfaces of macroforms and are typically planar to convex upward (for example, larger channels and lateral- and downstream-accreting macroforms). A macroform represents a large-scale depositional feature that reflects the cumulative effect of many dynamic events over periods of several years to thousands of years (Jackson, 1975). Soegaard (1990) showed, however, that fourth-order surfaces are not restricted to the upper surfaces of macroforms, and he redefined fourth-order contacts as those that encase a "complex" of stacked, similar depositional elements. His definition is used in this study. A grouping or assemblage of sedimentation units that are genetically related by facies and (or) paleocurrent direction define a "complex" (Allen, 1983). Complexes occur singly or can be composed of stacked "storys" (Friend and others, 1979). For example, a channel element can consist of several discrete channel fills and is thus considered a complex.

Table 1. Summary of lithofacies, characteristics, and depositional interpretations.
[Modified from Miall (1978)]

Facies	Abundance	Grain size and texture	Structure	Geometry	Interpretation
Gm	Common	Granule to cobble size; poorly sorted	Massive to crude horizontal stratification; clast-supported fabric	Lens to narrow sheet; wedge shaped; flat scoured base	Longitudinal bars; intraformational channel-lag deposits.
Gt	Common	Granule to cobble size; poorly sorted	Trough cross-stratified; reactivation surfaces	Lensoid; fills concave-upward base	Channel fills; transverse bar.
Gp	Rare	Granule to cobble size; poorly sorted	Planar cross-stratified	Lens to narrow tabular sheets; scours fills; flat-scoured or slightly irregular base	Transverse to linguoid bars.
St	Common	Granule to sand size; well to poorly sorted	Trough crossbedded; reactivation surfaces	Lens to narrow sheet; ribbons	Dunes; lower flow regime.
Sp	Common	Granule to sand size; well to poorly sorted	Planar tabular cross-stratified	Narrow to broad sheets; slightly irregular top; flat base	Sand waves; linguoid or transverse bars; lower flow regime.
Sr	Abundant	Fine to very fine sand; well sorted	Ripple stratification of all types including climbing-ripple crosslamination	Narrow to broad sheets; associated with mudstone	Crevasse-splay and overbank deposits; waning flood flow.
Sh/ Sl	Abundant	Fine to medium sand size; well sorted	Horizontal and low-angle laminations; parting lineations	Broad sheets	Plane-bedded simple bars and bar tops; shallow upper flow regime.
Ss	Rare	Granule to sand size; well to poorly sorted	Scour-fill sand	Symmetrical hollows; sand file conforms to base of scour	Bar-top dissection by fluctuating discharge.
Se	Common	Sand size with mudstone intraclasts; muddy matrix	Erosional scours filled with massive to crude crossbedding	Erosional base of individual sand bodies	Intraformation lag deposits of flashflood derivation.
Fm	Abundant	Fine-grained mud and silt	Massive; structureless; color mottling; pedogenic modification; mud drapes	Thin tabular sequences; thin lenses in sand bodies	Floodplain deposits; lacustrine; bar tops; abandoned channels.
Fl	Abundant	Very fine sand, silt, or mud	Thin horizontal laminations or heterolithic interbedding of sandstone, siltstone, and mudstone; minor bioturbation, pedogenic modification, and color mottling	Thin to thick tabular sequences to podlike or lensoid bodies	Ponding in abandoned channels; overbank or waning flood deposits; mud drapes; upper flow regime in channels and lower flow regime on floodplains.

Fifth-order surfaces are laterally extensive and bound major sand bodies or groupings of complexes. These surfaces encompass entire depositional environments characterized by genetically related architectural elements specific to that environment (Miall, 1985; Soegaard, 1990). They are generally planar to slightly concave upward and may be marked by local scours and basal lag gravels. Sixth-order surfaces envelop large-scale depositional features, such as groups of channels or paleovalleys, that typically constitute a mappable stratigraphic unit such as a formation, member, or submember.

ALLUVIAL ARCHITECTURE AND PALEOENVIRONMENTAL ANALYSIS

The Chinle deposits in the study area are dominated by four narrow to broad sheet sand bodies. These sand bodies correlate to the lower, middle, and upper Kane Springs units of Blakey and Gubitosa (1983) and the Black Ledge (fig. 7). The Kane Springs sand bodies are designated herein the Kane Springs 1, 2, and 3 sand bodies. Sand bodies were examined in both transverse (perpendicular to paleoflow

Table 2. Architectural elements common to deposits of the Chinle Formation in the study area.
[Architectural elements modified from Miall (1985, 1988). Lithofacies are described in table 1]

Element	Lithofacies assemblage	Geometry	Relationships
Channel (CH)	Gt, Gm, Gp, St, Sp, Sh, Sl, Sr, Se, Ss, Fl	Lens or narrow sheet; broad to steep, concave-upward erosional base	Can contain any of the other elements; smaller channels defined by third-order surfaces; a channel complex contains similar stacked elements floored by fourth-order surfaces.
Lateral-accretion macroform (LA)	St, Sp, Sh, Sl, Sr, Se, Gm, Gt, Fl	Wedge or sheet	Characterized by third-order lateral-accretion surfaces; can compose the fill of a channel complex.
Downstream-accreting macroform (DA)	St, Sp, Sh, Sl, Sr, Se, Ss	Lenticular; flat base	Tabular bedforms having convex-upward second-order internal erosion surfaces and upper bounding surfaces.
Gravel bars and bedforms (GB)	Gm, Gt, Gp, Se	Tabular to sheet	Tabular bodies 1–3 m thick; commonly interbedded with sandy bedforms.
Sandy bedforms (SB)	St, Sh, Sl, Sp, Sr, Se	Lens or sheet	Fining-upward from lithofacies Se or St at base to lithofacies Sh/Sl; lithofacies Sp rare; third-order flat to undulose erosional surfaces with lags separate similiar stacked sandy bedform or gravel bar and bedform elements.
Laminated sand sheets (LS)	Sh, Sl, minor St, Sp, Sr	Sheet, 0.5–5 m thick	Bound by flat to undulose third-order erosional surfaces; lithofacies Se ubiquitous at base; lacks fining- or coarsening-upward facies sequences.
Overbank fines (OF)	Fl, Fm	Thin to thick tabular sequences	Rhythmically or massively bedded; abundant subaerial features; may fill abandoned channels; may lack bedding due to burrowing or pedogenesis.

direction) and longitudinal sections (parallel with paleoflow direction) (fig. 2). Sand bodies are classified as single-story or multistory (fig. 9). Those having width to height ratios of less than 15 are ribbons, and those having ratios of greater than 15 are sheets (Friend and others, 1979). Blakey and Gubitosa (1984) defined narrow sheets as having width to height ratios of less than 100 and regionally extensive sheets as having width to height ratios of greater than 100.

For each of these four sand bodies, I present description and a paleoenvironmental interpretation. Architectural characteristics are summarized in table 3. Only those aspects that are crucial to understanding the environment of deposition are included herein; for other details and profiles, see Hazel (1991). Profiles used to examine the sand bodies consist of a photomosaic, a facies and bounding-surface map, and a stratigraphic unit interpretation. Architectural elements are labeled according to codes listed in table 2 and lithofacies according to codes listed in table 1. Circled numbers define the order of the erosional bounding surface. Paleocurrents within individual facies are denoted by arrows oriented relative to the outcrop (for example, paleocurrents that parallel the outcrop are represented by horizontal arrows).

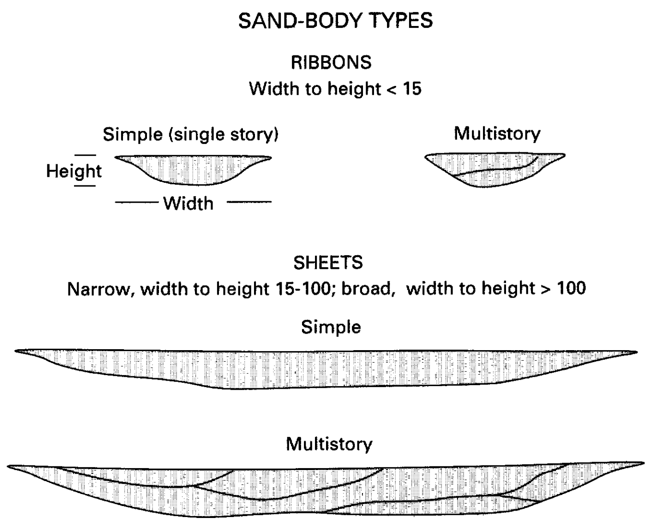


Figure 9. Classification of fluvial sand bodies. Modified from Blakey and Gubitosa (1984).

KANE SPRINGS 1 SAND BODY—LOW- AND HIGH-SINUOSITY STREAMS

DESCRIPTION

The Kane Springs 1 sand body crops out as a ledge and forms a prominent laterally extensive single-story to multistory sheet sandstone. Its width to height ratio is almost invariably greater than 100. The thickness of the sand body ranges from 0 to 30 m and averages 10 m. The base, which has as much as 5 m local relief on the underlying Moenkopi or on the mottled strata of the Chinle, is a sixth-order erosion surface (fig. 10). This surface is defined by its erosive nature and by the presence of overlying intraformational facies Gm or facies Gt. The surface is easily discernible due to the truncation of colorful paleosols of the Moenkopi or mottled strata. The Kane Springs 1 sand

Table 3. Architectural characteristics of units of the Chinle Formation in the study area.
[Units are listed top to bottom from youngest to oldest. Lithofacies are described in table 1, and architectural elements are described in table 2]

Unit	Geometry	Elements	Relationships
Black Ledge	Narrow multistory ribbon or sheet; width to height >50–100:1; 10–30 m thick	Laminated sand sheets, sandy bedforms, gravel bars and bedforms, channels	Exceptionally thick in Long Canyon; separated from the Kane Springs 3 sand body by an extensive, planar to undulose, sixth-order erosional surface; multistory stacking composed of offlapping elements that taper against and truncate each other; third-order erosional surfaces defined by lithofacies Se separate individual elements; little or no evidence of coarsening- or fining-upward cycles; more quartzarenite than in Kane Springs 3 sand body, but conglomerate is almost entirely calcirudite; unit thins or is missing over the Cane Creek anticline.
Kane Springs 3 sand body	Single-story to multistory broad sheet; width to height >100:1; 5–20 m thick	Laminated sand sheets, gravel bars and bedforms, sand bedforms, channels, lateral-accretion macroforms, rare downstream-accreting macroforms	First appearance of laminated sand sheets; sand bodies commonly splay and thin into proximal floodplain deposits similar to narrow sheets in Kane Springs 2 sand body; extensive sixth-order erosional surface at base can be traced for several kilometers; undulatory third-order bounding surfaces separate individual elements; channel elements are broad and have low-relief margins.
Kane Springs 2 sand body	Ribbon; narrow sheet, tabular mudstones; width to height <10–50:1; 10–30 m thick	Channels, lateral-accretion macroforms, sandy bedforms, overbank fines	Thins or is absent over Cane Creek anticline; lateral-accretion macroforms are filled with heterolithic facies assemblage and have steeply dipping channel margins; superimposed stacking of channel elements commonly forms lateral-accretion complexes; third-order lateral-accretion surfaces in heterolithic facies assemblage have preferential dip toward the northeast; ribbon and narrow sheet sand bodies have low-relief margins; base is a fifth-order erosion surface draped by lithofacies Gm.
Kane Springs 1 sand body	Narrow to broad sheet; width to height <20–100:1; 0–30 m thick	Channels, lateral-accretion macroforms, gravel bars and bedforms, sandy bedforms	Thins or is absent over Cane Creek anticline; lateral accretion is present only at the base; the base, having as much as 5 m of relief, is a highly erosive sixth-order surface; third-order lateral-accretion surfaces have preferential dip toward the northeast.
Mottled strata	Ribbon to sheet?; width to height <25:1	Channels, gravel bars and bedforms, sandy bedforms	Scattered as lenses and isolated channel fill; 0–10 m thick; rose quartz pebbles; friable exposures; dense pedogenic nodule horizons.

body is present in the western part of the study area (Hell Roaring Canyon, Mineral, and Horsethief Bottoms) but thins or is absent over the crest of the Cane Creek anticline (Long and Kane Springs Canyons) (Hazel, 1991, pl. 1).

Most of the sand body is composed of several architectural elements: channel forms composed of sandy bedforms and gravel bars, and lateral-accretion complexes (table 3). The elements are discrete sheetlike to broadly lenticular in geometry and 1–7 m thick. Lateral-accretion complexes are at or near the base (figs. 11, 12). Stacked elements are separated by undulose third- and fourth-order surfaces. The sandy bedforms and gravel bars are characterized by a complex, heterolithic facies association, primarily facies Gt, Gm, St, Sp, Sh/Sl, and Sr, and numerous internal second-order scour surfaces (fig. 10). Fining-upward associations consist of a basal lag of facies Gt that passes upward through facies St/Sp or Sh/Sl; facies Sr is preserved near the tops of some elements. The lithofacies typically are randomly interbedded and exhibit no discernible fining- or coarsening-upward trends. Sequences of stacked cosets of tangential-planar crossbedding (facies Sp) are common (fig. 13). The foreset dip of Sp sets tends to diverge from that of trough-crossbedded (facies St) sets (fig. 14). Mudstone drapes, dense carbonaceous horizons containing logs, and 1-meter-thick intraclast lags within a muddy matrix (facies Se) are associated with second- and third-order surfaces. The lithology of the sand body includes calcirudite and hybrid conglomerate composed of siliciclastic and carbonate clasts but is most commonly medium- to coarse-grained calcarenite and quartzarenite.

INTERPRETATION

The Kane Springs 1 sand body is the product of an alluvial meander-braidbelt setting that was subject to seasonal discharge. The streams that deposited the single-story to multistory sheet had a channel pattern intermediate between braided and meandering; probably both types of channels were present (fig. 15). Initially, the alluvial plain was characterized by coarse-grained meanderbelts partially confined between anticlinal upwarps. Unconfined braided channels developed as the system aggraded between, and eventually buried, the positive areas (Blakey and Gubitosa, 1984; Hazel, 1991). This is evidenced by increased thickness of fluvial sand bodies in synclinal areas, thinning or absence of sand bodies over anticlinal crests, and the limited preservation of lateral-accretion surfaces at or near the base of the Kane Springs 1 sand body (Hazel, 1991, pl. 1).

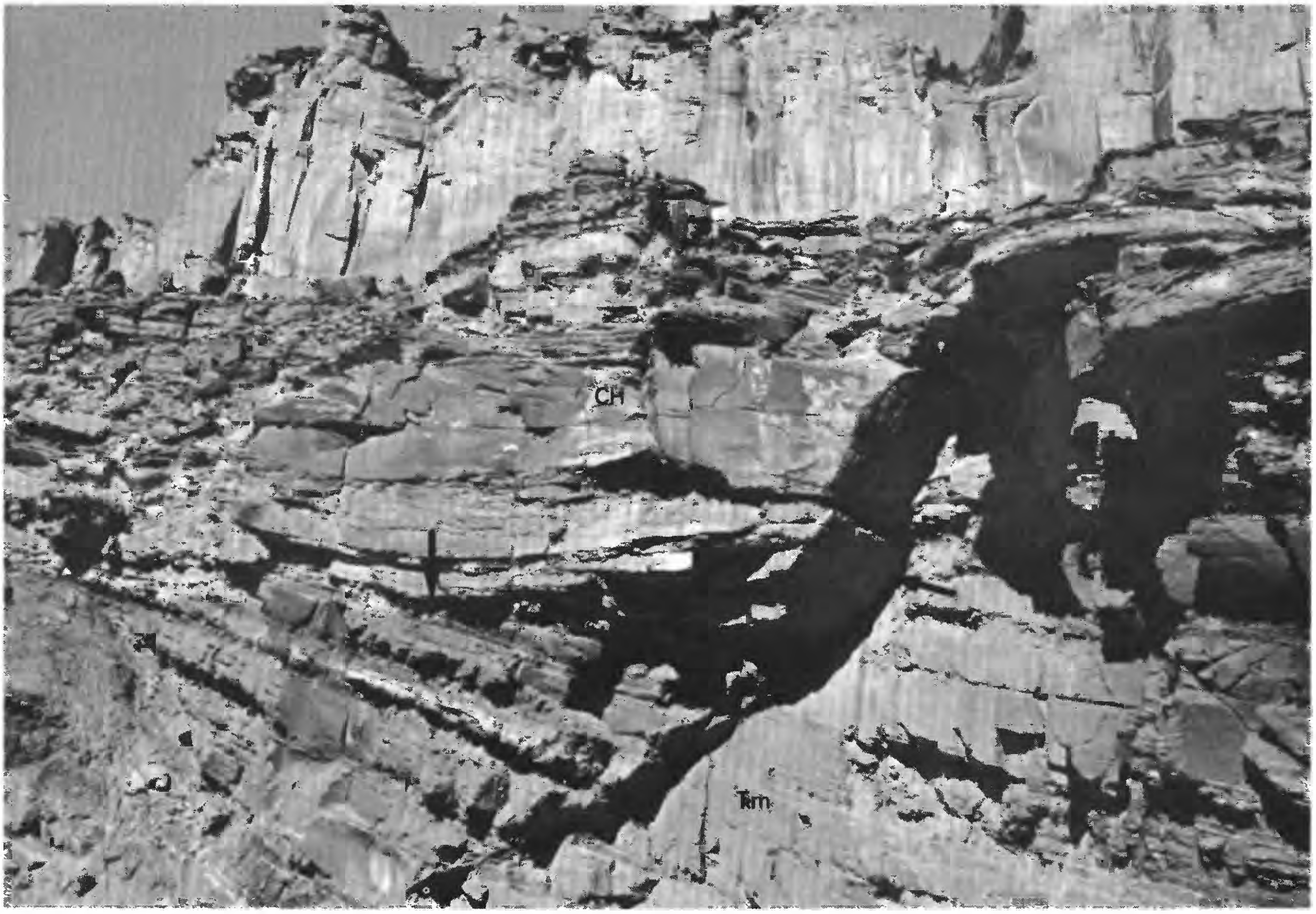


Figure 10. Channel element (CH), approximately 4 m thick, incised into beds of the Moenkopi Formation (Ꞧm). Note angular unconformity (left arrow) between locally uplifted Moenkopi strata and incised Chinle channel. The element displays a prominent cutbank (right arrow) and is filled with alternating heterolithic planar-tabular and trough-crossbedded sandstone and conglomerate (facies Sp/St and Gp/Gt), including a 1-meter-thick set of facies Sp. Location is shown in figure 2. Lithofacies and architectural elements are defined in tables 1 and 2, respectively.

The absence of persistent lateral accretion-bedding implies that the composite single-story to multistory sheet originated by aggradation of laterally adjacent and connected channel belts rather than by continuous lateral migration across the floodplain of sinuous streams with point bars (Allen, 1970). Lateral accretion is not confined to point or side bars but may be associated with mid-channel sand shoals and sand flats of low-sinuosity sand-dominated rivers (Cant, 1978; Rust, 1978; Allen, 1983; Bristow, 1987). The abundant planar and trough cross-stratified conglomeratic sandstone (facies Gp, Gt, Sp, and St) and the presence of symmetrical to lenticular sandstone-filled scours suggest that the Kane Springs 1 fluvial system was similar to modern coarse-grained point-bar deposits documented for the Amite River in Louisiana and the Colorado River in Texas (McGowen and Garner, 1970). Conversely, vertical stacking of similar bedform types that commonly overlie a gravel lag resembles migrating transverse bar deposits of braided rivers

Figure 11 (facing page). Sheet deposits of the Kane Springs 1 sand body of the Chinle Formation in Hell Roaring Canyon. The outcrop face trends N. 40° W. and is oriented oblique to paleoflow. This profile illustrates one of the few outcrops where sigmoidal epsilon erosion surfaces that have both offlapped upper terminations and downlapped lower terminations are preserved in a lateral-accretion complex (element 1–LA). The dip on third-order erosion surfaces is as steep as 40°; however, the average dip is closer to 20°. Each lateral-accretion unit consists of fine- to medium-grained bluish-gray ripple-laminated sandstone (facies Sr) that lacks either fining- or coarsening-upward trends. Shale partings as thick as 10 cm drape the third-order surfaces and extend to the base of the scour. Stacked on top of the lowermost lateral-accretion element are sandy bedform (SB) and gravelly bedform (GB) elements that together comprise a multistory sheet geometry. Location of outcrop is shown in figure 2. Lithofacies and architectural elements are defined in tables 1 and 2, respectively.

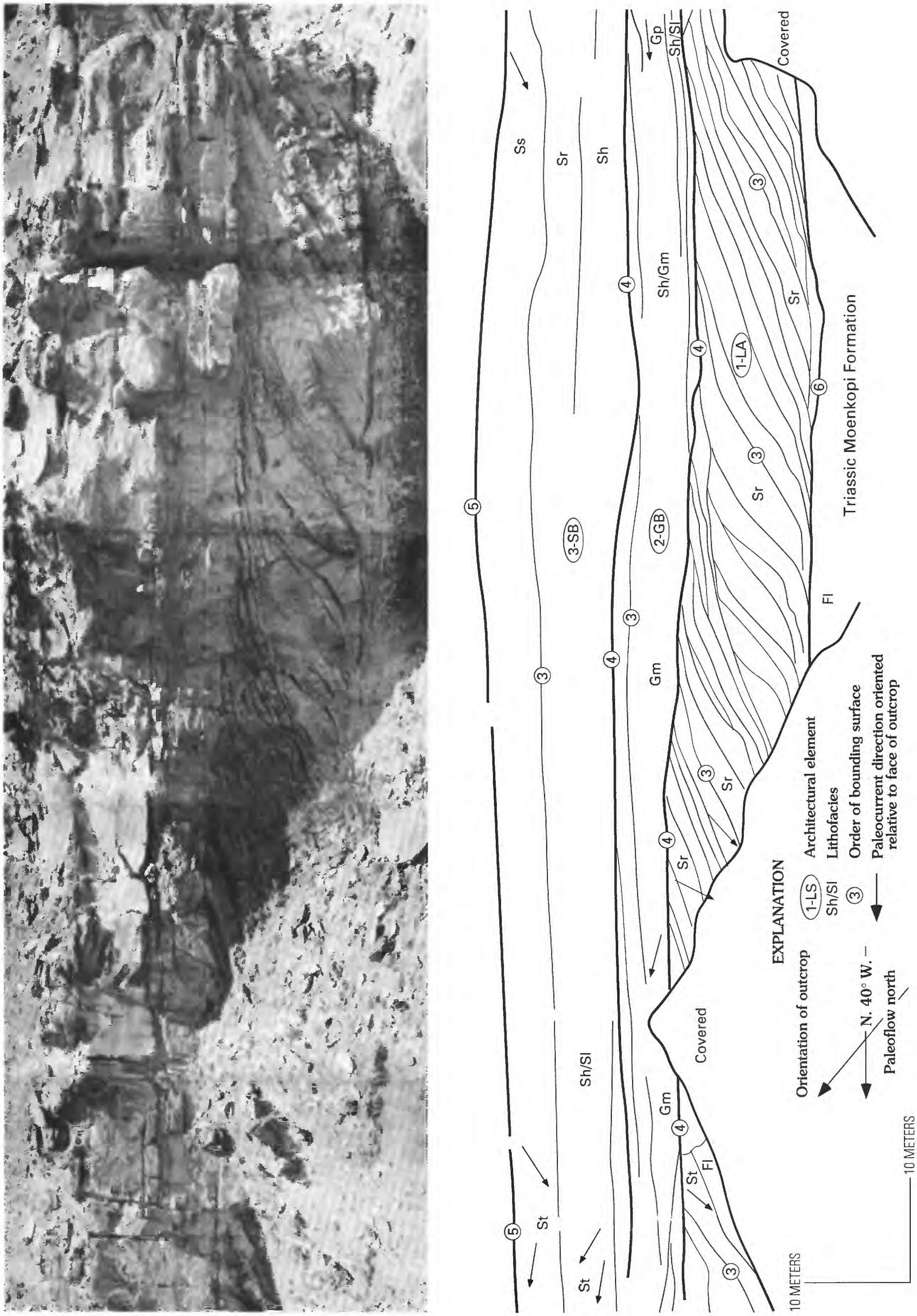

Ss
Sr
Sh
Sh/Gm
Gp
Sh/Sl
Covered
3-SB
2-GB
1-LA
Gm
Sr
Fl
St
Sh/Sl
Covered
Triassic Moenkopi Formation
EXPLANATION
1-LS Architectural element
Sh/Sl Lithofacies
3 Order of bounding surface
Paleocurrent direction oriented relative to face of outcrop
Orientation of outcrop
N. 40° W.
Paleoflow north
10 METERS

Figure 12. Lithofacies Sr (table 1) oriented almost perpendicular to a third-order lateral-accretion surface (arrows) in element 1–LA (table 2) in figure 11. Hammer is shown for scale.

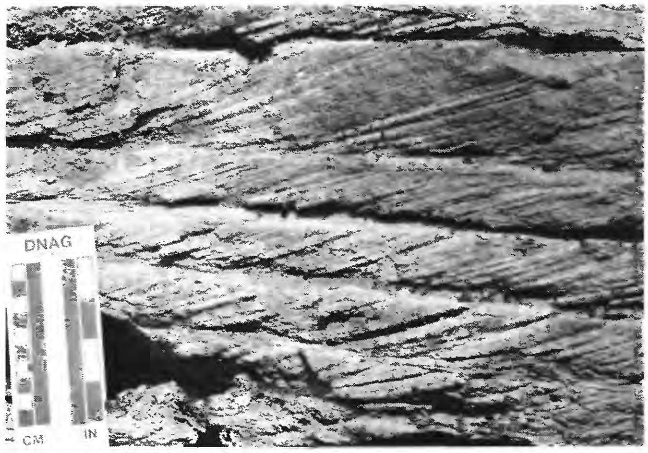


Figure 13. Stacked sets of tangential-planar cross-stratification (facies Sp, table 1) composed entirely of calcarenite in the Kane Springs 1 sand body of the Chinle Formation.

(Miall, 1985). Thickening-upward, stacked sets of facies Sp likely are the result of downstream-migrating transverse and linguoid bars in which the coarser bar top migrates over the finer base (Gustavson, 1978).

The size and geometry of the channel and lateral-accretion macroforms indicate that the streams were narrow (20–30 m) and deep (5–10 m). Only near the base is channelization evident. Point bars can extend two-thirds of the distance across the channel (Allen, 1970). A measure of bankfull channel depth for the lateral-accretion complex shown in figure 11 is given by the average thickness of the lateral accretion bedding (10 m), and the approximate bankfull channel width (35 m) is calculated by determining the average horizontal width of all lateral-accretion units and multiplying this value by 1.5 (Ethridge and Schumm, 1978). The resultant width to depth ratio of 4 is more indicative of a suspended-load river (<10) than a mixed-load river (<40) (Schumm, 1968). The steep slopes of the lateral-accretion surfaces are unusual in that similar suspended- to mixed-bedload deposits described from both the recent and ancient are not nearly as steep (less than 15°). The closest modern analog is the Barwon River in Australia where inclined bedding dips at angles as steep as 23° (Woodyer and others, 1979). A deep but narrow thalweg combined with bank stability can create steep lateral-accretion surfaces (Bridge, 1985). River banks may have been stabilized by minimal erosion on the concave bank of the stream due to the fine-grained cohesive sediment, vegetation, and possible early cementation. The third-order surfaces could also be oversteepened by differential compaction due to changing mud-drape thickness; however, the origin of such steep surfaces remains somewhat problematic.

Evidence for flashy, episodic flow includes upper flow regime plane beds and chute channels that indicate high-stage discharge fluctuation and reactivation surfaces and complete set-bounding mud drapes that indicate low-stage discharge. The lack of muddy channel fill and the predominance of sandstone and conglomerate suggest that chute cutoff was common and neck cutoff was not significant. McGowen and Garner (1970) found that the generation of chutes on the upper point bar is indicative of seasonal fluctuations in flow and that chute bars form during high-flow stage. If chute cutoff was an important abandonment mechanism in channels of the Kane Springs 1 depositional system, the rivers were probably characterized by a seasonally fluctuating, perennial flow regime. The third-order mudstone-defined lateral-accretion surfaces extending to the basal scour (fig. 11) are similar to features observed in both ancient and recent rivers that have been attributed to highly variable and possibly seasonal flow regimes (Puigdefabregas and van Pliet, 1978; Stewart, 1981; Campbell and Hendry, 1987).

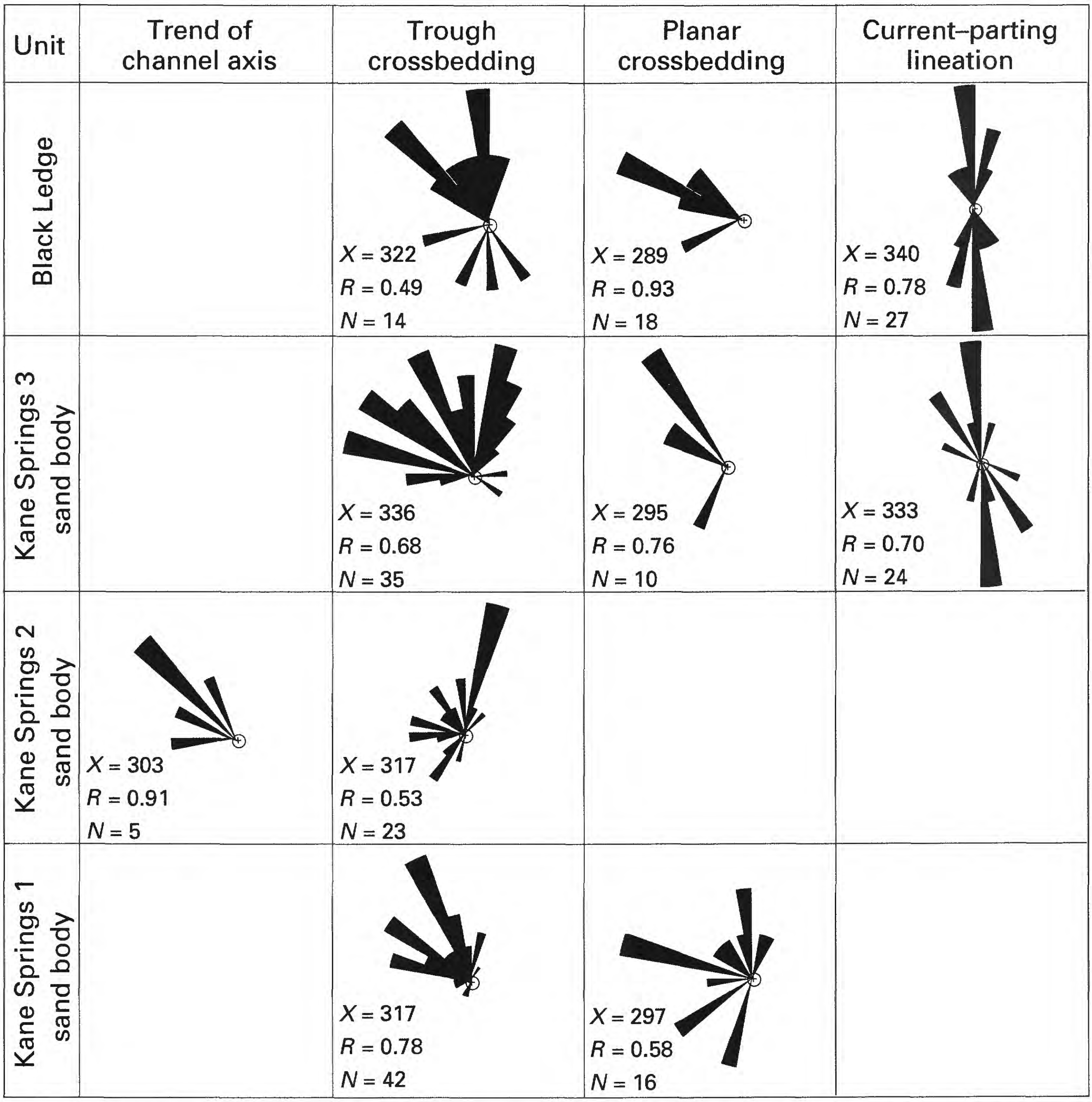


Figure 14. Paleocurrent plots for sedimentary structures of units of the Chinle Formation in the study area. Most structures were invariably small to medium scale and therefore were not differentiated into size classes. The rose diagram petal width is 10°; *X*, vector mean paleoflow direction; *R*, vector magnitude (1.0=0 percent dispersion); *N*, number of measurements. Range of vector magnitudes is greater than 50°, indicating low variance between sequences and the individual depositional systems.

KANE SPRINGS 2 SAND BODY—HIGH-SINUOSITY STREAMS

DESCRIPTION

The Kane Springs 2 sand body typically crops out as a slope-forming unit of bluish-gray to greenish-gray mudstone, and its sandstone to mudstone ratio contrasts sharply with that of the lower Kane Springs 1 sand body (Blakey and Gubitosa, 1984). In the study area, the mudstone deposits contain two distinctly different types of amalgamated channel deposits that combine to build a complex sand body (Hazel, 1991). The thickness of the sand body ranges from 10 to 30 m and averages 20 m; the sand body is thicker in synclinal areas but is present over the crest of the Cane Creek and Moab anticlines (Hazel, 1991, pl. 1). The base is a

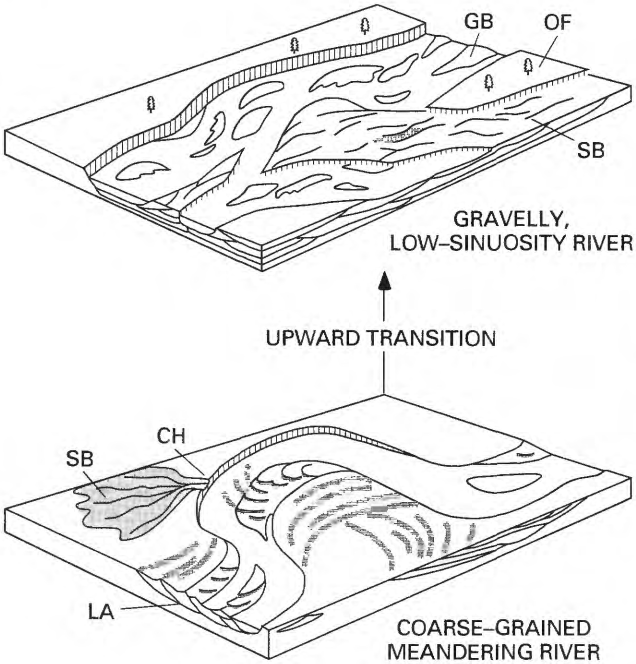


Figure 15. Inferred models for deposition in the Kane Springs 1 sand body of the Chinle Formation. It is probable that both types of channels coexisted on the alluvial plain. Architectural elements are defined in table 2. Modified from models 2, 3, and 5 of Miall (1985).

poorly defined fifth-order erosion surface that is commonly defined by weathered calcirudite (facies Gm).

The Kane Springs 2 sand body is characterized by a complex interbedded heterolithic facies association (fig. 16) and discrete, ribbon or tabular sand bodies. Multistory narrow sheets are present in Kane Springs and Long Canyons (fig. 17). Sand bodies range from thin ripple-laminated sequences generally less than 2 m thick to ribbon or narrow single-story sheets as thick as 5 m. Channel elements and lateral-accretion complexes are the dominant macroforms in the heterolithic facies, and sandy bedforms and laminated sheets predominate in the sand bodies.

The basal 5–10 m of the Kane Springs 2 sand body is dominated by mudstone (facies Fm). Above the mudstone, ribbon or sheet sand bodies typically alternate with equal to greater amounts of heterolithic facies. The sand bodies are single story, range from 1 to 10 m in thickness, and can be traced for 200 m perpendicular to paleoflow. They have low-relief channelized margins represented by concave-upward, fourth-order bounding surfaces. The lithology of the sand bodies ranges from fine-grained sandstone to calcirudite. The heterolithic facies association displays laterally accreted paleochannels that have prominent cutbanks and high-relief erosional bases. The average trend on preserved cutbank channel margins is N. 60° W. (fig. 14). The heterolithic facies association typically is composed of bluish silty mudstone interbedded with sandstone or conglomerate

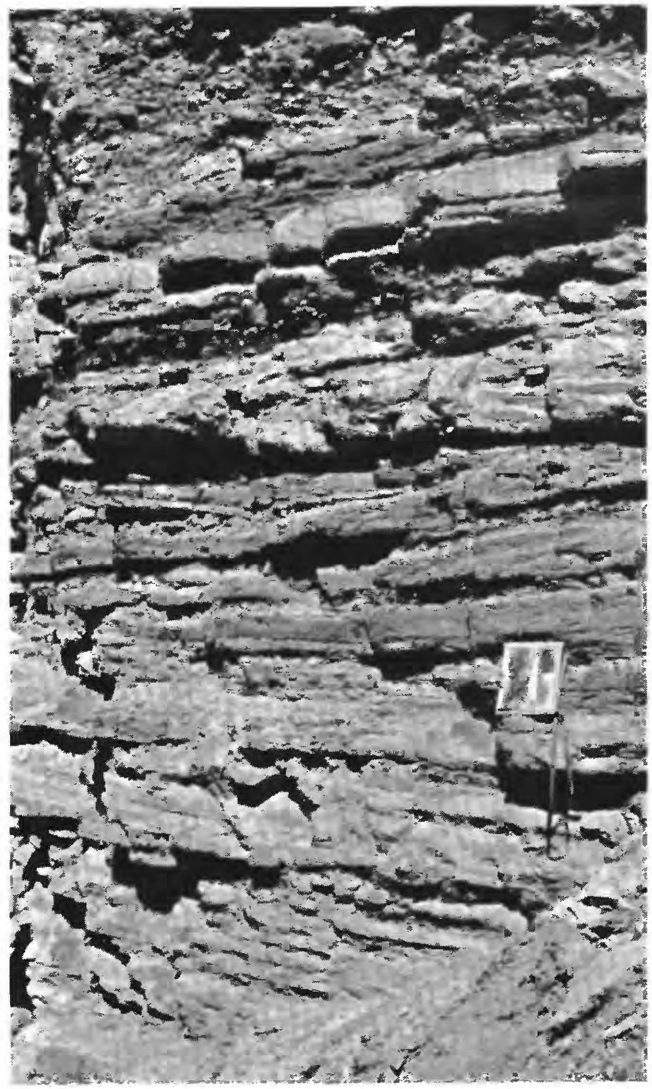


Figure 16. Interbedded fining-upward couplets composing the heterolithic facies association in the Kane Springs 2 sand body under protective ledges in Kane Springs Canyon. Map board (35 cm long) is shown for scale. Location is shown in figure 2.

in discrete fining-upward couplets 0.5–2 m thick (fig. 18). The couplets are composed of small-scale facies St and ripple-laminated facies Sr or lack any discernible stratification. The base of each couplet is marked by a thin intraclastic mudstone veneer of the third-order lateral-accretion surface (fig. 18). Mudcracks with interstitial sandstone infilling and color mottling are common desiccation features.

A unique feature of the heterolithic facies association is lateral-accretion macroforms contained within large concave-up fourth-order scours (fig. 19). These scour-bounded lateral-accretion macroforms truncate and erosively overlie one another to form stacked superimposed complexes. The third-order surfaces in individual macroforms dip preferentially to the northeast. The fill of each scour displays fining-upward bundles of heterolithic interbedding of mud, sand,

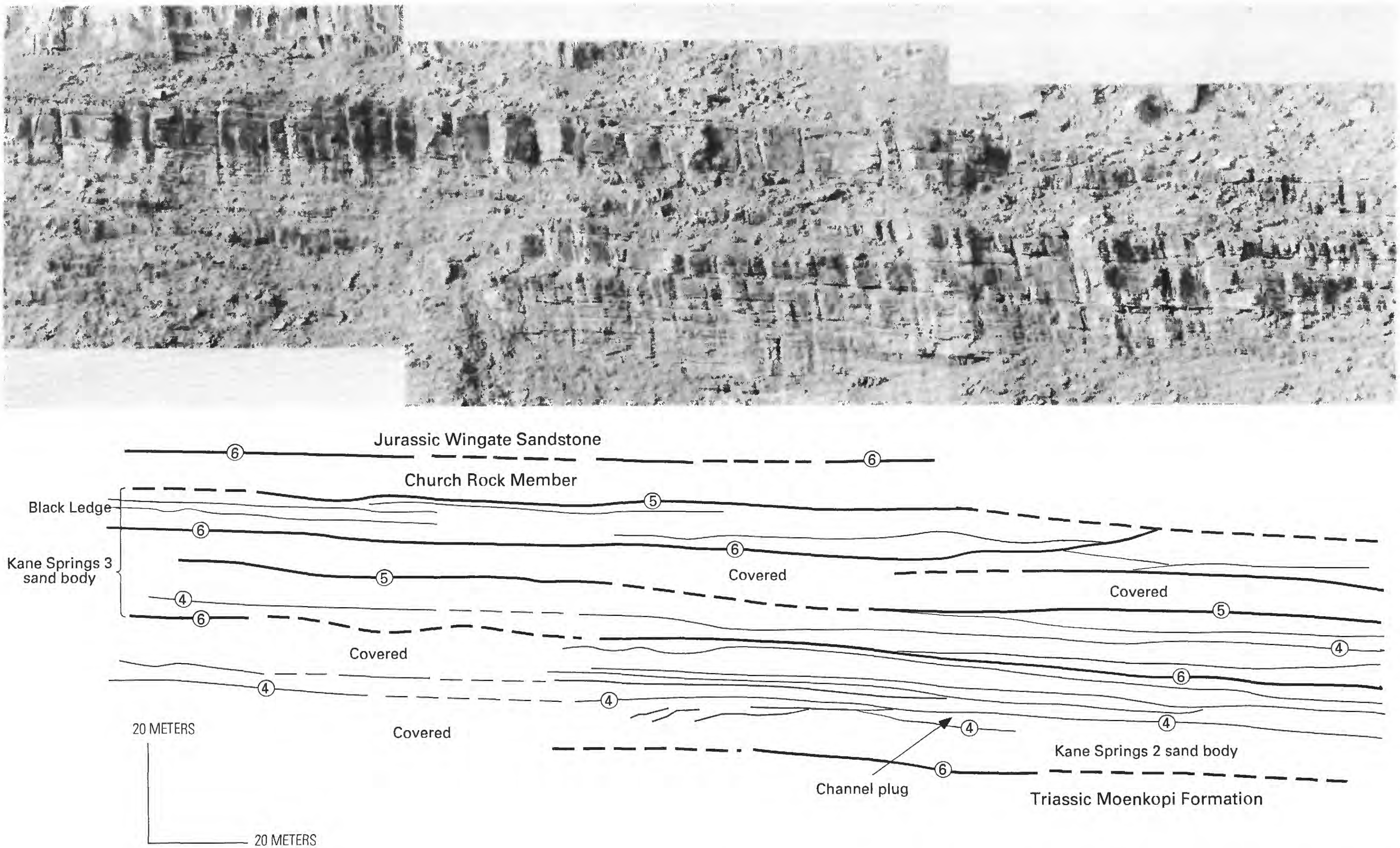


Figure 17. Tabular and narrow sheet sand bodies and associated floodplain deposits in the Kane Springs 2 and 3 sand bodies of the Chinle Formation in Long Canyon. Note channel plug of facies Fl (table 1) in the heterolithic facies association of the Kane Springs 2 sand body. Circled number indicate order of bounding surface. Axis of Cane Creek anticline is to the left of the profile. Location of outcrop is shown in figure 2.

Figure 18. Individual fining-upward lateral-accretion bundle within the heterolithic facies association of the Kane Springs 2 sand body of the Chinle Formation. Hammer is shown for scale.

and gravel facies, third-order bounding surfaces that dip at steep angles (20°–30°), and a surface that commonly is defined its entire length by thick mudstone intraclast layers (facies Se).

INTERPRETATION

Lateral profiles for the Kane Springs 2 sand body detail a fluvial-dominated delta plain constructed by low-gradient, tightly meandering and perhaps anastamosing streams (fig. 20). Evidence for a coalescing fluvial-deltaic sequence was provided by Blakey and Gubitosa (1984), who documented lacustrine and deltaic deposits in the Kane Springs 2 sand body along the Green River, just west and northwest of the study area (fig. 21).

The heterolithic facies association resembles floodplain, levee, and point-bar deposits common to suspended- and mixed-load meandering-stream deposits described from both the ancient and the recent (Jackson, 1976; Nanson, 1980; Stewart, 1981, 1983; Nanson and Page, 1983). This interpretation is based on the presence of well-developed lateral-accretion surfaces, rare ridge and swale topography, fining-upward accretionary bundles, paleoflow indicators oriented oblique to perpendicular to the lateral-accretion surfaces, preserved cutbanks, and channel plugs. Low paleocurrent dispersion, lateral-accretion macroforms, overlying concave-upward erosional scours, and cutbank preservation suggest that lateral sweeping of the fluvial system was limited and that the system was confined to partly confined. The crosscutting channel and lateral-accretion macroforms and rare abandoned channel plugs indicate that the streams were narrow (15–25 m) and deep (2–5 m). It is possible that a network of channels coexisted and were anastamosed. Smith and Smith (1980) showed that, despite the more characteristic vertical mode of accretion, lateral accretion may occur on channel bends in anastamosing rivers

The unique channel fill contained within the concave-upward fourth-order scours likely was the product of a single flood event in which vertical filling led to avulsion and incision of new courses. The accretionary units both conform to the fourth-order erosional cour and climb at a supercritical angle without merging or truncating the surface; this suggests extreme cutbank stability and rapid vertical aggradation of the channel floor and point bar. The repetition of lateral-accretion bundles composing channel fills requires a mechanism of progressive abandonment whereby floods of prolonged duration, and rapid fluctuations in discharge, allowed continuous aggradation to occur. The system would then avulse by diversion of flow through an adjacent distributary as former channels were rapidly plugged (Friend and others, 1979) rather than by neck or chute cutoff. Avulsed distributaries incised the pre-existing delta or floodplain deposits resulting in the concave-upward fourth-order scours. This type of avulsed distributary is distinct from distributaries that form by levee accretion and progradation (Brown, 1979). The preferential dip of third-order surfaces, contained within the fourth-order scours that bound individual lateral-accretion macroforms, suggests preferential downslope erosion and meander cutoff. This style of lateral migration is similar to channel-belt processes that resulted in preferentially abandoned meander loops in the South Fork of the Madison River (Bridge and Leeder, 1987).

The narrow sheet and lenticular sand bodies in the Kane Springs 2 sand body are interpreted as superimposed systems of fluvial channels and overbank splays formed by multiple periods of lateral and vertical accretion. Indistinct lateral-accretion bedding suggests that each episode of channelization was followed by asymmetrical infilling and lateral shifting of channels that were 2–5 m deep and as wide as 200 m. The rivers were localized within one area of the floodplain for an extended period of time and either coexisted or alternated depositional episodes with the finer grained, higher sinuosity streams. The low-relief channel margins, sandy bedforms, and lack of systematic lateral-accretion bedding in any of the individual paleochannels suggest deposition during overbank flow or sheetflooding. The color mottling and minor nodule development is typical of deposition in a floodbasin environment experiencing periodic wetting and drying (Bown and Kraus, 1981). Stear (1983) has attributed the origin of extensive sheet splays to poor levee development, which allows crevasse splays to coalesce as overbank flooding spreads outward from the channels. This process is analogous to many flood events (Williams, 1971; Tunbridge, 1984) and accounts for the high degree of sand-body interconnectedness.

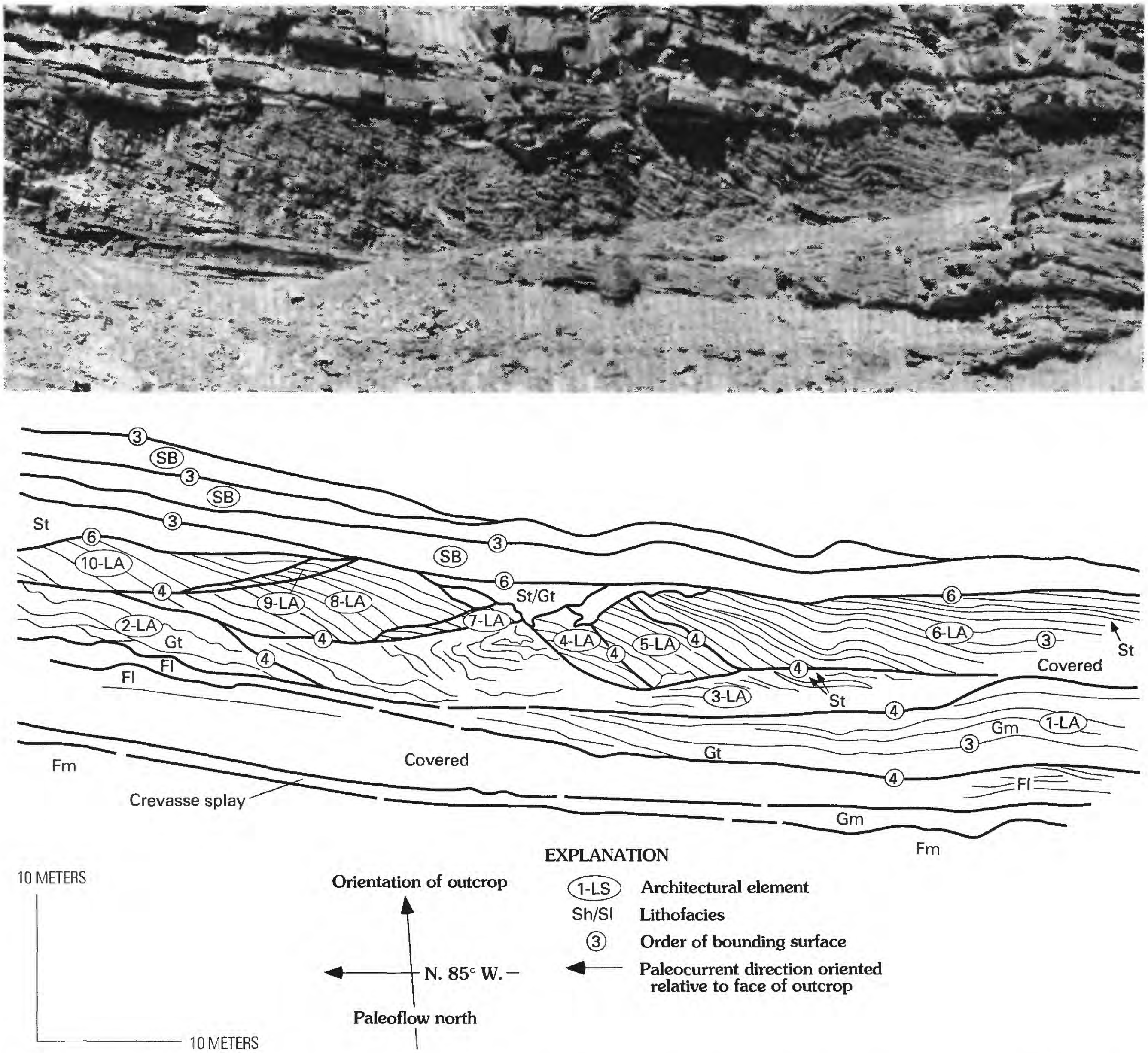


Figure 19. Stacked superimposed lateral-accretion macroforms (LA) in lower Long Canyon. The profile is approximately 90 m long, and the Kane Springs 2 sand body of the Chinle Formation is 30 m thick. A sixth-order surface, extending the length of the outcrop, separates the Kane Springs 2 sand body from the Kane Springs 3 sand body. The view is slightly oblique from below. The third-order lateral-accretion surfaces that comprise the channel fill of large concave-up fourth-order erosional scours preferentially dip toward the northeast. The undulatory third-order accretionary surfaces in element 6–LA resemble the "concave bench complexes" of Nanson and Page (1983). Lenticular-shaped sandy units in elements 6–LA and 7–LA are likely scroll bars that were embedded in the muddy sediments of swales. Location of outcrop is shown in figure 2. Lithofacies and architectural elements are defined in tables 1 and 2, respectively.

KANE SPRINGS 3 SAND BODY— LOW-SINUOSITY STREAMS

DESCRIPTION

The Kane Springs 3 sand body forms a resistant ledge throughout the study area. It typically is a single-story to multistory broad sheet that has width to height ratios of greater than 100 (Hazel, 1991, pl. 1). To the west of the study area, Blakey and Gubitosa (1984) showed that the sand body is intercalated with delta foresets and complex limestone and mudstone sequences of the Owl Rock Member. Thickness of the Kane Springs 3 sand body ranges from 5 to 20 m and averages 15 m. The base of the sand body is a well-defined sixth-order surface but is ambiguous where underlying sand bodies of

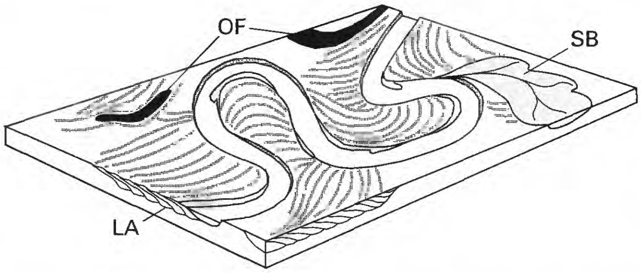


Figure 20. Suspended- to mixed-load meandering stream model inferred for the Kane Springs 2 sand body. The streams were subject to flooding and extensive sheetlike crevasse splays. Architectural elements are defined in table 2. Modified from models 6 and 7 of Miall (1985).

Figure 21. Foreset interbedded mudstone-sandstone facies (arrows) within the Kane Springs 2 sand body of the Chinle Formation interpreted by Gubitosa (1981) to be deltaic in origin. Location is across the Green River from Mineral Canyon.

the Kane Springs 2 sand body form similar stacked sequences. Where the overlying Black Ledge rests directly on the Kane Springs 3 sand body, the two are difficult to discern from one another. Detailed examination along the length of Long Canyon and Kane Springs Canyon reveals an extensive, undulatory sixth-order erosional surface that separates the two units. Sand-body interconnectedness decreases in the vicinity of the anticlinal crest (fig. 17).

Architectural elements and facies associations of the Kane Springs 3 sand body are similar to those in Kane Springs 1 and 2 sand bodies except that laterally extensive, horizontally laminated sand sheets, and downstream-accreting macroforms are also present in the Kane Springs 3 sand body (fig. 22, table 2). The multistory sheet is dominated by stacked channel and gravel-bar architectural elements and sand sheets. Discrete sandstone bodies either superimpose locally to form a thick multistory sequence or bifurcate into separate, thinner sheets of sandstone and interbedded mudstone (fig. 23). Complex-bounding fourth-order surfaces are flat and parallel or gently inclined and extend laterally as much as 100 m. Steeply dipping channel margins are rare. Both coarsening- and fining-upward sequences are present, but typically there is no systematic grain-size variation or change in facies sequence either within a single depositional element or within a complex (fig. 24). Flame structures are common in sand-sheet elements.

INTERPRETATION

The Kane Springs 3 sand body represents unconfined low-sinuosity streams that developed on the underlying fluvial-lacustrine sequence (Gubitosa, 1981; Blakey and Gubitosa, 1983, 1984; Hazel, 1991). The lateral profiles reveal some interesting architectural variations and suggest differences in depositional setting from the underlying sand bodies.

The architecture and internal bedform organization of the Kane Springs 3 sand body are most consistent with deposition on a extensive sandy braidplain subject to sheetfloods and highly flashy discharge (fig. 25). The deposits are similar to channel sandstone bodies in the Permian lower Beaufort Group (Stear, 1983). For the lower Beaufort Group, Stear detailed an ancient ephemeral stream-playa lake complex typified by channel patterns that were straight or highly sinuous or transitional between these two types. Similar to the lower Beaufort, in the Kane Springs 3 depositional environment channelization and subsequent confined stream floods were accompanied in most cases by extensive sheetflooding outside the channels. For example, the third-order bounding surfaces in element 1–LA shown in figure 23 climb and form the lowermost erosional bounding surface of the overlying element (2–LS). This relationship suggests that initial lateral migration of the channel was followed by unconfined sheet flow. The paucity of silt clasts above erosional surfaces indicates nonchannelized flow, and the lack of major channel features excludes a crevasse-splay origin.

The multistory sandstone sequences of the Kane Springs 3 sand body display features suggestive of seasonal flash flows. The preponderance of internal scour surfaces and intraformational conglomerate indicates fluctuating hydrodynamic conditions. Flame structures are thought to be a type of water-escape structure that results from rapid aggradation (Boggs, 1987). The predominance of facies Sh indicates that upper flow regime sheetflooding was common. Bar-top and scour-fill facies (Ss) are lacking; their presence has been related to discharge variations resulting in minor channels eroding preexisting sediments at low discharge flow (Abdullatif, 1989). The absence of this facies and other low-flow regime structures are suggestive of extremely episodic flash-flow events. The thick laminated sheet elements are likely the result of catastrophic sheetfloods of prolonged duration but could also have resulted from a simple series of

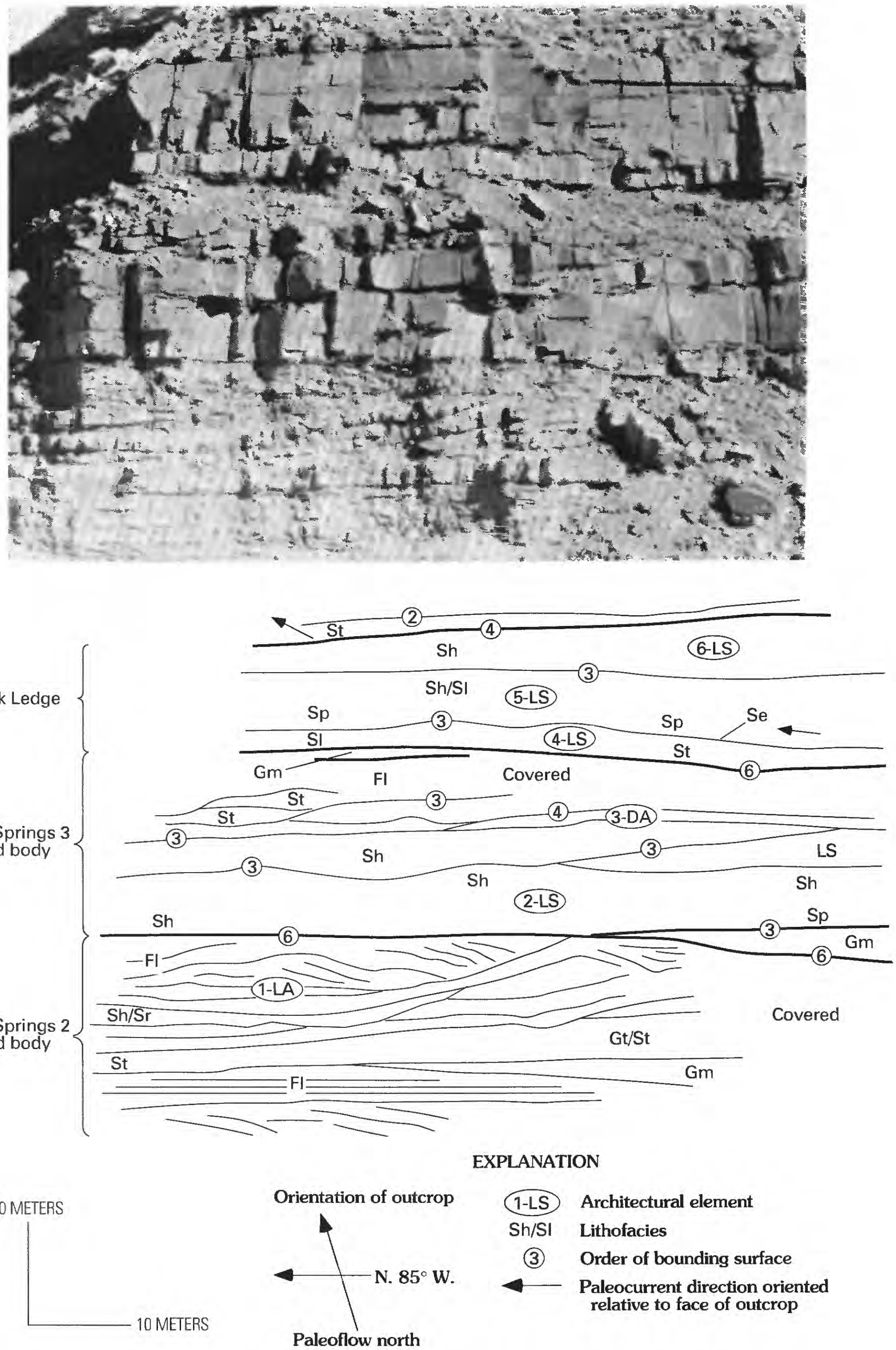


Figure 22. Multistory sandstone bodies in the Kane Springs 3 sand body and the Black Ledge, both of the Chinle Formation, in Hell Roaring Canyon. View looking northeast. The Kane Springs 3 sand body truncates heterolithic facies of the Kane Springs 2 sand body with an abrupt sixth-order surface and is separated from the overlying Black Ledge by a covered mudstone interval. Location of outcrop is shown in figure 2. Lithofacies and architectural elements are defined in tables 1 and 2, respectively.

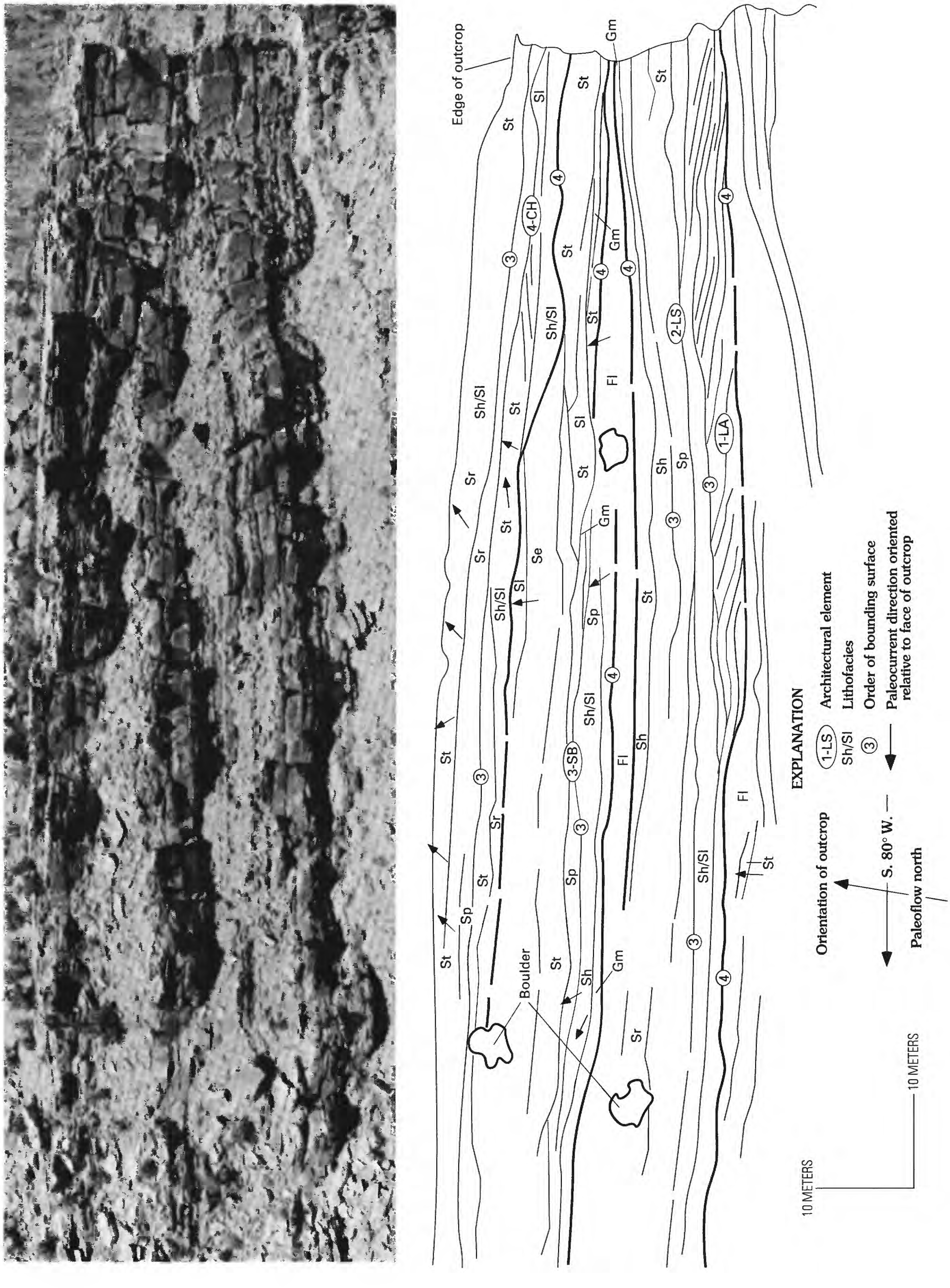

Edge of outcrop
Boulder
EXPLANATION
1-LS Architectural element
Sh/Sl Lithofacies
3 Order of bounding surface
Paleocurrent direction oriented relative to face of outcrop
Orientation of outcrop
S. 80° W.
Paleoflow north
10 METERS
10 METERS
4-CH
2-LS
1-LA
3-SB
St
Sl
Sh/Sl
Sr
Sp
Se
Sh
Gm
Fl

Figure 23 (facing page). Lateral bifurcation of the tabular sheet geometry of the Kane Springs 3 sand body of the Chinle Formation laterally into thin sheets separated by interbedded mudstone. The outcrop in Long Canyon is 20 m thick and 70 m long and is oriented almost perpendicular to paleoflow. Third-order bounding surfaces in element 1–LA climb and form the lowermost erosional bounding surface of the overlying element (2–LS). Initial lateral migration of the channel probably was followed by successive unconfined sheet flows. Location of outcrop is shown in figure 2. Lithofacies and architectural elements are defined in tables 1 and 2, respectively.

aggradational events and nonerosional periods. The lack of crosslaminated sequences above horizontal lamination (facies Sh) suggests rapidly waning flow at the end of the flood event. The laterally extensive and internally consistent sheet sand bodies are similar to units described from high-energy flood environments (McKee and others, 1967; Tunbridge, 1981, 1984; Stear, 1983).

BLACK LEDGE—LOW-SINUOSITY STREAMS AND SHEETFLOODS

DESCRIPTION

The Black Ledge is a narrow complex multistory ribbon or sheet sand body that is present throughout the study area. It thins or is absent over the Cane Creek anticline (Hazel, 1991, pl. 1) (fig. 26). The thickness of the Black Ledge ranges from 5 to 20 m and averages 10 m. The typical exposure is a blackish-maroon, sheet sand body that truncates and erosively overlies the Kane Springs 3 sand body along an undulose to planar sixth-order surface (fig. 26). The complex multistory ribbon geometry is best displayed in the western part of the study area in Mineral and Hell Roaring Canyons (fig. 27).

The major architectural elements composing the Black Ledge are laminated sand sheets and gravel and sandy bedforms (fig. 22, table 2). Lateral-accretion surfaces and downstream-accreting macroforms are rare. Individual sandstone bodies are laterally extensive (50–100 m) and lack interbedded finer facies. Where sheet sandstone bodies are superimposed, they typically are offlapping elements that taper against and truncate each other with low-relief channelized margins (fig. 28). Intraclast-lined third-order erosional surfaces separate each element; they extend for tens of meters and are undulatory to flat. Intraclast horizons (facies Se) between elements are as thick as 1.5 m and contain carbonaceous fragments and whole logs.

Internally, the laminated sand sheets and sandy bedforms are characterized by few internal breaks and little or no evidence of fining-upward cycles. In vertical section, it is difficult to distinguish between low-angle facies St and Sh facies. The typical facies sequence is a basal lag scour associated with facies Se and an upward transition to Sh/Sl facies that may or may not grade into facies Sr. Intercalated with this general sequence are minor conglomeratic Sp/St facies. Stewart and others (1972) described the composition

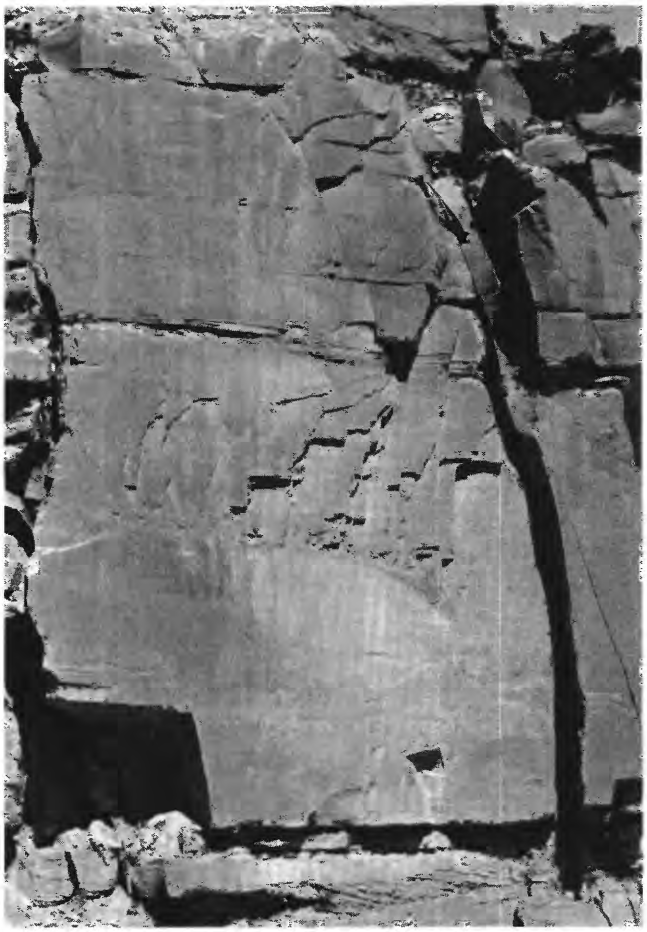


Figure 24. Laminated sand sheet element from the Kane Springs 3 sand body of the Chinle Formation. Uniform appearance of medium-grained sandstone and lack of second-order erosional surfaces suggest continuous supply of sediment. Hammer at base (arrow) is shown for scale. Location of outcrop is shown in figure 2.

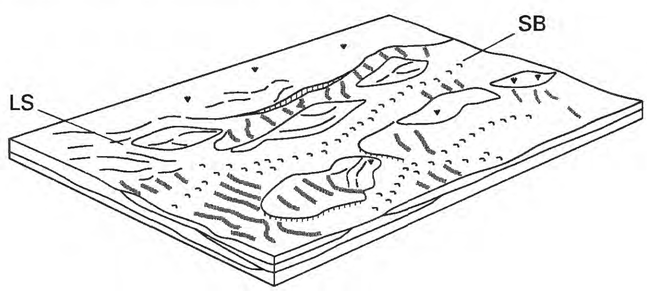


Figure 25. Distal braidplain model inferred for the Kane Springs 3 sand body of the Chinle Formation. The fluvial plain was subject to extensive sheetfloods as a result of highly flashy discharge. Architectural elements are defined in table 2. Modified from models 10 and 11 of Miall (1985).

Figure 26. Eastern view of Kane Springs Canyon near the Cane Creek anticlinal axis. Note the erosional and (or) depositional thinning of both the Kane Springs 3 sand body (KS–3) (lower arrow) and the Black Ledge (BL) (upper arrow), both of the Chinle Formation, along sixth-order surfaces. These unit-bounding surfaces are intraformational unconformities. Salt movement occurred at more than one interval prior to deposition of the Black Ledge. Rocks of the Triassic Moenkopi Formation (Ŧm), Jurassic Wingate Sandstone (Jw), and Kane Springs 1 and 2 sand bodies (KS–1, KS–2) are also shown.

Figure 27. Complex multistory ribbon geometry of the Black Ledge of the Chinle Formation in Mineral Canyon. The ribbon is approximately 9 m thick (between arrows). Location of outcrop is shown in figure 2.

of the Black Ledge as mostly quartzarenite; however, in the study area, calcirudite and calcarenite alternate with equal amounts of arenite-rich sandstone.

INTERPRETATION

Dominantly unidirectional paleocurrents, sedimentary structures, and low between-sequence paleocurrent variance in the Black Ledge suggest deposition by rapidly shifting, low-sinuosity sand and gravel bed rivers subject to widespread sheetfloods (fig. 29). The architectural style suggests a braidplain characterized by ephemeral runoff due to seasonal flow and is likened to the Si facies assemblage of Rust (1978).

Individual sheet sand bodies composing the multistory sequence probably formed in low-sinuosity channel networks that were incised during high-stage flows and then either were abandoned (mud draped) and later filled or were

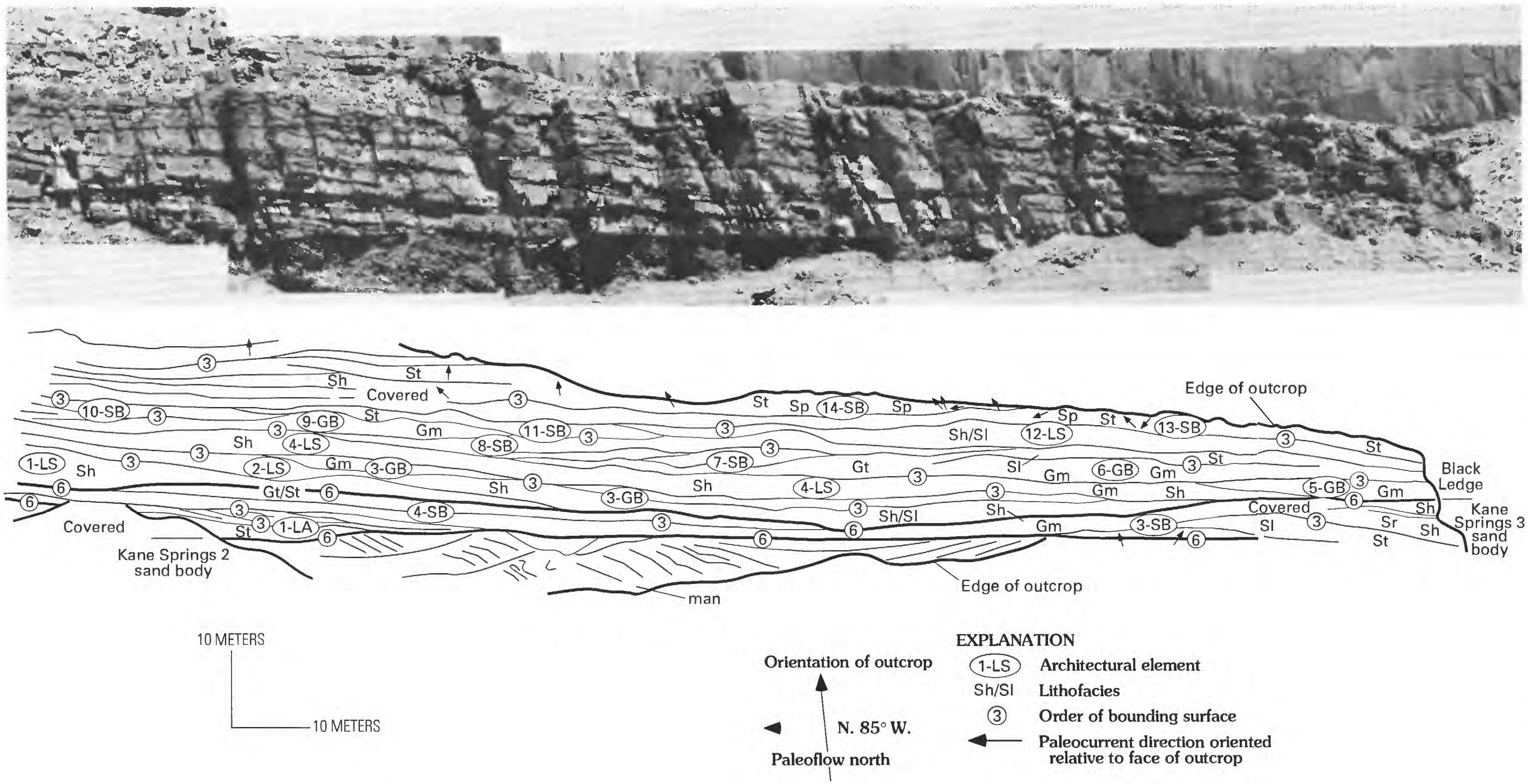


Figure 28. An exceptionally thick sequence of the Black Ledge incising into the Kane Springs 3 sand body, Long Canyon, forming a 30-meter-thick superimposed package. A laterally extensive sixth-order surface separates the two and can be traced up Long Canyon for almost a kilometer. Because the lateral profile is oriented almost perpendicular to paleoflow, channel outlines are accentuated. The complex multistory sheet is composed of intersecting tabular conglomerate and sandstone bodies constructed by CH, LS, and GB architectural elements. Location of outcrop is shown in figure 2. Lithofacies and architectural elements are defined in tables 1 and 2, respectively.

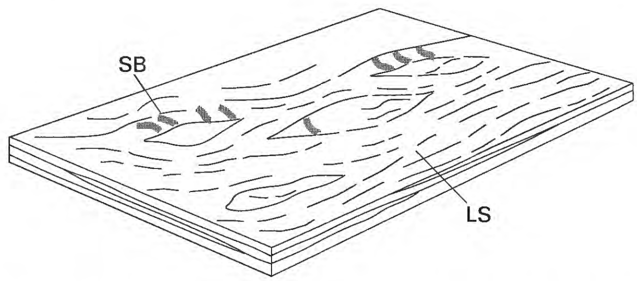


Figure 29. Inferred model for deposition of the Black Ledge of the Chinle Formation. An extensive braidplain is dominated by sheetfloods and highly flashy discharge. Architectural elements are defined in table 2. Modified from models 11 and 12 of Miall (1985).

filled during the same event by vertical accretion of plane bed sand (Tunbridge, 1981, 1984; Rust, 1984; Lawrence and Williams, 1987). The abundance of facies Se is indicative of extensive reworking and redeposition, within a short distance, of suspension-deposited muds during subsequent flooding (Karcz, 1969; Rust, 1984). Zones of large intraclasts are likely the result of active channel migration by bank collapse. Gibling and Rust (1984) attributed similar horizons in the Pennsylvanian Morien Group in Nova Scotia to deposition and subsequent reworking of cohesive fines due to alternating periods of flooding and subaerial exposure on high topographic levels of a braided and ephemeral river system. Because low-sinuosity streams tend to accrete vertically, they tend to avulse often and incise new channels in adjacent areas (Tunbridge, 1984). The constantly shifting channels erode adjacent overbank deposits and produce a broad sheet of crosscutting sand bodies. Sand-body dimensions in the Black Ledge indicate that the channels ranged from 0.5 to 5 m in depth and from 5 to 100 m in width. The laminated sand sheets were deposited as bank-to-bank sheets and as bar forms analogous to the plane-bedded simple bars of Allen (1983). This type of bar lacks a slipface. The predominance of facies Sh in the laminated sand sheets is attributed to deposition under extremely high flow velocities during short-lived, high-energy floods (McKee and others, 1967; Tunbridge, 1981; Sneh, 1983). Similar orientations in trough and planar cross-sets (fig. 14) suggest formation by downstream accretion of mid-channel bars (Lawrence and Williams, 1987) during periods of sustained flow. Indications of waning flow, such as sequences capped by facies Sp, St, or Sr, are rare.

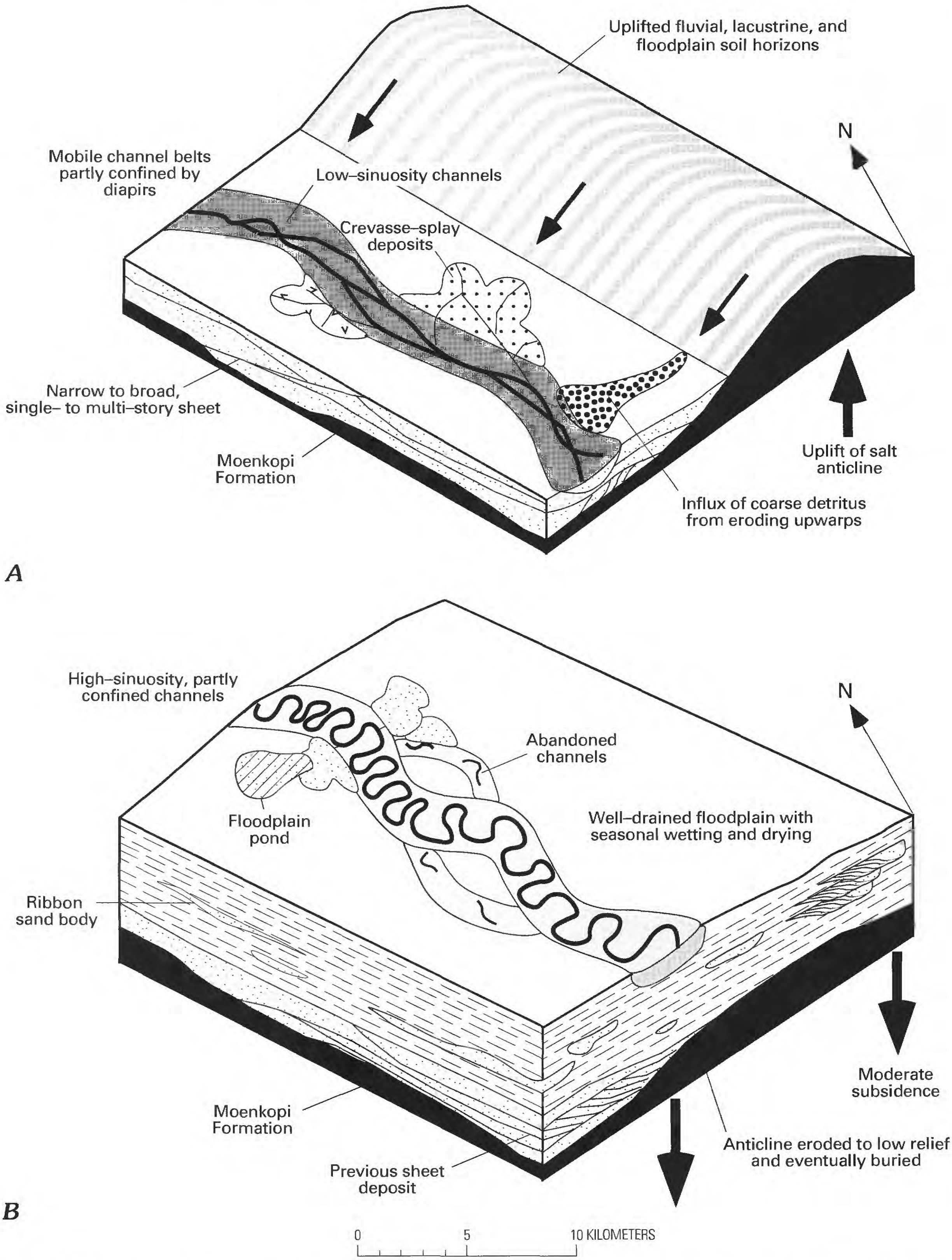
Possible modern analogs, such as the ephemeral terminal River Gash in Kassala, Sudan (Abdullatif, 1989), are present in semi-arid piedmont environments (Williams, 1970; Rust and Koster, 1984). Channelized streamfloods produced by torrential rainfall may, with decreasing slope and waning flow, pass downslope into unconfined sheetfloods (Tunbridge, 1981, 1984).

CONTROLS ON DEPOSITION

Tectonic control on alluvial architecture is common in many basin settings. In this discussion, the importance of intrabasinal tectonism in construction of Chinle stratigraphy is evaluated against expected architectural changes that in other studies have been attributed to extrabasinal (allocyclic) control. Basin subsidence (an allocyclic mechanism) has been suggested to have strongly influenced sediment accumulation rates (Blakey and Gubitosa, 1984), base-level changes (Kraus and Middleton, 1987), stream sinuosity (Deacon and Middleton, 1989), and, thus, alluvial architecture in the Chinle basin. It is difficult to separate the effects of processes operative across the entire Chinle basin from those of localized salt diapirism. Architectural patterns in the Salt Anticline region, however, contrast dramatically with sand-body geometry and architecture of time-correlative units elsewhere in the Chinle basin. Cyclic depositional sequences in the study area suggest that floodplain adjustment in this part of the basin was unrelated to allocyclic control and instead reflects a sedimentary response to intrabasinal salt tectonism (Hazel, 1991; Hazel and Blakey, 1992).

The study area contains a part of the Chinle alluvial plain that reflects the cumulative effects of several superimposed fluvial systems. The repetition of several coarse-grained progradational fluvial systems may indicate cyclic sedimentation common to an alluvial plain (Beerbower, 1964). Facies architecture analysis of each cycle (Kane Springs 1, 2, and 3 and Black Ledge) provides insight to the controlling mechanism of cyclicity. Hypothetical controls on the geometric distribution of lithofacies and sand-body interconnectedness are both allocyclic and autocyclic (Beerbower, 1964). Cyclic sedimentation may result from allocyclic controls such as changes in base level, gradient, discharge, and types of sediment load, controls that are directly related to such factors as climate, tectonism, and eustacy. Autocyclic controls, in contrast, result in redistribution of energy and sediment input within the depositional basin and require no net change in the total energy regime (Beerbower, 1964). These controls include channel migration and filling, avulsion, bar migration, crevassing, aban-

Figure 30 (facing page). Architectural styles in the Chinle Formation of the Paradox Basin. The contrasting styles of alluvial architecture resulted from different rates of basin subsidence and sedimentation accumulation. Rates of basin subsidence were altered in the Salt Anticline region by salt tectonism. *A*, Initial uplift of nearby anticlines, including the Cane Creek anticline, was followed by stability, erosion, and slow or negligible subsidence relative to regional subsidence rates. The narrow to broad, single-story to multistory Kane Springs 1 and 3 sand bodies and the Black Ledge sheet sandstone body were the result. The high sandstone to mudstone ratios resulted from increased



frequency of avulsion. Variations in sand-body story thickness and vertical stacking resulted from localized, differential subsidence in intervening synclinal areas. *B*, The Kane Springs 2 sand body and sand bodies in the fine-grained part of the Church Rock Member have ribbon to narrow sheet geometries that resulted when the area subsided at moderate regional subsidence rates. Frequency of avulsion and rates of subsidence were low to intermediate, and the resulting sandstone to mudstone ratio is low. Model loosely based on a model proposed by Blakey and Gubitosa (1984) for the Kane Springs strata.

donment, and erosion. The cause and effect of autocyclic controls have been discussed extensively by Allen (1978), Leeder (1978), and Bridge and Leeder (1979).

In order to understand and reconstruct the temporal and spatial evolution of Chinle fluvial systems across the Salt Anticline region, the following controls must be considered: salt tectonism, basin subsidence rate, avulsion frequency, sediment supply, and climate. Vertical superposition of the Kane Springs and Black Ledge sand bodies, each with different characteristic fluvial environments, allows potential controls to be tested against more than one depositional setting.

AUTOCYCLIC CONTROLS

CYCLIC SEDIMENTATION AND GEOMETRY

The fluvial succession described herein consists of small- to large-scale cycles that either fine upward or lack vertical grain-size trends. These cycles, represented by the Kane Springs 1, 2, and 3 sand bodies and the Black Ledge, are characterized by erosive bases that constitute fifth- and sixth-order bounding surfaces of limited regional extent (fig. 26). Each sand body is characterized by narrow to broad sheet geometry. The sheet sand bodies, however, are the products of different fluvial styles. Repeated vertical changes in fluvial paleoenvironment suggest that the basin slope was not constant during deposition.

The cycles probably are the result of successive migration of fluvial channels through the same area of the floodplain (autocyclic mechanism) (fig. 30). The sandbodies are a result of the influx of coarse detritus, in direct contrast to regional sediment supply, which diminished from Petrified Forest to Owl Rock time as evidenced by a decrease in sand and an increase in limestone in the Owl Rock Member (Blakey and Gubitosa, 1984). Because regional changes in sediment supply did not affect the Chinle systems proximal to the anticline area, other reasons for sediment influx must be considered.

Cyclicity in the study area was due to anticline development, which provided a local sediment source and slowed subsidence rates (fig. 30); elsewhere in the Chinle basin subsidence rates were moderately rapid (Blakey and Gubitosa, 1984). The narrow to broad sheet geometries and high sandstone to mudstone ratios north and northeast of the Moab-pinchout resulted from small, localized changes in frequency of avulsion versus rate of subsidence. Rapid anticlinal uplift produced local high-relief slopes. Monsoon-induced torrential rainfall resulted in high local rates of erosion and fluvial-deltaic sedimentation.

LOCALIZATION OF AXIAL DRAINAGE

The most obvious influence of salt tectonics on sediment dispersal patterns is the alignment of stream courses by the northwest-trending Cane Creek and Moab anticlines. Regional studies of Chinle fluvial systems show that flow was to the west and southwest off the Uncompahgre highland in western Colorado (Stewart and others, 1972; Blodgett, 1984). In contrast, paleocurrent indicators in the study area show a marked unidirectional trend toward the north-northwest (fig. 14) and exhibit low variance between sequences and individual depositional systems. Thus, fluvial systems most likely were deflected by anticline development and forced to flow northwest through the Salt Anticline region before draining into the northern part of the basin (fig. 3).

Channel belts tend to be concentrated in areas of greatest subsidence. In these areas, sand-body interconnectedness increases and sandstone to mudstone ratios are greater (Bridge and Leeder, 1979). For example, the multistory Black Ledge sequence thickens toward the Kings Bottom syncline (fig. 28). Conversely, all four sand bodies display a reduced interconnectedness and complex interfingering with mudstone over the Cane Creek anticlinal crest (figs. 17, 26). The increase in channel density and orientation of depositional axes is attributed to uplift of the Cane Creek and Moab salt anticlines and salt withdrawal related subsidence of the Kings Bottom syncline (fig. 30). In a similar setting, Tyler and Ethridge (1983) showed that geometry and orientation of the Slick Rock fluvial axis in the Salt Wash Member of the Upper Jurassic Morrison Formation was localized by the rising Gypsum Valley salt ridge.

DEVELOPMENT OF ASYMMETRIC MEANDERBELTS

There is evidence in the study area for autocyclic controls on stream behavior such as topographically triggered avulsion due to tectonic tilting, bedform-bar migration, and channel migration and filling. The preservation of predominantly northeast dipping third-order lateral-accretion surfaces in the heterolithic facies of the Kane Springs 2 sand body is likely the result of asymmetric channel-belt development and the resulting preferential abandonment of meanderloops. This preferred dip is especially obvious in Kane Springs and Long Canyons (fig. 19). Combined with paleocurrent values (fig. 14), such a skewed distribution is unlikely; a freely meandering river should show a more uniform distribution of dip directions around the regional paleoflow trend (Miall, 1985).

Alexander and others (1985) and Alexander and Leeder (1987) showed that on an alluvial plain undergoing active tectonic tilting in a direction perpendicular to regional paleoflow the channel belt may slowly migrate laterally by preferential downslope erosion and meander cutoff. Thus, meanders on the updip side have a higher preservation potential and produce a predominance of lateral-accretion surfaces facing up the paleoslope. The marked meanderbelt asymmetry in the heterolithic facies of the Kane Springs 2 sand body is suggestive of floodplain tilting due to uplift of the Moab salt anticline and salt-withdrawal subsidence between the Moab and Cane Creek anticlines (Kings Bottom syncline). Although it is likely that the system migrated down the tectonic tilt toward the southwest, the complex stacking of the preferentially abandoned meanderloops (fig. 19) is suggestive of confinement between the two anticlines.

Channel-fill deposits reflect the nature of the stream diversion process (Hopkins, 1985). Topographically triggered avulsion occurs when a river aggrades vertically until topography is increased such that the channel improves its profile by switching positions down a more suitable path (Alexander and Leeder, 1987). The rate of topographic growth in the study area may have been enhanced by preferential flooding of the downtilted side of the floodplain as the Kane Springs and Black Ledge channels moved toward the position of maximum subsidence; however, channel migration and filling could also have been a function of other parameters that control avulsion.

ALLOCYCLIC CONTROLS

Extrabasinal controls such as tectonic-induced subsidence, climate, and source terrane behavior are not likely controls given the depositional patterns in the study area; however, knowledge of these allocyclic controls on alluvial architecture is critical to a meaningful evaluation of Chinle architecture. The effects of eustacy on base-level changes were not examined in this study. The Chinle depositional system was in a partly landlocked basin, and the effects of sea-level changes are difficult to correlate without precise age constraints.

The Kane Springs units are time correlative to the Petrified Forest and Owl Rock Members, and one would expect similar cyclic thick floodbasin and soil deposits and widely scattered ribbon sand bodies in these units. Blakey and Gubitosa (1984) reasoned that preservation of these members involved a moderately rapid rate of basin subsidence relative to rate of sedimentation; however, Chinle strata in the study area do not contain well-developed paleosols. At best, only color mottling and minor nodule formation, indicative of a floodbasin environment experiencing periodic wetting and drying (Bown and Kraus, 1981), is present proximal to the Cane Creek anticline. The predominance of pedogenic nodules as clasts in Kane Springs strata and Black Ledge conglomerates indicates that soil-forming processes were operating nearby because such clasts do not survive long in a dynamic system (Karcz, 1969). It is reasonable to conclude that decreased rates of basin subsidence as a result of differential subsidence and uplift related to salt diapirism led to sheet sand-body development; however, because there was no constancy of river type or rate of avulsion, it is difficult to test the importance of subsidence rate on sandstone geometry in accordance to the models of Allen (1978) or Bridge and Leeder (1979).

Uplift of the ancestral Uncompahgre source area, providing pulses of coarse sediment (allocyclic mechanism), is not favored as a control on the depositional cycles. This mechanism would reflect progradation of an entire fluvial system and would be characterized by symmetrical cycles that coarsen and then fine upward (see, for example, Steel and others, 1977; Steel and Aasheim, 1978). The Kane Springs strata clearly do not represent basinwide facies variations and do not contain extrabasinal-derived detritus coarser than medium-grained quartz sand. There is no indication that size availability played any role in differing alluvial architectures. Source-rock control was locally important during deposition of the basal mottled strata when initial uplift of upper Paleozoic chert-bearing limestone provided an abundance of chert pebbles.

Variations in climate can cause changes in sediment grain size and in depositional style as a result of fluctuating discharge and weathering modes (Miall, 1980). The three Kane Springs sand bodies and the Black Ledge were deposited in fluvial systems subject to episodic flash floods. For example, laminated sheet elements indicate that flows traveled as sheetfloods on the order of tens of kilometers. The abundance of wood, carbonaceous horizons, and terrestrial fossils throughout the sequence suggests that water was abundant. The evidence for fluctuating water tables and the repeated occurrence of desiccation cracks suggest that the environment was episodically wet and the rainfall possibly seasonal. Clastic-carbonate couplets in Owl Rock laminated limestone have been attributed to seasonal influx of clastic sediment (Dubiel, 1989). These observations match the interpretation of Dubiel and others (1991) in which the Chinle climate in southeast Utah was tropical-monsoonal. The abundance of eolian sand-sheet deposits at the top of the Chinle suggests that the climate became drier at the close of Chinle deposition (Blakey and Gubitosa, 1984; Dubiel, 1989; Hazel, 1991), but the color mottling in the sand sheets is indicative of at least some seasonal flooding (Kocurek and Nielson, 1986).

Blakey and Gubitosa (1984) ruled out climate as a major factor in controlling deposition because the sheet-ribbon-sheet geometry in the Kane Springs strata indicates that controls on sand-body geometry changed at least twice during aggradation of the system. They noted that this large-scale cyclic change is not evident in immediately adjacent units of the Petrified Forest and Owl Rock Members. Dubiel (1989) suggested that long-term climatic fluctuations are represented by large-scale interbedding of siltstone and limestone in Owl Rock strata; however, purely seasonal climatic controls would tend to produce coarser, more immature grains in arid cycles and finer, more mature grains in humid cycles. This was not observed in the Kane Springs strata and Black Ledge. The sand bodies contain consistent indication of seasonal climate subject to flash-flood events throughout Chinle time, thus arguing for controls other than climatic variations.

CONCLUSIONS

Alluvial architecture of the Chinle Formation in the study area in the Paradox Basin of southeastern Utah was modified through time in response to salt diapirism rather than to changes in basinwide subsidence rates, climate, base-level, or sediment supply. Evidence presented herein shows that these important allocyclic controls were of secondary importance in generating the different styles of architecture in the Salt Anticline region. The Chinle sequence in the study area is dominated by broad multistory sheet sand bodies that are unique to this region. Channel morphology changed significantly within only one interval (single-story or weakly multistory ribbon and narrow sheet sand bodies in the Kane Springs 2 sand body), and the other units, although the products of different fluvial styles, have similar sand-body geometry and degree of interconnectedness. Only near the end of Chinle deposition is there evidence to suggest the development of vastly different climate or new source areas. Sheet development resulted from decreased rates of localized basin subsidence due to intrabasinal uplift of salt structures. Intrabasinal tectonic activity altered stream gradients and subsidence rates, which in turned governed sinuosity, flow regime, energy levels, and sediment distribution. During quiescence of the salt structures, moderately rapid basinwide subsidence produced ribbon sand bodies enveloped by overbank deposits.

Evidence supporting intrabasinal tectonism as the principal control on Kane Springs and Black Ledge depositional systems includes the following.

1. Cycles of fluvial sand-body sheet geometry are separated by extensive fifth- and sixth-order erosional bounding surfaces, some of which represent intraformational unconformities.

2. Abrupt changes in fluvial style are indicated by significant facies changes both within and between sand bodies. Highly variable lithologies resulted from changes in depositional environments in areas where salt movement occurred.

3. Fluvial sand bodies thin, display less interconnectedness, or are absent over the Cane Creek anticline. Post-depositional erosion between aggradational events was most likely caused by localized diapiric uplift of the Cane Creek and Moab salt anticlines.

4. The depositional axis of each fluvial sand body was controlled by gentle surface flexing between salt anticlines. Paleocurrents in all the fluvial units are remarkably consistent with one another and parallel the trend of the salt anticlines.

5. Variations in fluvial sand-body story thickness are a result of differential subsidence in intervening synclinal areas.

6. Heterolithic facies in the Kane Springs 2 sand body provide evidence for topographically triggered avulsion via tectonic tilting. The preferential preservation of bounding surfaces and the unique channel fill are attributed to tilting of the depositional slope.

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