THE CALICO BED,
UPPER CRETAEOUS, SOUTHERN UTAH:
A FLUVIAL SHEET DEPOSIT IN THE
WESTERN INTERIOR FORELAND BASIN
AND ITS RELATIONSHIP TO EUSTASY AND TECTONICS

by
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The Calico bed, Upper Cretaceous southern Utah: a fluvial sheet deposit in the Western Interior Foreland Basin and its relationship to eustasy and tectonics.
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The Calico bed of the Straight Cliffs Formation is a coarse-grained fluvial deposit that spread across the Western Interior Foreland Basin in southwestern Utah during Late Turonian - Early Coniacian time. The sheet-like Calico is composed of coarse-grained sandstone and sandy conglomerate and interfingers with underlying alluvial plain deposits. The composition of the sandstones and conglomerates and paleocurrents to the northeast indicate that two structurally different thrust belts, the Sevier Orogenic Belt and the Mogollon Highlands contributed sediment to the Calico bed.

The Calico bed consists of multilateral, multistoried sheet-sandbodies. An upward decrease in grain size, different suites of sedimentary structures, and different alluvial architectures in the three lithofacies of the Calico bed record rising base level during deposition, the end of which is marked by a marine transgressive lag in the eastern study area; in the west the Calico interfingers with overlying fluvial strata of the John Henry Member of the Straight Cliffs Formation. Rising base level was probably controlled by regionally rising sea level in the Cretaceous seaway during latest Turonian-Coniacian time.
The Calico bed was deposited by braided then meandering rivers and forms a broad sheet of channel deposits, devoid of overbank fines, across the Kaiparowits Plateau. This architecture and the presence of coarse conglomerate in the braided river deposits at the base of the Calico bed indicate the predominant influence of tectonics. Decreased rates of subsidence, and possibly basinal rebound, during a period of post-thrusting quiescence in the hinterland, resulted in a greater supply of coarse-grained sediment to the distal part of the foreland basin and reworking of fluvial deposits as sediment bypassed to the northeast.

The Calico bed can be considered in the context of sequence stratigraphy. The base of the Calico bed is the type 2 boundary of a sequence that developed in response to foreland basin tectonics. The lower part of the Calico bed represents the lowstand systems tract of this tectonically controlled sequence. The upper part of the Calico bed represents the transgressive systems tract of a larger, eustatically controlled sequence, which began in the late Middle Turonian before deposition of the Calico bed.
This work is dedicated to my mother, Mollie Cook, whose zeal for learning has inspired me. Truly, the completion of this thesis would not have been possible without the time she gave generously and with love to care for my children.
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CHAPTER I

INTRODUCTION

The Calico bed of the Smoky Hollow Member of the Straight Cliffs Formation was first described by Peterson (1969a, 1969b) in the southern region of the Kaiparowits Plateau (Figure 1). It is a fluvial, pebbly sandstone unit at the top of the Smoky Hollow Member and was deposited sometime between the late Middle Turonian and Early Coniacian. Although similar lithologies exist lower in the Smoky Hollow Member, the Calico bed was identified as a separate unit based on its unique white coloration, generally coarser texture, and sheetlike geometry (Peterson, 1969a).

The Calico bed is unusual and intriguing because it represents coarse-grained deposition far from the thrust front in a foreland basin. Deposition of the Calico bed has been interpreted as the result of orogenic pulses in the source area (Peterson and Kirk, 1977). Peterson and Kirk correlated the Calico bed with the upper sandstone member of the Toreva Formation in Black Mesa, Arizona (Figure 1) and concluded that both are remnants of a widespread fluvial sandstone that was an integral part of a major shoreline regression. They inferred that deposition of the orogenic, fluvial pebbly-sandstone beds
Figure 1. Index map showing major Cretaceous outcrops in parts of Utah and Arizona. The study areas of this investigation and of Peterson (1969a) are identified in the Kaiparowits Plateau (modified from Fisher, Erdmann, and Reeside, 1960, Figure 1).
coincided with the time of greatest uplift and erosion in the source region, transportation of sediment from it, and maximum retreat of the shoreline.

Since the work of Peterson and others in the late 1960's and mid-1970's, new concepts have been developed in stratigraphy, basin analysis, and alluvial architecture which are directly applicable to the interpretation of the Calico bed. Sequence stratigraphy (Vail et al., 1977; Posamentier et al., 1988; Van Wagoner, et al., 1988) is particularly appropriate to the problem, as deposition of the Calico bed was undoubtedly related to the migrating western shoreline of the Cretaceous Interior Seaway.

New ideas about foreland basin subsidence due to loading of thrust sheets and the related timing of progradation can also be applied to the Calico bed; its two source areas were the Sevier fold-thrust belt and a thrusted area known as the Mogollon Highlands. Several authors (Beck and Vondra, 1985; Heller et al., 1988; Blair and Bilodeau, 1988) have proposed that active thrusting in a source area is recorded in its foreland basin by rapid subsidence and the deposition of fine-grained sediment. When a subsequent period of tectonic quiescence occurs and the rate of basin subsidence slows, then the rate of sediment supply may equal or surpass the rate of subsidence and a coarse-grained progradation occurs. This hypothesis is the inverse of traditional views which hold that tectonic activity in the source
area is directly correlable to progradation, as proposed by Peterson and Kirk (1977) for the Calico bed.

In addition, much new progress has been made in the analysis of fluvial sediments. Models of fluvial sedimentation have become more sophisticated in the interpretation of paleochannel pattern. The method of architectural-element analysis has enabled geologists to focus on interpreting fluvial processes and sub-environments by considering the individual lithosomes in a fluvial unit (Miall, 1985, 1988). By considering the large-scale fluvial architecture (Allen, 1978, Bridge and Leeder, 1979) of a depositional sequence, the allocyclic factors which controlled its deposition may be interpreted.


The purpose of this study is to reexamine the Calico bed regionally and in greater detail and to reassess its depositional history in terms of the advances in sequence stratigraphy, basin analysis, alluvial architecture analysis, and updated regional stratigraphy. A second goal is to determine the regional significance of the Calico bed as it relates to source-area and basin tectonics and to sea-level changes.
Towards these goals the following approaches are taken. Major problems exist in the definition of the Calico bed and its identification as a single stratigraphic unit. The nature of the basal and upper contacts are discussed, as are the sedimentological criteria by which the unit is identified. Sandstone petrography was analyzed and is discussed as it pertains to stratigraphic problems. The fluvial environments and sub-environments of the Calico bed and bounding strata are interpreted, and temporal and spatial trends of environmental setting within the study area are discussed. Finally, these lines of evidence are synthesized and placed in a regional perspective, which is vital to interpretations involving sea level and tectonics.

Location

The Kaiparowits Plateau lies in the southwestern part of the Colorado Plateau Province in Kane and Garfield Counties, Utah (Figure 1). Most of the plateau has an irregular outline, but the northeast side is a nearly straight erosional escarpment named the Straight Cliffs. This escarpment is 1,000-2,000 feet high and extends 50 miles southeast from Escalante, Utah. The study area lies in the northern Kaiparowits Plateau (Figure 1). Twenty-six stratigraphic sections were measured and described in the study area. Two stratigraphic sections were measured and described in the southern Kaiparowits Plateau for purposes of comparison.
to Peterson's (1969a) stratigraphic framework (Figure 2). For purposes of regional correlation, two stratigraphic sections were measured and described in the Markagunt Plateau (Figure 2) and the temporally equivalent Ferron Sandstone Member of the Mancos Shale was examined in the Henry Mountains area (Figure 1).

**Geological Setting**

Deposition of the Calico bed was related to 1) the tectonic zones active during Cretaceous time (i.e. the Sevier Orogenic Belt in western Utah and eastern Nevada, the Mogollon Highlands in south central Arizona, and their associated foreland basins) and 2) the eustatic fluctuations of the Cretaceous Western Interior Seaway.

**Structural Setting**

During the Cretaceous, an orogenic belt stretched from northern Canada to southern Mexico (e.g. Armstrong and Oriel, 1965; Price and Mountjoy, 1970; Jordan, 1981; Wiltschko and Dorr, 1983). The Sevier Orogenic Belt is the segment that stretched from southern Idaho, across eastern Nevada and western Utah, to southern Nevada (Armstrong, 1968). The neighboring segment to the south is the Mogollon Highlands (Harshbarger et al., 1957) (Figure 3).
Figure 2. Map of the study area, showing locations of stratigraphic cross-sections presented in Plates 1, 2, 3, and 4. Twenty seven measured sections comprise the cross-sections. Sections TC and SM were examined in Peterson's (1969) study area and sections NG and NFV were examined in the Markagunt Plateau (modified from Doelling and Graham, 1972).
Figure 3. The tectonic zones active during Cretaceous time that affected deposition of the Calico bed: the Sevier Orogenic Belt and the Mogollon Highlands.
Sevier Orogenic Belt. This structural feature consists of a belt of north-northeast-trending, east- and southeast-verging low-angle thrust plates and folds extending to the western edge of the Colorado Plateau. It was a foreland thrust belt that developed behind a plutonic-volcanic arc of Andean type. As the North American and Farallon plates converged, underthrusting of the Farallon oceanic plate caused east-directed intracontinental thrusting behind the arc along the Sevier Orogenic Belt (Coney, 1972; Dickinson, 1974).

The location of the thrusts are believed to be controlled by a paleogeographic and paleotectonic element, referred to as the hinge line or Wasatch Line (Coney, 1972; Burchfiel and Davis, 1972; Stokes and Heylmun, 1963). This was a site of possible Precambrian rifting (Stewart, 1972) and has marked a fundamental tectonic boundary or flexure between areas of miogeosynclinal and stable platform or cratonic sedimentation throughout Phanerozoic time. Thrust faults bring thick, basinal, miogeosynclinal Paleozoic marine deposits over thinner, marginal platform or platform deposits.

Mogollon Highlands. Whereas much is known about the Sevier Orogenic Belt, which formed the western margin of the Cretaceous Western Interior Basin, very little is known about the southwestern margin of the basin or the area lying in south-central Arizona.
In southern Nevada, the Sevier Orogenic Belt departs from its characteristic north-northeast strike and trends southeast across southwestern Arizona (Figure 3). Major changes in tectonic style occur within this region. Thrusts of the Sevier Orogenic Belt in Utah and western Nevada are regarded as decollement type thrust faults, whereas in southern Nevada and southwestern Arizona thrusts involve Precambrian crystalline and Jurassic plutonic rocks (Burchfiel and Davis, 1972; Hayes, 1970) and are thus Laramide type thrust faults. This major change in thrusting style is coincident with the intersection of the Mesozoic plutonic-volcanic arc with the Colorado lineament, transcontinental arch, and Walker-Texas lineaments.

The causes and timing of deformation along the northwest-southeast-trending Mogollon Highlands are poorly understood. A period of rifting is proposed to have occurred from the Jurassic through latest Early Cretaceous (Bilodeau, 1982, 1986; Dickinson et al., 1986). The tectonic style then changed to thrusting. Burchfiel and Davis (1982) suggested thrusting began in southern Nevada and western Arizona during late Early Cretaceous (Albian) time, whereas Hayes (1970), Nydegger (1982) and Drewes (1982) believed thrusting began in southeastern Arizona in Late Cretaceous time (between 90 and 75 ma). Coney (1972) proposed that the relative motion of the North American plate switched from west-northwest to west-southwest, thereby radically switching an extensional tectonic region to a compressional region. The timing is poorly
understood, but Coney suggested an early Late Cretaceous switch. This agrees with the most recent dates on these thrusts (Nydegger, 1982; Drewes, 1982) which indicate thrusting occurred between 90-75 ma.

**Cretaceous Western Interior Foreland Basin and Seaway**

The thickest accumulation of Cretaceous strata in the Western Interior of North America occurs in a belt that parallels the eastern margin of the Sevier Orogenic Belt. This belt occurs along the axis of a highly asymmetrical foreland basin.

Jordan (1981) and Beaumont (1981) suggested that the tectonically generated foreland basin allowed flooding of the Western Interior and was the primary cause of the Cretaceous seaway. Although Cretaceous sea-level rises might have caused widespread flooding of the craton in the absence of an orogenic belt, they believed that the subsidence and sedimentation that characterized the western margin of the seaway were shaped by lithospheric loading in the thrust belt.

Kauffman (1977, 1984, 1985) reconstructed a relative sea-level curve for the Cretaceous Western Interior (Figure 4). He recognized ten (second-order) transgressive-regressive cycles, which had an average duration of 9 to 10 ma. This transgressive-regressive history has been determined to be relatively synchronous to those of the eastern United States, Europe, Australia, the North Sea, Africa, India, China, and Japan.
Figure 4. Generalized stratigraphy and ages of Cretaceous rocks of the Kaiparowits Plateau compared with transgressive-regressive cycles of the Cretaceous Western Interior Basin (modified from Eaton, 1987, and Kauffman, 1977).
(Vail et al., 1977; Hancock and Kauffman, 1979; Haq et al., 1988) and therefore indicates a eustatic control.

Kaiparowits Plateau

During Late Cretaceous time, the Kaiparowits area lay on the western, active side of the Western Interior foreland basin. The Cretaceous sea transgressed and regressed across the Kaiparowits area during the Late Cretaceous and produced two main cycles of marine, coastal, and continental sedimentation; superimposed on these cycles are many minor transgressive-regressive movements of the shoreline. The two large cycles in the Kaiparowits Plateau represent the Greenhorn and Niobrara second-order cycles of Kauffman (1977) (Figure 4) and third-order-scale cycles of Vail and others (1977).

The generalized stratigraphy and ages of Cretaceous rocks of the Kaiparowits Plateau are shown in Figure 4, and Figure 5 is a schematic diagram showing depositional environments of the Turonian-Coniacian strata. The oldest Late Cretaceous unit is the Dakota Sandstone, which was deposited unconformably on Jurassic rocks in the Cenomanian (Peterson, 1969a; Gustason 1989). Lithologic units in the Dakota show vertical changes from fluvial to coastal marine; these units record the first transgression of the Cretaceous sea into the Kaiparowits area. The Dakota Sandstone is overlain by the Tropic Shale, which was deposited in an offshore marine environment when the Cretaceous sea covered the entire Kaiparowits area during latest
Figure 5. Schematic drawing of the Turonian-Coniacian strata in the Kaiparowits Plateau. Progradational Tibbet Canyon Member shoreline deposits climb stratigraphically to the northeast. They interfinger with and are overlain by coastal and alluvial plain deposits of the Smoky Hollow Member. Marine shoreline and offshore deposits of the John Henry Member in the eastern part of the plateau are temporally equivalent to, and interfinger with, paludal and alluvial deposits to the west.
Cenomanian to Middle Turonian time (Peterson 1969a). As the Cretaceous sea regressed from the Kaiparowits area for the first time during Middle to Late Turonian time (Peterson, 1969a; Eaton, 1987), the Tibbet Canyon Member (shoreline sandstones) and Smoky Hollow Member (coastal and alluvial plain deposits) were deposited (Figure 5). A second, less extensive transgression and regression of the Cretaceous sea, spanning Early Coniacian to Early Campanian time (Peterson, 1969a; Eaton, 1987), is recorded by the John Henry and Driptank Members of the Straight Cliffs Formation (Figure 4). That the sea transgressed only halfway across what is the present day Kaiparowits Plateau is indicated by facies changes within the John Henry Member. On the western side of the plateau, facies represent paludal and alluvial environments; on the east are coastal and marine deposits. The interfingering of these facies was documented by Peterson (1969a, 1969b). During this transgression, the relatively stationary position of the shoreline resulted in the vertical stacking of John Henry shoreline sandstones with marine shales lying to the east-northeast (Figure 5). This stratigraphy is responsible for the existence of the geomorphic "Straight Cliffs escarpment" (Figure 2), which mimics the orientation of the John Henry shoreline (Peterson, 1969a), and from which the formation derives its name. The Wahweap, Kaiparowits and Canaan Peak formations were deposited after the Cretaceous sea had regressed northeast of the Kaiparowits region (Figure 4).
CHAPTER II

STRATIGRAPHY

Southern Kaiparowits Plateau, Previous Work


Peterson (1969a, 1969b) described the Calico bed in the southern Kaiparowits Plateau (Figure 1) as the uppermost of three informal units in the Smoky Hollow Member (Figure 6). The coal zone, the lowermost unit, contains carbonaceous mudstone and coal beds that range in thickness from 0 to about 1.2 m but generally are less than 0.6 m thick. These facies
Figure 6. Relations of members and informal units in the Straight Cliffs Formation, southeastern Kaiparowits region, Utah (Peterson, 1969b).
reflect lagoonal and paludal environments. The middle barren zone contains bentonitic mudstone interbedded with pale-yellowish-brown and grayish-orange, very fine-grained to medium-grained, horizontally stratified and cross-stratified sandstone. Scattered quartz and chert granules and chert pebbles occur locally in the sandstone beds. Strata in the barren zone were deposited in fluvial and floodplain environments. Fluvial sandstones contain abundant low- and high-angle, medium-scale trough cross-stratification and cut-and-fill structures. Ripple cross-laminae are common in the upper 0.3 - 0.9 m of many of these beds. Mudstone and interbedded laminated or ripple cross-laminated sandstone were interpreted as floodplain deposits (Peterson, 1969b).

The Calico bed was distinguished from barren zone sandstones based on color and texture. It consists of white to very light gray, fine- to coarse-grained, poorly sorted, cross-stratified sandstone. Locally, scattered quartz and chert granules, chert pebbles, or conglomerate lenses are common. The Calico bed contains deposits with bedding structures similar to those in the fluvial sandstones of the barren zone, although the Calico bed generally lacks ripple cross-lamination and is composed of more poorly sorted sandstone. These features suggest that the Calico bed was deposited mainly in fluvial channel environments (Peterson, 1969b).

In the southern Kaiparowits Plateau, the Calico bed interfingers with the underlying barren zone, and the bed is
unconformably overlain by the John Henry Member. The bed reaches a maximum thickness of 20.4 m in the central part of Peterson's study area (Figure 1), but is missing in the southwestern and northeastern parts where it was removed by erosion prior to deposition of the John Henry Member (Peterson, 1969b) (Figure 6).

The age of the Smoky Hollow Member is constrained by marine fossils in the underlying Tibbet Canyon Member and overlying John Henry Member. Peterson (1969b) dated the Tibbet Canyon Member based on the late Middle Turonian index fossil *Inoceramus howelli* White. Peterson had originally reported a Middle Coniacian age for the fauna from the base of the John Henry Member, but later revised it to Early Santonian (Peterson and Kirk, 1977, p.178). The fossils recovered from the John Henry Member have been re-examined by E.G. Kauffman (Eaton, 1987, p. 88 and 91 and Plate 1) and an early Coniacian age appears to be more likely based on more recent zonal schemes.

**Northern Kaiparowits Plateau, Revised Stratigraphy**

In the northern Kaiparowits Plateau, *Inoceramus howelli* White occurs in the Tibbet Canyon Member (Stephens, 1973; Zeller 1973). *Cremnoceramus erectus* Meek was recovered from 2 m above the base of the John Henry Member, and was identified by E.G. Kauffman (pers. comm., 1989) as representing
middle Early Coniacian time. These ages are the same as those reported from the southern Kaiparowits Plateau, and indicate that the Smoky Hollow Member was deposited in the Middle to Late Turonian, and possibly into the earliest Coniacian.

Four stratigraphic cross-sections, A-A', A'-A'', B-B', and C-C' (Plates 1, 2, 3, and 4; locations shown in Figures 2 and 7), show the Smoky Hollow Member in the study area. The three informal units in the Smoky Hollow Member, the coal zone, the barren zone, and the Calico bed, are identifiable, although the coal zone and barren zone are combined as the "lower Smoky Hollow Member" on the cross-sections. Stratigraphic sections were measured to the scale of centimeters with a Jacob's staff. The presence of continuous exposure and access determined the locations of the 28 measured sections on the Kaiparowits Plateau.

Cross-sections A-A', A'-A'', and C-C' (Plates 1, 2, and 4) are constructed parallel to depositional strike (Figure 2) as indicated by paleocurrent analysis and by the strike of the Tibbet Canyon Member shoreline as interpreted by Peterson (1969a).

In A-A', the six sections on the southeast end of the section utilize a coal bed near the base of the Smoky Hollow Member as a datum. To the northwest, no reliable datum is identifiable and the base of the Calico bed is used as the datum for convenience.

There are no identifiable chronostratigraphic surfaces for cross-section A'-A'', which consists of four widely spaced
Figure 7. Detailed map showing locations of measured sections in cross-sections A-A', B-B', and the middle of C-C'.
stratigraphic sections. For convenience, the datum is the top of the Calico bed; this is not known to be an isochronous horizon.

Cross-section C-C' lies along the eastern side of the Kaiparowits Plateau and the transgressive disconformity between the Smoky Hollow and John Henry Members was used as a datum. Because the cross-section is parallel to the paleoshoreline trend of the John Henry Member (Peterson 1969a), the transgressive disconformity is assumed to have been formed as an essentially horizontal surface along C-C'.

Cross-section B-B' is parallel to paleoslope, as determined from paleocurrent trends within the Smoky Hollow Member. It is constructed on a coal datum in the lower Smoky Hollow Member, assumed to have been deposited near-isochronously. Plate 3 only shows the upper part of the Smoky Hollow Member along B-B', but the datum can be seen in Figure 8.

Few additional chronostratigraphic units or surfaces exist within the Smoky Hollow Member upon which to base correlation. Six samples of pollen were analyzed, but yielded ages no more definitive than Upper Cretaceous (Terence Okumura, 1989, personal comm.). Ironstone concretion (siderite) horizons were correlated with thin coal units in the mudstone in the lower Smoky Hollow Member in cross-section A-A'. Ironstone concretions are the result of penecontemporaneous precipitation where the supply of ferrous iron is large or a reducing environment is maintained by abundant organic matter (Krauskopf, 1967; Morris, 1967). Thus,
Figure 8. Cross-section B-B'-C': stratigraphy within the Smoky Hollow Member. Transition facies are characterized by coarse-grained, pebbly sandstone. The Calico bed is subdivided into three lithofacies; Facies 1 is conglomeratic sandstone, Facies 2 is coarse-grained sandstone, and Facies 3 consists of medium- to fine grained sandstone and mudstone. Figure 7 shows horizontal scale.
ironstones are correlative with laterally adjacent carbonaceous mudstone/claystone and coal beds.

Identification of the Calico Bed

The Calico bed was distinguished from the barren zone by Peterson (1969a, 1969b) based on three coexisting criteria: 1) lithology: a dominance of sandstone and conglomerate with virtually no mudstone, 2) texture: coarser-grained more poorly-sorted sandstones and the presence of distinct conglomerate lenses, and 3) color: white to very light gray sandstone in contrast to yellowish-brown and grayish-orange sandstones in the barren zone. These features were found to occur independently of one another in the Smoky Hollow Member in the study area, and rarely occur together to uniquely define the Calico bed, as it occurs in the southern Kaiparowits Plateau. This problem was resolved by identifying three lithofacies in the Calico bed and by identifying a lithofacies in the barren zone, the "transition facies", which closely resembles the basal facies in the Calico bed.

Transition facies. The transition facies is composed of channel-belt deposits in the upper barren zone. It is characterized by the presence of conglomerate and coarse-grained sandstone and is commonly grayish-orange (Figure 9) and less commonly white (Figure 10). Thus, this facies satisfies two Calico bed criteria. However, the Calico bed, as defined by
Figure 9. Outcrop along Upper Valley between sections UVC and UV4. Of all the units at this location, the transition facies (TF) shows the most distinctive color: grayish-orange. The Calico bed (CB) lies directly upon the transition facies with no intervening mudstone (contact marked with arrow), and mostly exhibits the same color as sandstone in the Tibbet Canyon (TC) and John Henry (JH) Members, although white lenses are present near the base.
Figure 10. Outcrop of the Smoky Hollow Member at section JC2. The transition facies (TF) is the most prominently white unit at this outcrop and is separated from the Calico bed (CB) by 3.7 m of mudstone (M). The reddish horizon in the mudstone below the transition facies is the ironstone concretion horizon (IC) used to correlate with adjacent sections.
Peterson (1969a, 1969b), lacks mudstone and the transition facies is commonly separated from the Calico bed by mudstone deposits greater than 6.1 m thick, (cross-sections A-A' and C-C' in Plates 1 and 4, Figures 8 and 10). For this reason, the transition facies is not included in the Calico bed, but is considered part of the barren zone. Where the base of the Calico bed lies upon transition facies with little to no intervening mudstone, as in Cross-section B-B' (Plate 3, Figure 8, and Figure 9), the two units are distinguished by gravel size. At any stratigraphic section, the base of the Calico bed is characterized by the coarsest gravel in the Smoky Hollow Member. Gravel in the transition facies is rarely greater than 2 cm (intermediate axis). In contrast, the Calico bed has a basal conglomerate with average maximum clast sizes ranging from 2.6 to 5.7 cm.

Distinguishing the transition facies from the Calico bed can be difficult, as illustrated in cross-section A-A' (Plate 1). At section JC2, the coarsest conglomerate in the Smoky Hollow Member was preliminarily assigned to the Calico bed. However, this unit is overlain by 3.7 m of mudstone (Figure 10). Using an ironstone concretion horizon, these strata were correlated to sandstone bodies that lack conglomerate and are clearly part of the barren zone.

**Calico bed.** The Calico bed is not a single bed as the name implies; rather, it is laterally and vertically coalesced channel-belt deposits. The name is appropriate, however, in the
southern Kaiparowits Plateau, where it is a relatively thin unit, no more than 20.4 m thick, and consistently white, and thus appears to form a discreet "bed" (Figure 11).

In contrast, the Calico bed in the study area is thicker than in Peterson's type area, up to 48.8 m, and consists of three facies, which in ascending order are: a conglomeratic sandstone, a coarse-grained sandstone, and a medium- to fine-grained sandstone and mudstone facies. The last is present only in the northeast part of the study area.

The base of the Calico bed is identified based on two of the three original criteria: (1) where lithology is dominated by sandstone and conglomerate, with virtually no mudstone and (2) where the coarser texture of gravel within the Calico bed distinguishes it from underlying transition facies. The third characteristic of the Calico bed in the type area, white color, is unreliable for identifying the Calico bed in the study area. The Calico bed can be the same color as under- and overlying strata (Figure 9) and the white color is not ubiquitous throughout the unit (Figures 9 and 12).

**Basal Contact of the Calico Bed**

This study reinforces the original interpretation of Peterson (1969a, 1969b) that the Calico bed interfingers with the underlying fluvial sediments of the Smoky Hollow Member.

The presence of the transition facies indicates that the contact between the barren zone and the Calico bed is
Figure 11. The Smoky Hollow Member in the southern Kaiparowits Plateau (near section SM, Figure 2). The Calico bed (CB) appears to form a distinct white "bed" at the top of the member and is approximately 4 m thick. The Tibbet Canyon Member is marked "TC", the lower Smoky Hollow Member is "LSH", and the John Henry Member is "JH". A coal bed, the dark horizon marked with an arrow, lies 2.5 m above the contact between the Smoky Hollow and John Henry Members.
Figure 12. Calico bed, at section PL, displays variable color; isolated white lense within a generally pale, orange-brown outcrop. Staff for scale is 1.5 m.
gradational, and not disconformable. The transition facies and the basal facies of the Calico bed were deposited in similar fluvial environments (discussed in the next chapter). Thus, there is no displacement of facies, which would be an indication of an unconformity.

Furthermore, the base of the Calico bed in the study area is not an isochronous surface, but lies at different stratigraphic levels within the Smoky Hollow Member. For example, the base of the Calico bed at section LD correlates with a major erosion surface within the Calico bed at section NC2, and the base at NC2 lies approximately 6 m below that erosion surface, as shown in cross-section C-C' (Plate 4). Likewise, the base along the trend of cross-section B-B' lies 15 m lower than along C-C' (Figure 8).

Upper Contact of the Calico Bed

Peterson (1969a, 1969b) described an unconformity separating the Smoky Hollow Member from the overlying John Henry Member. The most convincing evidence is the truncation of strata and thickness variations in the Smoky Hollow Member across Peterson's study area (Figure 6). In the southwestern part of the area, the Calico bed was eroded before deposition of the John Henry Member, and channel and floodplain deposits of the John Henry Member rest directly on the barren zone of the Smoky Hollow Member. In this area the Smoky Hollow is only 7.3 m thick. In the central part of the region the Calico bed is present and the Smoky Hollow Member is 37 m thick. In the
northeastern part of Peterson's study area, the Calico bed was also eroded, and marine strata of the John Henry Member rest on 9.5 m of the Smoky Hollow Member (Peterson, 1969b).

In the northern Kaiparowits Plateau, no such unconformity was found. Although a transgressive disconformity, or ravinement surface (Swift, 1968), separates the Smoky Hollow Member from marine strata of the John Henry Member in the eastern part of the study area, no evidence of truncation was found: a gradual transition of lithofacies from the base of the Calico bed to the disconformity records a continuous base-level rise which culminated in the marine transgression (Figure 8, Plates 3 and 4).

The transgressive surface is dramatically marked by a sheet of quartzite cobble conglomerate, a transgressive lag deposit, which is typically 0.15 to 0.3 m thick and may be up to 0.6 m thick. The average maximum cobble size, 6.7 cm, is larger than maximum clast sizes in the Calico bed and thus indicates that the lag was not derived from erosion of the Calico bed alone. Rather, the conglomerate was probably sourced and transported from north of the study area by strong, storm-generated, coast-parallel currents (Ericksen and Slingerland, 1990). The conglomerate is interbedded with and overlain by a tabular unit of predominantly hummocky cross-stratified (HCS) sandstone, which has an average thickness of 2 m (Figure 13). *Inoceramus* shells, shell fragments, and sharks teeth in the HCS sandstone confirm that it was deposited in a marine
Figure 13. Transgressive disconformity separating the Smoky Hollow and John Henry Members in the eastern part of the study area. Contact is marked with an arrow. Quartzite cobble conglomerate occurs at the base of, and interbedded with, a hummocky cross-stratified sheet sandstone unit. Tape for scale is 2.0 m.
environment. The HCS sandstone is laterally continuous across the east half of the study area and is overlain by marine shale and sandstone of the John Henry Member. Where the lag conglomerate is absent, as at sections LD, StC 2, and LD (cross-section C-C', Plate 4), the transgressive disconformity is still easily identifiable by the presence of this HCS sandstone unit.

In the western part of the study area (Plates 1 and 2), an abrupt, yet conformable, contact exists between the Calico bed and the John Henry Member. The base of the John Henry Member in this area, like the barren zone of the Smoky Hollow Member, consists of interbedded sandstone and mudstone, representing channel and floodplain deposition. The Calico bed is distinguished from the muddy, fine- to medium-grained John Henry sandstones because it is dominated by medium- to coarse-grained sandstone. The Calico bed is characteristically cliff-forming in this area, whereas the John Henry sandstones and mudstones usually weather into slopes. In the few areas where cliff-forming John Henry sandstones directly overlie the Calico bed, the contact is difficult to identify. No distinct differences in sedimentary structures or alluvial architecture serve to distinguish sand bodies of the two units (Figure 14).

In fact, the Calico bed probably interfingers with the John Henry Member in this region. In cross-section A-A' (Plate 1), the upper 3.0 to 4.5 m of the Calico bed at PF and BV was identified on grain size. However, a 4 ft interval of mudstone below the upper unit at PF suggests that it could legitimately be
Figure 14. Abrupt, conformable contact between the Calico bed and John Henry Member in the western part of the study area; a) South of section PC, cliff-forming John Henry sandstone (JH) in contact with Calico bed (CB); b) At section JC2, contact between Calico bed (CB) and John Henry Member (JH) marked by arrow. Calico bed is 11.5 m thick.
included in the John Henry Member. At BV, a major bounding surface at the base of the upper unit suggests that it could be considered part of the John Henry Member.

Concurrent with this study, Shanley and McCabe (1990, 1991) correlated fluvial deposits of the John Henry Member on the western side of the Kaiparowits Plateau with the marine strata in the east. Ten meters above the base of the John Henry, tidally-influenced river deposits, which contain a brackish trace fossil assemblage, were interpreted to be temporally equivalent with the maximum flooding surface, which is marked by a thin sharks-tooth lag and lies 6 m above the transgressive disconformity at section L.C. (Shanley and McCabe, 1989, 1991) (Figure 2, Plate 3). Their correlation can be extrapolated downward to suggest that the abrupt contact between the Calico bed and John Henry fluvial strata represents a deepening event correlative with the transgressive disconformity.

The presence of an unconformity separating the Calico bed and the John Henry Member in the southern Kaiparowits Plateau is unquestioned by this study. However, the absence of a correlative unconformity in the northern Kaiparowits Plateau indicates that the processes that formed the unconformity in the south did not affect the entire region. Cross-section A'-A" (Plate 2) includes Peterson's (1969a) study area and shows that the Calico bed is much thinner in the south at section SM (4.0 m), than in the north at section H (11.4 m). While erosion may be partially responsible for this difference, there is also evidence
that subsidence was greater in the northern Kaiparowits region during deposition of the Smoky Hollow Member and resulted in greater thicknesses of preserved strata throughout the member. The lower Smoky Hollow interval is thicker in the north, with a thickness of 42.7 m at H compared to 32.0 m at SM (Plate 2). Cross-sections A-A' and C-C' (Plates 1 and 4) also illustrate this trend. Locations with a thick Calico bed have thick sections of lower Smoky Hollow; relatively thin Calico bed sections occur with relatively thin sections of lower Smoky Hollow. Lower Smoky Hollow thickness trends are inferred to be independent of interfingering with the underlying Tibbet Canyon Member, as cross-sections A-A' and C-C' parallel the orientation of the Tibbet Canyon paleoshoreline (Peterson, 1969a).

Eaton (1987) also concluded that thickness trends are the result of differential subsidence across the region. His measured sections of the entire Straight Cliffs Formation indicate a thickness of 354 m at Tibbet Canyon in the southern part of the plateau (section TC in Figure 2) and 493 m at Pardner Canyon in the northwest (Figure 2, section PC in Figure 7).

Petrography

Detrital Modes

Differences in modal mineralogy support the stratigraphic distinction between the transitional facies in the barren zone and the Calico interval. To determine modal mineralogy, three
hundred points were counted on thin sections of 21 sandstone samples, using the Gazzi-Dickinson (1966-1970) method (Zuffa, 1985). All slides were stained with sodium cobaltinitrite for the identification of potassium feldspars. Compositional comparisons of samples are considered valid, as all are of medium to coarse-grained sandstone (mean grain size 0.5 mm) (Dickinson, 1985).

The use of quantitative detrital modes to infer the tectonic settings of provenance terranes is well established (Dickinson and Suczek, 1979). Toward that purpose, in this study detrital modes were recalculated to 100 percent as the sum of Qm (monocrystalline quartz; grain size > 0.0625 mm), Qp (polycrystalline quartz; chert or metaquartzite), P (plagioclase grains), K (potassium feldspar grains), Lv (volcanic/metavolcanic lithic fragments), and Ls (sedimentary/metasedimentary lithic fragments), following the methodology of Dickinson (1985). Data are plotted on a triangular diagram with apexes Qm, F (total feldspar grains; P+K), and Lt (total lithic fragments; Qp+Lv+Ls).

Detrital modes indicate that the transition facies is mineralogically distinct from the Calico bed. Calico bed sandstones plot in the quartzofeldspathic field (Figure 15a). In contrast, transition facies sandstones have a higher ratio of sedimentary lithic fragments (including much chert) to monocrystalline quartz and plot more towards the lithic field. To test the assumption that differences in modal mineralogy are indeed temporal, and not spatial, Calico bed sandstones were
Figure 15. Detrital modes of sandstone in this study compared to standard distributions indicative of provenance. (a) Sandstone from this study plotted in groups according to geography and stratigraphic position. (b) Actual reported distribution of mean detrital modes for sandstone suites derived from different types of provenances (from Dickinson, 1984, Figure 2)
subdivided geographically; samples from the west side of the Kaiprowits Plateau in cross-sections A-A', and A'-A" were plotted together vs. those from the east in cross-sections B-B' and C-C'. Calico bed sandstones from all, widely separated, geographic locations fall along the same compositional trend, verifying this assumption (Figure 15a).

Dickinson and others (1983) suggested that detrital modes of sandstone suites primarily reflect the different tectonic settings of provenance terranes and that evolutionary trends in sandstone composition commonly reflect changes in tectonic setting through time. Comparison of detrital modes to Dickinson's (1985) triangle diagram (Figure 15b) shows that transitional facies sandstones lie in the "mixed" provenance field. It indicates that transition facies were sourced by both "continental block" and "recycled orogen" provenances. This is consistent with what is known about the tectonic zones active during deposition of the Smoky Hollow Member. Thrusts in the Mogollon Highlands involve Precambrian crystalline and Jurassic plutonic rock and this is regarded as the "continental block" provenance. Decollement-style thrusting in the Sevier Orogenic Belt constitutes the "recycled orogen" provenance.

In contrast, Calico bed sandstones lie entirely within the "continental block" provenance field indicating a stronger contribution of sediment from the Mogollon Highlands, although pebbles of chert and metaquartzite in the Calico bed indicate that the Sevier Orogenic Belt still supplied sediment to this unit.
This evolutionary trend in sandstone composition, from the transition facies to the Calico interval, reflects a change through time in the relative contributions from the different tectonic settings.

Three sandstone samples were collected from the upper, fluvial unit of the Ferron Sandstone Member of the Mancos Shale in the Henry Mountains region (Peterson and Ryder, 1975). They were examined and plot in the same field as the Calico bed, suggesting that the upper Ferron is correlative with the Calico bed. This lithostratigraphic correlation is supported by biostratigraphy, discussed more fully in Chapter 5.

Clay Mineralogy

In previous studies, the white color of the Calico bed was a key characteristic in the identification of the unit. This color was attributed to large quantities of authigenic kaolinite in the Calico bed. The kaolinite presumably was formed by the alteration of feldspar grains during subaerial exposure and erosion after Calico bed deposition and before deposition of the John Henry Member (Peterson, 1969a, 1969b; Vaninetti, 1979). This study concludes that the alteration took place during post-burial diagenesis or during recent exposure in outcrop, because no subaerially-formed unconformity exists at the top of the Calico interval in the study area.

In this study, thin section analyses and X-ray diffraction confirmed that authigenic kaolinite imparts white color to the
Calico interval. That the clay is authigenic is indicated by grain-size areas of monomineralic clay (Figure 16, and 17). Furthermore, thin sections were stained for both potassium feldspar and plagioclase, and point counts revealed that only the potassic variety is present (Figure 17). This suggests that plagioclase grains were altered and provided the elements for the clay pseudomorphs. Pseudomorphs of granitic rock fragments, composed of clay and potassium feldspar confirm this inference (Figure 17). Low birefringence and high relief indicate that this clay is mostly kaolinite. In some areas, the clay occurs in its characteristic vermiform habit (Figure 16). X-ray diffraction of six samples indicated that illite is present, but that most of the clay is kaolinite.
Figure 16. Photomicrograph of sandstone from section SM a) in plain light and b) under crossed nichols. Kaolinite (k), showing high relief, low birefringence and characteristic vermiform habit, fills grain size area. Detrital grains are potassium feldspar (Ksp), quartz (Q), and sedimentary rock fragments (SRF).
Figure 17. Photomicrograph of sandstone from section LD a) in plain light and b) under crossed nichols. Detrital grains are potassium feldspar (Ksp), quartz (Q), and sedimentary rock fragments (SRF). The pseudomorph of a granitic rock fragment (GRF) is composed of kaolinite and potassium feldspar. Kaolinite and minor illite fill intergranular areas.
CHAPTER III

SEDIMENTOLOGY

Introduction

Detailed sedimentology of the Smoky Hollow Member is shown in cross-sections A-A', A'-A", B-B', and C-C' (Plates 1, 2, 3, and 4). Stratigraphic sections along B-B' were described in the greatest detail and that cross-section is presented at twice the scale of the others. Because the various kinds of sedimentary structures are drafted to scale in Plates 1 through 4, a written description of the stratigraphic columns is not provided. Figure 18 illustrates the lithologies and sedimentary structures described in the measured sections. Sedimentary structures are matched with the lithofacies codes of Miall (1977, 1978). This interpretive lithofacies scheme is shown in detail in Table 1.

In this study, the term "conglomerate" is restricted to rocks composed of extrabasinal clasts, dominantly of chert (grayish, black, and banded) and quartzite (red, purple, gray, and white). It is distinct from the "intraclast" facies which consists of gravel-sized clasts of intrabasinally derived shale, claystone, and mudstone. Subdivision between sandstone and conglomerate into "pebbly sandstone" and "sandy conglomerate"
Figure 18. Key to stratigraphic sections in Plates 1-4.
# Table 1. Lithofacies Classification (from Miall, 1978)

<table>
<thead>
<tr>
<th>Facies Code</th>
<th>Lithofacies</th>
<th>Sedimentary Structures</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gm</td>
<td>Massive or crudely bedded...gravel</td>
<td>Horizontal bedding, imbrication</td>
<td>Longitudinal bars, lag deposits, sieve deposits</td>
</tr>
<tr>
<td>Gt</td>
<td>Gravel, stratified</td>
<td>Trough cross beds</td>
<td>Minor channel fills</td>
</tr>
<tr>
<td>Gp</td>
<td>Gravel, stratified</td>
<td>Planar cross-beds</td>
<td>Longitudinal bars, deltaic growths from older bar remnants</td>
</tr>
<tr>
<td>St</td>
<td>Sand, medium to very coarse, may be pebbly</td>
<td>Solitary or grouped trough cross-beds</td>
<td>Dunes (lower flow regime)</td>
</tr>
<tr>
<td>Sp</td>
<td>Sand, medium to very coarse, may be pebbly</td>
<td>Solitary or grouped planar cross-beds</td>
<td>Lingoid, transverse bars, sand waves (lower flow regime)</td>
</tr>
<tr>
<td>Sr</td>
<td>Sand, very fine to coarse</td>
<td>Ripple marks</td>
<td>Ripples (lower flow regime)</td>
</tr>
<tr>
<td>Sh</td>
<td>Sand, very fine to very coarse, may be pebbly</td>
<td>Horizontal lamination, parting or streaming lineation cross-beds</td>
<td>Planar bed flow (upper flow regime)</td>
</tr>
<tr>
<td>Sl</td>
<td>Sand, very fine to very coarse, may be pebbly</td>
<td>Low-angle (&lt;10°) cross-beds</td>
<td>Scour fills, washed out dunes, antidunes</td>
</tr>
<tr>
<td>Se</td>
<td>Erosional scours with intraclasts</td>
<td>Crude cross-bedding</td>
<td>Scour fills</td>
</tr>
<tr>
<td>Ss</td>
<td>Sand, fine to very coarse, may be pebbly</td>
<td>Broad, shallow scours</td>
<td>Scour fills</td>
</tr>
<tr>
<td>Fl</td>
<td>Sand, silt, mud</td>
<td>Fine lamination, very small ripples</td>
<td>Overbank or waning flood deposits</td>
</tr>
<tr>
<td>Fm</td>
<td>Mud, silt</td>
<td>Massive, desiccation cracks</td>
<td>Overbank or drape deposits</td>
</tr>
<tr>
<td>C</td>
<td>Coal, carbonaceous mud</td>
<td>Plant, mud films</td>
<td>Swamp deposits</td>
</tr>
</tbody>
</table>
was made according to the field classification scheme of Folk (1954); pebbly sandstone contains 5-30 percent gravel size clasts, sandy conglomerate contains 30-80 percent, and conglomerate contains greater than 80 percent gravel size clasts. Maximum clast size within significant conglomerate units is indicated next to the measured sections, and was determined by averaging the intermediate axes of the ten largest pebbles or cobbles.

Paleocurrent directions were reconstructed based on measurements taken from large scale cross-strata and the areal distribution of maximum clast size in conglomerate facies at the base of the Calico bed.

**Architectural-Element Analysis**

In addition to detailed vertical descriptions of sedimentology, the method of architectural-element analysis (Miall, 1985 and 1988) was used to interpret the fluvial systems that deposited the Calico bed. This method is based on two dimensional analyses of well-exposed lateral profiles and involves the identification of 1) architectural elements, which are lithofacies assemblages with distinctive internal relationships and external geometries, and 2) the bounding surfaces which define the architectural elements internally and which separate them from one another.
The Calico bed was analyzed in terms of architectural-element analysis utilizing photomosaics of outcrops as suggested by Miall (1985). However, complete application of this method was not possible because measured outcrops were commonly unsuitable for photomosaic analysis and clear vertical exposures were generally inaccessible. This logistical problem was dealt with by describing some bounding surfaces when measuring vertical sections. Bounding surfaces identified are indicated in the measured sections presented in cross-sections A-A', A'-A", B-B', and C-C' (Plates 1, 2, 3, and 4). Thus, although lateral descriptions of the geometries of lithofacies assemblages were not always possible, some architectural elements could still be assigned to vertical sections even in the absence of photomosaics. Where photomosaics were made, the bounding surface hierarchy was applied with only high-order surfaces identified. Although complete two-dimensional lithofacies identification was not possible, typically at least one detailed vertical section was described in each lateral profile and facilitated the identification of architectural elements.

Bounding Surface Hierarchy

Allen (1983) demonstrated that bounding surfaces in fluvial deposits could be classified into several main types, reflecting their areal extent and the significance of the time break they represent. A revised and expanded classification taken from Miall (1988) is used in this study. Because the higher
numbered bounding surfaces are more areally extensive and
time-significant, only third- through sixth-order bounding
surfaces (Miall, 1988) are specifically identified in photomosaics
and measured sections. Architectural elements are bounded by
surfaces of fourth order or higher rank (Miall, 1985).

Third- and fourth-order surfaces. These surfaces reflect
the presence of macroforms, which include complex bars, major
channel reaches, and meander belts (Miall, 1988). Macroforms
represent long-term accumulation of sediment in response to
major tectonic, geomorphic, and climatic controls.

Third order surfaces are cross-cutting erosion surfaces
that indicate stage changes, but no significant change in
sedimentary style or bedform orientation. These are common
within macroform bar complexes and represent surfaces of
growth, either by lateral or downstream accretion.

Fourth-order surfaces include the upper bounding surfaces
of macroform complexes, such as downstream accreting
macroforms (DA) and lateral accretion deposits (LA). These
surfaces typically are flat to convex-upward. Underlying
bedding planes may be locally parallel to the upper bounding
surface, and the form of this surface is commonly echoed by that
of internal third-order surfaces within the underlying
macroform element, indicating that the fourth order surfaces
originate as macroform growth surfaces.
A second type of fourth-order surface marks the basal scour surface of minor channels, such as chute channels. Where major channels are present these would be bounded by higher order surfaces.

**Fifth-order surfaces.** In this category are surfaces bounding major sand sheets, including channel-fill complexes. They are generally flat to slightly concave-upward, but may be marked by local cut-and-fill relief and by basal lag gravels.

**Sixth-order surfaces.** These surfaces define groups of channels, or paleovalleys. Mappable stratigraphic units such as member or submembers, are bounded by sixth order surfaces

**Architectural Elements**

Characteristic lithofacies assemblages with distinctive internal relationships and external geometries comprise architectural elements (Allen, 1983), of which eight common, basic types have been defined (Miall, 1988). Six have been recognized in the Smoky Hollow Member and the characteristics of these six elements are summarized below.

**Channels (CH).** These are bounded by flat or concave-upward scour surfaces. They may occur on more than one scale in any fluvial system, and the larger channels commonly contain complex fills consisting of one or more of the other element
types. Any combination of lithofacies (Table 1) may occur in this architectural element.

**Gravel bars and bedforms (GB).** Tabular or cross-beded gravels originate as simple longitudinal or transverse bars. Element GB usually occurs in tabular bodies composed of lithofacies Gm, Gp, and Gt.

**Sandy bedforms (SB).** This element represents channel fills, bar tops, minor bars, and crevasse splays, and occurs in lense, sheet, and wedge geometries. Sandy bedforms may be a combination of lithofacies St, Sp, Sh, Sl, Sr, Se, and Ss.

**Downstream accreting macroforms (DA).** The lithofacies assemblages in this element are similar to those in the sandy bedforms (SB); however, this element is distinguished by the presence of convex-upward third-order internal and upper bounding surfaces. Component bedforms are dynamically interrelated, with indicated paleoflow directions parallel or subparallel to the dip of the third- and fourth-order bounding surfaces, indicating that bar growth occurred by downstream accretion.

**Lateral accretion deposits (LA).** This element represents the familiar point bar, in which paleoflow, as indicated by sedimentary structures, is at a high angle to the dip of the internal accretion surfaces, indicating growth by lateral
accretion. Gradation may occur between LA and DA element types, particularly in multiple-channel rivers.

**Overbank fines (OF).** Included here are mudstone, siltstone and minor thin sandstone beds formed in interchannel areas. These intervals may form as thin drapes, as thick blankets, or as U-shaped fills of abandoned channels, and are commonly interbedded with element SB.

**Depositional Environments**

The sedimentary structures and alluvial architecture of the Smoky Hollow Member indicate that fluvial style changed through time. In each of the stratigraphic units introduced in the previous chapter, depositional sub-environments are distinguished and used to interpret the paleo-channel morphology using the terminology of Schumm (1981) (Figure 19), which characterizes rivers by channel pattern and the type of load carried. In a general sense, fluvial style in the Smoky Hollow evolved from meandering, to braided, and back to meandering.

**Barren Zone of the Smoky Hollow Member**

**Lower barren zone description.** Fluvial sandstone bodies in the lower barren zone consist of fine- to medium-grained, muddy sandstone. The dominant sedimentary structures are
Figure 19. Channel classification based on pattern and type of sediment load with associated variables and relative stability indicated (from Schumm, 1981; Schumm and Meyer, 1979).
trough cross-stratification (lithofacies St) and ripple cross-lamination (lithofacies Sr) as illustrated in cross-section B-B'. The alluvial architecture consists of fining-upwards sequences with erosional basal bounding surfaces. Intraclasts, 1 cm average diameter, line the basal erosion surfaces. Within the stories, set thicknesses of trough cross-stratification generally decrease upwards (Figure 20). Such sequences are commonly capped by ripple cross-lamination

**Lower barren zone interpretation.** Although no architectural elements are discernable in the sandstone units, the sedimentary structures present indicate deposition by meandering rivers. The dominance of trough cross-stratification over other sedimentary structures and upward decrease in the thickness of cross-stratified sets are characteristics of point bar deposits; the fining upward trend reflects bar growth by lateral accretion (Allen, 1965, 1970; Jackson 1976). Intraclasts overlying erosional surfaces at the bases of stories are also part of the familiar point bar facies model (Walker and Cant, 1984), and indicate that cohesive overbank deposits existed, which is characteristic of the meandering system (Schumm, 1977).

**Transition facies description.** Conglomeratic sandstone units which lie in the uppermost part of the barren zone of the Smoky Hollow Member comprise the transition facies. At a given outcrop, transition facies may consist of a single story or be multistoried. This facies is distinguished from other fluvial
Figure 20. Fining-upwards sequence in barren zone sandstone at section UV1. Trough cross-statification set thicknesses decrease upwards from 0.75 m at base to ripple cross-lamination at top. Staff for scale is 1.5 m.
sandstones in the lower Smoky Member by the abundance of pebbles, 1.5 to 2.0 cm (intermediate axes) (Figure 21), and by the dominance of coarse-grained sandstone.

A variety of sedimentary structures are present. Large-scale trough cross-stratified sandstone, pebbly sandstone, and conglomerate (lithofacies Ss and Gt) typically occur at the bases of stories. Interbedded horizontally stratified conglomerate and sandstone (lithofacies Gm and Sh) are common at the base of the unit. Large sets (0.5 to 1.5 m thick) of planar cross-stratified pebbly sandstone and conglomerate (lithofacies Sp and Gp) are common (Figure 22). Most common are thick intervals of medium scale (0.1 to 0.4 m thick sets) planar cross-stratified sandstone and pebbly sandstone. Similar scale planar cross-stratified sandstone also occurs interbedded with horizontally stratified pebble-conglomerate sheets which are typically one to two pebbles thick. Thick sequences of medium scale (0.1 to 0.6 m thick sets) trough cross-stratified sandstone and pebbly sandstone (lithofacies St) are common along cross-section B-B'. There is no fining-upwards trend in these sequences and in some cases such sequences are coarser at their tops (e.g. sections UV1 and UV2, Plate 3).

Because the transition facies usually is friable and slope-forming, alluvial architecture is difficult to determine. Fourth- and fifth-order bounding surfaces subdivide the facies into stories, but at only one location were third-order surfaces identifiable. In the uppermost story at section UV3 (Plate 3),
Figure 21. Typical pebble size in the transition facies, seen at section UV3. Ruler for scale is 15 cm long.
Figure 22. Large-scale, planar cross-stratified, pebbly sandstone and sandy conglomerate in the transition facies at section UV3. Cross-strata are inclined 15° in the direction of N65E. Staff for scale is 1.5 m.
third-order bounding surfaces dip in the same direction as the sets of planar cross-stratified sandstone which they separate (Figure 23), indicating a downstream accreting macroform (architectural element DA). The DA element is capped by ripple cross-lamination, suggesting complete preservation of the bar. Gravel bars and bedforms (architectural element GB) are indicated by the presence of lithofacies Gm, Gp, and Gt. Sandstone lithofacies (St, Sp, Sh, and Ss) are non-diagnostic, as they all occur in architectural elements SB, DA, and LA.

**Transition facies interpretation.** Transition facies record deposition by braided streams. This interpretation is based on the identification of the types of bar and channel deposits common to braided rivers and evidence of discharge fluctuations characteristic of braided rivers.

Large sets of planar cross-stratified pebbly sandstone and conglomerate (Sp and Gp) indicate foreset deposition on simple bars in shallow, multichannel rivers. Sp and Gp commonly overlie broad, thin conglomerate sheets, which are interpreted as channel floor lag deposits (the "diffuse gravel sheets" of Hein and Walker, 1977); this association confirms the interpretation that those Sp and Gp are main channel bars and not secondary, chute bars across the tops of point bars (McGowen and Garner, 1970; Jackson, 1976; Levey, 1978). The planar cross-stratified sets are similar in scale to those produced by "cross-channel bars" in the South Saskatchewan River (Cant and Walker, 1978)
Figure 23. Downstream accreting macroform deposit (DA) in the transition facies at section UV3. Sets of planar cross-stratification dip in the same direction (to the right; due east) as inclined third-order bounding surfaces (marked by small arrows). The base of the DA (marked by large arrow) is a fifth-order bounding surface. The DA is 6 m thick.
and lingoid bars in the Platte River (Blodgett and Stanley, 1980; Crowley, 1983)

Stacked sets of smaller scale planar cross-stratification (lithofacies Sp) are similar to mid-channel, compound bar deposits described in the braided South Saskatchewan River (Cant and Walker, 1978; Cant, 1978) and in the ancient record (Haszeldine, 1983). This interpretation is supported by the presence of third-order bounding surfaces in one such sequence of Sp (Figure 23), which confirms that it is a downstream accreting macroform (architectural element DA). The 6 m thickness of this sequence indicates the minimum water depth.

Channel deposition is represented by two scales of trough cross-stratification. Sequences of medium scale pebbly St reflect dune migration within main channels and resemble the "in-channel deposits" in the South Saskatchewan River (Cant and Walker, 1978; Cant, 1978). Scour and fill structures (lithofacies Ss) and large scale trough cross-stratified conglomerate (lithofacies Gt) reflect numerous small channels, indicative of a braided morphology.

Discharge fluctuations characteristic of a braided river morphology are indicated by sequences of interbedded channel lag deposits (thin sheets of conglomerate) and simple foreset bar deposits (Sp). Also, interbedded Gm and Sh suggest fluctuating flow strength; i.e. conglomerate deposited along the base of the channel during high flow, and sandstone deposited in the upper flow regime during waning flow.
The two scales of bar and channel deposits suggest deposition in braided rivers which were in places relatively deep, and in others, broad and shallow. A block diagram from Miall (1985) (Figure 24) shows how the various sub-environments were probably related.

**Calico Bed**

The Calico bed is multistoried. Vertically through this unit, there are consistent changes in grain size, types and sizes of sedimentary structures, and architectural elements. These changes are here discussed by contrasting the three different facies of the Calico bed, each of which is multistoried. Facies 1, the conglomeratic sandstone facies, always occurs at the base of the Calico bed. Facies 2, the coarse sandstone facies, always overlies Facies 1. Facies 3, the fine sandstone and mudstone facies, is present only in the northeast Kaiparowits Plateau where it overlies Facies 2, as shown in cross-section B-B' (Plate 3). Facies 1 and 2 occur everywhere in the study area, but the subdivision is informal, as Facies 1 grades into Facies 2; at most measured sections there is not a definite contact. A notable exception to this is along the west side of cross-section B-B' where Facies 1 is distinctly separated from Facies 2 by a fifth order bounding surface (Plate 3).

**Facies 1 description.** Pebbles within the lower portion of the Calico bed range in maximum diameter from 2.6 to 5.7 cm
Figure 24. Model of rivers that deposited the transition facies. The relatively deep, low sinuosity rivers had downstream accreting macroforms (DA) and isolated lingoid and transverse bars (SB) (from Miall, 1985, Figure 16).
(Figure 25). The coarsest pebbles at any given stratigraphic section lie at the base of Facies 1, and pebble sizes decrease upwards. Coarse- to very coarse-grained sandstone dominates this facies. Large intraclasts, b-axis as large as 13cm, occur, but are generally restricted to the base of the facies where they are associated with the basal sixth-order bounding surface of the Calico bed.

Horizontally stratified framework-supported conglomerate (lithofacies Gm) is variable in occurrence. Tabular bodies of Gm, up to 1.5 m thick, occur at the base of the facies (Figures 26 and 27). Intervals, up to 3.5 m thick, of interbedded horizontally stratified conglomerate and very coarse-grained pebbly sandstone (lithofacies Gm and Sh) also occur in a basal stratigraphic position. Thin tabular and lenticular lithosomes (0.1 to 0.3 m thick) of framework supported conglomerate commonly occur in the middle of Facies 1. These units occur 0.6 to 1.6 m apart, vertically, and separate intervals of coarse-grained pebbly sandstone (Figure 27). The lenticular lithosomes are convex downwards and are 0.2 to 0.3 m thick at the central, thickest portions. These units thin laterally, and pass into single-pebble thick (0.02 m) conglomerate beds over distances of 3.7 to 4.6 m.

An abundance of large scale planar cross-stratified sandstone, pebbly sandstone, and conglomerate (lithofacies Sp and Gp) is unique to Facies 1 (Figure 28 and 29). These sets range in thickness from 0.6 to 1.9 m and are commonly
Figure 25. Typical-sized pebbles at the base of Facies 1, seen at AW. Average maximum pebble size at this locality is 3.5 cm (b-axis). View looking down onto top of a thin (4-5 cm) sheet of horizontally stratified conglomerate.

Figure 26. Horizontally and low-angle stratified, framework-supported conglomerate (Gm) (at base and in lower left) in Facies 1 at section RV. Staff for scale is 1.5 m.
Figure 27. Thin, broad sheets of framework-supported conglomerate (Gm) interbedded with coarse-grained pebbly sandstone (in upper part of picture). Staff (1.5 m long) sits on thick (1.5 m) tabular unit of sandy conglomerate, which is framework supported at base. Top of staff points to a thin conglomerate sheet. (Facies 1, section UV3).
Figure 28. Large scale planar cross-stratified sandstone (Sp) in Facies 1, at section NC2. Staff (1.5 m long) rests on base of set and hand holds staff at top of the 1.1 m thick set.
Figure 29. Planar cross-stratified sandy conglomerate (Gp) overlain by horizontally stratified sandy conglomerate (Gm) in Facies 1, at section UV1; staff is 1.5 m long.
associated with lithofacies Gm. As in the transition facies, intervals of stacked sets of medium scale (0.2 to 0.6 m thick) planar cross-stratification (lithofacies Sp) are abundant and are commonly interbedded with thin (2 to 6 cm thick), tabular sheets of conglomerate. Medium scale (0.3 m thick sets) trough cross-stratified sandstone and pebbly sandstone (lithofacies St) is less common.

Large scale trough cross-stratified sandstone, pebbly sandstone and conglomerate (lithofacies Ss and Gt) and low-angle cross-stratified sandstone and pebbly sandstone (lithofacies Sl) are also abundant in Facies 1, and occur at all stratigraphic levels in this interval. Gt sets range in thickness from 0.4 to 1.2 m. Ss units are typically 1.5 to 2.0 m thick (Figure 30), although a 4.6 m thick unit was observed at the base of section UVA (Plate 3).

The architecture of Facies 1 is dominated by tabular bodies of sandstone and conglomerate (Figure 31). These tabular bodies are bounded by fourth- and fifth-order surfaces. Gravel bars (architectural element GB) are indicated by lithofacies Gm and Gp. Downstream accreting macroforms (architectural element DA) are documented towards the top of Facies 1 at sections UVA and UVB (Plate 3) (Figure 32).

Numerous fourth- and fifth-order bounding surfaces and lithofacies Gt, Ss, and Sl indicate an abundance of small channel elements (architectural element CH) (Figure 33). Most of the channels in Figure 33 are filled by lithofacies Ss, but some
Figure 30. Scour-fill structures composed of pebbly sandstone (Ss) and sandy conglomerate (Gt) at base of Facies 1 at section UV3. Hammer is 28 cm long.
Figure 31. Photomosaic showing architecture of the Calico bed at section UVC. Line drawing of this interval is on the next page.
Figure 31. Line drawing of the outcrop at section UVC in photograph on previous page. Facies 1 is dominated by tabular bodies (stories) of elements GB, SB, and minor DA. In contrast, Facies 2 is composed of more lenticular sand bodies, many of which are composed of element LA. More complete preservation of element LA occurs at the top of the facies.
Figure 32. Downstream accreting macroform deposit (DA) at the top of Facies 1, at section UVB. DA unit is 2.5 to 3 m thick and rests on a 30 cm thick sheet deposit of conglomerate (marked with arrow). Within the DA unit, planar cross-strata and third-order bounding surfaces dip to the right (northeast).
Figure 33. Sketch of architecture at section UV2. The architecture of Facies 1 is dominated by nested, small channel elements (CH), most of which are filled by lithofacies Ss. Some appear to have been filled by the mechanism of lateral accretion (LA).
appear to have also been filled by the mechanism of lateral accretion (architectural element LA). Channel elements occur at all stratigraphic levels in Facies 1 and are interbedded with element GB and sandy intervals of lithofacies Sp, which probably represent bar-formed elements SB and DA.

Facies 1 interpretation. Architectural elements and mesoform sedimentary structures in Facies 1 are similar to those in the underlying transition facies. As discussed in the transition facies, the sedimentological record of simple foreset bars, mid-channel compound bars, and numerous small channels indicates deposition by braided, bed-load streams. Particularly convincing evidence that braided streams deposited Facies 1 is the abundance of mesoform planar cross-stratified sandstone, pebbly sandstone, and sandy conglomerate, indicative of simple foreset bars (Figures 28) and avalanching on the downstream ends of longitudinal bars (Figure 29) (Smith, 1974; Hein and Walker, 1977). These structures are associated, both laterally and vertically, with horizontally stratified conglomerate, indicative of longitudinal bars. In some sections, conglomerate occurs in lenticular lithosomes, and is interpreted as channel lag deposits. The argument might be made that meandering rivers repeatedly migrated across the same area, and that the interbedded pebbly sandstone and conglomerate represent only the basal portions of coarse-grained point bars (Jackson, 1978). However, the braided river interpretation is
preferred based on the size and stratigraphic position of the planar cross-stratified sets. Large sets, typically 0.6 m thick, suggest deposition on the fronts of simple bars, rather than sandwave deposition along the sides of point bars. On coarse-grained point bars, large-scale Sp represents chute bar deposition, and as such it occurs at the top, not at the base, of the point bar sequence (McGowen and Garner, 1970; Jackson, 1976; Levey, 1978).

The alluvial architecture in Facies 1 supports the interpretation that braided rivers deposited it. Tabular bodies bounded by fourth- and fifth-order surfaces indicate high width to depth ratios typical of braided streams (Figure 31). Deposits of gravel bars, downstream accreting macroforms, sandy bedforms, and small channels are intricately interbedded, indicative of the variability of sub-environments in braided rivers. The presence of lateral accretion elements (Figure 33) is consistent with a braided interpretation. Lateral accretion occurs in braided rivers (Cant and Walker, 1978; Bluck, 1979; Allen, 1983; Ramos and Sopena, 1983; Bristow, 1987). The lateral accretion deposits in Figure 33 are very limited in lateral extent, relative to the thickness of the story they compose. Low "extent-to-thickness" ratios are characteristic of lateral accretion deposits in braided streams (Allen, 1983).

The abundance of foreset bar deposits at the base of Facies 1 indicates broad, shallow rivers like the Platte (Figure 34). Deposits of compound bars are more common in the upper part
Figure 34. Model of the broad, shallow, low sinuosity rivers that deposited Facies 1. Channels were filled with fields of large lingoid bars with avalanche faces on the downstream end. Not illustrated, but also present in the rivers that deposited Facies 1, were longitudinal bars, which accreted vertically and produced horizontally stratified conglomerate (from Miall, 1985, Figure 15).
of Facies 1, and suggest that later rivers were deeper, similar to those that deposited the transition facies (Figure 24).

**Facies 2 description.** In contrast to Facies 1, Facies 2 contains little to no gravel-size sediment. Scattered granules and small (1 cm) pebbles occur in the sandstone, but no conglomerate is present in this facies, with the exceptions of sections RV and GA (Plate 2) and UVA (Plate 3). Mean sandstone grain-size is medium to coarse. There is a notable increase in the abundance of intraclasts, which is most apparent in sections BV (Plate 1) and UV3 (Plate 3).

Facies 2 is dominated by trough cross-stratification (lithofacies St), in contrast to Facies 1, which has an abundance of planar cross-stratification (lithofacies Sp and Gp). St in Facies 2 is smaller than in Facies 1, and set thicknesses average 0.1 m and rarely exceed 0.2 m. Planar cross-stratified sandstone (Sp) also occurs at a smaller scale than in Facies 1. Sets rarely exceed 0.3 m, and average set thicknesses are 0.1 to 0.2 m. Large scale Sp (1.5 to 1.9 m thick) is rare, and at the three locations where it was observed, it occurs at the tops of stories (sections DM, H in Plate 1, and UV4 in Plate 3).

The architecture of Facies 2 is dominated by lateral accretion deposits (architectural element LA). This is discernable at individual measured sections, where lithofacies St is associated with inclined third-order bounding surfaces. Paleocurrent measurements from the St are oriented
perpendicular to the direction of dip of the bounding surfaces and thus confirm that these are lateral accretion deposits. LA elements are most clearly seen in photomosaics associated with cross-sections A-A' and B-B'.

In cross-section A-A', on outcrops near sections JC1 and PC (Plate 1) lateral accretion elements are clearly visible (Figures 35 and 36).

In cross-section B-B', LA elements are abundant and occur at all stratigraphic levels of the multistoried Facies 2, but are most completely preserved at the top (Figures 31, 37, 38, and 39). Figure 31 best illustrates the contrast between the architectures in Facies 1 and 2. Tabular architectural elements (GB, SB, DA) dominate in Facies 1 compared to lenticular LA elements in Facies 2.

Small channel elements are less common, but at sections UV 3 and BV they comprise nearly all of Facies 2 (Figure 40).

**Facies 2 interpretation.** Facies 2 is interpreted to have been deposited by meandering rivers. This interpretation is based on the dominance of lateral accretion elements, indicative of point bar deposition (Figure 41). In stratigraphic sections, this is manifested by a dominance of trough cross-stratified sandstone that is commonly associated with intraclasts. The abundance of intraclasts suggests the presence of cohesive banks at the time of deposition, a characteristic of meandering rivers. However, no overbank mudstones are preserved in
Figure 35. Lateral accretion element at base of the Calico bed near section PC. LA surfaces dip to north.

Figure 36. Lateral accretion element in center of photo, in middle of Calico bed. LA surfaces dip to the right (to the east). Location near section PC.
Figure 37. Photomosaic showing the architecture of Facies 2 at section UV2. Two distinct lateral accretion deposits are present and are separated by a fifth-order bounding surface. The upper LA element is more completely preserved than the lower. Arrow points to the fifth-order bounding surface between Facies 1 and 2 in the line of section. Facies 2 is 10 m thick.
Figure 38. Photomosaic showing the architecture of Facies 2 at section UV1. Lateral accretion deposits are abundant, but are relatively thin, suggesting only partial preservation of originally much thicker bar deposits. Arrow points to the fifth-order bounding surface between Facies 1 and 2 in the line of section. Facies 2 is 8.5 m thick.
Figure 39. Photomosaic showing the architecture of Facies 2 at a location 150 m east of section UV1. Lateral accretion elements comprise all of Facies 2. Arrow points to the fifth-order bounding surface between Facies 1 and 2.
Figure 40. Nested, small channel elements comprise all of Facies 2 at section BV; arrows point to channel bases. Top of cliff is the top of Facies 2 and the top of the Calico bed.
Figure 41. Model of rivers that deposited Facies 2 and 3; the classic sandy, meandering river. Channel deposits consist mostly of lateral accretion elements from point bar migration and overbank deposits are mudstone and siltstone (element OF) interbedded with crevasse splay deposits (element SB) (from Miall, 1985, Figure 12).
Facies 2, which occurs as a sheet sandstone complex of vertically and laterally coalesced channel-belt deposits.

At some locations, sedimentary structures in Facies 2 suggest deposition in a braided river. Thick stories of planar cross-stratification, such as at sections DM (Plate 1) and UV4 (Plate 3), may indicate deposition on mid-channel, compound bars as described by Haszeldine (1983) (although they could represent coarse-grained point bars; a large set of Sp at the top of each of these sequences suggests chute-bar deposition across the top of a point bar). Nested small channel elements at sections UV4 and BV (Figure 40) suggest deposition in a multi-channel river. It may be more conservative to conclude that Facies 2 was deposited by highly sinuous, multi-channeled rivers. This would explain the abundance of lateral accretion deposits as well as the typical braided-river characteristics. Another explanation could be that Facies 2 was deposited by both braided and meandering streams. Some modern rivers show a variable morphology, wherein some reaches are braided and some reaches are meandering (Leopold et al., 1964), such as the Izhma River in the U.S.S.R. (Figure 42), and as shown experimentally by Ouchi (1985).

While the exact morphology of the rivers which deposited Facies 2 is debatable, these rivers were certainly different from those which deposited Facies 1. Lateral accretion deposits indicate greater sinuosity of Facies 2 rivers. Although scattered pebbles are present in Facies 2, no conglomerate is present, and
Figure 42. Landsat infrared photograph of the Izhma River, USSR, which is braided in some reaches and meandering in others.
the mean grain-size of the sandstone is smaller than in Facies 1. Cross-stratification sets are smaller in Facies 2. These characteristics indicate that Facies 2 rivers were less competent and less energetic than those of Facies 1. Intraclasts indicate the presence of cohesive overbank areas at the time of deposition and suggest that Facies 2 rivers were mixed-load rivers, in contrast to the bed-load rivers of Facies 1.

**Facies 3 description.** Along Cross-section B-B' (Plate 3 and Figure 8) Facies 3 is distinguished from Facies 1 and 2 both texturally and architecturally. The sandstone is mostly fine-grained and significant intervals of mudstone comprise up to 40 percent of the section.

Sedimentary structures are dominated by horizontal stratification (Sh), ripple cross-lamination, and some climbing ripple cross-lamination (Sr) (Figure 43). Small trough cross-stratification (St) (0.1 to 0.15 m thick sets) also is present. Intraclasts are abundant, especially at the erosional bases of massive sandstone bodies (Figure 44).

Facies 3 is generally a slope-forming unit and commonly covered by talus, consequently bounding surfaces and architectural elements are difficult to define. Some inclined third-order bounding surfaces associated with lithofacies St were discernable at sections NC2 and UV2 (Plate 3), and indicate lateral accretion deposits. Lithofacies Sh and Sr comprise element SB. This interpretation is supported by the association
Figure 43. Ripple cross-lamination in Facies 3, at section UV1. Pencil points to sets of climbing ripple cross-lamination.

Figure 44. Abundant, large intraclasts of laminated overbank mudstone and claystone at base of a massive sandstone channel element in Facies 3, at section UVC. Hammer head is 15 cm long.
with architectural element OF, (overbank fines, indicated by the mudstone), which, by definition, is commonly interbedded with SB. Channel elements also occur in this interval, but, in contrast to those in Facies 2, may be encased in element OF (Figure 45).

Correlation of cross-section B-B' with C-C' (Figure 8) indicates that Facies 3 in B-B' lies at the same stratigraphic level as Facies 2 in C-C' and suggests that they were deposited contemporaneously. Interfingering of these units can be seen in outcrops along cross-section B-B'. The top of Facies 2 is not a single surface, but lies at different stratigraphic levels correlative with the tops of individual stories (Figure 46).

**Facies 3 interpretation.** Facies 3 is interpreted to have been deposited by meandering rivers. Although a few lateral accretion elements were found to support this, the interpretation is mainly based on the abundance of overbank deposits. Interbedded with floodplain mudstone (element OF) are tabular sandstone bodies (element SB) composed of horizontally bedded and ripple cross-laminated sandstone, which are interpreted as crevasse splay deposits (Figures 41 and 43). Lenticular sandstone units within OF are interpreted as channel deposits (Figure 45). The low width-to-depth ratio of these units and the abundant intraclasts within them (Figure 44) indicate deposition by meandering streams. Hettinger and others (1990) have suggested that these streams were tidally influenced based on their observations of sigmoidal bedding.
Figure 45. Lenticular channel element (CH) encased in overbank fines (OF) in Facies 3, at location east of section UV1 (see also Figure 39). Arrows mark bottom and top of Facies 3.

Figure 46. Interfingering between Facies 2 and 3 along outcrop between sections UV4 and UVC. Arrows point to different elevations of the top of Facies 2, coincident with the tops of separate stories. Large arrow points to the top of Facies 3.
geometries, wavy and lenticular bedding, multiple reactivation surfaces, and double mud drapes (Figure 47).

Sandstone within the channel deposits is even more fine-grained than in Facies 2, indicating that these meandering streams were less competent than those of Facies 2. The large amount of mudstone suggests that the rivers of Facies 3 were mixed- and suspended-load rivers.

**Calico bed summary.** The Calico bed records the change from deposition by highly competent, energetic rivers to deposition by less competent, less energetic rivers. This is indicated by decreasing grain sizes and decreasing cross-stratification set-thickness upward through the interval. Facies 1 was deposited by gravelly, bed-load braided rivers; Facies 2 was deposited by coarse-grained sandy, mixed-load meandering rivers; and Facies 3 was deposited by fine-grained sandy, mixed- and suspended-load, meandering rivers.

**Summary of fluvial depositional environments**

A gradual evolution of fluvial style is recorded in the Smoky Hollow Member. The barren zone was deposited by meandering rivers, followed by relatively deep, braided rivers (Figure 24). The Calico bed was deposited successively by shallow, braided rivers (Figure 34), deep braided rivers (Figure 24), and meandering rivers (Figure 41).
Figure 47. Trough cross-stratification (St) in Facies 3, at an outcrop between sections UV4 and UVC. St sets are 7 to 30 cm thick and comprise a 6 m thick story. Sigmoidal bedding geometries, multiple reactivation surfaces, and double-mud drapes have been interpreted by other workers as indicative of tidally influenced rivers.
Architectural-elements reflect fluvial style, but on a larger scale, fluvial architecture is also a function of preservation. Chapter 4 discusses the Smoky Hollow in terms of stratigraphic fluvial architecture and all of the controlling variables.

**Paleocurrents**

Paleocurrent measurements were taken on planar-tabular and trough cross-stratification in the transition facies and in Facies 1 and 2 of the Calico bed. Although fluvial morphology evolved over the period of time that these facies were deposited, paleocurrent measurements show that the rivers consistently flowed in the same direction, to the northeast.

In the petrography section, the transition facies was shown to be compositionally distinct from the Calico bed. Also, the Calico bed is distinctive because its base is characterized by the first occurrence of large pebbles in the Smoky Hollow Member. These differences between the transition facies and the Calico bed suggest that there might have been a change in the flow directions of the rivers which deposited them; a change which may have been responsible for the differences. To test this hypothesis, the paleocurrents of the transition facies were plotted separately from those in the Calico bed. There was no change in paleocurrent direction; rivers which deposited the transition facies flowed N57E and Calico rivers flowed N55E (Figure 48).
Regional Paleocurrents;
All measured sections

A. Transition facies and Calico bed;
Trough and Planar-tabular cross-stratification

B. Transition facies;
Trough and Planar-tabular cross-stratification

C. Calico bed;
Trough and Planar-tabular cross-stratification

Figure 48. Paleocurrent measurements from cross-bedding in transition facies and the Calico bed (total data) expressed as rose diagrams.
Paleocurrents were plotted by cross-section (Figures 49, 50, 51) and more significant differences in paleoflow exist between different regions of the study area than between the transition facies and the Calico bed. On the west side of the study area, in cross-section A-A' (Figure 49), rivers which deposited both the transition facies and the Calico bed flowed more strongly to the north than rivers in the east, as recorded from locations in cross-sections B-B' and C-C' (Figures 50 and 51). Vector means for paleocurrent measurements in the west are N38E, N25E, and N32E (Figure 49). In contrast, vector means in the eastern region are N65E, N62E, N50E, and S89E (Figure 50) and N73E and N65E (Figure 51).

The grand vector mean, N56E, indicates that rivers which flowed across the Kaiparowits Plateau were parts of drainage systems with headwaters in both the Sevier Orogenic Belt and the Mogollon Highlands (Figure 52). This paleocurrent pattern is consistent with sandstone compositions from the transition facies and Calico bed, as discussed in the petrography section.

The maximum clast size distribution (based on conglomerate in Facies 1) is not an accurate representation of the northeasterly paleoflow, because grain sizes increase to the west and northwest (Figure 53). Rather, the distribution of chert and quartzite pebbles and cobbles indicates proximity to, and sediment contribution from, the Sevier Orogenic Belt. The Mogollon Highlands were an igneous source terrane, and this less resistant composition resulted in smaller clasts being
Figure 49. Paleocurrent measurements from cross-bedding in transition facies (A) vs. the Calico bed (B and C) at measured sections in cross-section A-A' (Plate 1); expressed as rose diagrams.
Cross-section B-B’

A. Transition facies
Planar-tabular cross-stratification

B. Transition facies
Trough cross-stratification

C. Calico bed
Planar-tabular cross-stratification

D. Calico bed
Trough cross-stratification

N = 40
Vector Mean = 65
Circle = 33%

N = 48
Vector Mean = 62
Circle = 25%

N = 57
Vector mean = 50
Circle = 16%

N = 39
Vector mean = 91
Circle = 38%

Figure 50. Paleocurrent measurements from cross-bedding in transition facies (A and B) vs. the Calico bed (C and D) at measured sections in cross-section B-B’ (Plate 3); expressed as rose diagrams.
Cross-section C-C'

A. Transition facies
Planar-tabular cross-stratification

B. Calico bed
Planar-tabular cross-stratification

Figure 51. Paleocurrent measurements from cross-bedding in the transition facies (A) vs. the Calico bed (B) at measured sections in cross-section C-C' (Plate 4); expressed as rose diagrams.
Figure 52. Paleogeographic map for Late Turonian time, with fluvial paleocurrents indicated (based on this study, Ryer and McPhillips, 1983, and Franczyk, 1988). Rivers flowed 56° across the Kaiparowits area, with source areas in both the Sevier Orogenic Belt and in the Mogollon Highlands. Rivers flowing immediately to the north of the Kaiparowits area supplied sediment from the southwest to shorelines in the Castle Valley area of central Utah.
Figure 53. Maximum grain size from the Calico bed (Facies 1) plotted with rose diagrams of paleocurrent measurements from cross-bedding in the Calico bed. Small rose diagrams indicate local paleocurrent at measured sections and the large, central rose diagram indicates total paleocurrent data, from both the transition facies and Calico bed.
available for transport to the Kaiparowits region. No igneous gravel clasts were recovered from the study area; rather, the signature of Mogollon contribution is the feldspathic composition of sandstone in the Calico bed.
CHAPTER IV

FLUVIAL ARCHITECTURE

Introduction

Fluvial architecture addresses the geometry, internal arrangement, and relative distribution of channel and overbank deposits in fluvial sequences (Allen, 1978; Friend et al, 1979; Friend, 1983).

Sand bodies, which are produced by channel deposition, can be classified by their width/height ratios and by the internal relationships of individual channel deposits (stories). Sand bodies may be ribbons (width/height ratios < 15), narrow sheets (width/height ratios between 15 and 100), or wide sheets (width/height ratios > 100) (Blakey and Gubitosa, 1984).

Internally, sand bodies can be simple or complex. Simple sand bodies may have diastems, but not major scours, within them. Complex, or multistory, sand bodies contain at least one major scour-surface internally. Stories are bounded by surfaces of fourth-order or higher rank (Miall, 1988).

Computer simulation models by Allen (1978), Leeder (1978), and Bridge and Leeder (1979) have shown that sandstone/mudstone ratios, and sandstone body geometry and
interconnectedness, are controlled by channel type and magnitude, rates of lateral channel migration, avulsion and vertical floodplain accretion. These variables depend upon those climatic and geological factors that determine discharge magnitude and distribution, valley slope, and sediment supply. Those independent variables are climate, tectonics, and base level (Schumm, 1977; Leeder, 1978; Blakey and Gubitosa, 1984; Kraus and Middleton, 1987).

This chapter describes the fluvial architecture of the Smoky Hollow Member and interprets the independent variables which controlled the development of the architecture. In the following chapter, the Smoky Hollow architecture is considered in the context of regional stratigraphy.

Description

Sand bodies in the barren zone are mostly single- and multistory sheets, although immediately above the Tibbet Canyon Member are some multistory ribbon sand bodies (Figure 54). At the top of the barren zone, sandstone/mudstone ratios increase and the transition facies occurs as interconnected sheets. These sheets are multistoried, and average story thickness is 3.7 m.

The Calico bed is distinguished from the barren zone because Facies 1 and 2 are sheet sand bodies with practically no mudstone. Together, they form a multistory sheet which
Figure 54. Multistory, ribbon sand body in the lower Smoky Hollow Member. Arrows point to the bases of four distinct stories; large arrow points to the top of the Tibbet Canyon Member.
extends across the entire Kaiparowits Plateau. Average story thicknesses are 1.5 m in Facies 1 and 2.3 m in Facies 2. In Facies 2, stories are, in general, thicker towards the top, and this is reflected in the more complete preservation of lateral accretion deposits in the upper part of this facies (Figure 37).

Facies 3, which occurs only in the northeastern part of the study area, contains ribbon sand bodies within mudstone and siltstone (Figure 45) but also occurs locally as a multistory sheet. Story thicknesses were not averaged because of the lack of bounding surface data, and because not all sand bodies in this facies represent channel deposition as they do in the other facies. Significant amounts of overbank, crevasse-splay sandstone compose Facies 3 and thus the sand-body architecture is not directly comparable to that in the other facies.

**Interpretation**

Fluvial architecture is fundamentally controlled by sediment supply and the degree of preservation of that sediment. Climate and tectonics influence sediment supply; base level, which determines preservation, is influenced by tectonics and eustasy.

Climatic fluctuations have been documented in the Cretaceous. Studies by Barron and others (1985), Fischer and others (1987), and Kauffman and others (1987) link regular interbedding of limestone and shale that occur in deposits of the
central part of the Cretaceous epeiric sea (the Greenhorn and Niobrara limestones) to variations of climate in response to Milankovitch cycles (particularly the 20-25, 40-45, and 100-115 thousand year cycles; a 400ky cycle may also be recorded, but has not been documented). The Smoky Hollow Member represents roughly one million years of deposition (Figure 55) and therefore the presence of the Calico bed in the middle of the member could represent the 400ky cycle. However, if this climatic cycle was the controlling factor, it should be expressed throughout the Cretaceous Western Interior. The Calico bed, to the contrary, is unique to southern Utah; no such coarse-grained deposit is present in central Utah, suggesting that a more localized cause, such as tectonics, was the controlling factor. Deposits that reflect the 100-115ky cycle have been recorded in the Cretaceous Western Interior, and it seems reasonable that such climatic fluctuations would have affected the discharge and load of the rivers that deposited the Calico bed. However, because of the fluvial architecture of the Calico bed, time is represented in a lateral sense, across the Kaiparowits Plateau. Distinguishing cycles in fluvial deposits preserved in this way is virtually impossible, especially when combined with the inherent problem of distinguishing allocyclic (e.g. climatic) from autocyclic (e.g. avulsion) mechanisms.

Tectonics in a foreland basin exert control on subsidence rates and sediment supply. Thrusting and the resultant crustal loading cause rapid basin subsidence (Beaumont, 1981; Jordan,
Figure 55. Correlation diagram; eustatic sea level relative to regional stratigraphy.
1981). The thrust sheet that causes basin subsidence simultaneously erodes and contributes sediment that fills, partially or entirely, the space created by the subsidence.

Base-level is considered to be the "equilibrium surface" above which deposition of sediment is temporary and below which preservation is possible (Sloss, 1962). The space below base level, in which sediment may be preserved, has recently been referred to as "accommodation space" (Jervey, 1988). Base level and accommodation space are functions of tectonically controlled subsidence and eustatic changes in sea level (Leopold and Bull, 1979; Jervey, 1988; Ryer, 1991).

Because the effects of climate are indistinguishable, and because base level is a function of subsidence and eustasy, the two fundamental controls of Smoky Hollow architecture were 1) tectonics and 2) eustasy. These variables controlled the channel patterns of the rivers that deposited the Smoky Hollow Member and the differential preservation of the deposits.

**Channel Pattern**

Leeder (1978) and Galloway (1981) suggested that the proportions of coarse channel deposits and overbank fines are controlled by channel type. Interconnectedness of sand bodies increases progressively in deposits from suspended-load to mixed-load to bed-load channels because of the relative proportion of fines transported.
Sand-body geometry is also controlled by channel type. Braided rivers have a higher width/depth ratio than meandering rivers, which is reflected in higher width/height ratios of their channel deposits. Also, in contrast to meandering channels, braided channels are marked by greater instability which causes both short term channel migrations of a continuous nature and avulsions with a much shorter recurrence interval than meandering rivers. There is a greater tendency for braided channel deposits to overlap both laterally and vertically, thus forming sheets (Leeder, 1978).

In the Smoky Hollow Member, fluvial architectural style generally correlates to paleochannel type. The lower barren zone and Facies 3 were deposited by mixed- and suspended-load, meandering rivers, and both are characterized by ribbon sand bodies, low interconnectedness, and large amounts of overbank deposits. The bed-load, braided river deposits of the transition facies and Facies 1 consist of sand sheets, which are highly interconnected, and associated with little overbank fines.

The architecture of Facies 2 does not correlate well to paleochannel type. Deposited by meandering rivers, it forms a wide sand sheet, as does Facies 1. This is attributed to differential preservation of channel deposits and is discussed later.

Channel pattern changes are interpreted as having been driven by eustatic changes and variations in sediment supply. Deposition of the Smoky Hollow Member correlates to a period
of high magnitudes of sea-level change. Figure 55 shows that the Tibbet Canyon Member was deposited during falling sea level. Because the youngest fossil recovered from the Tibbet Canyon Member indicates the *Prionocyclus hyatti* (late Middle Turonian) ammonite zone, it is probable that the lower part of the Smoky Hollow Member was also deposited during this sea-level fall, which occurred through the *Coilopoceras colleti* ammonite zone. The Calico bed lies in the upper portion of the Smoky Hollow Member and thus correlates with a time of rising sea level (Figure 55). Rising sea level triggers increases in channel sinuosity, decreases in stream energy, and decreases in transported grain size (Schumm, 1977). The depositional sequence of Facies 1, 2, and then 3 in the Calico bed records all of these responses of the fluvial system to rising sea level. Examples of similar responses to sea-level rise, with vertical facies transitions like those in the Calico interval, have been documented by Nami and Leeder (1978), Gordon and Bridge (1987), Franczyk (1988), and Rust and Gibling (1990).

Braided river deposits in the middle of the Smoky Hollow Member indicate that an increase in sediment supply occurred. Braided rivers are associated with coarse-grained sediment loads and steep valley gradients; thus, they are usually considered to occur in the upstream reaches of river systems, proximal to the source area (Schumm, 1968; Schumm and Kahn, 1972, Schumm, 1977). The proximity of the source area to the study area did not change (the Sevier and Mogollon orogenic
areas were essentially stationary in location during the Turonian); therefore, increases in the quantity and coarseness of sediment transported to the Kaiparowits area caused the rivers to braid.

**Differential Preservation**

Different fluvial architectures in the Smoky Hollow Member are also ascribed to differential preservation, controlled by changing accommodation space.

Rising sea level increases accommodation space. Therefore, the change in fluvial architecture between Facies 2 and 3, specifically the increased proportion of overbank deposits, could be a function of increased preservation, as well as a function of changing fluvial style.

Low accommodation space is indicated by Facies 1 and 2 of the Calico bed. Together they form a sand sheet that extends across the entire Kaiparowits Plateau. A sheet of this scale could not have been produced by channel pattern alone; this would have required an active channel-belt width of 80 km. Rather, the wide sheet is indicative of extensive reworking and winnowing of overbank material resulting in preservation of only the channel deposits. The models of Allen (1978) and Bridge and Leeder (1979) suggest that broad sheet sandstones like the Calico bed were deposited during times of rapid lateral shifting with very low rates of aggradation.
Comparison of the internal architectures of the transition facies, Facies 1, and Facies 2 supports the interpretation that accommodation space was lowest during deposition of the lower Calico bed. If preservation of a channel deposit is complete, the thickness of a story should reflect the channel depth. The transition facies and the upper part of Facies 1 were deposited by the same kind of deep, braided river (Figure 24). Average story thickness in the transition facies is 3.7 m, compared to 1.5 m in Facies 1, indicating that accommodation space was greater during deposition of the transition facies. The low average story thickness of Facies 1 could be interpreted as a reflection of the shallow depth of the depositing rivers, however, the average in Facies 2, 2.3 m, supports the hypothesis that accommodation space was the control. Facies 2 was deposited by rivers at least as deep as those that deposited the transition facies. A complete lateral accretion deposit at the top of Facies 2 (Figure 37) and a complete downstream accreting macroform deposit in the transition facies (Figure 23) both indicate paleo-water depths of 6 m. That stories in Facies 2 average 1.4 m thinner than stories in transition facies indicates lower preservation in Facies 2.

Thus, the Calico bed (Facies 1 and 2) indicates low accommodation space and very slow rates of base-level rise. This could have been a function of reduced subsidence rates or a sea-level fall superimposed on foreland basin subsidence; regional evidence suggests that it was both. The Late Turonian Juana Lopez Member of the Carlile Shale in Colorado contains a
regional disconformity indicative of sea-level fall, which is dated by the ammonite Prionocyclus wyomingensis that occurs between the Prionocyclus macombi and Scaphites warreni ammonite zones (Figure 55) (Kauffman and Pratt, 1985). Thus, this eustatic fall seems likely to correlate with the lower Calico bed. However, it was a minor fluctuation of sea-level superimposed on the general sea-level rise of the Late Turonian; alone it cannot explain the architecture of the lower Calico bed. A reduced rate of subsidence in the foreland basin is considered the most significant factor in the low rates of aggradation and extensive lateral reworking indicated by the Calico bed. Other workers, such as Blakey and Gubitosa (1984) and Kraus and others (1982), have studied broad sandstone sheets which they attributed to periods of reduced subsidence rates. Regional correlations support this interpretation for the Calico bed and are discussed in the following chapter.
CHAPTER V

REGIONAL CORRELATIONS AND SEQUENCE STRATIGRAPHY

Introduction

Unraveling eustatic from tectonic controls of deposition requires a regional perspective. This chapter compares the stratigraphy of the Kaiparowits Plateau with those of Black Mesa in Arizona, Castle Valley in central Utah, the Henry Mountains region in south central Utah, and outcrops in southwestern Utah (Figures 1 and 55).

Recently, Shanley and McCabe (1989, 1991) and Gardner (1990, 1991) placed the Turonian-Coniacian strata of the Kaiparowits Plateau and Castle Valley in a sequence-stratigraphic framework. For purposes of comparison with their work, the results of this study are presented in terms of sequence stratigraphy. Sequences and their stratal components are interpreted to form in response to the interaction among the rates of eustasy, subsidence, and sediment supply (Van Wagoner et al, 1988; Posamentier and Vail, 1988), the same controls which govern fluvial architecture. Therefore, it is appropriate to interpret the fluvial architecture of the Smoky Hollow Member in terms of sequence stratigraphy.
Regional Correlations

Correlation of the Cretaceous rocks in the Kaiparowits Plateau, Black Mesa, and the Henry Mountains area has been previously discussed by Peterson and Kirk (1977) and Franczyk (1988). This study builds on those interpretations in greater detail and with some departures, and incorporates the Turonian-Coniacian rocks of Castle Valley (Figure 55).

The Tibbet Canyon Member was deposited by a regressive shoreline that prograded to the northeast across the Henry Mountains area, where it correlates with the lower unit of the Ferron Sandstone Member of the Mancos Shale (Peterson and Ryder, 1975). The lower sandstone member of the Toreva Formation, a deltaic-shoreline deposit (Figure 56) (Franczyk, 1988), and the lower Ferron Sandstone Member in Castle Valley, deposited in delta-front and shelf environments (Figure 57) (Ryer and McPhillips, 1983), formed during the same regional regression in Middle Turonian time. All of the above units contain fossils of the Prionocyclus hyatti faunal zone (Peterson and Ryder, 1975; Franczyk, 1983; Gardner, 1991). The lower Ferron in the Henry Mountains also contains juvenile specimens of Prionocyclus macombi (Peterson and Ryder, 1975).

Because the Smoky Hollow Member was deposited in continental environments and contains no age diagnostic fossils, it is correlated physically with rocks in Black Mesa and the Henry Mountains. In Black Mesa (Figure 56), the middle
Shoreline deposits are the lower sandstone member of the Toreva Formation and the Rough Rock Sandstone; the Mancos Shale consists of offshore marine deposits; coastal and alluvial plain deposits comprise the middle and upper members of the Toreva Formation and the Wepo Formation (based on Franczyk, 1988).
Figure 57. Stratigraphy of the Ferron Sandstone Member in Castle Valley, schematically shown. The lower Ferron consists of the lowest (un-numbered) shoreline deposits and sheetlike, shelf sandstone deposits named the Clawson and Washboard units. The upper Ferron consists of seven vertically-stacked, progradational, delta-front sandstone units and laterally equivalent delta-plain/alluvial plain deposits (in which the lettered units are coal) (courtesy of T. Ryer).
carbonaceous member of the Toreva Formation is composed of fluvial channel and overbank deposits and is analagous to the lower barren zone of the Smoky Hollow member. The fluvial, upper sandstone member of the Toreva Formation is composed of sheetlike sandstone bodies that form the lower part of the member, and ribbonlike sandstone bodies in the upper part of the member. The lower carbonaceous member of the Wepo Formation gradationally overlies the Toreva Formation and is composed of fluvial channel and overbank deposits. Thus, the architecture and depositional environments of the upper sandstone member of the Toreva Formation and the lower carbonaceous member of the Wepo Formation together resemble those of the Calico bed (Figure 5).

In the Henry Mountains area, shoreline deposits of the lower Ferron Sandstone Member are overlain by paludal and alluvial plain deposits of the upper Ferron (Peterson and Ryder, 1975). Sandstone composition of channel deposits at the top of the upper Ferron is feldspathic like that of the Calico bed (Figure 15), and suggests that these two units are correlative. The coarsest-grained channel deposits are pebbly and are thus correlated with Facies 1 of the Calico bed. Because the pebbly deposits occur at the top of the upper Ferron, the upper Ferron is correlated with only a portion of the Smoky Hollow Member: the coal and barren zones, and the lower part of the Calico bed. A subaerial unconformity at the top of the Ferron was reported by Peterson and Ryder (1975), and strata equivalent to the
upper part of the Calico bed was probably eroded during the formation of the unconformity.

In Castle Valley (Figure 57), the upper Ferron accumulated as a series of seven vertically-stacked, progradational, delta-front sandstone units and laterally equivalent delta-plain/alluvial plain deposits, and is correlated with the Calico bed. Faunal ages (from the Tibbet Canyon and John Henry Members) indicate the Calico bed was probably deposited during the Late Turonian; *Scaphites warreni* and *Scaphites ferronensis*, of the Late Turonian (Figure 55), were collected from the upper Ferron (Gardner, 1991).

The transgressive disconformity that separates the Smoky Hollow Member from the John Henry Member in the eastern part of the study area (Figure 5) is part of a regional transgressive surface found at the top of Smoky Hollow equivalent units in Castle Valley, Black Mesa, and the Henry Mountains. Marine deposits that overlie the disconformity in Castle Valley (the Bluegate Shale Member of the Mancos Shale) contain fossils indicative of the Turonian-Coniacian boundary (Ryer and McPhillips, 1983). This supports the correlation of the upper Ferron with the Calico bed, which is overlain by John Henry strata of the middle Lower Coniacian. Fossils collected from the Wind Rock Tongue of the Mancos Shale in Black Mesa indicate latest Turonian (Kirkland, 1990, p. 641). In the Henry Mountains, Upper Coniacian strata of the Bluegate Shale Member of the Mancos Shale overlie the unconformity. Molluscan faunas
in the Blue Gate Shale, originally reported as Santonian age by Peterson and Ryder (1975), were reexamined by E.G. Kauffman (pers. comm., 1990), and identified as belonging to the Upper Coniacian. Together with the correlation of the upper Ferron with the lower Calico bed, this indicates that erosion in the Henry Mountains area probably took place in the latest Turonian and Early to Middle Coniacian.

Above the transgressive disconformity, progradational shoreline deposits in the John Henry Member correlate with those of the Rough Rock Sandstone of the Mancos Shale (lower Middle Coniacian; Kirkland, 1990, p. 641) in Black Mesa. Temporally equivalent fluvial deposits of the John Henry Member on the west side of the Kaiparowits Plateau, thus correlate with the upper carbonaceous member of the Wepo Formation in Black Mesa (Figures 5 and 56).

The correlations of this study depart from the interpretations of Peterson and Kirk (1977) in the identification of unconformities. Peterson and Kirk (1977) proposed a regional, subaerially formed unconformity spanning the upper Turonian and Coniacian in the Kaiparowits Plateau, Black Mesa, and the Henry Mountains. The unconformity appears to exist in the Henry Mountains, but detailed sedimentology and new faunal data presented in this study and by Franczyk (1988) indicate that no such unconformity exists in the Kaiparowits Plateau or in Black Mesa. Deposition in these two areas was uninterrupted during the Late Turonian to Coniacian, although
rates of aggradation slowed during the time of deposition of the Calico bed and upper sandstone member of the Toreva Formation, as indicated by fluvial architecture (Figures 5 and 56).

In Black Mesa, Franczyk (1988) interpreted an unconformity of Middle to early Late Turonian age within the Toreva Formation. She documented great variation in the thickness of the middle carbonaceous member and observed that the upper sandstone member is thickest where the middle carbonaceous member is thin or absent (Figures 55 and 56). This unconformity does not exist in the other regions correlated in Figure 55, but it may correlate with an unconformity in the proximal foreland basin deposits of southwestern Utah.

Moir (1974) conducted a comprehensive study of the Cretaceous rocks in the southwestern corner of Utah (Figure 1) and discussed correlation of these rocks with those of the Kaiparowits Plateau. He suggested that the 'Middle' fluvial facies of his study represents a thickened Smoky Hollow Member, and that the 'Upper' fluvial facies is an upper floodplain equivalent of the lower part of the John Henry Member (Moir, 1974, p. 248). The 'Middle'-'Upper' fluvial facies contact is an unconformity, and the base of the 'Upper' fluvial facies is marked by a thin conglomeratic horizon that is recognizable in varying degrees of prominence throughout the study area. West of the Hurricane fault (Figure 2) in the Iron Springs Formation, the Cretaceous rocks are uniformly fluvial in
origin and no natural subdivisions exist, with the exception of
the basal conglomerate of the 'Upper' fluvial facies. In the most
thrust-proximal positions of the study area, the conglomerate is
a thick piedmont fan deposit (Moir, 1974, p. 58).

The conglomerate at the base of the 'Upper' fluvial facies is
inferred to correlate with the Calico bed. Two sections were
examined in the eastern part of Moir's study area, the
Markagunt Plateau, one along the North Fork of the Virgin River
(section NFV) and the other north of the town of Glendale
(section NG) (Figure 2).

Correlation with the Calico bed was based on stratigraphic
position and lithology. At NFV and NG, the conglomerate
examined is the first coarse-grained deposit above a thick
section of mudstone and fine-grained sandstone overlying
regressive littoral sandstone units equivalent to the Tibbet
Canyon Member (Moir, 1974, p. 247) (Figure 58). At NFV
(Figure 59) the 9 m thick sandy conglomerate is under- and
overlain by 9 m thick units of sandstone; thick intervals of
mudstone lie above and below. Average maximum clast size of
the cobbles is 6.3 cm. At NG (Figure 60), the 3 m thick sandy
conglomerate directly overlies mudstone, and is overlain by 7.5
m of sandstone and pebbly sandstone; mudstone lies above the
sandstone. Average maximum clast size of the pebbles and
cobbles is 4.1 cm. The grain-sizes at NFV and NG, when
compared with those of the Calico bed in the Kaiparowits Plateau
(Figure 53), are indicative of upstream increases in grain-size,
Figure 58. Regressive, littoral sandstone units (marked by arrows) in the Markagunt Plateau which are equivalent to the Tibbet Canyon Member View to the north, of Cogswell point in T39S, R9W (see Figure 2).

Figure 59. Outcrop of Calico bed-equivalent unit in the Markagunt Plateau at section NFV. The gray, slope-forming unit in the middle of the outcrop is the most conglomeratic.
Figure 60. Outcrop of Calico bed - equivalent unit in the Markagunt Plateau at section NG. (a) The unit (bracketed by arrows) is over- and underlain by thick sections of mudstone and fine-grained sandstone. (b) Interbedded conglomerate and sandstone at base of unit; hammer for scale.
suggesting that the Markagunt Plateau conglomerates are the upstream equivalents of the Calico bed. Composition of the pebbles and cobbles is similar to that of the conglomerate in the Calico bed, and includes quartzite (white, red, and gray), grayish, fossiliferous chert, black and banded cherts, and some sandstone. In addition, some cobbles are composed of sedimentary-lithic, pebble conglomerate.

**Sequence Stratigraphy**

Correlation of strata in the Kaiparowits Plateau with neighboring Turonian-Coniacian strata makes speculation of eustatic and tectonic controls more convincing. This is presented in the context of sequence stratigraphy, as defined by Vail and others (1977) and revised by Van Wagoner and others (1988).

Sequence stratigraphy has been defined as "the study of rock relationships within a chronostratigraphic framework of repetitive, genetically related strata bounded by surfaces of erosion or non-deposition, or their correlative conformities" (Van Wagoner et al., 1988, p. 40).

Two types of sequence boundaries have been defined. A type 1 sequence boundary is characterized by "subaerial exposure and erosion associated with stream rejuvenation, a basinward shift of facies, a downward shift in coastal onlap, and onlap of overlying strata" (Van Wagoner et al, 1988, p. 41). A type 1 boundary forms when the rate of eustatic fall exceeds the
rate of basin subsidence at the depositional-shoreline break, producing a relative fall in sea level at that position. A type 2 sequence boundary is marked by "subaerial exposure and a downward shift in coastal onlap landward of the depositional-shoreline break; however, it lacks both subaerial erosion associated with stream rejuvenation and a basinward shift in facies" (Van Wagoner et al, 1988, p. 42). A type 2 boundary forms when the rate of eustatic fall is less than the rate of basin subsidence at the depositional-shoreline break, so that no relative fall in sea level occurs at the shoreline position.

A sequence can be subdivided into systems tracts, which are associations of contemporaneous depositional systems (Brown and Fisher, 1977). Within a sequence, lowstand, transgressive, and highstand systems tracts are deposited sequentially during a cycle of relative sea-level change. The actual time of initiation of a particular systems tract is believed to be a function of the interaction among eustasy, sediment supply, and tectonics (Van Wagoner et al, 1988).

Interpretation: Eustatic and Tectonic Sequences
The sequence stratigraphy of the Smoky Hollow Member is presented in the larger context of deposition from the base of the Tibbet Canyon Member through the John Henry Member and is related to the correlative stratigraphy in the Henry Mountains Black Mesa, and Castle Valley. The definitions of sequence boundaries seem to stress the importance of eustacy as a
control. However, because the formation of sequences is also influenced by subsidence and sediment supply, it seems reasonable to identify tectonically controlled sequences.

The stratigraphy of the four regions previously correlated (Figure 55) consists of a large scale, eustatically driven sequence that is overprinted in the middle by a smaller sequence, driven mostly by tectonics but also influenced by a small fluctuation in sea level.

**Eustatic sequence.** The lowstand systems tract of a eustatically driven sequence is represented by the Tibbet Canyon Member, the shoreline deposits of the lower Ferron Sandstone Member in the Henry Mountains, the lower Ferron Sandstone Member in Castle Valley, and the lower sandstone member of the Toreva Formation in Black Mesa. Gardner (1991) referred to this as the "Hyatti sequence", as all of these units contain *Prionocyclus hyatti* or age equivalent molluscs. The base of this sequence is a type 2 boundary, as all of the above units interfinger with underlying marine shale.

**Calico sequence.** The Calico sequence was initiated by the coincidence of tectonics and a small-scale eustatic fall, and was overprinted in the upper part by rising sea-level of the larger scale eustatic sequence. In Black Mesa, the base is a type 1 sequence boundary, but in all other areas the base is gradational with underlying facies and is a type 2 boundary. The lowstand
systems tract (in response to increased sediment supply and reduced basin subsidence, as well as to sea-level fall) is represented by Facies 1 of the Calico bed, the upper unit of the Ferron Sandstone Member in the Henry Mountains region, the lower two progradational shoreline deposits of the upper Ferron Sandstone Member in Castle Valley (Gardner, 1991), and the lower part of the upper sandstone member of the Toreva Formation.

The transgressive systems tract (in response to eustatic rise in late Turonian time), is represented by Facies 2 and 3 of the Calico bed, the upper five progradational shoreline deposits in the upper Ferron in Castle Valley (Gardner, 1991), and the upper part of the upper sandstone member of the Toreva Formation and the lower carbonaceous member of the Wepo Formation in Black Mesa. Local uplift and erosion during late Turonian and Coniacian time removed the transgressive systems tract in the Henry Mountains area.

Rates of sea-level rise in the Coniacian exceeded sediment supply as indicated by major transgressive disconformities in all four regions. Marine deposits of the basal John Henry Member and correlative fluvial deposits, the basal Bluegate Shale Member of the Mancos Shale in Castle Valley and in the Henry Mountains area, and the Wind Rock Tongue of the Mancos Shale, the Rough Rock Sandstone and the upper carbonaceous member of the Wepo Formation in Black Mesa, all represent the youngest
parts of the transgressive systems tract and the highstand systems tract.

Discussion

This interpretation is different from that put forth by Shanley and McCabe (1989, 1991) and Hettinger, McCabe, and Shanley (1990). They interpreted the Tibbet Canyon Member as recording progradation of the late highstand systems tract. They identified "coarse, laterally amalgamated fluvial clastics of the Smoky Hollow Member" (the Calico bed) as "sharply" overlying the Tibbet Canyon Member, and interpreted this to "represent an abrupt seaward shift in facies tracts due to increased rates of sea-level fall" (Shanley and McCabe, 1989, p. 410). The base of the Calico bed was characterized as a major third-order, type 1, sequence boundary unconformity (Hettinger, et al., 1990; Shanley and McCabe, 1991). The lower Ferron Sandstone Member in Castle Valley was interpreted to be the resulting lowstand system tract.

The interpretation of this study differs from that of Shanley and McCabe (1989, 1991) by correlating the lower portion of the Ferron Sandstone Member in Castle Valley with the Tibbet Canyon Member as part of the same lowstand systems tract. The Tibbet Canyon Member and the lower Ferron were deposited during the late middle Turonian and the progradation of these units appears to have occurred in response to a large-scale sea-level fall, which occurred at that
time (Figure 55). That this regression is recognized throughout the Western Interior of the U.S is good evidence that the regression was forced by sea-level fall. Some examples of regressive units which are part of this lowstand systems tract are the Oyster Ridge Sandstone Member of the Frontier Formation in north-central Utah and southwestern Wyoming, and the Codell Sandstone in central and northern Colorado (Ryer, 1991).

In contrast to the interpretation of Shanley and McCabe (1989, 1991), there is no evidence for a type 1 sequence boundary at the base of the Calico bed in the northern Kaiparowits Plateau. There is no displacement of facies tracts in the Smoky Hollow Member; the deposits represent a complete transition from paludal to meandering to braided river depositional environments in the lower part of the member. More importantly, as argued in the stratigraphy chapter, there is no evidence of an unconformity at the base of the Calico bed. The base of the Calico bed lies at different stratigraphic levels, which is part of the evidence that the Calico interfingers with the underlying, and adjacent, finer-grained deposits of the Smoky Hollow. Although these different stratigraphic positions could represent different elevations along the base of an incised paleovalley associated with a type 1 unconformity, this possibility is refuted by the facies architecture in the Calico bed and by the stratigraphy of the Ferron Sandstone Member in the Henry Mountains area.
Facies patterns within the Calico bed do not resemble typical paleovalley-fill sequences. Coarse-grained fluvial facies are typically restricted to the central, deepest parts of incised drainages and are overlain by progressively finer-grained facies, which lap out against the paleovalley walls (e.g. Fisk, 1944, Plates 4-8; Curry, W.H., 1985; Wilkinson and Byrne, 1977; Morton and Price, 1987, p. 193; Suter et al., 1987). For example, regarding deposits in the paleovalley of the Mississippi River, Fisk (1944, p. 17) noted that "the upper surface of the graveliferous deposits is fairly uniform; the principal variation in thickness of the graveliferous unit results from irregularities in the depth of the entrenched valley". If the base of the Calico bed represented the outline of a paleovalley, the "graveliferous" facies, Facies 1, should be restricted to the deepest part of the valley. In reality, Facies 1 occurs at various stratigraphic levels; the upper surface is not uniform (Figure 8).

Shanley and McCabe (1989; 1991, Figure 3) interpreted the lower portion of the Ferron Sandstone Member in Castle Valley as part of the lowstand system tract, correlative with the formation of paleovalleys at the base of the Calico bed; however, Calico rivers did not flow towards Castle Valley. Rather, paleocurrent data from the Calico bed indicate that the rivers flowed towards, and across the Henry Mountains region (Figure 52). Paleoflow to the east is confirmed by the paleocurrent data for the Calico bed presented by Shanley and McCabe (1991, Fig. 7).
Depths of valley entrenchment due to sea-level fall are greatest at the shoreline and gradually decrease landward (Schumm, 1977; Morton and Price, 1987; Butcher, 1990). Consequently, the depth of an incised surface at the base of the Calico bed should increase in a seaward direction, to the northeast, and thus towards the Henry Mountains. Detailed work on the Ferron Sandstone Member in the Henry Mountains region (Peterson and Ryder, 1975; Peterson et al., 1980) indicates that deposition within the member was continuous and uninterrupted; there is no evidence of an incised paleovalley. Middle and Upper Turonian age shoreface deposits are progressively overlain by deposits of paludal and alluvial plain environments, and there is no truncation of the shoreface deposits. The lack of an incised paleovalley in a seaward position supports the interpretation that no incision occurred in the Kaiparowits Plateau as a result of sea-level fall.

Although the evidence indicates that the base of the Calico bed is not a type 1 sequence boundary, it is reasonable to identify this surface as a type 2 sequence boundary, which could have formed in response to a eustatic fall. This type of sequence boundary should be expected on the active sides of foreland basins, where high rates of subsidence counter the effects of falling sea-level on sedimentation.

Sequence stratigraphy in a foreland basin. An inherent problem in applying sequence stratigraphy in a foreland basin
setting is that the concepts were originally developed for deposits of passive margins. In foreland basins, the basic geometry of the basin is reversed with respect to source area; sediment supply is from the side with maximum subsidence. Jordan and Flemings (1991) demonstrated through stratigraphic modelling that the location of the sediment source relative to the subsidence geometry is of critical importance to sequence development. In their model, subaerially eroded sequence boundaries did not form in foreland basins in response to eustatic changes.

In the Cretaceous Western Interior foreland basin, Nummedal and Swift (1987) suggested that sequence-bounding unconformities would develop on the cratonic eastern side of the basin in response to sea-level fall. These might grade into conformities within the clastic wedges of the rapidly subsiding western margin.

Comparison of the strata of Black Mesa with that of the Kaiparowits Plateau illustrates the variable nature of the same sequence boundary in different parts of the foreland basin. Franczyk (1988) interpreted the unconformity in the Toreva Formation and the overlying deposits as indicative of paleovalley formation during an erosional event and gradual infilling of the erosional topography accompanied by a regional decrease in gradient, reflected by the upsection changes in lithology and architecture (Figure 56)
Thus, a type 1 sequence boundary is present in Black Mesa at the base of the unit correlative with the Calico bed. Black Mesa is more distal in the foreland basin than the Kaiparowits Plateau (Figure 52), and probably experienced lower rates of subsidence. Stratigraphic thicknesses indicate different rates of subsidence between the two areas. The Smoky Hollow member at section UV2 (Plate 2) is 68 m thick and the correlative interval in Black Mesa at Chilchinbito Canyon (Franczyk, 1988), which includes the middle carbonaceous and upper sandstone members of the Toreva Formation and the lower carbonaceous member of the Wepo Formation, is 46 m thick. The greater thickness in the Kaiparowits Plateau is interpreted to be the result of higher rates of subsidence.

Thus, the lower part of the Calico bed and the upper sandstone member of the Toreva Formation fit the model proposed by Nummedal and Swift (1987). The sequence boundary at the base of these units is a type 1 unconformity in the basinward position of Black Mesa and it grades into a conformable surface, a type 2 boundary, in the more proximal foreland basin position of the Kaiparowits Plateau. This sequence boundary appears to correlate with a minor sea-level fall in the early Late Turonian that is recorded in deposits in Colorado (discussed in the previous chapter).

There is evidence, however, that this sequence was influenced by tectonic forces, governing sediment supply and subsidence, which were then overprinted in the upper part by
eustatic rise. Evidence that tectonism played a major role in
Turonian deposition is present in the study area, Black Mesa,
Castle Valley, and in foredeep deposits in southwestern Utah.
Grainsize, composition, and quantity of sediment supplied all
indicate that sediment supply to the foreland basin increased.
Architecture of the Calico bed and an associated unconformity to
the southwest suggest that reduced foreland basin subsidence,
more than sea-level fall, influenced accommodation space.

The presence of coarse conglomerate at the base of the
Calico bed suggests that tectonics played a role in its deposition;
eustasy alone is not sufficient to explain the presence of this
conglomerate, even if the base of the Calico bed were a type 1
sequence boundary. In the event of a eustatically induced base-
level fall, a nickpoint develops on the river profile (Butcher,
1990). The nickpoint represents the point below which the
profile is adjusting to the newest base level and above which the
profile is not adjusting (Begin et al, 1980; Gardner, 1983). Below
the nickpoint, incision occurs and fluvial gradients increase;
above the nickpoint, incision may or may not occur as the
river responds independently of the most recent base-level fall
(Schumm et al., 1987; Butcher, 1990).

Theoretically, the nickpoint should migrate to the head of
the profile (Mackin, 1948; Begin et al, 1981); however, recent
work has shown that nickpoints migrate only part of the
distance up the profile (Leopold and Bull, 1979). Additionally,
flume experiments have shown that there is a significant lag
time between eustatic lowstand and transmittal of the impulse of incision upstream (Wood et al., 1990). Based on this information, it is inferred that Turonian rivers responding to a minor eustatic fall would not have incised headward far enough to tap the coarse conglomerate being deposited in the proximal foreland basin. Furthermore, rates of subsidence in a foreland basin increase towards the thrust front. As the rivers incised headward, the effects of eustatic fall would probably have been counteracted by the high rates of subsidence and incision would have ceased.

The abundant feldspathic and subfeldspathic sediment supplied to the Kaiparowits Plateau area during deposition of the lower Calico bed is considered evidence of a tectonic influence. As discussed in the petrography section, higher feldspar-to-sedimentary lithic ratios of sandstones in the Calico bed, as contrasted with the transition facies, indicate an increased contribution of sediment from the Mogollon Highlands. During Late Turonian time, the first phases of compressional tectonism and uplift occurred in this source area (Hayes, 1970; Nydegger, 1982; Drewes, 1982). Abundant arkosic to subarkosic sediment was supplied to the Black Mesa area during deposition of the upper sandstone member of the Toreva Formation and was interpreted by Franczyk (1988) to indicate the onset of this uplift. A gradient increase associated with the uplift was speculated to have produced channel incisement, and thus the unconformity at the base of the upper sandstone member.
In central Utah, a tectonically caused increase in the rate of sediment input to the basin during Late Turonian time has been used to explain the upper part of the Ferron Sandstone Member (Ryer, 1991) (Figure 57). As discussed earlier, lowering of sea level during Middle Turonian time brought about an initial regression, resulting in deposition of the lower Ferron Sandstone. Rising sea level during Late Turonian time brought on a short-lived transgression whose marine shale deposits separate the lower and upper parts of the Ferron Sandstone. The upper Ferron accumulated as a series of river-dominated deltas which prograded basinward even as sea level continued to rise (Ryer and McPhillips, 1983; Ryer, 1983 and 1991). That such a progradation occurred and that the position of the upper Ferron shoreline prograded beyond that of the lower Ferron (Figure 57) (Ryer and McPhillips, 1983) is considered strong evidence that sediment supply increased greatly in Late Turonian time.

**Tectonic models.** An increase in sediment supply to the Kaiparowits Plateau during Calico deposition can be documented, but relating this sedimentation to tectonics is speculative. The deposition of a rapidly prograding clastic wedge in a foreland basin has traditionally been correlated with an episode of deformation. Thus, Armstrong and Oriel (1965) and Wiltchko and Dorr (1983) dated the times of movement of the Sevier thrust belt from the ages of conglomerates that were shed into
the adjacent foreland basin. Specific to this study, Peterson and Kirk (1977) correlated the Calico bed with the upper sandstone member of the Toreva Formation and inferred that deposition of these units coincided with the time of greatest uplift and erosion in the source region.

A more recent concept is that, during uplift, subsidence is assumed to be so rapid that coarse sediment is trapped adjacent to the thrust belt while fine-grained sediment is deposited in the rest of the basin; during quiescence, subsidence is slow and coarse strata prograde into the distal basin (Beck and Vondra, 1985; Blair and Bilodeau, 1988; Heller et al., 1988). Because the Kaiparowits Plateau lies east of the foreland basin axis, this model better explains deposition of the Calico bed.

The validity of this model for the Cretaceous Interior Foreland Basin relies on regional correlations. Comparison of strata from the proximal foreland basin in central and southwestern Utah with the Ferron Sandstone and Calico bed, respectively, provides evidence that coarse-grained progradation in the distal foreland basin occurred during tectonic quiescence in the hinterland.

In central Utah, Villien and Kligfield (1986) dated six major thrusting events using paleontologic data from synorogenic deposits at the toes of the thrusts (Figure 61). Synorogenic conglomerates, in the Indianola Group, are restricted to the thrust fronts, and are labeled A through F in Figure 58. It can be seen that a thrusting event occurred in the
Figure 61. Tectonostratigraphic framework of central Utah. The thrust belt in central Utah can be divided geometrically into four major thrust systems: the Canyon Range, Pavant, Gunnison, and Wasatch thrust systems, whose times of movement are indicated on the left side of the figure. Synorogenic conglomerates, in the Indianola Group, are restricted to the thrust fronts, and are labeled A through F. The wavy lines indicate the inferred latest movements of a thrusting event (from Villien and Kligfield, 1986).
Albian-Cenomanian, but that no thrusting occurred at the time of Ferron progradation in the Turonian.

The progradation of the lower Ferron in the Middle Turonian was most probably a response to eustatic lowering, but the progradation of the upper Ferron occurred even as sea-level rose (Figures 55 and 57). Because no thrusting event correlates with the latter progradation, the increased sediment supply at that time is interpreted as a function of decreased rates of foreland basin subsidence. Sediment no longer trapped in the rapidly subsiding foredeep could be transported to more distant positions in the basin. Late Turonian paleogeography of central Utah (Figure 52) (Ryer and McPhillips, 1983) shows that the upper Ferron progradated to the northeast, a change from southeasterly progradation of the lower Ferron. Although rivers that deposited the Calico bed in the Kaiparowits Plateau did not flow directly towards Castle Valley, the north-northeasterly paleoflow directions measured in the western Kaiparowits Plateau (Figure 49) suggest that rivers flowing immediately to the north of the Kaiparowits area supplied sediment from the southwest to the upper Ferron shorelines (Figure 52).

In southwestern Utah, stratigraphy in the proximal foreland basin (as documented by Moir (1974)) indicates that a more extreme version of the model, proposed by Heller and others (1988), may be applicable to deposition of the Calico bed. Their model suggests that, during quiescence, removal of the thrust by erosion results in flexural rebound of the thrust belt.
and proximal foreland basin. A regional unconformity then develops in the proximal part of the foreland basin and coarse-grained sediments derived from the eroding thrust sheet and reworked sediments from the proximal part of the foreland basin are deposited in the distal foreland basin (Figure 62).

Within the framework of this model, the proximal fan deposits in the Iron Springs Formation west of the Hurricane Fault (Figure 2) represent the synorogenic sediment trapped adjacent to the thrust belt during thrusting by the associated rapid rates of basin subsidence. The unconformity between the 'Middle' and 'Upper' fluvial facies is interpreted to represent the period of erosion resulting from flexural rebound of the thrust belt and proximal foreland basin. The thin conglomerate, which correlates with the Calico bed, that overlies this unconformity represents a period of sediment bypass as coarse-grained sediments derived from the eroding thrust sheet and reworked sediments from the proximal foreland basin were deposited in the more distal foreland basin. This interpretation is supported by the composition of the conglomerate in the Markagunt Plateau (Figure 2). There, cobbles composed of sedimentary-lithic, pebble conglomerate indicate reworked proximal fan deposits.

The architecture of the Calico bed is also consistent with the second phase of the tectono-stratigraphic model. As discussed in the fluvial architecture chapter, the broad, sheet-
Figure 62. Depositional models for synorogenic (A) and postorogenic (B) phases of foreland-basin evolution (from Heller et al., 1988, Figure 1)
like geometry of the Calico bed could be interpreted to indicate deposition during a period of reduced subsidence.

Although the two-phase model discussed is the preferred interpretation for the Calico bed, it is recognized that one-to-one correlation of stratigraphy in the basin with tectonics is probably not a realistic expectation. The Kaiparowits region was subject to the tectonic effects of two thrust belts, the Sevier and the Mogollon, of differing structural styles and oriented at opposing angles (Figure 52). One stratigraphic record records the influence of two tectonic belts, which may, at times, have acted out of phase with one another.

Summary

Two scales of eustatic change influenced the deposition of the Calico bed. The sheetlike geometry of Facies 1 was influenced by a small-scale drop in sea-level, which reduced accommodation space. This small-scale drop is recorded most clearly in the Juana Lopez Member in central Colorado (Kauffman and Pratt, 1985) and by the unconformity at the base of the upper sandstone member of the Toreva Formation in Black Mesa (Figure 55). Sea-level rise of a larger-scale eustatic cycle is recorded by the architecture of and paleochannel patterns reflected in Facies 2 and 3.

Foreland basin tectonics were the primary driving mechanism of the Calico sequence. Low rates of foreland basin subsidence allowed an increase in sediment supply to the distal
foreland basin. Such an increase is most clearly recorded by the progradation of the Upper Ferron Sandstone Member in Castle Valley and by the coarse grain size of the lower Calico bed. At the same time, low rates of subsidence reduced accommodation space. This is reflected in Facies 1 and 2 and in the upper sandstone member of the Toreva Formation in Black Mesa, which have in common a sheetlike geometry, devoid of mudstone.

Eustasy and tectonics thus worked in concert in the early Late Turonian to produce the Calico sequence.
CHAPTER VI

CONCLUSIONS

1. The Calico bed interfingers with under- and overlying fluvial strata in the study area. In the eastern part of the study area, it is separated from overlying marine strata of the John Henry Member by a transgressive disconformity; however, there is no evidence of significant erosion and removal of Calico bed deposits as in the southern Kaiparowits Plateau. This is attributed to higher rates of subsidence in the northern Kaiparowits Plateau.

2. The Calico bed is subdivided into three lithofacies based on grain size, and a Calico-like facies, the transition facies, is identified in the barren zone. The transition facies was deposited by relatively deep, braided rivers, Facies 1 by shallow then deep braided rivers, and Facies 2 and 3 by mixed- then suspended-load rivers. These facies reflect the influence of tectonics and eustasy on the deposition of the Calico bed.

3. The evidence of tectonics are increased sediment supply and the fluvial architecture of Facies 1 and 2. The coarse size of the conglomerate in Facies 1 and the increased proportion of feldspar in Facies 1 sandstone, relative to that in the transition facies, indicates that the rate of sediment supply
increased from both the Sevier Orogenic Belt and the Mogollon Highlands. Facies 1 and 2 together form a broad, multistoried sheet across the Kaiparowits Plateau; this is interpreted as a function of extremely low rates of subsidence at the time of deposition. Together, these lines of evidence suggest that, during a phase of post-thrusting quiescence, subsidence in the foreland basin slowed and coarse-grained sediment from the eroding thrust belt and reworked sediments from the proximal part of the foreland basin were deposited in the distal foreland basin.

4. The evidence of eustasy is the correlation of the sequence of facies in the Calico bed with regional sea-level changes. Because the Calico bed lies in the upper half of the Smoky Hollow Member, whose age of deposition is bracketed as between the late Middle Turonian and Early Coniacian, the Calico bed is inferred to have been deposited in the Late Turonian. A minor eustatic fall in the Late Turonian contributed to the low accommodation space reflected by the architecture of Facies 1. Eustatic sea-level rise in the latest Turonian triggered increases in channel sinuosity, decreases in stream energy, and decreases in transported grain size, which were recorded in the sequence of Facies 1, 2, and 3. Different fluvial architectures in the three facies indicate increased rates of vertical accretion through time; thus, rising sea-level increased accommodation space.

5. Through regional correlation, the Calico bed can be described and interpreted in terms of sequence stratigraphy.
The base of the Calico bed is the type 2 sequence boundary of a sequence initiated largely by tectonics, but also by a minor sea-level fall; Facies 1 is part of the lowstand systems tract of this sequence. Facies 2 and 3 represent the transgressive systems tract of a larger-scale, eustatically controlled sequence, the base of which lies in the Middle Turonian.
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