CRETACEOUS AND EARLY TERTIARY GEOLOGY OF CEDAR AND PAROWAN CANYONS, WESTERN MARKAGUNT PLATEAU, UTAH UTAH GEOLOGICAL ASSOCIATION FIELD TRIP ROAD LOG SEPTEMBER, 2001

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INTRODUCTION

Cedar and Parowan Canyons are incised into the western margin of the Markagunt Plateau (figure 1) at the transition between the Basin and Range and the Colorado Plateau. This area was also transitional in the Cretaceous when sediments were being shed from the Sevier orogenic belt a short distance to the west and epicontinental seaways were flooding in from the east. The region underwent high rates of subsidence during much of the Late Cretaceous (Cenomanian-Campanian?) with renewed subsidence in the Paleocene and Eocene. The two canyons described in this field log are only separated by 12 miles (20 km) and represent the remarkable complexity of this region in the Cretaceous and early Tertiary.

Initial geologic studies in these canyons were undertaken by Gregory (1944, 1950) establishing a basic framework for further study. Averitt (1962), in a study of coal resources, furthered our understanding of the problems and complexities of working in the Cedar Canyon area. A significant development in regional research resulted from Gustason's (1989) study of the Dakota Formation. Field trips to this area led by Gus Gustason and his major professor, Erle Kauffman (then at the University of Colorado, Boulder), have resulted in a fascination with the region by several of Kauffman's students (four of whom are authors on this paper) and many of them have in turn introduced their students and others to the complex geology of the region. This, coupled with the U.S. Geological Survey's BARCO (Basin and Range to Colorado Plateau Transition Study) project, has rejuvenated research in the region. Many presently doing research in the area have made contributions to this field guide. Other workers, whose contributions to this field guide would have been significant, were unable to contribute (e.g. Timothy White, who at the time of manuscript preparation is on a leg of the Ocean Drilling Project, but see his dissertation (1999) which focuses on geochemistry of fine-grained lower to mid-Turonian marine sediments in the region; and Florian Maldonado, who has mapped extensively in the region, but current obligations at the U.S. Geological Survey precluded his participation).

The work on these canyons is a work in progress as this paper clearly demonstrates. As there is not always agreement between the authors on all points presented in this paper, an effort is made to identify which authors contributed which data and interpretations.

STRATIGRAPHY

Cedar Canyon

Dakota Formation

Averitt (1962) abandoned the use of the Dakota Formation in the Cedar Canyon area and assigned the lower 1,075 feet (349 m) of Cretaceous rocks in the region to the Tropic Shale (from the base of the section through the "Upper Culver Coal Zone"). However, Gustason (1989) and all subsequent workers have assigned most of this section to the Dakota Formation. The upper boundary of the Dakota has been variously placed at the top (Elder, et. al., 1994; Kirkland; Tibert, see figure 2) or bottom (Tibert et al., in press) of a sandstone overlying a distinctive marl (base of unit 10, figure 2). This places the boundary at the base or top of a sandstone of lower Turonian age (containing the inoceramid *Mytiloides puebloensis* collected by Tibert, identified by Cobban; see Kennedy et al., 2000).

The stratigraphic succession of the Dakota Formation into the lowermost part of the Straight Cliffs Formation (figure 2) records long-term base-level rise, which was punctuated by shorter-term base-level changes (relationships between the stratigraphic patterns observed and base-level changes are discussed in Laurin and Sageman, this volume). The long-term change in accommodation is recorded by superimposition of shallow-marine clastics of the uppermost part of the Dakota Formation (and the overlying Straight Cliffs Formation; units 8 -11, figure 2) over coastal deposits of the lower part of the upper member of the Dakota Formation (units 1-7, figure 2), which in turn overlie estuarine to fluvial deposits of the middle and lower Members of the Dakota Formation. The short-term base-level cycles are represented by packages of depositional units (figure 2) bounded by surfaces of fluvial incision or subaerial exposure whose formation can be attributed to base-level falls (see Laurin and Sageman, this volume).

The Dakota Formation is Cenomanian in age in Ce-

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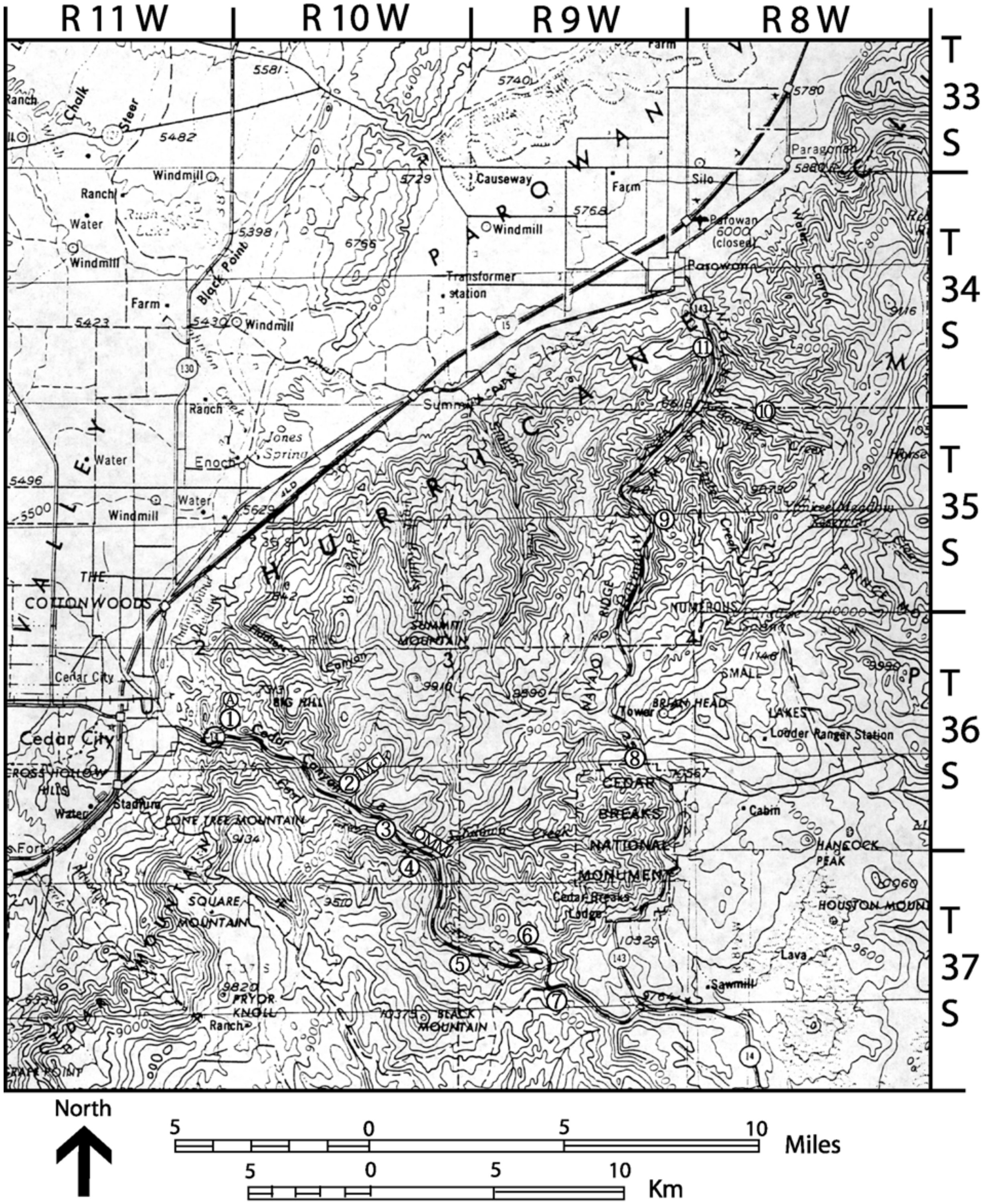

Figure 1. Map of the Cedar-Parowan Canyons area showing position of stops and sections of Laurin and Sageman (figure 2: A = section A; MC = Maple Canyon; OMM = Old MacFarlane Mine). Based on U.S.G.S. Cedar City 1:250,000 topographic map NJ 12-7, 1971 (revised).

dar Canyon if the upper boundary is placed at the base of the sandstone overlying the "marl" (base of unit 10, figure 2), but if the lithostratigraphic boundary is placed above the sandstone, the Dakota Formation ranges into the lower Turonian in this area unlike the Dakota to the east in the Kaiparowits region where the formation is entirely Cenomanian in age. The middle nonmarine member of the Dakota Formation yielded its first fossil during the 2000 field season and pro-

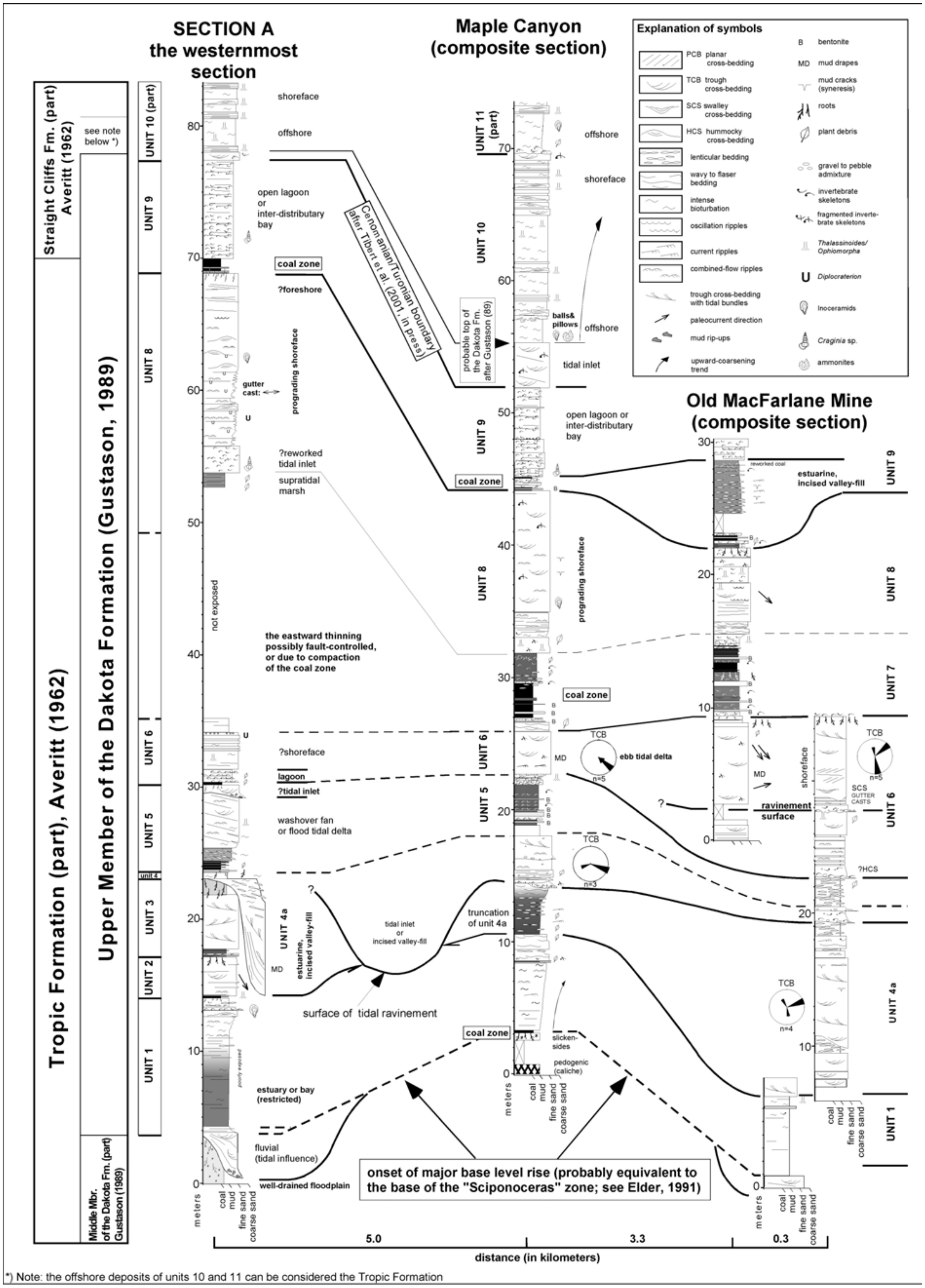

Figure 2. Sedimentary structures and correlation of units of the upper member of the Dakota Formation and the lowermost part of the Straight Cliffs Formation of three sections in Cedar Canyon; see figure 1 for location of sections. Interpretation of depositional environments is indicated. Units1 to 11 refer to informal stratigraphic units defined by Laurin and Sageman (this volume). Proposed position of the sections in the regional stratigraphic context is indicated (cf. Elder et al., 1994). Formal lithostratigraphy after Averitt (1962) and Gustason (1989).

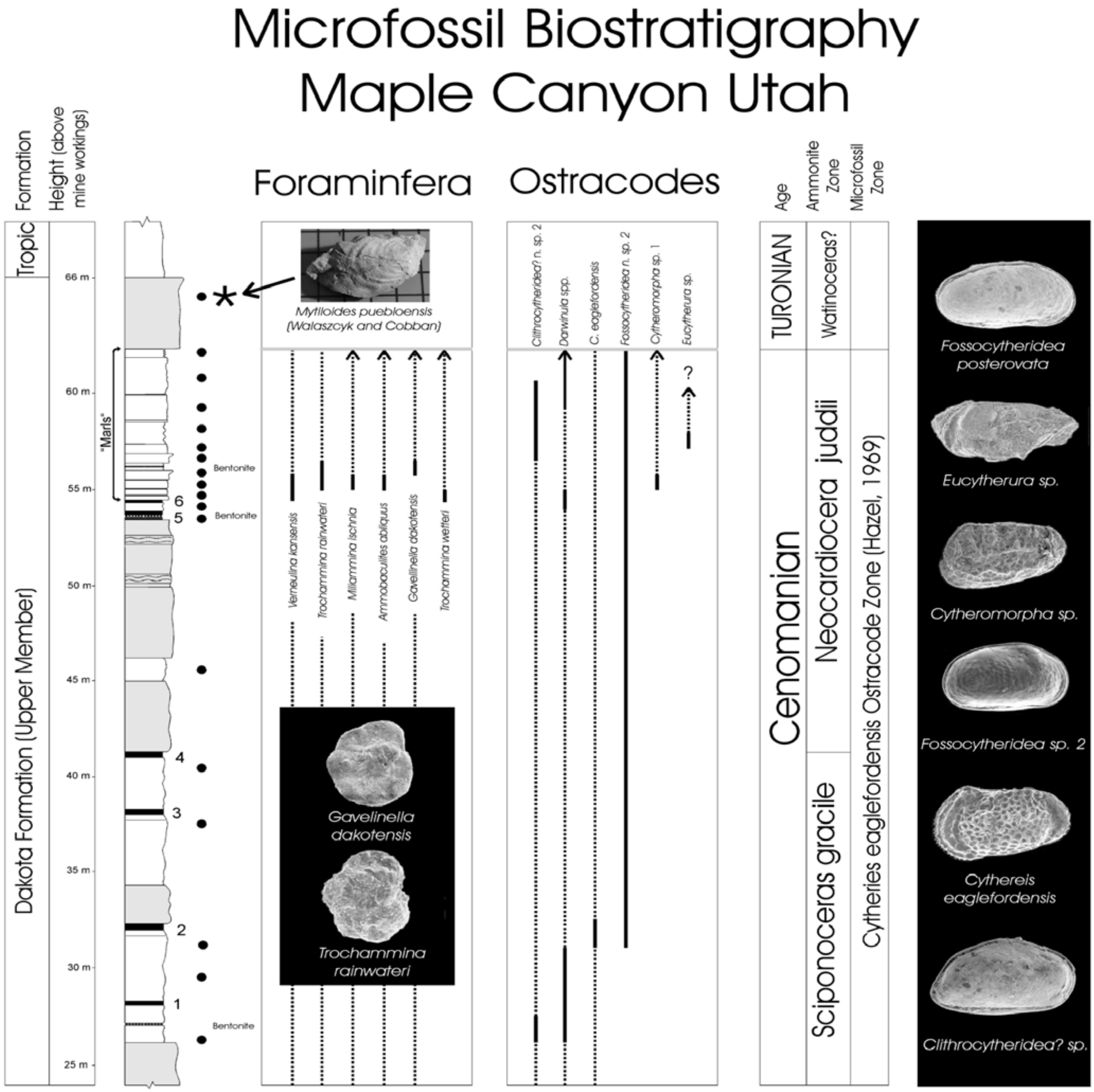

Figure 3. Microfossil biostratigraphy of the Dakota Formation in Maple Canyon (modified from Tibert et al., in press). Solid lines indicate actual taxon occurrences. Dotted lines indicate that the taxon has been identified within stratigraphic equivalents at nearby localities (Table Bench, Old MacFarlane Mine, and Bigwater). Note the coal seams are labeled 1-6 and the presence of bentonites (smectite-rich with minor kaolonite claystone beds 1-3 cm thick, clay analyses by Bob Newton, Smith College) are indicated next to the column. Solid black circles indicate beds sampled. The *Sciponoceras gracile* ammonite zone is characterized by the First Occurrence (FO) of *Cythereis eaglefordensis* and *Clithrocytheridea*? n. sp. 2. The *Neocardioceras juddii* ammonite zone is characterized by an abundance of *Fossocytheridea* n. sp. 2 and the FO of *Cytheromorpha* n. sp. 1 and *Eucytherura*. In addition, the abundance of *Trochammina rainwateri*, the FO of *T. wetteri*, and the presence of *Gavelinella dakotensis* mark the approximate position of bentonite TT2 (of Leithold, 1994, equivalent to Bentonite B of Elder, 1988, 1991) as observed in prodeltaic strata in the Kaiparowits Plateau region. Note that we place the base of the *Sciponoceras gracile-Neocardioceras juddii* boundary below the medial sandstone unit (as indicated by Elder et al., 1994). The bivalve *Mytiloides puebloensis* (Walaszczyk and Cobban) marks the first lower Turonian unit. Finally, *Fossocytheridea posterovata* (Lankford *in* Peterson et al., 1953) was identified at approximately 246 feet (75 m) above the uppermost *Neocardioceras* marlstone within the undifferentiated Straight Cliffs Formation.

duced vertebrates including fish and crocodilian teeth, as well as ostracodes (Eaton, research in progress). In a sandstone below the marl (unit 8, figure 2), *Inoceramus ginterensis* has been found. An abundant brackish-water molluscan fauna has been recovered from the marl (unit 9, figure 2) including *Craginia coalvillensis* (Meek), *Levicеritherium funicula* (Meek), *Admentopsis* n. sp. B (not *A. gregorium* (Meek) (White, 1877)), *Crassostrea soleniscus* (Meek) (not *Ostrea uniformis* (Meek) (White, 1877)), and *Veloritina securis* (Meek). See the discussion below by Kirkland on brackish-water molluscan faunas.

Until recently, the foraminifera and ostracodes contained within the mudrocks of the Dakota Formation and the Tropic Shale on the western Markagunt Plateau have gone unstudied. Tibert et al. (in press) provide the first report of microfossil populations and this road log contains a brief summary of their findings. Detailed descriptions of the taxa are currently in progress (Tibert, Leckie, and Colin, manuscript in preparation); therefore, the proposed names of the new taxa remain in open nomenclature. Tibert, Leckie and others have used foraminifera and ostracodes to interpret the biostratigraphy of the upper member of the Dakota Formation at Maple Canyon (STOP 2) (figures 3 and 12). They consider the lower 127 feet (41.5 m) of the member to represent the *Sciponoceras gracile* biozone based on the occurrence of the ostracode *Cythereis eaglefordensis*. They assign the rest of the upper member of the Dakota Formation (with the exception of the uppermost sandstone) to the *Neocardioceras juddii* biozone based on an association of agglutinated foraminifera (see figure 3 and STOP 2).

Stops 1, 2, and 4 of this field trip are designated to demonstrate basic facies assemblages and corresponding depositional environments of the upper member of the Dakota Formation and the lowermost part of the Straight Cliffs Formation, as well as the concept of depositional cyclicity outlined above. The regional stratigraphic context of the Cedar Canyon succession is outlined in Laurin and Sageman (this volume; their figure 5). STOP 2 will specifically deal with the biozones recognized by Tibert and Leckie on the basis of foraminifera and ostracodes.

Tropic Shale

With the placement of most of Averitt's (1962) Tropic Shale into the Dakota Formation, the Tropic Shale virtually disappears as a mappable unit. This is reflected in Laurin and Sageman's interpretation of sections (figure 2) in which the Dakota Formation is directly overlain by the Straight Cliffs Formation. Kirkland, Tibert, Leckie and Cobban prefer to use the term Tropic Shale for a thin sequence between the sandstone that overlies unit 9 (figure 2) and the base of the prominent cliff-forming sandstones of the Straight Cliffs Formation. This interval includes about 30 feet (10 m) of dark olive-gray sandy mudstone and muddy sandstone. Although it is questionable if this interval is mappable, it makes a significant marker unit in the region. The sequence represents normal marine deposits and includes lower Turonian taxa such as *Mytiloides kossmati*, as well as *Watinoceras* sp. (possibly *W. coloradoense*) and *Fagesia catinus*.

Straight Cliffs Formation - Tibbet Canyon Member

Above the Tropic Shale or uppermost Dakota Formation, depending on terminology employed, is an approximately 600 foot (190 m) thick succession of cliff-forming oyster-bearing sandstones. The sequence was considered by Averitt (1962) to be equivalent to the entire Straight Cliffs Formation of the type area (Kaiparowits Plateau region); however, most subsequent workers (e.g. Eaton, 1999; Moore and Straub, this volume) consider these cliff forming sandstones to be equivalent to the Tibbet Canyon Member. This is an overall regressive sequence but it records, as the result of rapid subsidence, minor eustatic fluctuations in remarkable detail. The equivalent sandstone ranges between 65-173 feet (21-56 m) thick in the type area (Peterson, 1969). Abundant fossils of the oyster *Crassostrea soleniscus* are found at many levels, often in biostromal accumulations. Other brackish-water taxa recovered from the unit are *Craginia whitfieldi* (White), *Brachiodonte* sp., *Barbatia* sp., and *Carycorbula* sp. Some of the sandstones also preserve open marine taxa including "*Inoceramus labiatus*" (probably = *Mytiloides mytiloides* and *M. subhercynicus*), indicating an age of late early Turonian. In the upper part of the Tibbet Canyon Member, *Inoceramus cuvieri* and *Collignoniceras woollgari* are present indicating middle Turonian (Kirkland, specimens at the Museum of Northern Arizona). Also reported from the Tibbet Canyon Member by Cobban (this paper) are *Mytiloides kossmati*, *M. mytiloides*, *Collignoniceras woollgari*, and *Pholadomya coloradoensis.*

Comments on Cenomanian-Turonian Brackish Water Mollusks (Kirkland)

The "Wasatch Line" extends from near Coalville, Utah, in the north, to Cedar Canyon and the Pine Valley Mountains to the south, and is represented by a thick sequence of marginal-marine strata deposited near peak sea-level rise on the western margin of a rapidly subsiding foreland basin. Many brackish-water molluscan taxa were named from these strata during the latter third of the 19th century (Meek, 1873; White, 1877, 1879; Stanton, 1893), but were often described from poorly preserved material at poorly documented sites. Though evolutionarily conservative, brackish-water mollusks generally display a great deal of morphologic variation, resulting in a number of previously described species being lumped together (Stanton, 1893). Therefore, little systematic research on these faunas was undertaken except for comparing the Utah taxa to fossils described from similar-age strata to the south in Texas and Arizona (Stephenson, 1952; Fursich and Kirkland, 1986; Kirkland 1996). Recently these brackish-water taxa have been found to be useful as a proxy for determining ancient substrate conditions, as well as paleoturbidity and paleosalinity gradients (Fursich and Kirkland 1986; Fursich, 1994). Field research in southwestern Utah has revealed that abundant, well-preserved molluscan fossils commonly characterize brackish-water strata in this area. Distinct species can be recognized at different stratigraphic levels, indicating the potential for some of these brackish-water taxa to have local biostratigraphic utility.

To test their utility, collections of brackish-water fossils have been made at many stratigraphic levels at different localities by Kirkland. These sites are tied into the standard ammonite biostratigraphy established for the Cretaceous Western

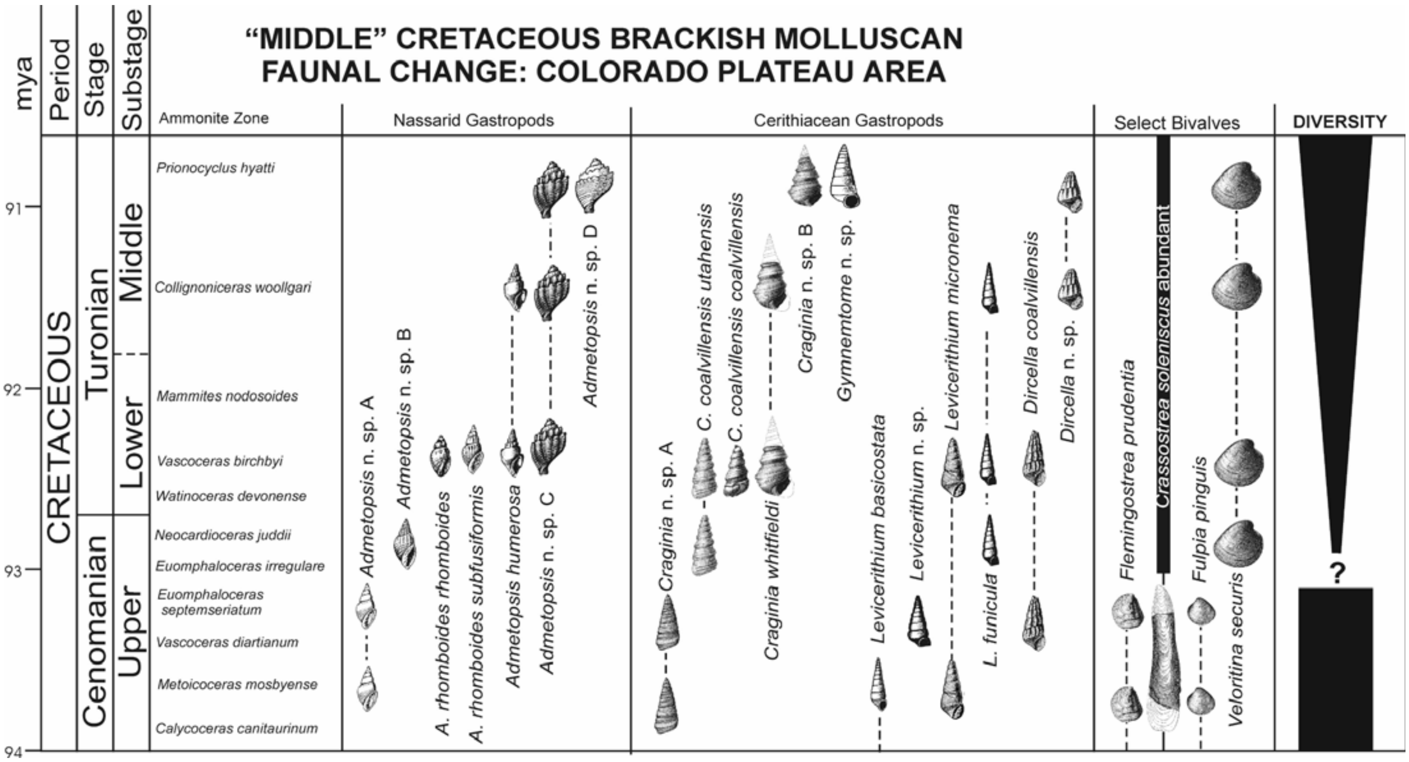

Figure 4. Ranges of brackish-water molluscan taxa plotted against standard ammonite zones. Mollusk illustrations from White (1877, 1879), Stanton (1893), and Brad Wolverton, as modified by Kirkland (Kirkland, research in progress).

Interior Seaway through intertonguing and onlap and offlap relationships between the brackish-water facies and marine, ammonite-bearing strata. The type collections of the original taxa from the 19th century were borrowed from the Smithsonian Institution and the type localities have been revisited and fossils collected when feasible. Many type specimens were originally collected from the now closed coal mines near Coalville (the "Wasatch" coal sites) in the lower Turonian Coalville Member of the Frontier Formation and are no longer accessible (Meek, 1873; Ryer, 1975, 1977).

Results of these studies indicate that several brackish-water species needlessly had been lumped together, and that a number of undescribed forms historically lumped with these species represent new undescribed species. While the systematic research of these collections continues, the preliminary results are clear: a number of brackish-water molluscan lineages reveal speciation rates much higher than would be considered typical of brackish-water taxa. Gastropods of the *Craginia coalvillensis*-*C. whitfieldi* (Mennissier, 1984; Cleevely and Morris, 1988) and *Admetopsis rhomboides* lineages appear to have the most potential for biostratigraphy, but other gastropods and a number of bivalves also appear to have potential (Kirkland, research in progress). Using all of these taxa, two “zones” per substage may be discerned. Although it is unlikely that this system will have utility outside the Colorado Plateau region, it appears to work well in that region. It is often difficult to correlate the lithologically similar stratigraphic succession of marginal marine Cretaceous strata along the “Wasatch Line” due to poor exposures and structural complexity. The brackish-water taxa found in strata along the “Wasatch Line” (belt) provide a new tool on which to base these correlations (figure 4).

Straight Cliffs Formation - Smoky Hollow Member

Overlying the cliff-forming sandstones of the Tibbet Canyon Member is a series of fine-grained, organic-rich, brackish-water deposits. Averitt (1962) assigned 987 feet (320 m) of strata overlying the Tibbet Canyon Member to the Wahweap Sandstone. His measured section of the "Wahweap Sandstone" includes frequent occurrences of oysters and coal beds. Neither oysters or coal have been observed in the type area of the Wahweap Formation (Kaiparowits region), and the entire unit is considered nonmarine in origin (Eaton, 1991). At least the organic-rich brackish deposits overlying the cliff-forming sandstones can be correlated both lithostratigraphically and paleontologically to the brackish coal-bearing parts of the lower part of the Smoky Hollow Member in its type area. The brackish horizons produce abundant benthic foraminifera, ostracodes, the sawfish *Ptychotrygon*, and a molluscan brackish-water fauna consisting of *Craginia whitfieldi* (White) (the type locality for this species is from this horizon on the Markagunt Plateau), *Levicerithium funicula* (Meek), *Admetopsis* n. sp. C, *Dircella* n. sp. (= *Chemnitzia* sp. of White (1879) and Stanton (1893)), and *Veloritina securis* (Meek) (see discussion of brackish-water faunas above). Many of these same brackish-water mollusks are found in the Smoky Hollow Member in the Kaiparowits region. Brackish-water deposits are 177 feet (54 m) thick in the area (Eaton, Diem et al., 1999).

The top of the Smoky Hollow Member is indistinct in this area. There is a sandy conglomeratic unit 350 feet (107 m) above the base of the Smoky Hollow beds that may be equivalent to the Calico Bed that marks the top of the Smoky Hollow Member in the Kaiparowits Region (see Moore and Straub, this volume). The conglomerate is exposed in a stream cut along the west side of Utah Highway 14. Eaton has not been able to locate the unit while measuring sections elsewhere, but vegetative cover is extensive in the area making the measurement of complete sections impossible.

The Smoky Hollow Member will be examined at STOP 5.

Straight Cliffs Formation - "Upper"

The uncertainty of the uppermost boundary of the Smoky Hollow Member, as well as uncertain lithostratigraphic correlation (see Moore and Straub, this volume), makes equivalence of the remainder of the Straight Cliffs Formation in Cedar Canyon uncertain. Much of the section consists of variegated floodplain mudstones and meandering river sandstones that are probably upland equivalents of the John Henry Member in the type area. These deposits are better drained and lack the coal and organic-rich mudstones of the Kaiparowits region.

A vertebrate locality 793 feet (222 m) above the top of the Tibbet Canyon Member contains abundant freshwater sharks (rare in localities above and below this site), which are most common in streams during episodes of transgression (Eaton et al., 1997) and suggest this may represent the Coniacian transgression recorded in the type area of the John Henry Member (Eaton, 1991; Eaton et al., 1999). A euhedral biotite bed present above this locality has produced an $^{40}Ar/^{39}Ar$ age of 86.72 ± 0.58 Ma (Eaton, Maldonado, and McIntosh, 1999) very near the Coniacian-Santonian boundary (85.8 in Gradstein et al., 1994). This strongly supports a John Henry Member equivalency for this part of the section.

A conglomerate (STOP 6, figure 15) is present 1,157 feet (375 m) above the top of the Tibbet Canyon Member (Eaton, Diem et al., 1999). In the section of Moore and Straub (this volume), the same conglomerate occurs 1408 feet (457 m) above the Tibbet Canyon Member. Their section used calculated true thickness while the section in Eaton, Diem et al., 1999, was measured with Jacob staff. These differences highlight the problems in working in heavily vegetated areas. Moore and Straub (this volume) consider this conglomerate (as does Edward Sable, USGS, Denver, retired, pers. comm., 1994) to possibly represent the uppermost member of the Straight Cliffs Formation, the Drip Tank Member. This may well be the case as a vertebrate fauna recovered 65 feet (21 m) above the conglomerate contains a mammalian fauna similar to that recovered from the Wahweap Formation in its type area (Eaton, Diem et al., 1999). However, it must be cautioned that the faunal study was preliminary and at that time there was little to compare to from the John Henry Member in its type area. Eaton is currently processing sediment from a productive locality near the type section of the John Henry Member and plans a comparative study of the faunas from the John Henry Member and that recovered above the conglomerate in Cedar Canyon. Until that study is complete, Eaton remains cautious about the correlation of this conglomerate. The conglomerate will be examined at STOP 6.

Wahweap (?) Formation

There is another 895 feet (290 m) of variegated muds and meandering river sandstones above the possible Drip Tank equivalent conglomerate. These beds look very much like those that underlie the conglomerate, and no clear lithologic differentiation of these sequences is evident. Also, this sequence resembles neither the Wahweap or the Kaiparowits Formations either in gross appearance or petrology (Moore and Straub, this volume).

The Wahweap Formation in its type area correlates to the Masuk Formation in the Henry Mountains region, and based on palynomorphs (identified by Douglas Nichols, USGS, Denver) the Masuk Formation is Campanian in age (Peterson and Kirk, 1977). Also, latest Santonian-early Campanian mollusks (*Scaphites* sp., *Baculites asper* Morton, *Baculites ovatus* Say or *B.* sp. smooth) were recovered from the Mancos Shale 300 feet (100 m) below the Masuk Formation (Eaton, 1990). This strongly suggests a Campanian age for the Masuk Formation and the lateral equivalent Wahweap Formation (Eaton, 1990). However, Nichols (1997) reports no palynomorphs younger than Santonian from the Wahweap(?) Formation of Cedar Canyon. If this is the case, it is possible that the so-called Drip Tank conglomerate is Cedar Canyon in actually a local conglomerate and the entire upper sequence is equivalent to the John Henry Member.

It must be pointed out that the preliminary review of the vertebrate fauna from just above the possible Drip Tank equivalent beds (Utah Museum of Natural History - UMNH VP Locality 10) correlated best to faunas known from the Wahweap Formation, but as stated earlier in this paper relatively little is known of vertebrate faunas from the John Henry Member (Eaton, Diem et al., 1999). Further confusing the issue is that these faunas (both those from Cedar Canyon [Eaton, Diem et al., 1999] and the Wahweap Formation [Eaton, Cifelli et al., 1999] in the type region) were compared to those recovered from localities in the Milk River Formation of Canada that have generally been considered to be of early Campanian age. However, several workers (e.g. Leahy and Lerbeckmo, 1995) think these localities may actually be upper Santonian. Another locality at the very top of the Wahweap(?) equivalent sequence (UMNH VP Locality 11) contains a marsupial molar of an unnamed genus and species almost identical to a molar collected from the Wahweap Formation in Bryce Canyon National Park as well as another unnamed marsupial taxon known from the Wahweap Formation. As such, there are some contradictory biostratigraphic correlations made from palynomorph assemblages and vertebrate faunas.

The base of the Wahweap(?) Formation will be examined at STOP 6 and the uppermost part at STOP 7.

"White Sandstone"

At the top of the sequence there is a distinctive 250 foot (76 m) thick sandstone that is principally a quartz arenite. The sandstone was assigned to the Kaiparowits Formation by Gregory (1950). Moore and Straub (this volume) also use the term Kaiparowits(?), although they recognize that the unit does not appear to be lithologically similar to the typical "salt and pepper" sandstones of the Kaiparowits Formation. This same sandstone was correlated by Goldstrand (1991, 1992) to the middle sandstone member of the Grand Castle Formation (of Goldstrand and Mullett, 1997) based both on petrology and lateral walking of the unit to its type area. Pollock (1999) assigned the sandstone to the capping member of the Wahweap Formation (terminology of Eaton, 1991) based on petrologic comparisons. No biostratigraphically datable materials have been recovered from the sandstone.

The term "white sandstone" is used here, although this unit most likely represents the middle sandstone member of the Grand Castle Formation reflecting that the most detailed studies of the unit were undertaken by Goldstrand (1991, 1992).

The "white sandstone" is overlain by the upper Paleocene - middle(?) Eocene Claron Formation. The white sandstone will be examined at Stop 7.

Parowan Canyon

Iron Springs Formation

The oldest rocks in Parowan Canyon represent approximately the upper third (1,123 feet; 364 m) of the Iron Springs Formation. Although the base of the Iron Springs Formation is not present in Parowan Canyon, the next major canyon to the south (Summit Canyon) contains the lower part of the formation which includes an oyster-rich interval that probably represents the early Turonian transgression in the area. Above this brackish interval is a several-meter-thick conglomerate containing drab colored chert pebbles; this the only conglomerate observed in this area (although outcrops in many places are poor). Leckie recovered a specimen of *Mytiloides kossmati* in the Iron Springs Formation in Parowan Gap (west of Parowan Canyon) indicating the presence of normal marine lower Turonian deposits. In general, the Iron Springs Formation in Parowan Canyon consists of thick tabular sets of sandstone with thinner overbank mudstones, probably representing fine-grained braid plain deposits. No detailed study of the sediments of the Iron Springs Formation has been undertaken in this area and the closest study area is well to the southwest near the town of Gunlock (Fillmore, 1991).

Only recently have fossils been recovered from the Iron Springs Formation (Eaton, 1999). Near the base of the section exposed in Parowan Canyon, a locality (UMNH VP Locality 6) contains abundant remains of a turtle associated with brackish water, cf. *Naomichelys* sp., in a black, organic-rich mudstone. This same turtle is commonly found in the John Henry Member of the Straight Cliffs Formation (Santonian part) and is at least suggestive of correlation to that member. A blue biotite ash (similar to that in Cedar Canyon) 410 feet (133 m) above the base of the canyon floor and 712 feet (231 m) below the top of the section yielded an $^{40}Ar/^{39}Ar$ age of 83.0±1.1 Ma (Eaton, Maldonado, and McIntosh, 1999), close to the Santonian-Campanian boundary (83.5 Ma in Gradstein et al., 1994). A locality (UMNH VP Locality 64) is present at the very top of the section has produced vertebrates (under study by Eaton) and ostracodes (under study by Elizabeth Brouwers, USGS, Denver). Other localities have been found in the uppermost part of the Iron Springs Formation in the next two canyons to the north (Red Creek and Little Creek canyons), which have produced vertebrates and ostracodes as well as palynomorphs (currently under study by Nina Baghai-Riding at Delta State University). It appears likely that with further study of these faunas and palynofloras, an upper limit for the age of the Iron Springs Formation in the region will be attained.

The lower part of this section (still within the upper part of the Iron Springs Formation) will be examined at STOP 11.

Grand Castle Formation

The Grand Castle Formation consists of about 590 feet (194 m) of conglomerates and sandstones of Paleocene age (Goldstrand, 1991, 1992; Goldstrand and Mullet, 1997). The formation is divided into three members: a lower conglomeratic member, a middle sandstone member, and upper conglomeratic member. The formation represents a braided fluvial system. No unreworked datable fossils have been recovered from the Grand Castle Formation, although Eaton has found a bed of freshwater mollusks in the middle member along Little Creek near the town of Paragonah. The formation's age is constrained by palynomorph dates generated from units above and below the Grand Castle Formation (Goldstrand and Mullet, 1997). The Grand Castle Formation is overlain by the Claron Formation and will be observed at STOPS 9 and 10.

Other Common Units in Parowan Canyon

Parowan Canyon is geologically complex as is well demonstrated by the excellent geologic map of the area by Maldonado and Moore (1995). Abundant faults throughout the area have brought the Tertiary Claron, Brian Head, and Isom Formations in lateral contact with both the Iron Springs and Grand Castle Formations.

Comparison of Stratigraphic Sections in Cedar and Parowan Canyons (Eaton)

The Parowan Canyon section is floored by the Iron Springs Formation which appears to be mostly time equivalent to the remarkably different sequence or Dakota Formation, Tropic Shale, and Straight Cliffs Formation found in Cedar Canyon only 12 miles (20 km) to the south (figure 5). This remarkable lateral change in depositional style is puzzling. Somewhere between these two canyons we may transect the maximum axis of subsidence of the foreland basin in this area (at least in the Cenomanian-Santonian). It may be (Eaton, research in progress) that the side of the axis associated with the Sevier orogenic thrust belt had a slightly higher gradient and greater sediment supply than on the side of the axis associated with the advancing epicontinental seaway. This may have resulted in a complex interplay of gradient, sediment supply, and subsidence such that a marked axis (that of maximum subsidence) divided braid plain from lower gradient coastal-margin and floodplain depositional styles. Much work remains to be done on the depositional systems represented by these two canyons.

FIELD TRIP ROADLOG

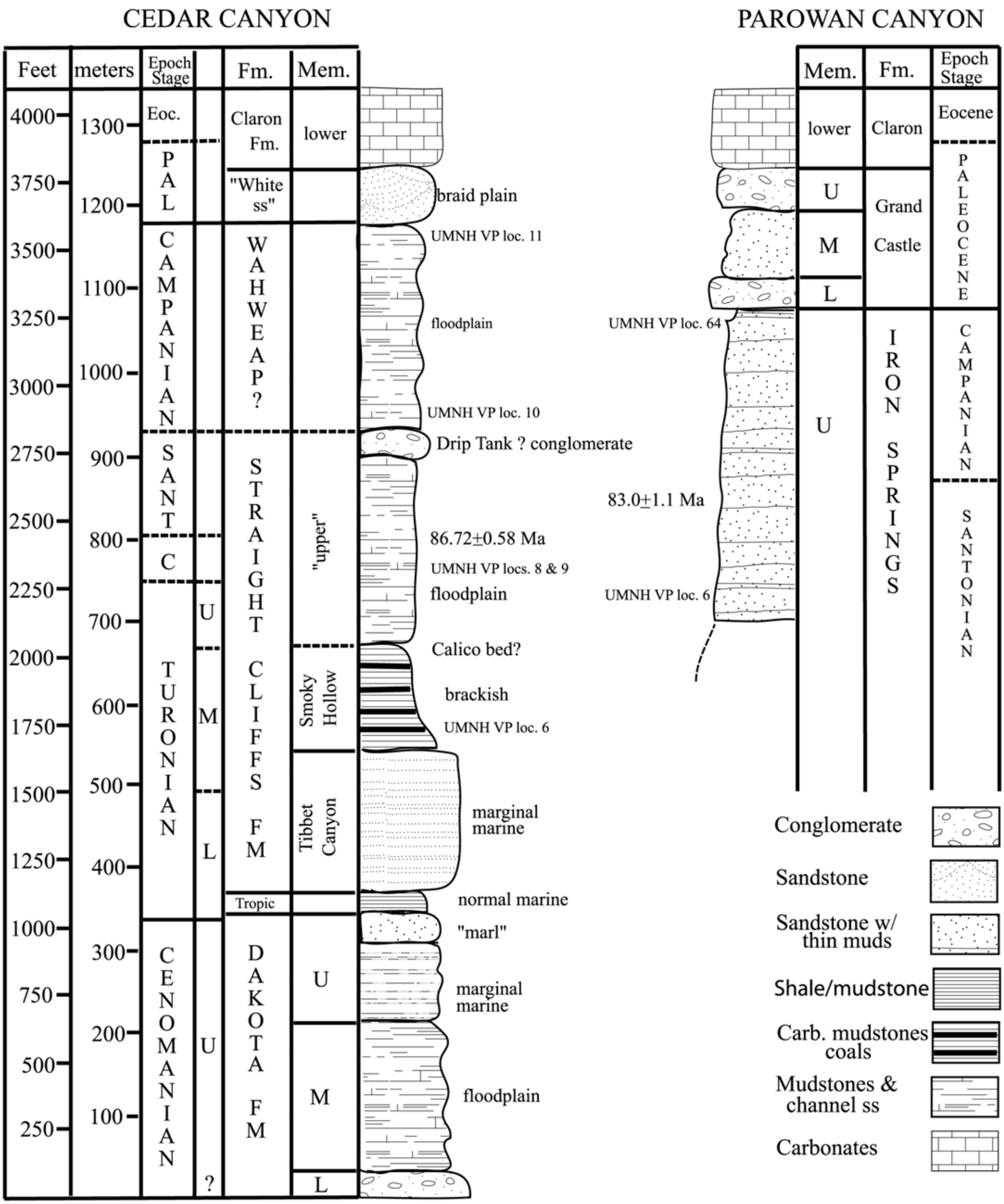

Figure 5. Comparison of Upper Cretaceous and lower Tertiary stratigraphy in Cedar and Parowan Canyons. The Parowan section is hung on the contact between the Claron and Grand Castle Formations.

MILEAGE Interval	Cumulative	DESCRIPTION
0.0	0.0	BEGIN. TURN EAST off Main Street onto Center Street, Utah Highway 14, in Cedar City.
0.6	0.6	Pass the Moenkopi Formation (Triassic) on the left (this part of the road log is based on Averitt and Threet, 1973).
0.3	0.9	Chinle Formation (Triassic) on left.
0.1	1.0	Moenave Formation (Jurassic) on left.
0.1	1.1	Kayenta Formation (Jurassic) on left.
0.1	1.2	Navajo Sandstone (Jurassic) on left.
0.4	1.6	Cedar City Tongue of the Kayenta Forma tion (Jurassic) on left.
0.1	1.7	Navajo Sandstone (Jurassic) on left.
0.2	1.9	Limestone Member of the Carmel Forma tion (Jurassic).
0.1	2.0	Mile 2 signpost. Banded member of the Carmel Formation (Jurassic) into the gyp siferous member.
1.2	3.2	Banded member of the Carmel Formation (Jurassic) visible on low hills.
0.2	3.4	STOP 1. WESTERNMOST CRETA-

CEOUS SECTION (Laurin and Sageman, Section A of figure 2). "Section A" offers the most continuous exposure of the Dakota Formation and the lowermost part of the Straight Cliffs Formation in the area. Even though the section is covered by debris in places, relative lateral continuity of the exposed parts makes this site the most significant source of information on lithology and internal architecture of the fluvial to shoreface deposits of the Dakota Formation locally (figure 6).

Lower and Middle Members of the Dakota Formation

The base of the Cretaceous deposits can not be identified precisely at this site, but the lower and middle members of the Dakota Formation are estimated to be at least 555 feet (180 m) thick. The lowermost outcrops are mostly gray, green or reddish mudstones (including nonmarine mollusks, in places; more or less pedogenically modified in certain zones: roots, slickensides, caliche), which are interbedded on decimeter to meter scale with tabular sandstone bodies (climbing ripples and water-escape structures are common; see figure 6 for general appearance of these deposits). Small, several-meter-wide channels (with clay rip-ups at their bases) are encased in these deposits. At certain levels, relatively thick (several meters), laterally extensive, and internally complex sand-filled channels occur. Locally, lateral accretion internal geometry of the channels is distinguishable. In the uppermost part of the middle member of the Dakota Formation, the channel fills are weakly bioturbated (*Thalassinoides*) and contain mud-draped foresets. The drapes locally bundle. No coal zone has been found within or at the top of these deposits (the Willow Creek Coal Zone of Averitt, 1962, which probably correlates to the boundary between the middle and upper members of the Dakota Formation, is absent at this site).

The interbedded mudstones and tabular sandstones are interpreted as floodplain deposits (the sandstone bodies representing crevasse splays). The small channels encased in these deposits probably represent distributary channels of an anastomosed fluvial system. The larger and more complex sandstone packages probably represent distributary channels of high-sinuosity rivers. Bundled mud drapes and crustacean burrows (*Thalassinoides*) in the uppermost part of the succession suggests tidal influence on the fluvial sedimentation (i.e.

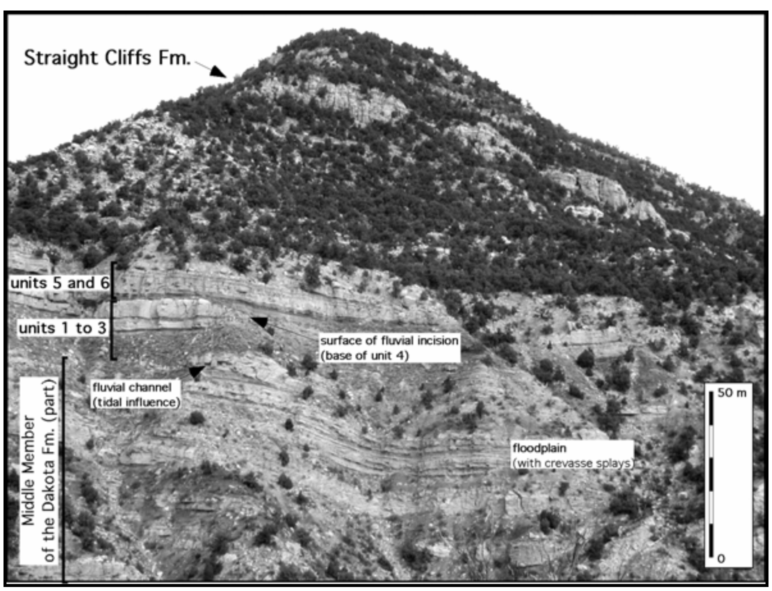

Figure 6. Photomosaic of part of the Cretaceous succession exposed in the westernmost part of the area of Cedar Canyon. The middle and upper members of the Dakota Formation (lower third of the figure) are at least 590 feet (180 m) thick at this site. Units 1 to 6 (lower part of the upper member of the Dakota Formation) refer to informal stratigraphic units defined by Laurin and Sageman (this volume). Subdivision of the Dakota Formation is after Gustason (1989). Recognition of the Straight Cliffs Formation is after Averitt (1962); see figure 2 for detailed relationships of the stratigraphic concept of Averitt (1962) and that of Gustason (1989).

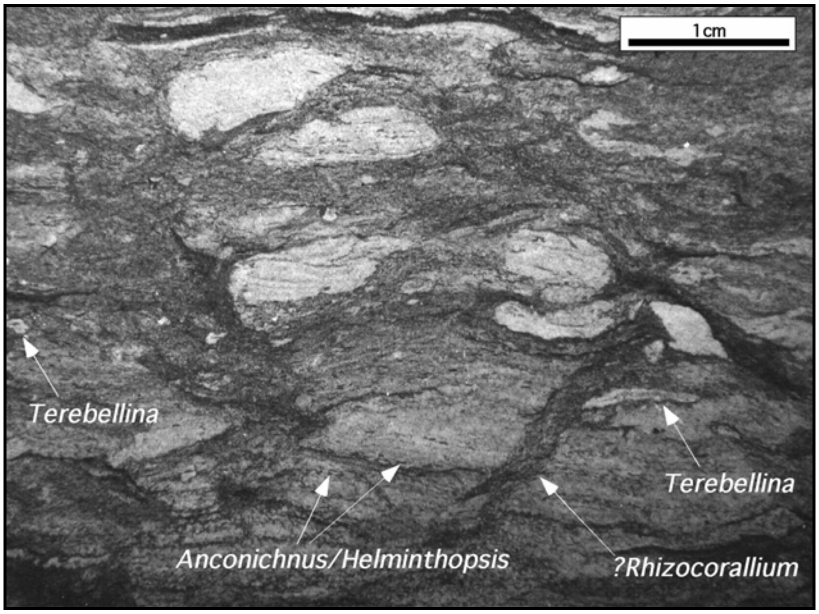

Figure 7. Typical appearance of heterolithic deposits forming the uppermost part of unit 1 (figure 2). Note relatively strong bioturbation (ichnofabric index *sensu* Bottjer and Droser, 1991, is variable; locally as high as ii = 4) and moderate diversity of biogenic structures (the highest diversity throughout the succession). Combined-flow ripples are typical for this facies. This facies is capped in places (e.g. west of the Maple Canyon) by a distinct skeletal lag with fragments of inoceramids and shark teeth.

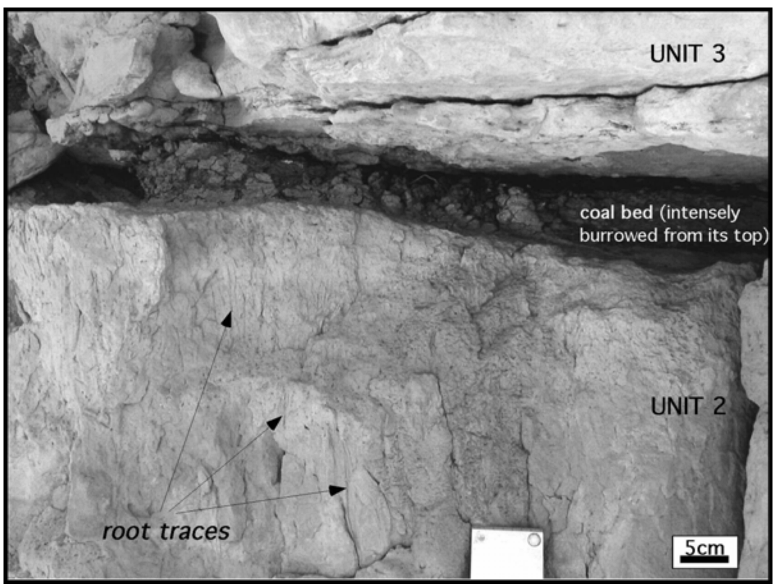

Figure 8. Rooted zone and coal bed forming a boundary between units 2 and 3 at "Section A" (figure 2). Note intense burrowing (*Thalassinoides*) at the top of the coal bed. Underlying sandstones of unit 2 as well as overlying sandstones of unit 3 are interpreted as having been deposited in a tidal-inlet/delta setting.

deposition in proximal zones of an estuary).

Upper Member of the Dakota Formation

The base of the upper member of the Dakota Formation is marked by superimposition of finely laminated carbonaceous mudstones upon greenish, pedogenically modified floodplain deposits and, locally, upon fluvial/estuarine sandstones (figure 6). The carbonaceous mudstones coarsen upward into heterolithic, weakly to moderately bioturbated sandy mudstones (facies similar to that on figure 7). These deposits are capped by a skeletal lag that contains a relatively diverse assemblage of bivalves including fragmented inoceramids; farther east, between Maple Canyon and Old MacFarlane Mine, shark teeth were found in correlative deposits. The lag is immediately overlain by moderately to strongly bioturbated sandy mudstone (ichnofabric index *sensu* Bottjer and Droser, 1991, ii =3 to 4). This deposit is overlain by non-bioturbated, gray mudstones, which pass upward into

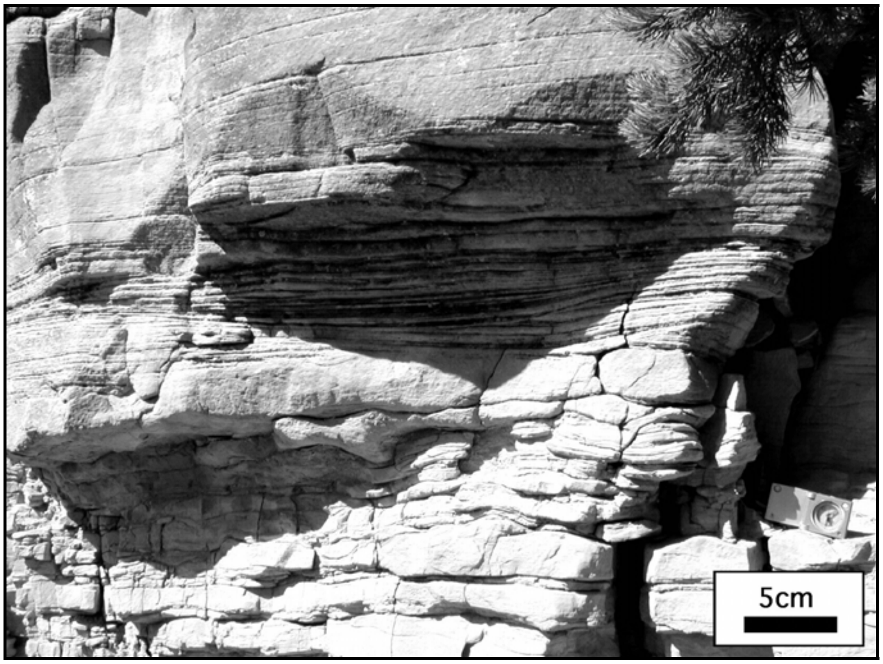

Figure 9. Hummocky cross-stratification in the uppermost part of the Dakota Formation (unit 8) in the westernmost part of the area (Section A - STOP 1).

a thin coal bed (figure 2).

The overlying part of the Dakota Formation can be briefly described as a set of sandstone (mostly fine to medium-grained) packages, which are punctuated at several levels by rooted surfaces, overlying coals and sometimes, fossiliferous (brackish) mudstones (figures 2 and 8). Tops of the coal beds are typically strongly burrowed (monotypic assemblage of *Thalassinoides*). The sandstones forming the bulk of the succession (units 2-6, figure 2) are typically trough cross-bedded, or combined flow-rippled. Sets of the ripple-bedded sandstones are typically interbedded with thin beds of strongly or entirely (ichnofabric index, ii = 5) bioturbated muddy sandstones (high abundance, low diversity assemblage of *Thalassinoides* and/or *Planolites*). Internally, these slightly heterolithic deposits may display distinct inclined accretion surfaces (e.g. unit 5, figure 2; dominant direction of dip is towards the southwest in this case). As shown in figure 2 and 6, units 2 and 3 are locally truncated by a prominent incision surface. The erosional relief is filled with heterolithic mudstones and sandstones. Trough cross-bedded sandstones locally display reactivation surfaces and mud draped foresets.

The middle part of the upper member of the Dakota Formation (tentatively labeled unit 7, figure 2) is covered with debris and was not studied. Correlative deposits are, however, exposed farther east (see below).

The uppermost part of the upper member is formed by two distinct (and laterally traceable) units (labeled 8 and 9, figure 2). The bulk of the lower unit is represented by a succession of sandy mudstones (at the base), hummocky cross-stratified sandstones (in the middle part; figure 9) to sub-horizontally laminated sandstones (in the upper part). A relatively diverse ichnoassemblage can be found in the lower, heterolithic part of the succession including *Skolithos* and *Diplocraterion*. The unit is rooted at the top, and overlain by a basal coal zone of unit 9 (figure 2). The top of the Dakota Formation is formed mostly by bioclastic, strongly calcareous mudstone (up to the bioclastic limestone or "marl"). It is finely laminated on a millimeter-scale. No distinct bioturbation structures have been identified. Elongated gastropods are locally aligned in one direction, but this direction does not appear to be consistent throughout the unit.

The basal unit 1 (figure 2) represents restricted deposits of a central lagoon or estuary. The upward-coarsening trend associated with the increase in bioturbation suggests a shift to the outer zone of the lagoon/estuary with increased marine influence. The capping skeletal lag may represent a surface of wave ravinement or marine flooding. Locally, peak marine influence might have been reached right after the formation of this surface as suggested by marine faunas (inoceramids) and ichnofabrics. The superimposition of non-bioturbated (lagoonal) mudstones and a coal upon these deposits suggests renewed shallowing.

This part of the succession records the onset of a major “*Sciponoceras*” base-level rise (and accompanied transgression), which has been recorded throughout the Colorado Plateau (e.g. Elder, 1991; Kirkland, 1991; Elder et al., 1994). At this particular site, the actual base of the transgressive systems tract of the long-term sequence (*sensu* Van Wagoner et al., 1988) may coincide with the erosive basal surface of the fluvial/estuarine deposits of the uppermost part of the middle member of the Dakota Formation (figure 6), although further research is needed to answer this question.

The bulk of the overlying deposits are represented by back-barrier (e.g. washover) to barrier (mostly tidal-inlet) deposits. The incision surface shown on figure 6 is interpreted as a surface of fluvial incision. The incision probably took place in response to base-level fall (sequence boundary *sensu* Van Wagoner et al., 1988). The heterolithic deposits filling the incision topography formed in an estuarine (tide-dominated) setting.

Unit 8 is represented mostly by shoreface deposits (possibly overlying a relic of a tidal-inlet; figure 2). The uppermost part of this unit might have originated in a foreshore (beach) environment. The skeletal mudstones of unit 9 were probably deposited in an open lagoon or an inter-distributary bay. In any case, the depositional site did not receive any significant amount of terrigenous clastics. The contrasting depositional conditions, relative to the underlying clastic-dominated deposits, suggest that a significant change in drainage pattern took place right before deposition of unit 9. Indeed, as revealed by regional sequence stratigraphic correlation, the major depocenter was offset towards the south and southeast during that time (see Laurin and Sageman, this volume).

Lower part of the Straight Cliffs Formation

The Straight Cliffs Formation overlies a thin bed of strongly bioturbated, bioclast-rich sandstone. This bed is overlain by an upward-coarsening succession of muddy sandstones to fine-grained sandstones. The succession is characterized by vertical alternations of strongly bioturbated sandstones and compact, weakly bioturbated sandstone beds (figure 10). The sandstone beds are sharp-based and subhorizontally laminated (structures similar to low-amplitude hummocky cross-stratification are locally visible). The upper contacts are typically strongly bioturbated by *Ophiomorpha* (figure 10).

The bioclast-rich sandstone at the base may represent a relic of a tidal-inlet deposit, which formed during early phases of transgression (thicker tidal-inlet deposits of similar lithology outcrop farther east). The overlying succession of bioturbated sandstones and laminated sandstone beds is probably of shoreface origin (the laminated beds likely represent storm-generated deposits; the bioturbated layers probably formed during fair-weather conditions), and represents gradual shoreface progradation during relative sea level stillstand or initial fall.

0.3	3.7	Carmel Formation (Jurassic) at road level.
0.6	4.3	To south a complex slump topography.
0.6	4.9	Right Hand Canyon turnoff.
0.8	5.7	TURN LEFT onto dirt road, heading north towards Maple Canyon. Cretaceous section is well exposed.

0.1 5.8 TAKE RIGHT FORK, go 100 feet (32 m) and park. First part of Stop 2 here, and then hike up fault-controlled Maple Canyon for about 10 minutes looking on the left for evidence of an old coal mine.

STOP 2 - **Part 1. Physical Stratigraphy** (Laurin and Sageman)

The Maple Canyon section, as depicted on figure 2, is a composite section. The upper member of the Dakota Formation has been measured near the canyon entrance, on a southeast-facing slope. The Straight Cliffs Formation (as well as the uppermost part of the Dakota Formation which is not shown on the figure) has been measured to the northeast of the first section, up Maple Canyon, at the "classic" site measured by Gustason (1989) and coincident with the section of Tibert et al. (in press).

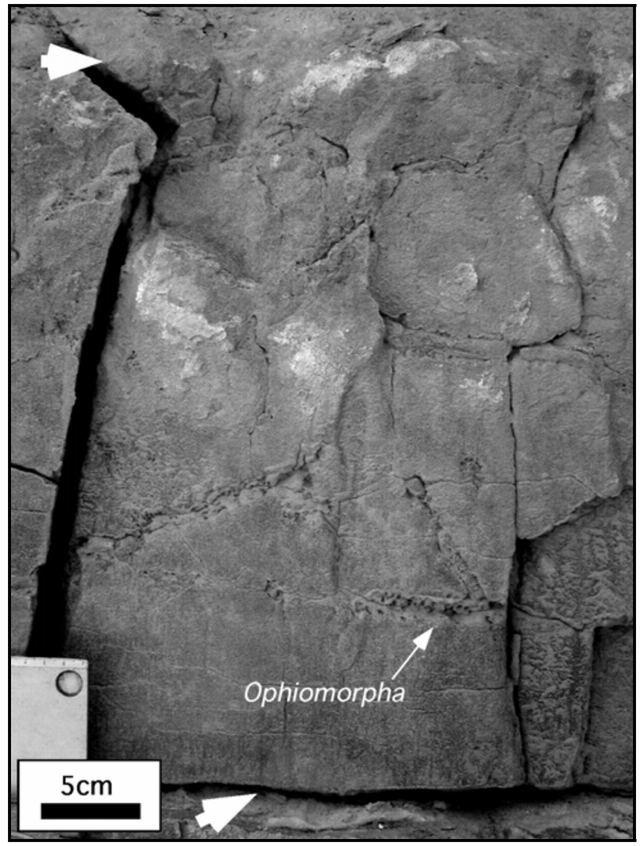

Figure 10. Lower part of the Straight Cliffs Formation is characterized by bioturbated muddy sandstones, which are interbedded on a scale of a few inches with sharp-based sandstone beds (arrows show the lower and upper contacts of one of these beds). Internally, the sandstone beds display subhorizontal lamination (locally, structures resembling low-amplitude hummocky cross stratification can be found). In contrast to the sharp basal contacts, the tops of the beds are typically disrupted by bioturbation (*Ophiomorpha*; the example marked by the arrow penetrates into the bed from the upper contact). These beds are interpreted as event, probably storm-generated, deposits.

Upper Member of the Dakota Formation

The lower and middle members of the Dakota Formation are poorly exposed throughout the area. Nevertheless, conglomeratic (probably braided-stream) deposits of the lower member of the Dakota Formation can be found approximately 1,200 feet (400 m) to the west from the canyon entrance (a few tens of meters above the highway).

The facies assemblages of the upper member of the Dakota Formation are very similar to those described from the "Section A" at STOP 1; however, there are a few distinctive features in the Maple Canyon area.

The upper member of the Dakota Formation rests on an approximately 1-foot-thick coal bed, which overlie pedogenically modified, greenish mudstones of middle member of the Dakota Formation. The position of this coal bed in the overall succession suggests that it is partly equivalent to the

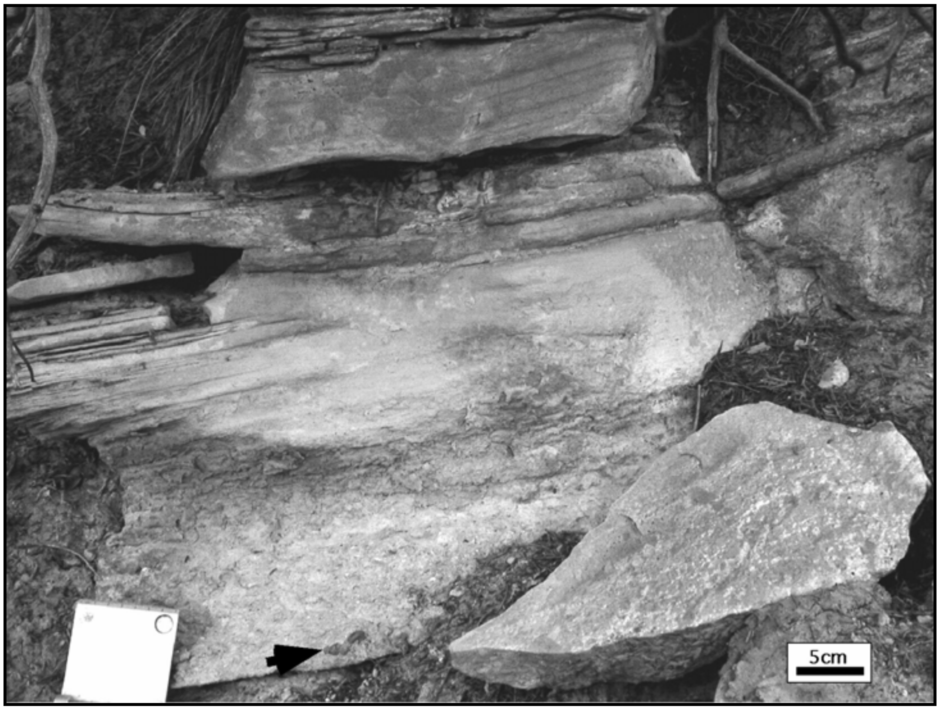

Figure 11. Base of a channel-fill unit 4 at locality "Maple Canyon-south" (figure 2). Note the subrounded limestone clasts (several centimeters in diameter) at the base of the channel. Sediments underlying the channel trough (not shown) are rooted. Stratification of the overlying sandstones suggests deposition under combined-flow conditions. These deposits are interpreted as tidal-inlet/delta deposits overlying a surface of tidal ravinement.

Willow Creek Coal Zone (*sensu* Averitt, 1962). Regional sequence stratigraphic correlation (Laurin and Sageman, this volume, figure 5; also see Gustason, 1989; Kirschbaum and McCabe, 1992) suggests that the coal bed is also partly correlative to the coal zone labeled "Smirl coal zone" by Doelling and Graham (1972), and "coal zone 4 and 5" by Kirschbaum and McCabe (1992), to the east of Cedar Canyon. The top of this coal zone marks the onset of the *Sciponoceras* biozone, in the area of eastern Markagunt Plateau (cf. Gustason, 1989; Elder, 1991; Kirschbaum and McCabe, 1992).

In the easternmost part of the exposure, heterolithic deposits of the central estuary (probably correlative to unit 4a, figure 2) are truncated by a channel filled with tidal-inlet/delta deposits (figure 2; base of channel fills shown in figure 11). This incision surface, paved with gravelly sandstone, represents a prominent marker throughout the eastern part of the Cedar Canyon area. Laurin and Sagemen (this volume) interpret this as a surface of tidal ravinement. The topographic lows of the incision surface are strongly rooted at some locations, including Maple Canyon, which points to a possibility that another base-level fall took place before the tidal reworking and the tidal-inlet/delta setting might have occupied an erosional depression previously formed by fluvial incision (cf. Oertel et al., 1991). However, further investigation is needed to make a reliable conclusion about the sequence stratigraphic significance of this surface.

Units 8 and 9 show little lateral change in lithology between "Section A" and Maple Canyon (figure 2), and can thus be used as reliable markers for correlation of the sections. Unit 8 is underlain in Maple Canyon by approximately 6 feet (2 m) of fossiliferous mudstone, which, in turn, is underlain by an approximately 8 feet (2.5 m) thick coal zone. Corresponding deposits are not exposed in the westernmost section; however, correlation based on data from another section (not shown here) suggests that the coal zone may be partly or completely missing from the western margin of the area (STOP 1). The observed westward thickening of this part of the Dakota Formation (figure 2) may be explained by compaction of the coal zone.

Lower part of the Straight Cliffs Formation

Unit 9 is overlain by poorly sorted, bioclast-rich, trough cross-stratified sandstones containing *Mytiloides puebloensis*. Bimodal paleocurrent directions together with mud draped-foresets suggest tide-dominated (probably tidal-inlet) setting. They are overlain with offshore mudstones (rich in marine fossils, including ammonites and inoceramids), which pass upward into lower shoreface sandstones (the same facies as the one described in "Section A" and STOP 1). These sandstones are, in turn, overlain by trough cross-stratified sandstones of possible upper shoreface origin. The upward-shallowing succession is capped with a bioclast lag, and overlain by another offshore-shoreface succession (unit 11, see figure 3 of Laurin and Sageman, this volume; this unit is best exposed in the eastern part of Cedar Canyon, near mile marker 8 of Utah Highway 14).

Part 2. Microfossil Trends (Tibert and Leckie)

Nearly 30 samples were taken from the fine-grained facies in Maple Canyon that span the upper member of the Dakota Formation into the basal Tropic Shale equivalent. Figure 3 shows an excerpt of the measured section that begins at approximately 25 m above the mine workings easily identified at the base of the section.

***Sciponoceras gracile* Biozone**

We assign the lower units that encompass the coal seams 1-4 (figure 3) to the *Sciponoceras gracile* ammonite biozone of Kauffman et al. (1993) based on the occurrence of *Cythereis eaglefordensis,* an important age indicator for the Western Interior (Hazel, 1969). An ornate ostracode, and very abundant in the offshore, this taxon indicates normal marine conditions. Its occurrence with the dominant *Fossocytheridea* n. sp. 2 (a low salinity indicator) suggests *C. eaglefordensis* was likely pushed landward during a significant marine flooding event. *C. eaglefordensis* has also been recovered from coeval strata below the "Marl-equivalents" at Coal Creek and Fiddlers Canyon. *Clithrocytheridea*? n. sp. 2 and *Cytheropteron eximium* are present with *C. eaglefordensis* at most nearby localities.

Hazel (1969) described *Cythereis eaglefordensis* from the latest Cenomanian Eagleford Shale of Texas in addition to stratal equivalents at New Mexico, Utah, Georgia, and Florida. Because of the common association with the ammonite *Sciponoceras gracile*, Hazel (1969) considered this taxon as an excellent late Cenomanian indicator. At Bigwater, Utah, Tibert et al. (in press) identify a sharp pulse of *C. eaglefordensis* several meters below bentonite TT2 (equivalent to Bentonite B of Elder 1988, 1991) that also corresponds to maximum flooding of Leithold's (1994) latest *Sciponoceras gracile* Cycle 1; Leithold (1994) and Leithold and Dean (1998) delineated three transgressive cycles based on increased total carbonate within the Tropic Shale. A second "pulse" of *C. eaglefordensis* occurs at or just below the Cenomanian-Turonian boundary (approximately 2-3 m below TT3 equivalent to the "Boundary Bentonite" or Bentonite C of Elder, 1988, 1991). At Bigwater, *C. eaglefordensis* rarely occurs without *Cytheropteron eximium*, and together they pro-

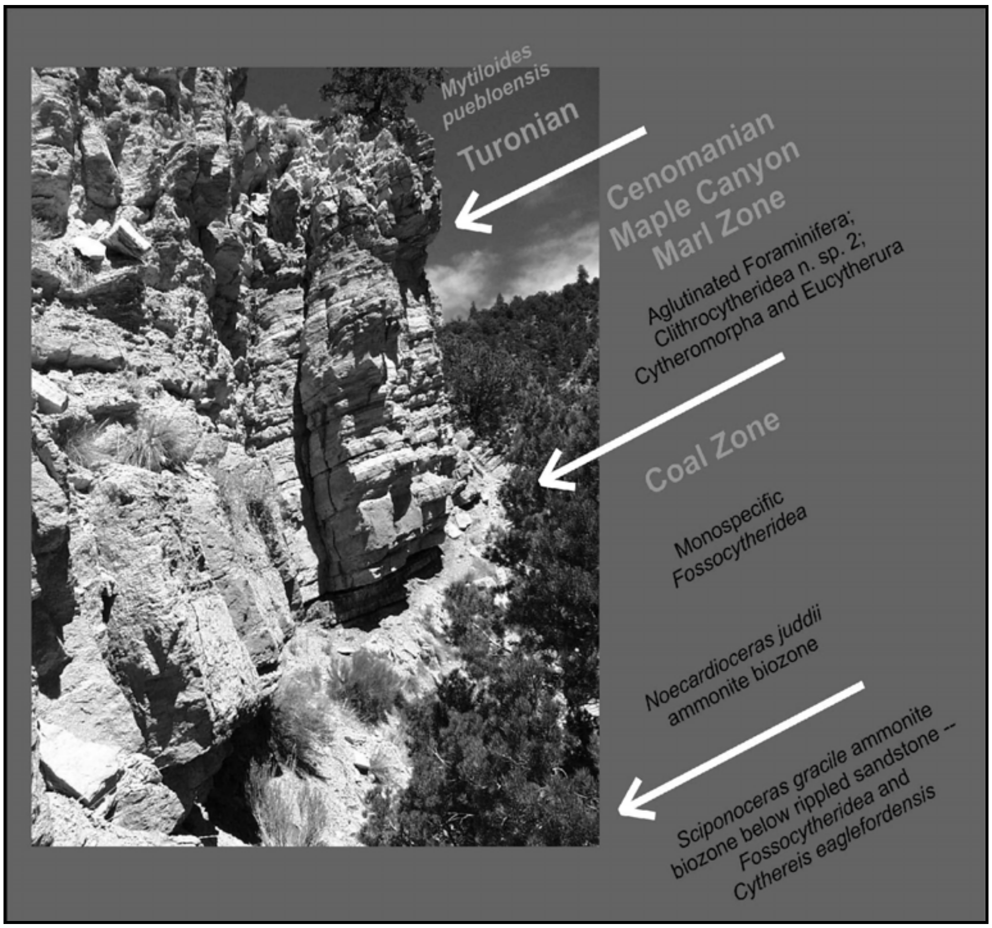

Figure 12. Strata belonging to the upper member of the Dakota Formation that mark the contact between the Cenomanian-Turonian stage boundary. The skeletal shell beds comprise hard, calcified oyster-rich marlstones that alternate with sandy-shale intervals. There are approximately nine marl-shale couplets above the thin coals. The marlstones are dominated by foraminifera, ornate gastropods, and ornate marine ostracodes. In contrast, the thinner shaly intervals contain a depauperate ostracode assemblage dominated by *Fossocytheridea* (synonymous with *Sarlantina* - Tibert, Colin, Babinot, and Leckie, manuscript in prep.) and *Darwinula*, both low-salinity genera (Babinot and Colin, 1976; Neale, 1988). The domination of these "brackish" genera indicate phases of increased freshwater runoff. The ostracode composition of the shaly intervals is similar to the shelly mudstones that characterize the underlying coal zone. Note also that the "marl" biotic association occurs at the Old MacFarlane Mine road cut (STOP 4).

vide a useful biostratigraphic marker for the *Sciponoceras* and latest *Neocardioceras* ammonite biozones.

Neocardioceras juddii Biozone

The change from generally siliciclastic mudrock and coal to carbonate-rich skeletal shell beds (marlstones) coincides with a rather distinct paleoecological shift with the appearance of several new ostracode taxa and foraminifera. We assign the uppermost coal seams (5 and 6) and the overlying marlstone beds (figures 3 and 12) to the *Neocardioceras juddii* ammonite biozone (Kauffman et al., 1993) for reasons discussed below.

Within the first marlstone bed (~54 m) (figures 3 and 12) there is an agglutinated foraminiferal association that includes *Trochammina rainwateri*, *Trochammina wetteri*, *Ammobaculites obliquus*, *Miliammina ischnia*, *Verneuilinoides kansensis*, and rare *Gavelinella dakotensis*. The generic composition of the foraminiferal assemblage is estuarine indicated by the dominance of *Trochammina* and *Ammobaculites* (e.g., Ellison, 1972; Scott et al., 1996). *Trochammina wetteri* is significant within the marlstones at Maple Canyon because it ranges from latest Cenomanian into the Turonian (Eicher and Worstell, 1970; Frush and Eicher, 1975). Tibert et al. (in press) report the first occurrence of *T. wetteri* at the approximate position of TT2 (Bentonite B of Elder, 1988, 1991) at Bigwater. In addition, the calcareous benthic *Gavelinella dakotensis*, although rare on the Markagunt Plateau, may signify an important biotic event (Leckie et al., 1998). West et al. (1998) report a *G. dakotensis* acme at approximately the position of the TT2 at Bigwater, Utah, Lohali Point, Arizona, and both Mesa Verde and Rock Canyon, Colorado.

The marlstones also mark the first occurrence of *Cytheromorpha* n. sp. 2 and *Eucytherura* n. sp. (figure 3), common in late Cretaceous strata; both genera range into the modern and are generally found in marginal marine settings (Neale, 1988). Both *Cytheromorpha* and *Eucytherura* occur with *Fossocytheridea* n. sp. 2., the most abundant ostracode of the uppermost Dakota Formation. At all localities and ages, *Fossocytheridea*-bearing units most commonly comprise monospecific populations and are considered excellent indicators of very low salinity (Swain and Brown, 1972; Babinot and Colin, 1976; Colin, 1983).

Regional Biostratigraphy and Cyclostratigraphic Trends

The first occurrence of *Cythereis eaglefordensis* at Cedar Canyon marks a significant flooding event within the *Sciponoceras gracile* biozone as observed in the Tropic Shale (Tibert et al., in press). The first occurrence of *Cytheromorpha*, *Eucytherura*, and agglutinated foraminifera in the carbonate-rich skeletal shell beds (marlstones) signify increasing marine influence on the Markagunt Plateau. The first occurrences of *Trochammina wetteri* and *Gavelinella dakotensis* within the lower marlstone beds are significant because both taxa occupy beds associated with TT2 at Bigwater; this bentonite is associated with a *Neocardioceras juddii* transgressive flooding event (Elder, 1991; Leithold, 1994). That an abundance of *Clithrocytheridea* n. sp. 2 occurs several marlstone beds above *T. wetteri* and *Gavelinella dakotensis*, is also reminiscent of the trends observed in the Tropic Shale (Tibert et al., in press). Considering these vertical biotic relationships and the significant increased carbonate content of the marlstone beds, we conclude that the marlstone beds at Maple Canyon were deposited coincident to the lower-to-middle *Neocardioceras juddii* ammonite biozone during early transgression of Cycle 2 as described from prodeltaic facies at Bigwater (Leithold, 1994; Tibert et al. in press). Finally, the contact between the marlstone beds and the overlying sandstone at Maple Canyon marks the Cenomanian-Turonian boundary as indicated by the first occurrence of the bivalve *Mytiloides puebloensis* (Walaszczyk and Cobban) known to range from the lower to middle Turonian (Kennedy et al., 2000).
RETURN TO Utah Highway 14.

0.1 5.9 TURN RIGHT (east) onto Utah Highway 14.
0.1 6.0 Cross fault.
0.2 6.2 Lower member of the Dakota Formation (Cretaceous) exposed in roadcut.
0.1 6.3 Mile 6 signpost.
0.1 6.4 To the north in the lower cliff face, in exposures of the upper member of the Dakota Formation, estuarine deposits of unit 4a resting on bioturbated (lagoonal to marine) heteroliths of unit 1 (figure 7) are excellently exposed. The contact, marked by a distinctive inoceramid fragment-bearing lag, has a complex history as suggested by the absence of units 2 and 3. The last reworking of the surface took place during relative sea-level fall (probably fluvial incision), and subsequent initial relative sea-level rise (filling of the estuary; Laurin and Sageman, in prep.).

0.4 6.8 Cross fault that down-drops Smoky Hollow Member of the Straight Cliffs Formation to road level.
0.2 7.0 Cross fault - Tibbet Canyon Member of the Straight Cliffs Formation exposed at road level.
0.3 7.3 STOP 3 at Mile 7 signpost (Eaton). Note the beautifully exposed fault on the south side of the highway in the roadcut. On the east side of the fault, the typical green and red mudstones of the nonmarine middle member of the Dakota Formation are flat lying. On the west (down) side of the fault, sandstones with interbedded thin coals of the Tibbet Canyon Member are dragged nearly vertical. Along the north side of the road, on the west side of the fault, the top of the Tibbet Canyon Member and the basal brackish-water deposits of the Smoky Hollow Member are present.

0.5 7.8 Slide area along south side of the road.
0.3 8.1 Along the north side of the road are the sections measured by Laurin for the Old MacFarlane Mine composite section (figure 2).
0.05 8.15 STOP 4 - Old MacFarlane Mine (Laurin and Sageman) - due to the instability of the slopes in the area parking is discouraged, so it may be necessary to drive farther and walk back to this stop. This site provides excellent exposures of the lower part of the upper member of the Dakota Formation (figure 13). An almost complete succession of the Straight Cliffs Formation is excellently exposed along the highway to the southeast from here. It should be noted that

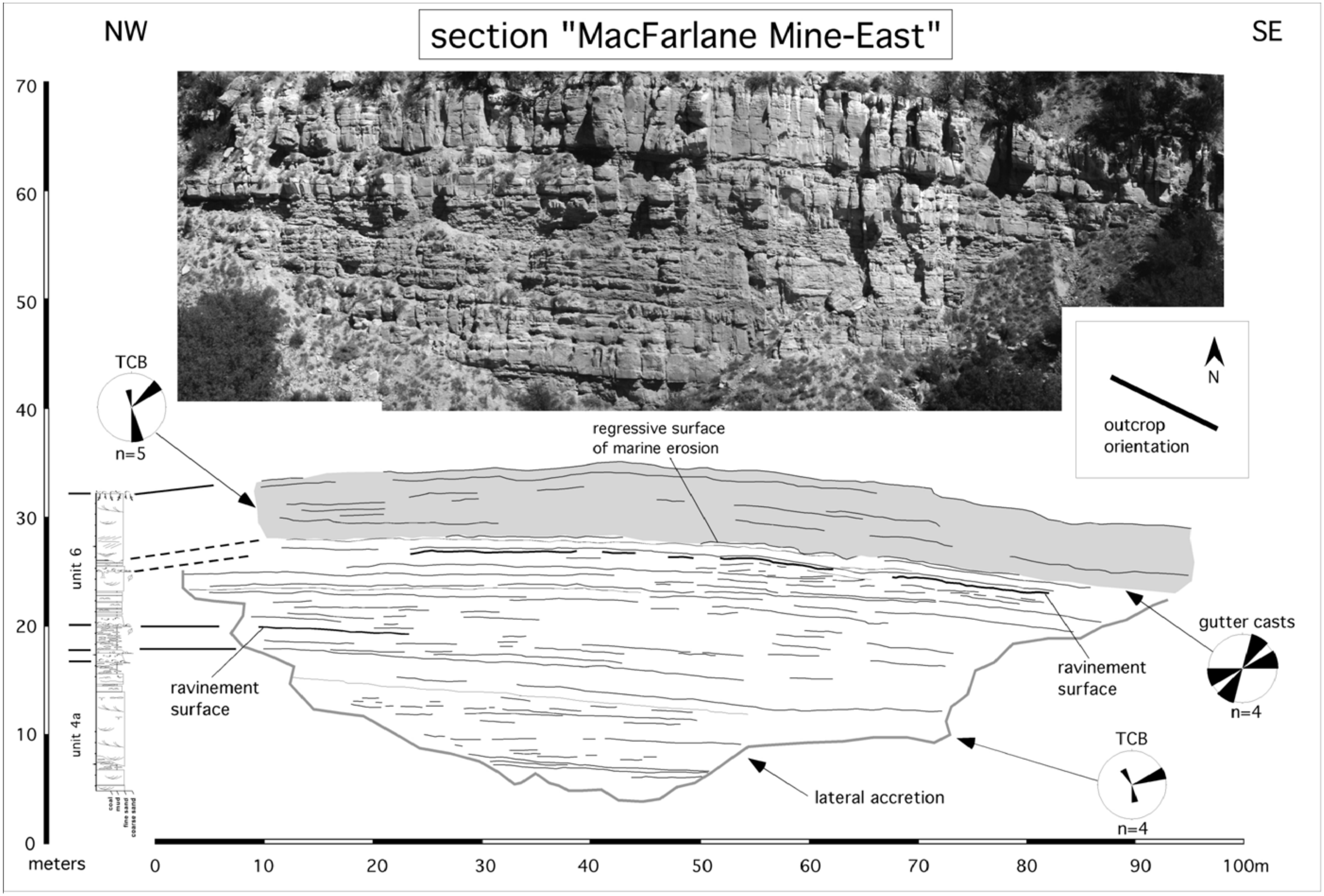

Figure 13. Photomosaic of the lower part of the upper member of the Dakota Formation (*sensu* Gustason, 1989) in the eastern part of the Old MacFarlane Mine section. Relationships of this succession to the other sections are shown in figure 2. The lower part of the succession (unit 3) is of estuarine origin. The upper portion of the cliff (unit 6) is formed mostly by shoreface deposits. Note the large gutter casts at the base of the uppermost sandstone package (grayed). This surface is interpreted to have formed in response to forced regression (cf. Plint and Nummedal, 2000). The partial misfit between the lithological log and the photomosaic is due to a composite origin of the log (the uppermost part was measured in the southeastern part of the exposure).

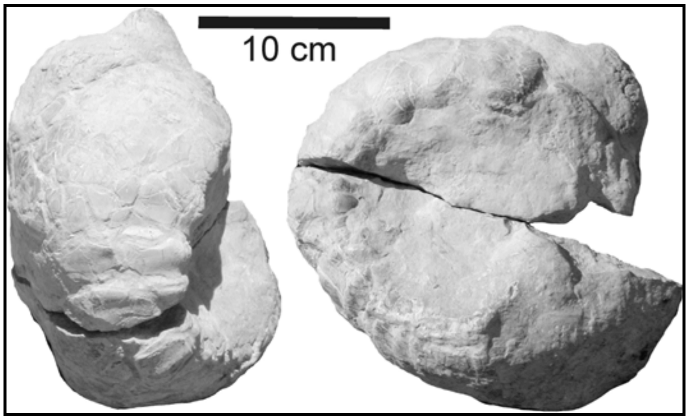

Figure 14. *Fagesia catinus* (Mantell), University of Utah UMNH IP 2252, the only three-dimensional ammonite specimen recovered from the "Tropic Shale" (or very low in the Straight Cliffs Formation depending on favored stratigraphic nomenclature) in Cedar Canyon.

much of the exposure along the road is slumped; however, a relatively intact section of marine mudstones containing abundant inoceramids and ammonites (*Mytiloides kossmati*, *Fagasia catinus* (figure 14), *Watinoceras*, and an unidentified aberrant ammonite) has been found here southwest of the cement culvert. The range overlap of *M. kossmati* and *F. catinus* is restricted to the middle lower Turonian *Vascoceras birchbyi* zone (= *Watinoceras coloradoense*, Kennedy et al., 2000).

Upper Member of the Dakota Formation

Basic facies assemblages are similar to those of the previous sections. Details on lithology are depicted in figure 2 (the section shown is a composite one).

The mudstones of unit 1 (exposed to the west from the section of figure 13) are abruptly overlain by sandstones (trough cross-bedded, with tidal bundles) and heterolithic beds of probable estuarine origin (the base of these deposits is reworked by ichnogenus *Gastrochaenolites,* suggesting firmground conditions; cf. MacEachern et al., 1992). Sequence stratigraphic correlation (figure 2) suggests that units 2 and 3 might be missing in this area due to fluvial incision at the base of unit 4a.

Bases of both units 4b and 6 are marked by distinct skeletal, gravel-bearing lags, which are interpreted herein as a ravinement surface. The wavy to flaser-bedded sandstones beneath the base of unit 6 might correspond either to unit 4 or to unit 5. Unit 6 (forming the upper part of the cliff face; see figure 13) can be subdivided into two depositional packages bounded by another bioclast and gravel-bearing lag (strongly burrowed). Each of the sub-units displays a distinct upward-coarsening trend: combined flow-rippled heterolithics of the lower parts pass upward into clean cross-bedded sandstones. Sandstones of the upper sub-unit are sharp-based (with gutter casts) and swaley cross-stratified in the lowermost part. The two sub-units of unit 6 are tentatively interpreted as prograding shoreface packages bounded by surfaces of wave ravinement. The sharp lower contact of the swaley cross-stratified sandstones probably represents a regressive surface of marine erosion (Plint and Nummedal, 2000).

Unit 6 is capped by a rooted surface and overlain by lagoonal mudstones and coals of the Upper Culver Coal Zone (Tibert reports the occurrence of *Cythereis eaglefordensis* from above the ledge-forming sandstone, unit 6 of figure 2). Overlying sandstones of unit 8 (probably of shoreface origin) are truncated in this area by a channel. The occurrence of *in situ* coal in the trough of the channel suggests fluvial (or tidal) incision, and subsequent abandonment of the drainage. The upper part of the channel is filled with tidal flat deposits. Only the lowermost part of the skeletal mudstones of unit 9 is preserved here.

Lower part of the Straight Cliffs Formation

The lowermost part of the Straight Cliffs Formation (unit 10) is poorly exposed in this area. Lithology and interpretation of the exposed part of the lowermost Straight Cliffs Formation is shown on figure 3 of Laurin and Sageman (this volume).

0.25	8.4	Mile 8 signpost. The dark sandy mudstones below the cliffs of Tibbet Canyon just to the west preserve abundant specimens of *Mytiloides kossmati* of early Turonian age and are considered to represent one of the westernmost outcrops of the Tropic Shale representing peak westward extent of open marine conditions in the area.
0.1	8.5	Oyster-rich sandstones and thin coals of the Tibbet Canyon Member by the road mark the regressive cycle of the Greenhorn Cyclothem.
2.4	10.9	Top of the Tibbet Canyon Member and base of section of Eaton, Diem et al., 1999 and Moore and Straub (this volume).
0.1	11.0	Just past the road to the south on the hillside is UMNH VP Locality 6, Cory's Brackish Water Locality, in the base of the Smoky Hollow Member.
0.3	11.3	TURN RIGHT into Southern Utah University's Mountain Center.

STOP 5 (Eaton). Down the road (to the west) along the south side, UMNH VP Locality 66 has produced diverse middle Turonian mollusks, benthic foraminifera, and ostracodes from the base of the Smoky Hollow Member. The brackish part of the member is 167 feet (54 m) thick in this area whereas in the type area this interval is typically only a 10-20 feet thick (Eaton, 1991). It is clear that subsidence rates were still very high here during the middle Turonian relative to subsidence rates in the Kaiparowits Plateau region.

To the east in a stream cut along the north side of the highway (not clearly visible from here, but can be seen when driving up the canyon) is a pale-yellowish sandstone about 40 feet (12 m) thick containing some pebbles. This the horizon Moore and Straub (this volume) considered to possibly represent the capping bed of the Smoky Hollow Member, the Calico Bed. If this is the case, then the Smoky Hollow Member is 364 feet (110 m) thick in this area.

Across the road (on Southern Utah University property) and in higher strata (600-900 feet [200-300 m] above the top of the Tibbet Canyon Member) the ridges are relatively barren

Figure 15. Biotite ash bed, 882 feet (286 m) above the top of the Tibbet Canyon Member, which yielded a Coniacian age 86.72±0.58 Ma. Florian Maldonado in the foreground, J. Howard Hutchison in the background (SE SE SE Sec. 1, T. 37 S., R. 10 W.).

of vegetation. The tops of these ridges expose fine-grained channel sandstones and floodplain mudstones which contain UMNH VP Localities 8 and 9 as well as the radiometrically dated biotite ash bed (86.72±0.58 Ma, figure 15) discussed in the section on Straight Cliffs Formation - "Upper." This part of the Straight Cliffs sequence appears to be Coniacian in age, making it equivalent to the lower part of the John Henry Member.

0.1	11.4	RETURN TO Utah Highway 14 and pro ceed east past milepost 11.
1.3	12.7	Entrance to USDA Forest Service Cedar Canyon Campground on left.
0.4	13.1	Begin switchbacks in variegated floodplain deposits and channel sandstone probably equivalent to the John Henry Member of the Straight Cliffs Formation.
0.5	13.6	Notice conglomerate low in roadcut on the right side of the highway. This is the hori zon considered to possibly be equivalent to the Drip Tank Member of the Straight Cliffs ormation (see Moore and Straub, this vol ume).
0.2	13.8	TURN LEFT at Crystal Springs turnoff (to north), drive a short distance up the road and take the fork to the left and park.

STOP 6 (Eaton). The flats south of the parking area consist of variegated mudstones and thin sandstones. The mudstones have produced an abundant record of vertebrates when screen-washed (UMNH VP Locality 10, Paul's locality; see Eaton, Diem et al., 1999 and discussion above under Straight Cliffs Formation - "Upper"). Preliminary study of these vertebrates suggested a Campanian age, but considerable work remains to be done before this correlation can be made with certainty.

Figure 16. Conglomerate 1,140 feet (375 m) above the Tibbet Canyon Member of the Straight Cliffs Formation, possibly representing the Drip Tank Member. Hammer for scale (N½ NE NE Sec. 18, T. 37 S., R. 9 W.).

The conglomerate can be examined by walking back to the highway and going a short distance west on the highway. The conglomerate is at the top of a 90 foot (27 m) thick sequence of sandstones and considered to possibly be the Drip Tank Member of the Straight Cliffs Formation by Moore and Straub (in this volume, their "Interval F"). The same conglomerate is well exposed (figure 16) on the next ridge north of the parking area. There, the clasts are as large as 3 cm in diameter and are dominantly quartzite with some black, brown, and tan chert pebbles and some silicified limestone clasts in a gritty matrix. The pebbles are only in the lower part and the unit fines upward into sandstone. If this unit does represent the Drip Tank Member, then the Straight Cliffs Formation is about 1,160 feet (380 m) thick in this region. As the formation ranges from 900-1,550 feet (300-500 m) thick in the type area (Eaton, 1991), it seems odd that in this region where there is evidence of high subsidence rates, evidenced by the accumulation of thick brackish-water and floodplain deposits, that the formation would not be significantly thicker than in its type area.

RETURN TO Utah Highway 14, TURN LEFT (east).

0.3	14.1	Notice there has been no significant lithologic change in the variegated mud stone-sandstone sequence relative to the floodplain sequences below the "Drip Tank Member" conglomerate.

0.3	14.4	Slumped Claron Formation.
0.1	14.5	Mile 14 signpost.
0.2	14.7	Road cuts are in another floodplain sequence with meandering stream sandstones and variegated floodplain mudstones.
0.4	15.1	The "white sandstone" variably assigned to the Kaiparowits, Wahweap, and Grand Castle Formations.
0.1	15.2	STOP 7 (Eaton and Goldstrand), Websters Flat Turnoff.

The uppermost fine-grained beds down the Websters Flat road, underlying the "white sandstone," have produced some enigmatic fossil mammals from UMNH VP Locality 11 (see Eaton, Diem et al., 1999), most of which are undescribed. One of these specimens is very similar to an undescribed specimen recovered from the Wahweap Formation in Campbell Canyon, Bryce Canyon National Park. Another taxon, *Iqualadelphis* sp., represents a genus unknown prior to the Campanian. This suggests that there may be some Campanian-age rocks that are time equivalents of the Wahweap Formation in the upper part of the Cedar Canyon section. However, unlike some of the lower members of the Straight Cliffs Formation, it is difficult to lithologically recognize the Wahweap Formation here (see discussion in Moore and Straub, this volume). If the "Drip Tank Member" conglomerate about 900 feet (300 m) stratigraphically lower in the section is not equivalent to the Drip Tank in the type area, then there may be a transition to Wahweap equivalent rocks without a distinctive marker unit. In that case it will be very difficult to ever determine a precise basis on which to correlate to the Wahweap Formation in its type area. However, ongoing analysis of vertebrates, ostracodes, and palynomorphs may resolve some of these problems in the near future.

The "white sandstone" in the roadcut correlates petrographically and stratigraphically with the middle sandstone member of the Grand Castle Formation of Goldstrand and Mullett (1997). This fluvial sandstone is cross-bedded and contains large convoluted bedding. Intraclasts of white claystone (possible kaolinite) and carbonized wood debris are also present.

A petrographic sample analyzed from this location has a quartz-feldspar-lithic percentage (QFL%) of 76-0-24 (Goldstrand, 1991). This QFL% is nearly identical to the mean QFL% of the Grand Castle Formation of 76-1-23 (n=34). Lithic fragments in this sandstone are 100% carbonate and silicified carbonate lithic grains (Goldstrand, 1991; 1992).

This sandstone can be traced below the Claron Formation northward where the sandstone thickens to 633 feet (193 m) in Ashdown Creek below Cedar Breaks (Goldstrand, 1991; Goldstrand and Mullett, 1997). South of Websters Flat, the Grand Castle Formation pinches out, and the Claron Formation directly overlies Cretaceous rocks. Except for a few locations, the Grand Castle Formation is absent on the south Markagunt Plateau, suggesting this region was upwarped between deposition of the early Paleocene Grand Castle Formation and the late Paleocene onlap of the Claron sediments. Goldstrand (1994) refers to this Laramide-style upwarp as the Markagunt-Paunsaugunt Upwarp.

Pollock (1999) took two samples from the "white sandstone" which he found to consist of 100% well-rounded monocrystalline quartz grains; this is very similar to composition of the capping sandstone member of the Wahweap Formation in the Kaiparowits region to which he correlated the unit. However, the lateral tracing of this unit into the type area of the Grand Castle Formation by Goldstrand casts doubt on this interpretation. Also, note the description of Interval V in Moore and Straub (this volume) which indicates that lateral to the very clean deposits at Websters Flat the "white sandstone" is variable in composition and more heterolithic throughout most of the 96 foot (29 m) thick unit.

0.1	15.3	More of the "white sandstone."
0.4	15.7	Claron Formation landslide debris.
0.5	16.2	More of the "white sandstone" in place.
0.2	16.4	Claron Formation in place.
0.1	16.5	Mile 16 signpost.
0.5	17.0	Scenic view of Kolob Terrace and down the Deep Creek drainage.
1.5	18.5	TURN LEFT onto Utah Highway 148 North to Cedar Breaks National Monument. Claron Formation on the sides of the road.
1.0	19.5	Mile 1 signpost.
1.5	21.0	Entering Cedar Breaks National Monument. Hatfield et al. (2000) should be used as a guide to the geology of the monument. The only stop in the monument (STOP 8) will be made where additional data are now available.
1.1	22.1	Cedar Breaks National Monument visitor center.
0.9	23.0	Sunset View turnoff.
0.3	23.3	Roadcut in the upper white member of the Claron Formation.
0.8	24.1	Chessman Overlook.
1.1	25.2	Alpine Pond trailhead.
0.2	25.4	Driving on the Brian Head Formation. Brian Head is visible to the north.
0.8	26.2	Continue on Utah Highway 143 West, past Utah Highway 143 East to Panguitch.
0.4	26.6	North View overlook.
0.9	27.5	Entering Dixie National Forest.

STOP 8 (Eaton). The Brian Head Formation.

The Brian Head Formation was originally named by Gregory (1944) for a series of volcaniclastic rocks, well exposed at Brian Head (figure 17) on the Markagunt Plateau, overlying the Claron Formation (see Sable and Maldonado, 1997; Hatfield et al., 2000, for more complete discussion of the history of this formation). The range of ages from the Markagunt Plateau and Red Hills area to the west of between 32-35 Ma (Sable and Maldonado, 1997), indicates the formation is late Eocene in age (Chadronian North American Land Mammal "Age"). However, charophytes (Feist et al., 1997) and vertebrates (Eaton, Hutchison et al., 1999) indicate a latest middle Eocene age of approximately 38 Ma (Duchesnean North American Land Mammal "Age") for the lowest part of the Brian Head Formation (the "variegated unit") on the Sevier Plateau 38 miles (60 km) to the northeast of the type section. The "variegated unit" overlies a basal conglomerate that is very thick and well developed on the Sevier Plateau. Although a conglomerate is present near the base of the Brian

Figure 17. View looking northeast at the type section of the Brian Head Formation.

Head Formation in the type section (unit 5 of Sable and Maldonado, 1997), it is unclear if this is a temporal equivalent of the one on the Sevier Plateau and in the type area there is no evidence of the "variegated unit" which is more than 150 feet (50 m) thick on the west side of the Sevier Plateau (Eaton, Hutchison et al., 1999). Furthermore, almost 275 feet (90 m) above the base of the Brian Head Formation on the Sevier Plateau, vertebrates and charophytes still yield a latest middle Eocene age. It may be that subsidence began earlier on the Sevier Plateau than on the Markagunt Plateau, leaving a more complete record of early Brian Head deposition.

Important to this dating of the Brian Head formation is the implications for the age of the underlying Claron Formation. The Claron is constrained between latest Paleocene (Goldstrand, 1991) and latest middle Eocene (Eaton, Hutchison et al., 1999) making the formation approximately coeval with other regional lacustrine deposits (e.g. Green River Formation).

0.5	28.0	Continue past junction with road (to right) that goes up onto Brian Head Peak.
0.3	28.3	Summit elevation of 10,400 feet.
0.2	28.5	Exposures of slumped Brian Head Formation.
0.1	28.6	Entering town of Brian Head.
0.6	29.2	Late Tertiary volcanic breccia in road cut, probably Baldhills Tuff Member of the Isom Formation (Oligocene).
2.1	31.3	Begin 13% grade and switchbacks.
0.4	31.7	Claron Formation exposed in road cuts.
0.3	32.0	Contact between the Claron Formation and tan-white sandstones and conglomerates of the Grand Castle Formation exposed in road cut.
0.1	32.1	Claron(?) landslide debris covering the Grand Castle Formation.
0.6	32.7	Straight ahead is a view (figure 18) of the middle and upper members of the Grand Castle Formation.

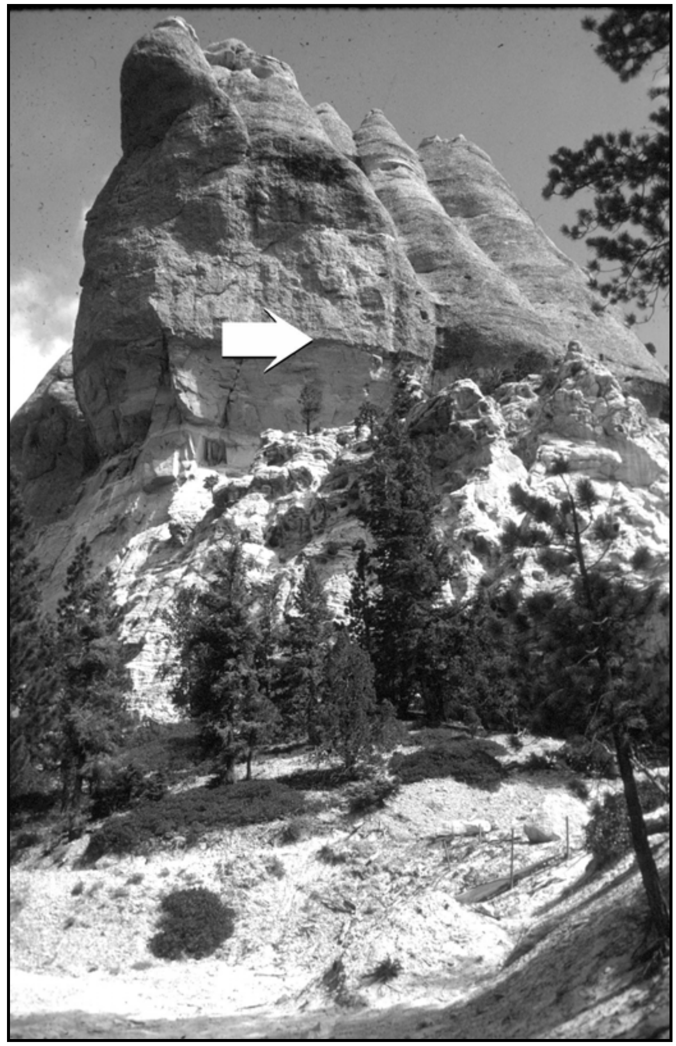

Figure 18. View to the west of the contact (shown by the arrow) between the middle sandstone and upper conglomerate members of the Grand Castle Formation.

0.1	32.8	Mile 12 signpost.
0.6	33.4	Volcanic landslide debris.
0.2	33.6	PULL OFF to the right by a small dirt road.

STOP 9. Hike up to contact between the middle sandstone member of the Grand Castle Formation and the overlying conglomerate member (see Goldstrand and Mullett, 1997, for detailed discussion).

CONTINUE WEST on Utah Highway 143.

0.3	33.9	Road cuts in middle sandstone member of the Grand Castle Formation.
0.3	34.2	Entering Parowan Quadrangle and geologic map of Maldonado and Moore (1995).
0.1	34.3	Crossing Quaternary landslide debris and fault (down to west).
0.1	34.4	Mostly Quaternary landslide debris to mileage 34.9.
0.5	34.9	Dry Lakes turnoff to left. Poorly exposed Baldhill Tuff Member of the Isom Formation (Oligocene) along the right side of road to mileage 35.1.

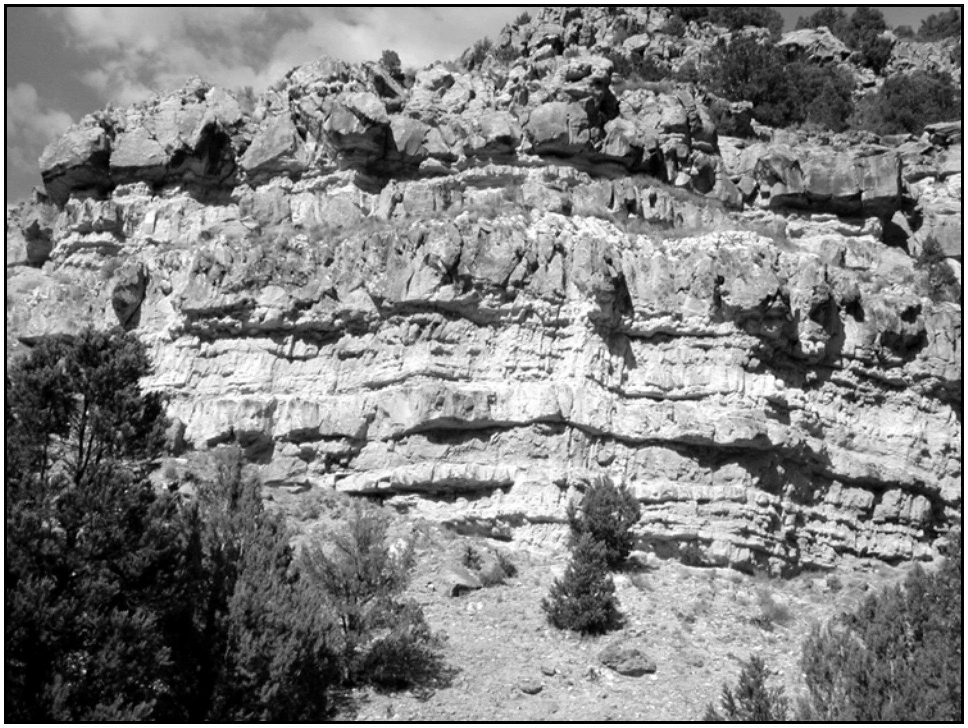

Figure 19. View looking east at outcrop of typical Iron Springs Formation 1,110 feet (364 m) below the top of the formation at Stop 11. The section is dominated by thick tabular sandstones and thin mudstone layers representing deposition on a fine-grained braid plain. The stratigraphic section of the Iron Springs Formation from here up to the Grand Castle contact probably only represent about the upper third of the formation in this area (estimated to be about 3,000 feet [1,000 m] thick by Maldonado and Moore, 1995).

0.2	35.1	Well exposed Baldhill Tuff Member (Oligocene) along road.
0.2	35.3	Straight ahead, visible on the high slopes, is the contact between the upper Iron Springs Formation (Cretaceous) and the Grand Castle Formation (Paleocene).
0.2	35.5	Exposures of the tan sandstones of the Iron Springs Formation on the northwest side of the road.
0.3	35.8	Faulted sliver of Claron Formation on the northwest side of the road. This is part of a complex northeast-southwest-trending fault system.
0.6	36.4	The volcanic knob to the east consists of mudflow and lava-flow breccias and tuffaceous sandstone (Oligocene) which overlies the Baldhills Tuff Member of the Isom Formation (Oligocene) which is exposed lower on the slope.
0.6	37.0	Note fault straight ahead that down-drops the Claron Formation.
1.3	38.3	Continue straight past turnoff to Second Left Hand Canyon (up this canyon are excellent exposures of the Claron, Grand Castle, and Brian Head Formations).
0.3	38.6	TURN RIGHT into First Left Hand Canyon to Vermillion Castle.
1.5	40.1	STOP 10 (Goldstrand). Type section of the Grand Castle Formation.

Here the Grand Castle Formation is 595 feet (181.4 m) thick, consisting of a lower conglomerate member, a middle sandstone member, and an upper conglomerate member (Goldstrand, 1991; Goldstrand and Mullett, 1997).

This sequence of conglomerate and sandstone had previously been mapped as the basal part of the Claron Formation in the Markagunt Plateau and Parowan Gap areas, but is now formally named the Grand Castle Formation (Goldstrand and Mullett, 1997). The Grand Castle differs significantly from the Claron Formation both compositionally and texturally. In contrast to the fine-grained carbonate rock and calcareous sandstone and shale of the Claron Formation, the Grand Castle Formation consists of siliciclastic conglomerate and sandstone. The contact between the Grand Castle and the Claron Formations is placed at the boundary where red, calcareous sandstone and siltstone of the Claron Formation become dominant. The contact of the Grand Castle with the underlying Upper Cretaceous Iron Springs Formation is marked by an abrupt change from sandstone or mudstone of the Iron Springs to conglomerate of the Grand Castle. In the western part of Parowan Gap, conglomerate of the Grand Castle Formation overlies fine-grained lithologies of various Cretaceous and Jurassic units with angular discordance.

The Grand Castle Formation is believed to be Paleocene in age. Along the western Markagunt Plateau, the Grand Castle grades upward into possible upper Paleocene strata of the Claron Formation. Strata correlative with the Grand Castle

Formation are present in the Table Cliff Plateau (northeast of Bryce Canyon), where they are stratigraphically overlain and underlain by lower Paleocene units.

The Grand Castle represents a Paleocene braided fluvial system, the sediments of which were derived from the inactive Wah Wah, Blue Mountain, and Iron Springs thrust sheets of the Sevier thrust belt (Goldstrand, 1991; 1992; 1994). Grand Castle conglomerates overlie the easternmost and youngest Sevier thrust faults (in Parowan Gap), indicating that the formation postdates the Sevier orogeny. Deposition of conglomerate and sandstone that compose the Grand Castle Formation was controlled initially by Laramide deformation within the Sevier foreland basin around a depocenter south of the town of Parowan (Goldstrand and Mullett, 1997). Continued Laramide deformation in the Table Cliff Plateau area tilted Grand Castle conglomerate prior to deposition of intermontane deposits of the Paleocene to middle Eocene Pine Hollow Formation (Goldstrand, 1991; 1994). Thus, strata of the Grand Castle Formation appear to have been deposited after the Sevier orogeny and before extensive partitioning of the foreland basin during the Laramide orogeny.
RETURN TO Utah Highway 143.

1.5	41.6	TURN RIGHT (west) on Utah Highway 143 toward Parowan. Iron Springs Formation is present along both sides of the road.
1.2	42.8	The ridge to the east is Lightning Ridge where the section in Eaton (1999) was measured. Along this ridge is a fossil locality (UMNH VP Locality 64) just below the Grand Castle contact, and the sample of biotite ash that was dated at 83.0±1.1 Ma (Eaton, Maldonado, and McIntosh, 1999) 712 feet (231 m) below the top of the Iron Springs Formation.
0.4	43.2	STOP 11 (Eaton). Pullout on left side of road (figure 19).

Outcrops here are typical of the Iron Springs Formation throughout the Parowan Canyon area. The sandstones form roughly tabular bodies interbedded with dark and thinner floodplain deposits. The sandstones, particularly at their bases, locally contain abundant fossil leaves. Vertebrates, mollusks (mostly gastropods), and ostracodes have been recovered by screen-washing the dark floodplain deposits. Vertebrates are currently under study by Eaton and ostracodes by Elizabeth Brouwers (USGS, Denver).

This part of the Iron Springs Formation represents a fine-grained braid plain deposited in a basin that was subsiding rapidly enough to preserve organic-rich floodplain mudstones.

0.1	43.3	Cross fault down-dropping Claron Formation on the west side of the fault.
0.3	43.6	Hills on east side of the road are the Brian Head Formation and on the west side of the road the Claron Formation. Oligocene mud flows and lava flow breccias and the Baldhills Tuff Member of the Isom Formation are also exposed at the mouth of the canyon along the south side of the road.
0.6	44.2	Fault to the east down-drops the Claron Formation.
0.3	44.5	Entering the town of Parowan.
0.8	45.3	Junction Highway 214 and I-15 business loop. TURN LEFT onto Main Street - South I-15 Business Loop.
0.2	45.5	TURN RIGHT following Utah Highway 143, Business Loop South to Cedar City.
2.4	47.9	TURN LEFT onto I-15 South freeway entrance.
15.5	63.4	EXIT 62 LEFT - Cedar City exit.
0.3	63.7	TURN LEFT at bottom of exit ramp towards Cedar City.
3.2	66.9	Intersection of Main Street and Utah Highway 14 (east) - starting point of field guide.

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