Tectonic and sedimentary controls, age and correlation of the Upper Cretaceous Wahweap Formation, southern Utah, U.S.A.

Zubair Ali Jinnah

A thesis submitted to the Faculty of Science, University of the Witwatersrand, in fulfilment of the requirements for the degree of Doctor of Philosophy

Johannesburg
2011

Declaration

I declare that this dissertation is my own, unaided work. It is being submitted for the Degree of Doctor of Philosophy in the University of the Witwatersrand, Johannesburg. It has not been submitted before for any degree or examination in any other University

(Zubair Ali Jinnah)
On the 10th day of October 2011

Abstract

The Wahweap Formation is an ~400 m thick clastic sedimentary succession of fluvial and estuarine channel sandstones and floodbasin mudrocks that was deposited in western North America during the Late Cretaceous. It preserves important mammal, dinosaur, crocodile, turtle and invertebrate fossils that have been the subject of recent palaeontological investigations. The Wahweap Formation can be divided into lower, middle, upper, and capping sandstone members based on sand:mud ratios and degree of sandstone amalgamation. Facies analysis reveals the presence of ten facies associations grouped into channel and floodbasin deposits. Facies associations (FAs) from channels include: (1) single-story and (2) multistory lenticular sandstone bodies, (3) major tabular sandstone bodies, (4) gravel bedforms, (5) low-angle heterolithic cross-strata, and (10) lenticular mudrock, whereas floodbasin facies associations include: (6) minor tabular sandstone bodies, (7) lenticular interlaminated sandstone and mudrock, (8) inclined interbedded sandstone and mudrock, and (9) laterally extensive mudrock.

The lower and middle members are dominated by floodbasin facies associations. The lower member consists dominantly of FA 8, interpreted as proximal floodbasin deposits including levees and pond margins, and is capped by a persistent horizon of FA 3, interpreted as amalgamated channel deposits. FAs 4 and 6 are also present in the lower member. The middle member consists dominantly of FA 9, interpreted as distal floodbasin deposits including swamp, oxbow-lake and waterlogged-soil horizons. FAs 1, 2, 5, 6, 7, 8, and 10 are present in the middle member as well, which together are interpreted as evidence of suspended-load channels. The upper member is sandstone-dominated and consists of FAs 1, 2, 3, 5, 7, and 8. FAs 5 and 7, which occur at the base of the upper member, are interpreted as tidally influenced channels and suggest a marine incursion during deposition of the upper member. The capping sandstone is characterized by FAs 3, 4, and 6, and is interpreted to represent a major change in depositional environment, from meandering river systems in the lower three members to a low-accommodation, braided river system. Combined results of facies and palaeosol analyses suggest that the overall climatic conditions in which the Wahweap Formation was deposited were generally wet but seasonally arid, and that

conditions became increasingly moist from the time of lower member deposition up to the time of middle member deposition.

Improved age constraints were obtained for the Wahweap Formation by radiometric dating of two devitrified ash beds (bentonites). This allowed for deposition to be bracketed between approximately 81 Ma and 76 Ma. This age bracket has two important implications: firstly, it shows that the Wahweap Formation is synchronous with fossiliferous deposits of the Judithian North American Land Mammal Age, despite subtle differences in faunal content. Secondly, it shows that the middle and upper members were deposited during the putatively eustatic Claggett transgression (T8 of Kauffman 1977) in the adjacent Western Interior Seaway. This is consistent with facies analysis which shows a marked increase in tidally-influenced sedimentary structures and trace fossils at the top of the middle and base of the upper members. Following recent alluvial sequence stratigraphic models, the middle member is interpreted as the isolated fluvial facies tract, while the upper member represents the tidally influenced and highstand facies tracts. Maximum transgression occurred during deposition of the lowest part of the upper member, synchronous with the Claggett highstand in other parts of the Western Interior Basin. The sequence boundary is placed at the base of the overlying capping sandstone member, diagnosed by a major shift in petrography and paleocurrent direction, as well as up to 4 m of fluvial incision into the underlying upper member. The capping sandstone member is interpreted as the amalgamated fluvial facies tract of an overlying sequence.

Analysis of the western-most exposures of the Wahweap Formation on the Markagunt and Paunsaugunt plateaus shows facies variations in the proximal and distal parts of the central Western Interior Basin. The inconsistent thickness and variations in fluvial architecture, as well as the presence of unconformities and generally poor exposure in the west, hinder correlation attempts and also prevent the subdivision of the Wahweap Formation into members. Only the capping sandstone, which can be positively identified west of the Paunsaugunt fault, has a consistent thickness and fluvial architecture across the west-east extent of the Wahweap Formation. The capping sandstone also bears remarkable lithological similarity to the Tarantula Mesa Formation which is exposed to the east in the Henry Mountains Syncline, and it is suggested that these two units be equated under the name “Tarantula Mesa Formation”, which has precedence.

Acknowledgements

I am indebted to many people who have helped me during the time I have spent on this dissertation. Firstly, to my supervisors, Eric Roberts, who gave me the opportunity to work on this exciting project and also provided a great deal of meticulous feedback, and Paul Dirks who administered financial support through the BHP Billiton Academic Fellowship, which has provided many exciting research opportunities to me. Lulu Khumalo and Nomfundo Mqadi from BHP Billiton are also thanked for facilitating generous research funding.

In the field I was lucky to spend time with palaeontologists from the Utah Museum of Natural History – Mike Getty, Mark Loewen, Scott Sampson, Bucky Gates, Eric Lund, Jelle Wiersma, Sue Beardsmore, Deanna Brandau and Joe Sertich, who provided both useful scientific discussion and excellent company. Pat O'Connor and Leif Tapanila have also had a very valuable, if brief, presence during fieldwork. Steve Roberts has also provided much support during my field seasons.

Alan Titus, Marietta Eaton and Doug Powell from the Bureau of Land Management, and Joe Rechsteiner from the Forest Service have all been tremendously helpful in providing research and collection permits.

In the School of Geosciences at Wits University I have been supported by Safia Cannell, Jamila Das, Lenah Mtiyane and Roger Gibson with all things budget-related. Alex Mathebula is acknowledged for thin section preparation and Janine Robinson for XRF analyses. Matt Kitching and Mitch Miles directed me towards all the laboratory supplies and equipment I needed. Dave Billing and his team of postgraduates in the School of Chemistry have also been invaluable in facilitating my XRD analyses.

Finally, I am grateful to Valerie Nxumalo for her support and the input she has provided during my write-up.

Table of Contents

List of Figures

List of Tables

1. Introduction

1.1. Background

The Kaiparowits Plateau in Grand Staircase-Escalante National Monument (southern Utah) preserves a 1200 m thick succession of Upper Cretaceous rocks deposited in marine, marginal marine and continental settings. This succession preserves important marine and terrestrial ecosystems and has been the focus of an increasing amount of palaeontological work in the past two decades. The terrestrially-deposited rocks (specifically the Wahweap and Kaiparowits formations) were initially the subject of palaeontological investigation by workers focusing on mammalian faunas (Eaton and Cifelli, 1988; Cifelli, 1990). More recently, in the past decade, palaeontologists from the Utah Museum of Natural History (e.g. Sampson *et al*. 2002, 2010), Bureau of Land Management (e.g. Titus *et al*. 2008) and the Utah Geological Survey (e.g. Kirkland and DeBlieux, 2005) have turned attention to the dinosaurs and associated fauna that is preserved in the Wahweap and Kaiparowits formations. This in turn has led to a number of related studies in geochronology, sedimentary palaeoenvironments, taphonomy and palaeofloral analysis, among other fields, to provide a more complete understanding of these Late Cretaceous ecosystems.

1.2. Project outline

In this project, new data on facies architecture, radiometric dates, thickness trends, sandstone composition and geochemistry are used to place the Wahweap Formation into a geochronological and sequence stratigraphic context, as well as reconstruct the Wahweap depositional system.

In light of the important Campanian mammal fossils recovered from the Wahweap Formation (Cifelli and Eaton, 1988; Eaton, 2002, 2006), as well as its rapidly emerging dinosaur fauna (Sampson *et al*., 2002; Kirkland and De Blieux, 2005, 2010), and trace fossil evidence of dinosaur and mammal behaviour (E. Simpson *et al.* 2010), detailed environmental reconstruction and age constraints are necessary to allow for faunal comparisons and test palaeoecological and palaeobiogeographic hypotheses. In particular, hypotheses such as latitudinal variation in Western Interior Basin faunas (Rowe *et al.*, 1992; Zanno, 2005; Sampson et al., 2010) require detailed age constraint to make valid comparisons. When comparing faunas from different palaeolatitudes in order to determine the extent of difference or similarity, as has been done for the north (Montana-Alberta) and south (Texas-New Mexico) of the Western Interior Basin, it is essential that the strata compared overlap temporally in order to ensure that differences are due to latitudinal variation rather than faunal turnover through time. High-resolution absolute ages are required for this sort of work. Although Utah occupies an important position in the central Western Interior Basin, until recently there were no absolute age dates available for the terrestrial Upper Cretaceous section, and thus correlation depended on biostratigraphy. The work of Roberts *et al.* (2005) provided the first ^{40}Ar-^{39}Ar age dates from devitrified volcanic ash units in the Kaiparowits Formation, and provided a correlation point for the Upper Cretaceous section of southern Utah, which could now be compared with other age-constrained units.

Additionally, the Wahweap Formation was deposited during an important time in the southern Western Interior Basin, just before or during the onset of Laramide Orogeny (see section 1.3.2.2., Tindall *et al*. 2010) and a detailed history of basin evolution during Wahweap time is essential to elucidate regional Late Cretaceous tectonic, climatic and eustatic events. For example, outcrop of the Wahweap Formation extends across several older fault zones (the Sevier fault and the East-Kaibab monocline) which may have been reactivated numerous times either syndepositionally or soon after deposition. Examination of Wahweap Formation thickness trends and sedimentation style can help elucidate whether Late Cretaceous movement on these fault systems partitioned the Western Interior Basin in southern Utah and led to the development of separate depocentres with differing accommodation space or whether Wahweap Formation deposition was limited to a single depocentre in which the geometry of the basin controlled accommodation space.

This thesis consists of seven chapters: chapter 1 contains an introduction to the Western Interior Basin and the study area in southern Utah. Chapter 2 contains the results of a facies associations study, which was published in the Journal of Sedimentary Research (Jinnah and Roberts, 2011). Chapter 3, which contains the first radiometric ages for the Wahweap Formation, was published in Cretaceous Research (Jinnah *et al*. 2009). Chapter 4, which deals with sequence stratigraphy is contained together with Chapter 2 in Jinnah and Roberts (2011). Chapter 5, which deals with lateral variability in the Wahweap Formation and basin development, has been accepted for publication in the edited book, "Late Cretaceous Geology and Paleontology of Southern Utah: A Critical Window into End-Mesozoic Ecosystems" (edited by A. Titus and M. Loewen; Indiana University Press). Chapter 6 deals with palaeosol development across the proximal-distal extent of the Wahweap Formation, and chapter 7 summarizes the important implications of this work.

1.3. Geological setting

1.3.1. The Western Interior Basin

Development of the Western Interior retroarc foreland basin began in the Late Jurassic as a result of subduction of the Farallon Plate under the North American Plate (Cross, 1986; White *et al.*, 2002). Sedimentation continued in the basin for over 100 million years but finally ceased in the Eocene before the onset of basin and range extension (Cross, 1986; DeCelles, 2004; Miall *et al*., 2008). At its greatest extent, the Western Interior Basin stretched from the Gulf of Mexico to Arctic Canada, and it hosted an enormous epicontinental sea that covered much of central North America (the Western Interior Seaway; Fig. 1.1.) during the Jurassic and Cretaceous (Kauffman, 1977; Miall *et al*., 2008).

The earliest sedimentary units within the Western Interior Basin are the Upper Jurassic Morrison Formation in the western U.S.A. and the equivalent Fernie and Kooteney formations in Canada. At this time, uplifted areas in the thrust belt to the west of the basin were established as the major sediment source. It has been suggested that the Jurassic fill of the basin was originally between 2 and 5 km thick (Royse, 1993), but that much of the Jurassic deposits were later caught up in, and recycled by, thrust-belt uplift and erosion during Cretaceous times, resulting in incomplete preservation (Miall *et al*., 2008).

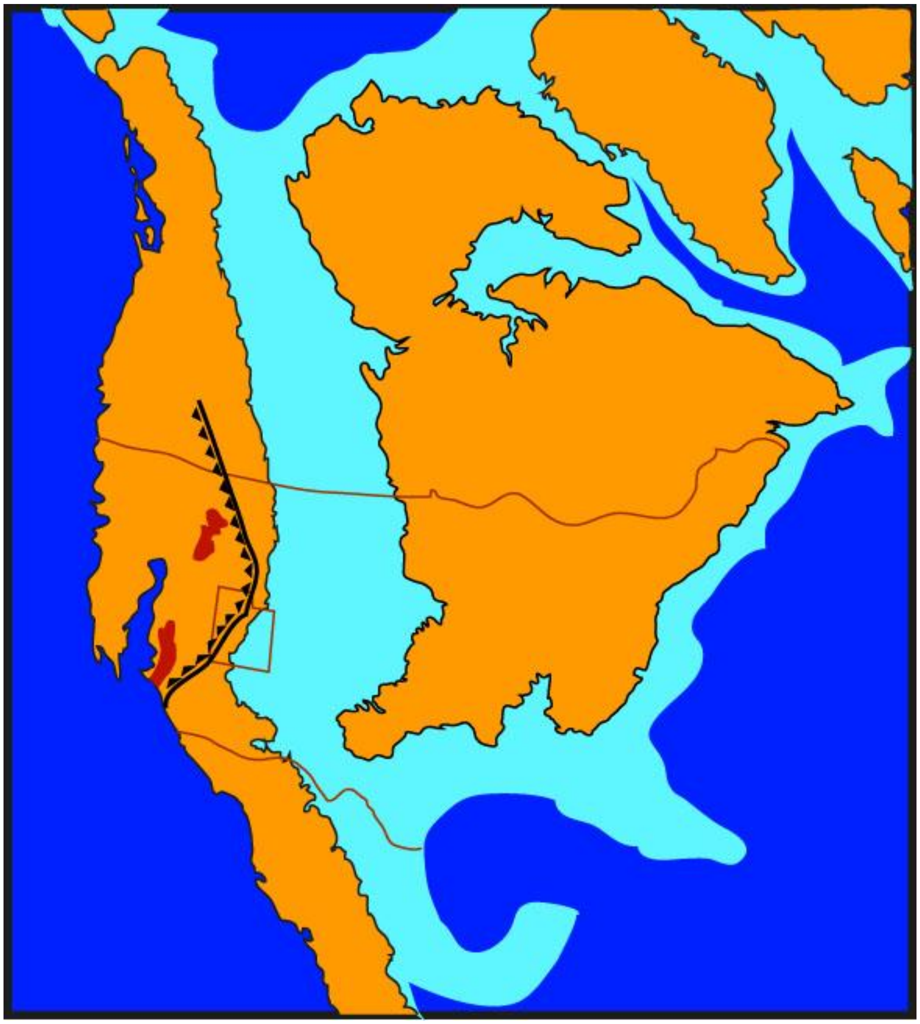


Figure 1.1. The Western Interior Seaway developed in Cretaceous North America (after Blakey, 1997). The Sevier fold-and-thrust belt (black line) and Cordilleran arc (red) are also shown.

The Morrison Formation is believed to have accumulated in the backbulge part of the basin (Miall *et al.*, 2008), which extended from eastern Utah to North Dakota. At this time the forebulge was located in central Utah (DeCelles and Currie, 1996). However, the forebulge of the Western Interior Basin migrated eastward as time went by (Catuneanu *et al.,* 2000; White *et al*., 2002), resulting in a regional unconformity through part of the Lower

Cretaceous succession (Dyman *et al.*, 1994). This unconformity is commonly overlain by Berriasian or Aptian age conglomerates believed to be related to uplift in the magmatic arc (Heller and Paola, 1989). At this time, the Western Interior Seaway, which had been restricted to the northern part of the basin in the earliest Cretaceous, expanded southwards (Blakey, 2008).

During the Cretaceous, nine cycles of transgression and regression occurred in the Western Interior Seaway (Kauffman, 1977; Kauffman and Caldwell, 1993; Fig. 1.2.), with those transgressions of the Early Cretaceous limited to the Canadian part of the basin (Miall *et al*., 2008). Much of our present understanding of the effect of base level change on fluvial systems has come from work in the northern and central part of the Western Interior Basin (Montana-Wyoming) where the study of laterally equivalent and inter-fingering terrestrial and marine deposits (Weimer, 1960; Gill and Cobban, 1973) has led to the recognition of four major transgressive-regressive events in the Western Interior Seaway during the Late Cretaceous (Kauffman, 1969, 1977; Weimer, 1986). Sequences deposited during these events are referred to as the Greenhorn, Niobrara, Claggett and Bearpaw cyclothems.

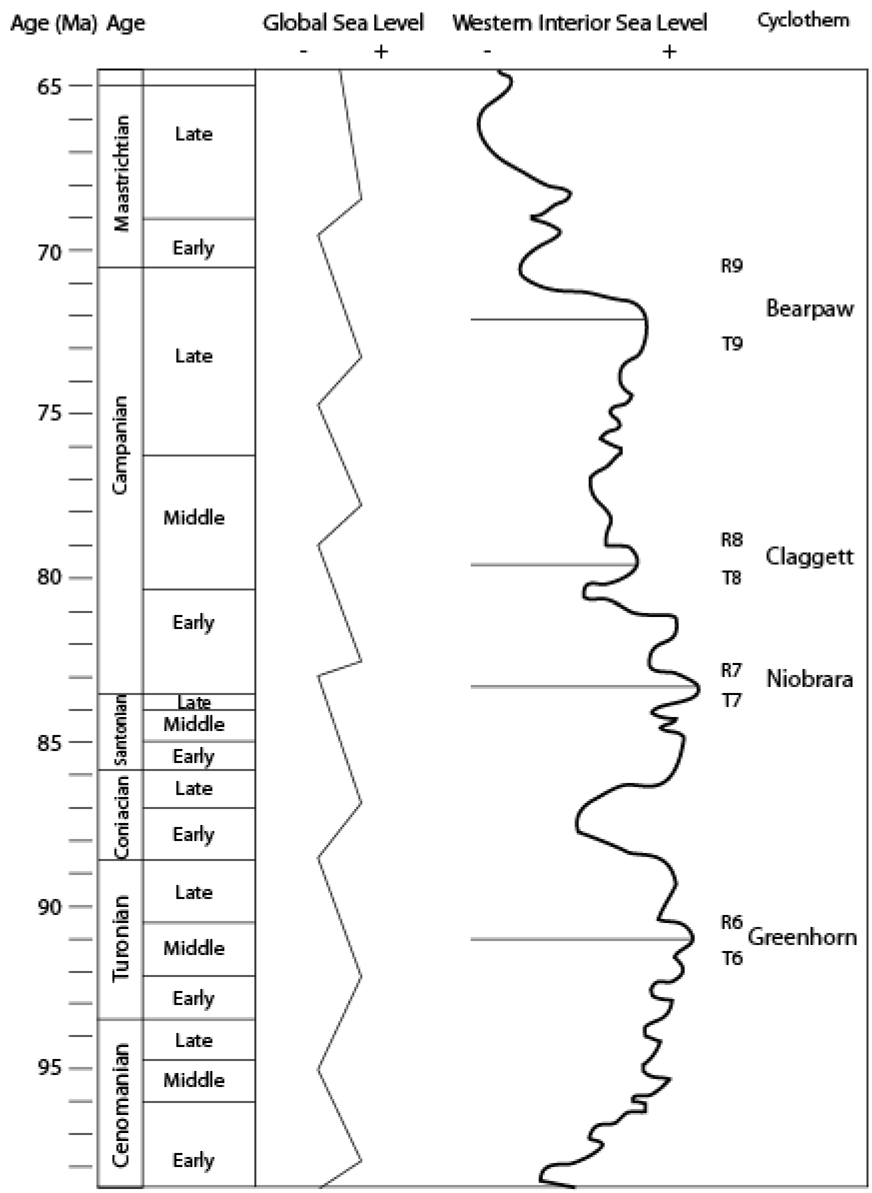


Figure 1.2. Cycles of the Western Interior Seaway. Timescale after Gradstein et al. (2004). Global sea level after Ogg and Ogg (2008). Cycles after Kauffman and Caldwell (1993).

Sedimentation in the Western Interior Basin terminated in the Eocene due to post-orogenic uplift, caused by the decoupling of subducted and overlying lithosphere (Cross, 1986; Miall *et al.*, 2008). During this uplift event, the Colorado Plateau, which was underpinned by subducted lithosphere and thus had a doubled crustal thickness, behaved as a rigid block

(Cross, 1986), while basin and range extension, which began in the Oligocene (Tristán-González *et al*., 2009) added an additional structural overprint to the sedimentary accumulations west of the Colorado Plateau.

1.3.1.1. Upper Cretaceous strata of the western margin of the seaway

Outcrop of Upper Cretaceous strata are more abundant in the Western Interior than any other age; hence Late Cretaceous sedimentation is better studied than any for other time period in western North America. Nonetheless, limited biostratigraphic correlation, incomplete continental-marine transects and ambiguous along-strike facies variations have resulted in a number of key hypotheses remaining unaddressed.

Upper Cretaceous terrestrial and marine strata from the western margin of the Western Interior Seaway crop out in Alberta and Saskatchewan, Canada, as well as Montana, Wyoming, Colorado, Utah, New Mexico and Texas in the U.S.A. Extensive studies in Canada and the northern United States (Gill and Cobban, 1973; Plint, 1991**;** Brinkman *et al*., 1998, Rogers 1998; Plint and Kreitner, 2007) identified that the relationship between inter-digitating marine and non-marine units, with eastward-thinning non-marine strata interbedded with westward-thinning marine "tongues", could be used to identify transgressive and regressive phases of deposition. Such studies have been undertaken in the Niobrara Formation, the Claggett Formation and its non-marine equivalents (Two Medicine and Judith River formations) and the Bearpaw Formation. This work, together with similar work in central Utah (the Book Cliffs) by Olsen *et al.* (1995), Van Wagoner (1995), McLaurin and Steel (2000), Willis (2000), Miall and Arush (2001), Adams and Bhattacharya

(2005) and many others, as well as southern Utah by Shanley and McCabe (1991, 1993, 1995) and Lawton *et al*. (2003) provide a basis for understanding sequence stratigraphy in both non-marine settings and foreland basins in general.

1.3.2. Stratigraphic and structural setting of the Upper Cretaceous of southern Utah

In southern Utah, the Western Interior Basin was flanked on the west and northwest by the Sevier fold-and-thrust belt (Fig. 1.3.), to the southwest by the Cordilleran magmatic arc and to the south by the Mogollan highlands (Bilodeau, 1986; Lawton *et al*., 2003; DeCelles and Coogan, 2006). These features, together with the Hurricane and Sevier faults, as well as the East Kaibab monocline (northeast-extending Precambrian faults which ran across the basin and have been reactivated many times throughout their histories [Huntoon, 1993; Tindall and Davis, 1999; Tindall *et al*. 2010]), all controlled aspects of sedimentation in the basin. Fluvial systems were established which drained from the highlands in the west and south, into the seaway to the east. The Upper Cretaceous record consists of near-shore, marginal marine and fully terrestrial depositional systems which were deposited from the Cenomanian to the Maastrichtian.

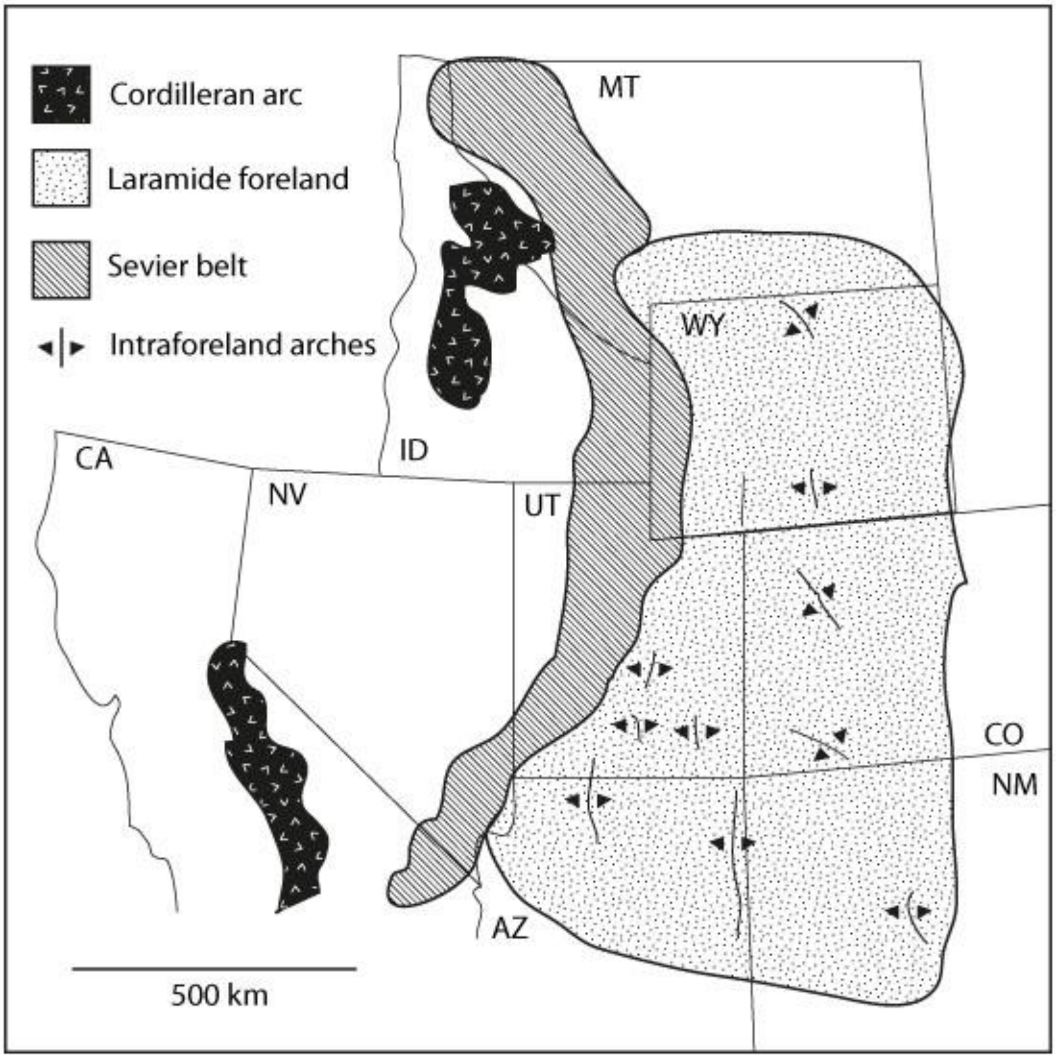


Figure 1.3. Position of the Sevier Belt and Cordilleran arc in Western Interior North America (after DeCelles, 2004)

Upper Cretaceous strata in southern Utah occur in several separate structural realms (Fig. 1.4.). The westernmost strata are exposed west of the Hurricane Fault (Goldstrand, 1992;

Lawton *et al*., 2003), whereas strata of the Markagunt Plateau occur between the Hurricane and Sevier faults.

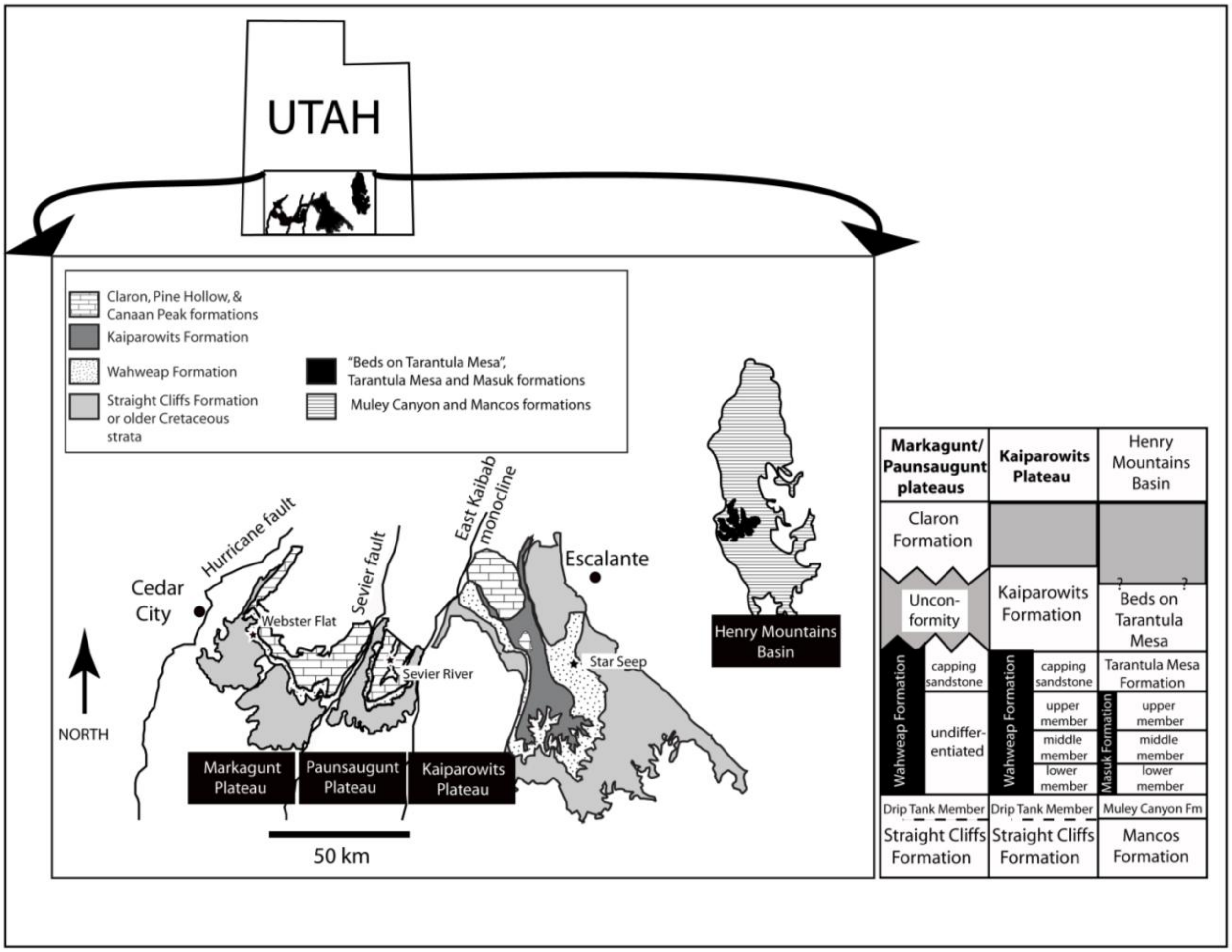


Figure 1.4. Upper Cretaceous outcrop of southern Utah. After Eaton (1990) and Lawton et al. (2003).

The Paunsaugunt Plateau, between the Sevier fault on the west and the Paunsaugunt fault on the east has a discontinuous Upper Cretaceous record whereas patches of Upper Cretaceous strata crop out between the Paunsaugunt fault and the East Kaibab monocline. The

Kaiparowits Plateau, east of the East Kaibab monocline has extensive and continuous exposures of Upper Cretaceous strata preserved in a large syncline (Fig. 1.5.). The easternmost Upper Cretaceous terrestrial strata in southern Utah are found close to the Henry Mountains.

1.3.2.1. Sevier-style orogeny

Sevier-style orogeny, or thin-skinned thrusting (Fig. 1.6.), refers to events in the thrust belt which involve thrusts that are present in sedimentary cover only and do not penetrate into crystalline basement rocks (Armstrong, 1968; Burchfiel and Davis, 1975). This specific style of thrusting is believed to be a result of the nature of subduction, and is caused by conditions where the subducted plate has a high angle of subduction (Cross and Pilger, 1982; Cross, 1986). The Sevier fold-and-thrust belt consists of a zone of décollement thrusts which shallow at depth, as well as associated folds (Armstrong, 1968; Cross, 1986; DeCelles and Coogan, 2006) and is 2000 km long (running roughly southwest to northeast) and 300 km wide. Sevier structures are found on the Palaeozoic-Mesozoic sedimentary prism at the craton-margin adjacent to the foreland system, but not on the craton itself (Cross, 1986).

In Utah, Sevier thrusts cut through Neoproterozoic quartzite and argillite, and Palaeozoic carbonates and shale (DeCelles and Coogan, 2006). In central Utah, the Gunnison and Paxton thrusts mark the easternmost extent of the Sevier thrust belt. Sevier-style thrusting began late in the Early Cretaceous (Cross, 1986; Dickinson and Gehrels, 2008).

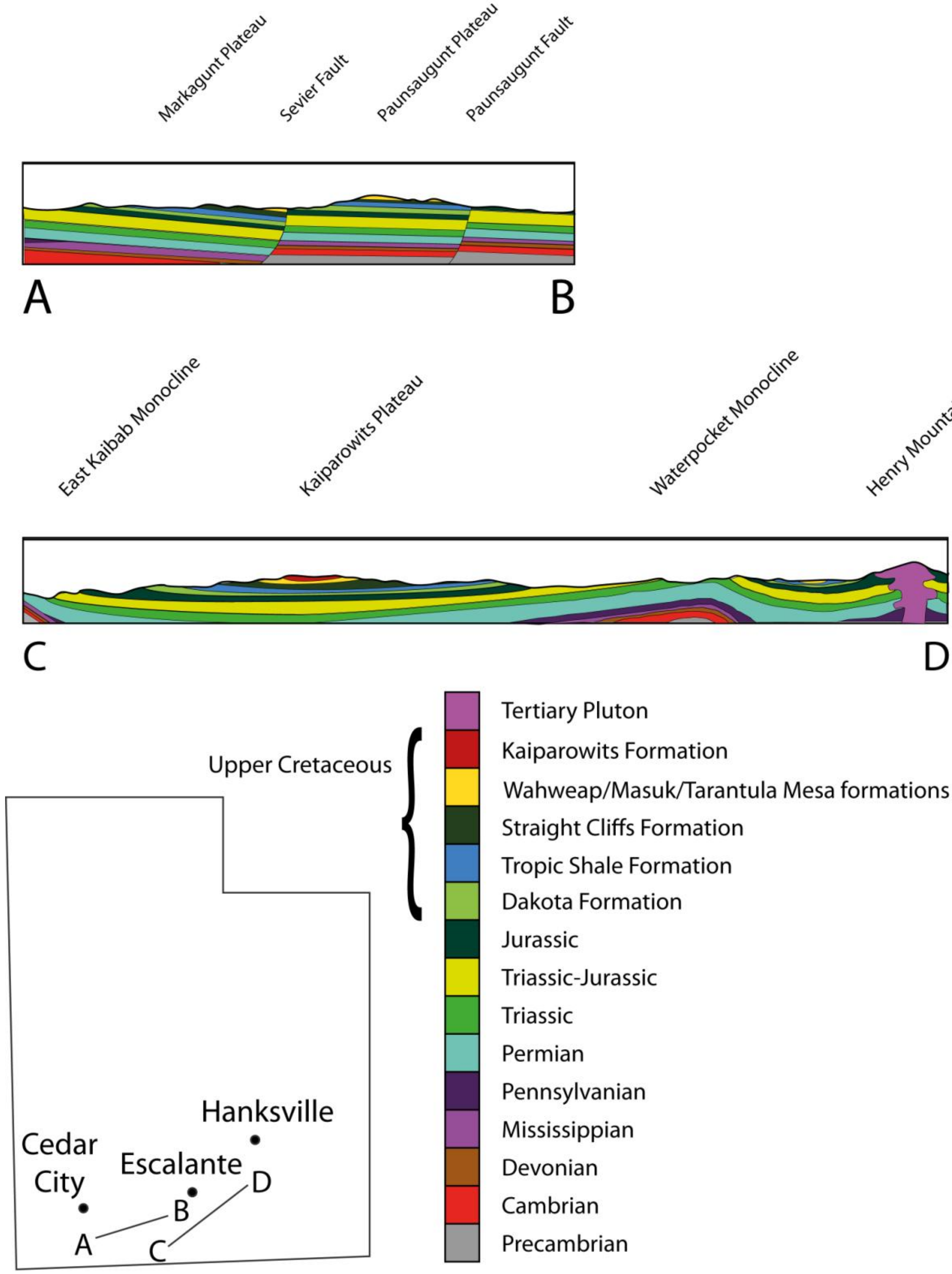


Figure 1.5. Cross-sections through southern Utah. Vertical exaggeration 2 times. After Hintze (1997).

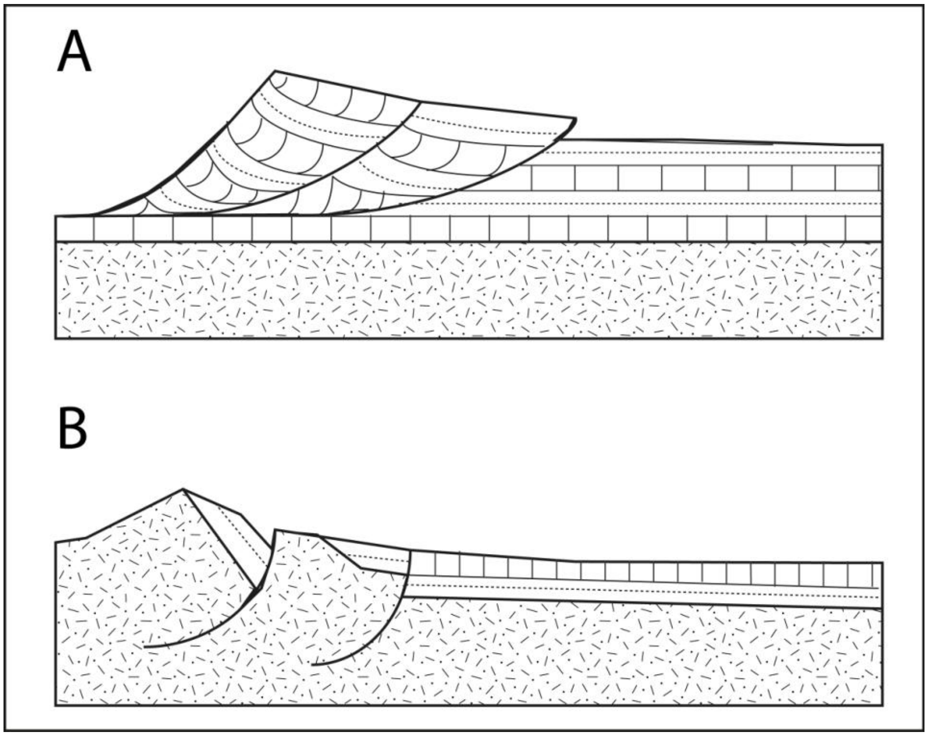


Figure 1.6. A. Sevier and B. Laramide style thrusting. Lower stippled unit represents igneous basement and upper stratified units represent sedimentary cover.

1.3.2.2. Laramide-style orogeny

In contrast to Sevier-style orogeny, Laramide style-orogeny, or thick-skinned thrusting, refers to thrust faults that cut through both the sedimentary cover and crystalline basement (Armstrong, 1968; Cross, 1986; Lawton, 1986). Movement along these structures resulted in basement-cored uplifts that are typically associated with basin partitioning and separation of depocentres in the Western Interior Basin (Fouch *et al.*, 1983; Goldstrand, 1992). Laramide-style orogeny in the Western Interior Basin resulted in the development of a number of basins separated by uplifted blocks of crystalline basement (Dickinson *et al.*, 1988; Goldstrand, 1994).

The transition from Sevier to Laramide style orogeny remains a widely debated topic, but is generally thought to have occurred in Utah and Wyoming in the Campanian or Maastrichtian (Lawton, 1986; Dickinson *et al*., 1988; Tindall *et al*. 2010) and is believed to have been driven by a shallowing of subduction angle caused by the subduction of a low-density aseismic ridge (Cross and Pilger, 1978). Laramide structures are found on the craton (Cross, 1986), and Laramide-style thrusting is limited to areas in the central and southern Rocky Mountains (northern New Mexico to southern Montana). Sevier-style thrusting continued contemporaneously to the north and south (Cross, 1986).

1.3.2.3. The Markagunt and Paunsaugunt plateaus

Sedimentary rocks of the Markagunt Plateau generally dip northeastward and terminate against the Sevier fault to the east (Sable and Hereford, 2004). The oldest sedimentary rocks on the Markagunt Plateau are the Triassic Moenkopi and Chinle formations which crop out adjacent to the Hurricane fault zone that extends along the western margin of the plateau. They are unconformably overlain by the Lower Jurassic Kayenta and Navajo formations that are in turn unconformably overlain by the Middle Jurassic Carmel Formation, which is the youngest part of the Jurassic succession (Table 1.1.). A major unconformity separates Jurassic rocks from the overlying Upper Cretaceous succession, which preserves a relatively continuous sedimentary record. The base of this succession is the Cenomanian Dakota Formation, followed by the Turonian-Coniacian Tropic Shale, Coniacian-Campanian Straight Cliffs Formation and capped by the Campanian Wahweap and Iron Springs formations (Sable and Hereford, 2004; Rowley *et al*., 2006).

Table 1.1. Stratigraphy of the Markagunt, Paunsaugunt and Kaiparowits plateaus, and the Henry Mountains syncline. After Sable and Hereford (2004), Titus et al. (2005) and Peterson and Kirk (1977).

Age	Markagunt Plateau	Paunsaugunt Plateau	Kaiparowits Plateau	Henry Mountains
TERTIARY	Claron	Claron	Claron	
	Grand Castle		Pine Hollow	
CRETACEOUS			Canaan Peak	
			Kaiparowits	beds on Tarantula Mesa
	Iron Springs/ Wahweap	Wahweap	Wahweap	Tarantula Mesa Masuk
	Straight Cliffs	Straight Cliffs	Straight Cliffs	Muley Canyon
	Tropic Shale	Tropic Shale	Tropic Shale	Mancos Shale
	Dakota	Dakota	Dakota	Dakota
			Cedar Mountain	Cedar Mountain
JURASSIC			Morrison	
			Entrada	
	Carmel	Carmel	Carmel	
	Navajo	Navajo	Navajo	Navajo
	Kayenta	Kayenta	Kayenta	Kayenta
			Wingate	Wingate
TRIASSIC	Chinle		Chinle	
	Moenkopi		Moenkopi	
PERMIAN			Kaibab	
			Toroweap	

Campanian-aged rocks are overlain by the Palaeocene Grand Castle and Eocene Claron formations (Goldstrand, 1992, 1994). Tertiary volcanics overlie the sedimentary record across much of the plateau (Sable and Hereford, 2004).

There are several competing nomenclatures for subdivision of the Straight Cliffs and Wahweap formations on the Markagunt Plateau. Sable and Hereford (2004) subdivide the Straight Cliffs into a lower and upper member and leave the Wahweap Formation undivided, whereas Nichols (1997) subdivides the Straight Cliffs into a lower and upper unit. Eaton *et al*. (2001) divide the Straight Cliffs Formation on the Markagunt Plateau into the basal Tibbet Canyon Member, overlain by the Smoky Hollow Member, which is in turn overlain by an “upper unit” in which they tentatively identified strata equivalent to the John Henry and Drip Tank members of the Kaiparowits Plateau.

The Paunsaugunt Plateau lies east of the Markagunt Plateau and is separated from it by the Sevier fault. Strata are preserved in a shallow syncline with an axis that extends northeast-southwest, the centre of which is eroded by the east fork of the Sevier River, so that older rocks are exposed in the syncline centre as well as on the flanks (Bowers, 1990). There is no exposure of Triassic rocks at the base of the succession on the Paunsaugunt Plateau; however the Jurassic and Upper Cretaceous succession are similar to that on the Markagunt Plateau. Only the Wahweap Formation is found at the top of the Upper Cretaceous section, not the equivalent Iron Springs as on the Markagunt Plateau. Additionally, the Wahweap Formation on the Paunsaugunt Plateau is much thinner than on the Markagunt Plateau and the upper unit is typically not present (Doelling and Willis, 1999; Pollock, 1999; Sable and Hereford, 2004). The Upper Cretaceous section is overlain unconformably by the Claron Formation, as the Grand Castle Formation is not present on the Paunsaugunt Plateau (Bowers, 1990; Goldstrand, 1994).

The eastern flank of the Paunsaugunt Plateau is bordered by the Paunsaugunt fault (Bowers, 1990), west of which there is a discontinuous Jurassic-Upper Cretaceous succession which bears more resemblance to the strata of the Paunsaugunt Plateau than it does to the eastward lying Kaiparowits Plateau.

1.3.2.4. The Kaiparowits Plateau

Strata of the Kaiparowits Plateau are preserved in the Table Cliffs syncline and the Upper Valley anticline. Although Cambrian to Permian units are known from subsurface studies, the oldest exposed units on the Kaiparowits Plateau are of Permian age, including the Toroweap Formation and overlying Kaibab Formation. They are overlain by the Triassic Moenkopi and Chinle formations, which are in turn overlain unconformably by the Jurassic succession, which includes: the lowermost Wingate, Kayenta, Navajo, Carmel and uppermost Entrada Formation. The Upper Jurassic Morrison Formation unconformably overlies the Entrada Formation and is the first of the sedimentary units that were deposited in a foreland basin setting (Miall *et al.*, 2008). The Morrison Formation is unconformably overlain by the Lower Cretaceous Cedar Mountain Formation, followed by the Upper Cretaceous Dakota Formation, the Tropic Shale, Straight Cliffs Formation, Wahweap Formation, Kaiparowits Formation and Canaan Peak Formation (Doelling and Willis, 1999). The Palaeocene to mid-Eocene Pine Hollow Formation unconformably overlies the Canaan Peak Formation (Goldstrand, 1994) and the succession is capped by the Eocene Claron Formation.

The Straight Cliffs and Wahweap formations are both subdivided on the Kaiparowits Plateau. The Straight Cliffs was broken down by Peterson (1969) into the lowermost Tibbet Canyon Member, followed by the Smoky Hollow, John Henry and Drip Tank members, whereas the Wahweap Formation was informally divided by Eaton (1991) into lower, middle, upper and capping sandstone members. While attempts have been made to correlate the members of the Straight Cliffs Formation from the Kaiparowits Plateau to the Markagunt and Paunsaugunt plateaus (e.g. see Eaton *et al*., 2001), lithostratigraphic variability and ambiguous biostratigraphy have prevented a robust correlation from being developed.

1.3.2.5. The Henry Mountains

In the Henry Mountains Syncline, exposures of Jurassic Wingate, Kayenta and Navajo formations are overlain unconformably by the Cedar Mountain Formation, Dakota Formation and Mancos Shale. The Mancos Shale is equivalent to the Tropic Shale that occurs on the Kaiparowits Plateau (Eaton, 1990). The Muley Canyon Sandstone overlies the Mancos Shale, and is thought to be equivalent to the Drip Tank Member of the Straight Cliffs Formation on the Kaiparowits Plateau, while the Masuk Formation, which overlies the Muley Canyon, is equivalent to the lower, middle and upper members of the Wahweap Formation on the Kaiparowits Plateau (Peterson and Kirk, 1977; Eaton, 1990). The Tarantula Mesa Sandstone overlies the Masuk Formation in the Henry Mountains Basin, and correlates to the capping sandstone of the Wahweap Formation. The uppermost unit in the Cretaceous system of the Henry Mountains Basin is called the ‘Beds on Tarantula Mesa’, and is considered equivalent to the Kaiparowits Formation in the Kaiparowits Plateau (Eaton, 1990).

1.3.3. Non-marine sequence stratigraphy

Many principles of non-marine sequence stratigraphy and fluvial response to base level changes have been deduced from extensive study of the Book Cliffs in central Utah, where there is continuous outcrop along a transition from fully fluvial strata of the Castlegate Formation proximal to the Sevier fold-and-thrust belt to marginal marine and fully marine strata where the Western Interior Seaway developed distal from the fold-and-thrust belt (e.g. Yoshida et al., 1996; Miall and Arush, 2001; McLaurin and Steel, 2007). Through bio- and litho- stratigraphic correlation and the identification of synchronous units, as well as detailed environmental reconstruction, Campanian shorelines have been mapped and periods of regression and transgression identified. By studying fluvial units deposited during periods of known regression and transgression, researchers have been able to define fluvial systems tracts according to the alluvial architecture and depositional style (aggrading vs. amalgamating) expected as a result of base level changes and variable accommodation space (Miall and Arush, 2001; Adams and Bhattacharya, 2005). Comparison of predicted systems tract stacking with the observed pattern can clarify the primary control on fluvial behaviour: eustatic control is likely where the changes in stacking pattern match that predicted from global sea level curves or important surfaces such as sequence boundaries are traceable over large distances along-strike in a basin, whereas tectonic control is plausible where the stacking pattern does not match that predicted, or sequence boundaries cannot be traced along-strike.

Identification of sequence boundaries in fluvial systems relies on changes in depositional patterns that are thought to occur as base level drops. It has long been thought that a transition from mud-rich meandering river systems with isolated channel bodies to sand-rich braided river systems with extensive amalgamated sandstone bodies occurs across the non-marine sequence boundary. However, work by Adams and Bhattacharya (2005) has shown that a change in stream type (meandering to braided) is not necessarily resultant from a drop in base level, and the best indicator of the non-marine sequence boundary is a change in the sand-mud ratio and architecture of sandstone bodies (i.e. those inferred to form by aggradation with well-preserved floodplains vs. those formed by amalgamation with little floodplain preservation).

1.3.3.1. Sequence stratigraphic models for the Upper Cretaceous of Utah

Shanley and McCabe (1991, 1993, 1995) developed a sequence stratigraphic framework for the Straight Cliffs Formation based on their study of the fluvial-marine transition preserved in the Straight Cliffs Formation on the Kaiparowits Plateau. In their work, which was a pioneering study in non-marine sequence stratigraphy, they divided the terrestrial part of the Straight Cliffs Formation into five facies tracts. The amalgamated fluvial facies tract (representative of lowstand), contains a sandstone-dominated succession with numerous erosional surfaces. The isolated fluvial facies tract (transgressive), contains a mud-rich succession with isolated lenticular sandstone bodies and is generally representative of increasing to high accommodation space. The tidally-influenced facies tract (initiation of highstand), contains sedimentary structures indicative of tidal influence and possible brackish water incursion. The alluvial plain facies tract (alluvial aggradation during highstand) contains a sandstone-dominated succession representative of reducing accommodation space. A coal-bearing facies tract (transgressive to highstand) may cap the sequence. Little (1995, 1997) proposed a different sequence framework for the Upper Cretaceous of the Kaiparowits Plateau, in which each sequence consists of a mudrock-rich unit, interpreted to form in a high-accommodation space setting, capped by an amalgamated sandstone, from a low-accommodation space setting. Upper Cretaceous strata of the Straight Cliffs (Drip Tank Member), Wahweap and Kaiparowits formations were also placed into two third-order sequence stratigraphic cycles, similar to those of Shanley and McCabe (1991) by Lawton *et al*. (2003).

1.3.4. The Wahweap Formation

This project focuses on the Wahweap Formation, which is a series of interbedded sandstone and mudrock that crops out on the Markagunt, Paunsaugunt and Kaiparowits plateaus. Equivalent strata in the Henry Mountains Basin (the Masuk Formation) are also dealt with here.

1.3.4.1. Stratigraphy and palaeoenvironment

There are two competing stratigraphic schemes used for the Wahweap Formation. Eaton (1991) divided the Wahweap into lower, middle, upper and capping sandstone members on the Kaiparowits Plateau. This subdivision cannot be carried across to the Markagunt and Paunsaugunt plateaus, where only the capping sandstone member can be differentiated from the lower part of the formation. For this reason, Eaton's members remain informal and are mostly applicable to the Kaiparowits Plateau (Fig. 1.7.).

Doelling (1997) divides the Wahweap into a lower and upper member, with his upper member equivalent to Eaton's capping sandstone. This scheme can be used satisfactorily throughout the extent of the Wahweap Formation from the Markagunt to the Kaiparowits Plateau; however, on the Kaiparowits Plateau the lack of differentiation in the lower part of the succession causes a loss of resolution that hinders stratigraphic correlation. This is problematic because the lower three members on the Kaiparowits Plateau attain a maximum thickness of more than 300 m (Eaton, 1991).

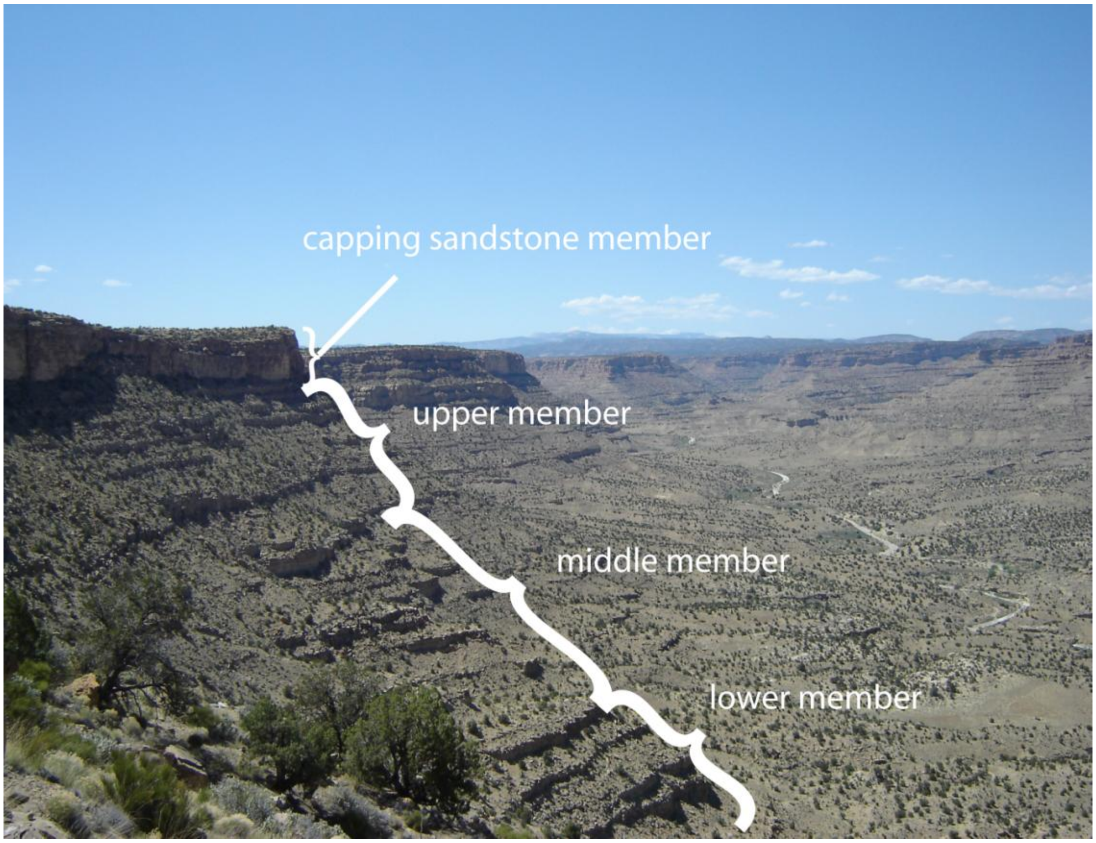


Figure 1.7. The type section of the Wahweap Formation at Reynold's Point on the Kaiparowits Plateau.

In this study the scheme of Eaton (1991) is followed, and the uppermost sandstone unit of the Wahweap Formation is always referred to as the capping sandstone. The Wahweap Formation strata which underlie the capping sandstone on the Markagunt and Paunsaugunt plateaus will be referred to as the lower unit when discussing the stratigraphy on these plateaus, but that does not imply it is equivalent to the lower member of the Kaiparowits Plateau; it is more likely equivalent to the lower, middle and upper members together although it is more uniform.

According to Eaton's (1991) subdivision, the lower member is found from the base of the Wahweap Formation, where the first laterally extensive mudrock overlies the Drip Tank Member of the Straight Cliffs Formation, up to and including a distinctive 5-10 m thick laterally extensive sandstone unit that occurs 20-60 m above the base of the formation. Eaton noted that the exact stratigraphic position of this sandstone unit was variable, although it is present throughout the outcrop exposure, and is therefore a useful stratigraphic marker.

The middle member is a mudrock-rich unit without any laterally persistent sandstone units. It characteristically forms slopes (Eaton, 1991; Pollock, 1999). The basal contact of the middle member is at the lowest mudrock which overlies the uppermost sandstone unit of the lower member, while the top of the middle member is located at the base of the first laterally extensive sandstone of the upper member.

The upper member consists of laterally extensive sandstone units interbedded with mudrock units. The basal contact of the upper member is at the base of the first laterally extensive sandstone which overlies the mudrock of the middle member while the top of the upper member is at a distinctive erosive surface where rippled or trough-cross bedded tan sandstones are truncated by orange or white sandstones or granulestones with large trough cross-bedding which form part of the capping sandstone.

The capping sandstone is distinctive in terms of both its colour and its alluvial architecture. It consists of orange or white sandstones with large cross-beds and gravel bars. Its lower contact is a distinct erosional surface while its upper contact is a gradational transition in which the succession eventually becomes more mud-rich. The upper contact is placed at the first extensive mudrock of the Kaiparowits Formation (Eaton, 1991; Pollock, 1999).

Work by Lawton *et al*. (2003) found that there is stratigraphic control on sandstone petrology. Sandstones of the lower and middle members have a quartzofeldspatholithic composition (averaging 61% total quartz, 19% feldspar and 20% lithic fragments) while sandstones of the upper member have a quartzolithic petrofacies (average 75% total quartz, 5% feldspar and 20% lithic fragments). Sandstones from the capping sandstone have a quartzose petrofacies (98% total quartz, 1% feldspar and 1% lithic fragments).

Equivalent strata of the Masuk Formation in the Henry Mountains Basin are divided into lower, middle and upper members (Eaton, 1990). The diagnostic criteria for the members of the Masuk Formation are remarkably similar to those of the Wahweap Formation on the Kaiparowits Plateau: a lower member capped by an extensive sandstone unit, a mudrock-rich, slope-forming middle member and an upper member that consists of interbedded sandstone and mudrock with several extensive sandstone bodies. The Masuk Formation is overlain by the Tarantula Mesa Sandstone Formation which is thought to be equivalent to the capping sandstone member (Eaton, 1990).

Previous work on the depositional environment of Wahweap Formation sandstones and mudrocks on the Kaiparowits Plateau has concluded that the lower three members were deposited in a meandering river palaeoenvironment while the capping sandstone was deposited by braided rivers (Eaton, 1991). Pollock (1999) suggested that the middle member was deposited by anastomosed river systems and that estuarine conditions developed for some period during Wahweap deposition, but he did not expand on either of these assertions. Simpson *et al.* (2008) described aeolian stratification in the capping sandstone and inferred that deposition occurred during a period of aridity (possibly seasonal). Further support for arid conditions during deposition of the capping sandstone was found by W. Simpson *et al*. (2010) who documented the preservation of biologic soil crusts typical of arid conditions. In the Masuk Formation, the entire succession was thought to be of marine origin until the work of Peterson *et al*. (1980) and Eaton (1990) who postulated a brackish-water depositional environment based on preserved faunal remains.

1.3.4.2. Source area and drainage model

Lawton *et al*. (2003) and Larsen (2007) described a drainage model for the Wahweap Formation in which the lower three members were deposited by northward and northeastward flowing rivers, whereas the capping sandstone was deposited by eastward and southeastward flowing rivers. Work on detrital zircon populations in the Wahweap

Formation by Larsen (2007 – Fig. 1.8.) and Dickinson and Gehrels (2008) has found that zircon populations from the lower and upper member are most likely derived from uplifts in the Mogollan highlands (Fig. 1.9.) to the south and Cordilleran magmatic arc to the southwest, whereas zircon populations from the capping sandstone are most likely derived

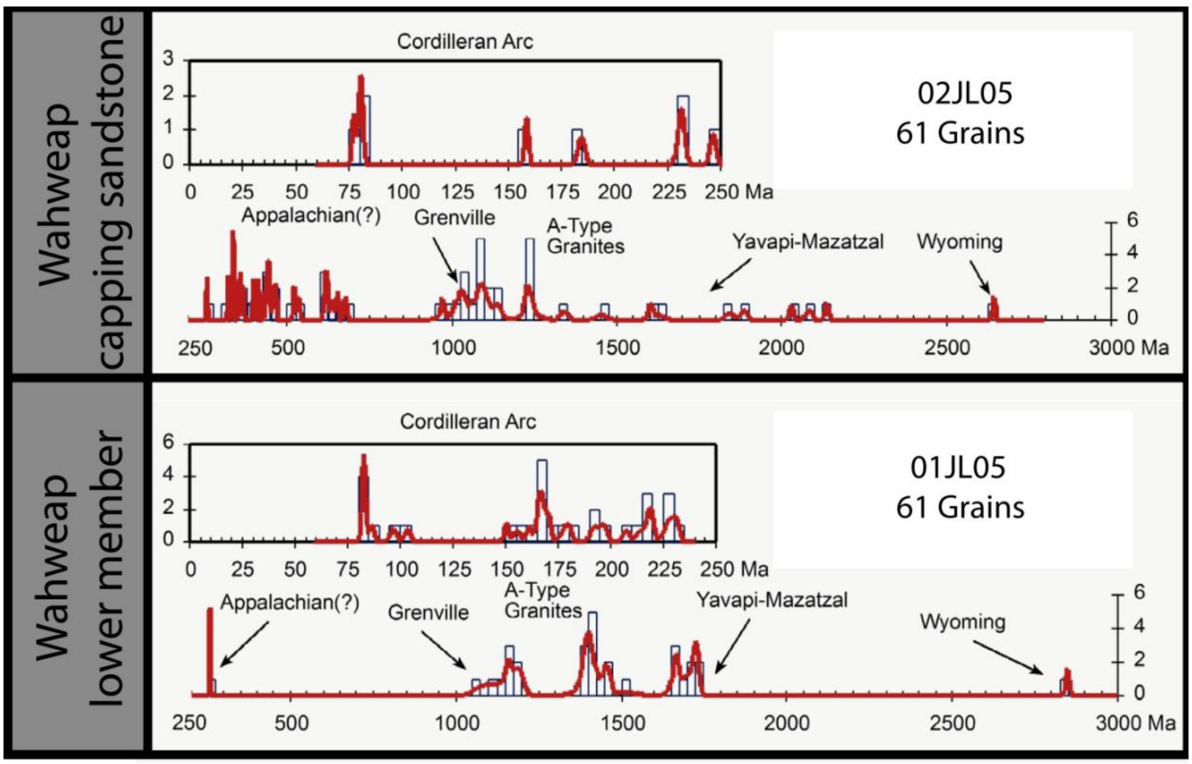


Figure 1.8. Age histogram for detrital zircon populations from the lower member and capping sandstone (after Larsen, 2007)

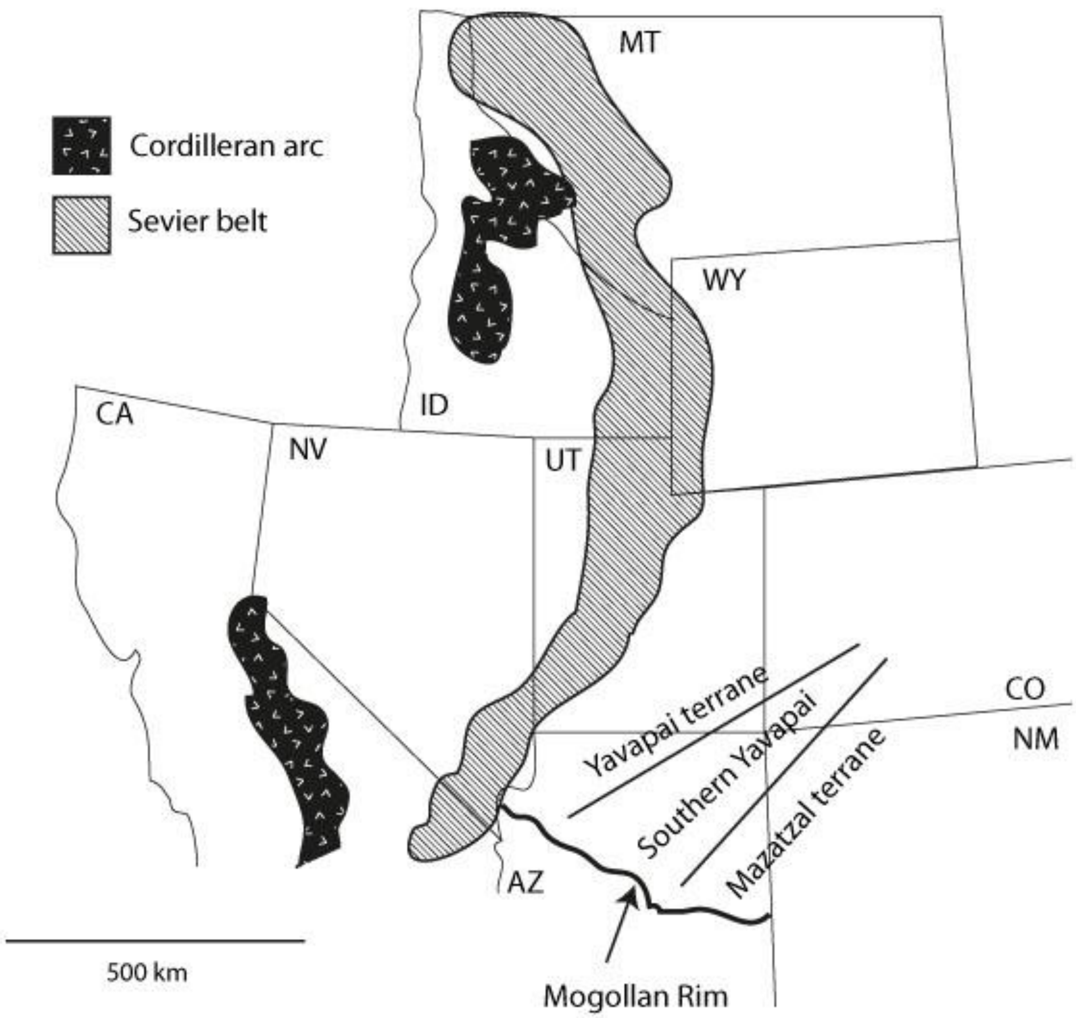


Figure 1.9. Source areas for the Wahweap Formation. After Bilodeau (1986), DeCelles (2004) and Gilbert et al. (2007)

from recycled sedimentary strata in the Sevier belt to the west. Specifically, zircons of the lower Wahweap Formation have large populations with ages between 1300 Ma and 1800 Ma (derived from the Yavapai and Mazatzal terranes), and Mesozoic magmatic zircons (derived from the Cordilleran arc). The capping sandstone contains a suite of zircons dominated by populations of 900 Ma to 1200 Ma (Grenville provenance), 300 Ma to 550 Ma (Appalachian provenance) and Mesozoic magmatic zircons of the Cordilleran arc. Together the zircon populations in the capping sandstone are very similar to a suite of zircons present in Mesozoic aeolian strata in the Sevier fold-and-thrust belt (Larsen, 2007; Dickinson and Gehrels, 2008).

1.3.4.3. Biostratigraphic age and correlation

Until recently, age constraint for the Wahweap Formation was based on mammalian biostratigraphy (Eaton, 1991, 2002) which was used to assign the formation an Aquilan North American Land Mammal "Age" (NALMA; Table 1.2.). The Wahweap Formation was thus important as one of very few pre-Judithian mammal-fossil bearing localities, along with the type-Aquilan Milk River Formation in Canada. Radiometric dating of the Milk River Formation by Payenberg *et al*. (2002) found that it was deposited between approximately 84.5 Ma and 83.5 Ma, and its deposition was followed by a hiatus in sedimentation. Although NALMAs are defined exclusively by fossil content, as opposed to inferred age, the Aquilan is therefore considered to be represented by a fauna that occurred during the Santonian and earliest-Campanian.

Table 1.2. North American Land Mammal Ages (after Lillegraven and McKenna, 1986; Kielan-Jaworowska et al. 2004 and Eaton, 2006)

NALMA	***Notable formations***	***Notes***
Lancian	Scollard (Alberta) Frenchman (Saskatchewan) Hell Creek (Montana) Lance (Wyoming) North Horn (Utah)	Extensively sampled. Restricted to localities around the central and northern Rocky Mountains
"Edmontonian"	St Mary River (Alberta) Kirkland/Fruitland (New Mexico) - *disputed*	Poorly documented due to contemporaneous transgression resulting in paucity of terrigenous deposits
Judithian	Oldman (Alberta) Judith River (Montana) Mesaverde Group (Wyoming) Kaiparowits (Utah)	Most extensive NALMA geographically
Aquilan	Milk River (Alberta) Wahweap (Utah) Straight Cliffs (Utah)	Unclear lower boundary due to older faunas being poorly preserved

Later work by Eaton (2006) found that parts of the Straight Cliffs Formation had faunas with Aquilan affinity and a likely-Santonian age, which would move the Wahweap Formation into the Campanian. Other workers have considered that the Wahweap Formation represents a time-slice not preserved in the northern Western Interior Basin (due to a depositional hiatus between the Milk River and Pakowki formations) and have suggested that it be the type area for a new Wahweapian NALMA (Kirkland and DeBlieux, 2010).

1. 4. Hypotheses

Several hypotheses related to the palaeoenvironment, tectonic setting and allocyclic controls on sedimentation were tested in this work.

1.) *The lower, middle and upper members of the Wahweap Formation were deposited in meandering fluvial systems with no appreciable change in style through the succession.*

-Changes in fluvial style or depositional environment are important in understanding changes in base level, and thus the amount of accommodation space in a basin. Robust palaeoenvironmental reconstructions are therefore essential for fully understanding basin evolution and the allocyclic controls on sedimentation. This hypothesis is addressed in chapter 2.

2.) *Tidal activity caused by marine transgression influenced sedimentation in the Wahweap and Masuk formations.*

-Eaton (1991) described the Wahweap as entirely fluvial, whereas Pollock (1999) found that some deposition may have occurred in an estuarine setting. Sedimentological evidence such as heterolithic stratification, clay and carbon draped foresets, herring-bone cross-stratification and tidal couplets are all criteria commonly used to indicate tidal activity. This hypothesis is addressed in chapters 2 and 4.

3.) *The Wahweap Formation can be divided into four members that are distinguishable across the east-west outcrop extent of the formation.*

-The four informal members of the Wahweap Formation were defined at Reynold’s Point and Pardner Canyon, and their applicability to other localities remains untested (Eaton, pers. comm.). This hypothesis is dealt with in chapter 7.

4.) *The Wahweap Formation falls into the Aquilam NALMA.*

-Indicator taxa for a NALMA assignment in the Wahweap Formation are ambiguous because the only definitive taxa recovered occur in both the Aquilan and Judithian. Radiometric dating can improve correlation with the type-localities for the Aquilan and Judithian NALMAs and thus refine the Wahweap Formation assignment. This hypothesis is discussed in chapter 3.

5.) *The Wahweap Formation was deposited in a foredeep setting before basin partitioning occurred along the East-Kaibab monocline.*

-Doubt exists as to whether foreland partitioning (and associated onset of Laramide thrusting) occurred during Wahweap time (e.g. Lawton *et al.*, 2003; Tindall *et al.*, 2010) or after Wahweap time (e.g. Goldstrand, 1992). In this project, sedimentary thicknesses across large faults (Paunsaugunt fault, East-Kaibab monocline) are used determine whether basin partitioning occurred during or after deposition of the Wahweap Formation. This hypothesis is discussed in chapter 5.

6.) *Subsidence was at a maximum in the Kaiparowits and Henry basins during deposition of the lower, middle and upper members, with less subsidence in the Markagunt and Paunsaugunt basins.*

-The thickness of the lower part of the Wahweap Formation increases from ~130 m on the Markagunt/Paunsaugunt plateaus up to a maximum of 320 m on the Kaiparowits Plateau. This change in thickness can either be attributed to unconformities, or differential subsidence. This hypothesis is addressed in chapter 5.

7.) *Palaeosol development in the Markagunt, Paunsaugunt and Kaiparowits plateaus is variable and reflects sedimentation in different environments under similar climatic conditions.*

-Differential palaeosol development in the Markagunt and Paunsaugunt plateaus vs. the Kaiparowits and Henry Mountains regions may either be related to different climatic conditions or, more likely, to different styles of sedimentation corresponding to different types of river systems and different accommodation space. The great lateral extent of the Wahweap Formation allows for palaeosol analysis across a transect of upland river to coastal plain environments. This hypothesis is discussed in chapter 6.

2. Facies Associations and Palaeoenvironmental Reconstruction

2.1. Introduction

Fluvial systems represent a significant proportion of the preserved continental sedimentary record (Gibling, 2006), and ancient fluvial systems are among the best-studied sedimentary deposits due to their accessibility and the clues they provide on tectonics, eustasy, climate, basin evolution, palaeontology and palaeogeography as well as numerous other topics (Miall, 1996; Bridge, 2003). The study of fluvial sediments has a long history and contributions have been made by geomorphologists, sedimentary geologists and engineers (Miall, 1978).

Different approaches have been used in the study of the ancient fluvial record; for example sedimentary structures, lithofacies, architectural element analyses and ichnology are commonly used to reconstruct palaeoenvironment, determine accommodation space and palaeohydrological conditions (Miall, 1985, 1996; Dalrymple *et al*., 1991; Heller and Paola, 1996; Fielding, 2006; Gibling, 2006). Initially, facies models for fluvial systems were based on vertical profiles (e.g. Miall, 1977) but it was later found that 3-dimensional studies of a suite of architectural elements or facies associations is more effective in understanding fluvial depositional elements (e.g. Miall, 1985; Reading, 1996; Walker 2006). A synthesis of the data collected by facies analysis, combined with other data sources (such as sandstone composition, palaeocurrents and mudrock geochemistry) can eventually result in a robust reconstruction of a depositional system (e.g. Allen, 1983; Rogers, 1998; Willis *et al*., 1999;

Miall and Arush, 2001; Komatsubara, 2004; Therrien, 2005; Ghosh *et al.*, 2006; Yan *et al*., 2006; Allen and Fielding, 2007; Roberts, 2007).

2.2. Aims of this chapter

In this chapter, detailed sedimentological work is used to reconstruct the palaeoenvironmental conditions in which the Wahweap Formation was deposited. A detailed understanding of the depositional environment will then be used to underpin palaeogeographic, sequence stratigraphic and geochronologic work in subsequent chapters.

2.3. Previous work

Eaton (1991) studied the stratigraphy and sedimentology of the entire Upper Cretaceous succession on the Kaiparowits Plateau, and regarded the lower, middle and upper members of the Wahweap Formation as the deposits of meandering streams, with a braided stream palaeoenvironment postulated for the capping sandstone. Little (1995, 1997) measured a stratigraphic section on the Kaiparowits Plateau and provided the first sequence stratigraphic interpretation for the Upper Cretaceous strata and also suggested that anastomosed streams played a role in deposition of the middle member of the Wahweap Formation. Pollock (1999) studied thickness trends and lithofacies in the capping sandstone member on the Markagunt, Paunsaugunt and Kaiparowits plateaus, and Pollock (1999) and Lawton *et al*. (2003) have also found that the capping sandstone has a quartzose composition, dissimilar to the quartzofeldspathic or quartzolithic composition of the lower part of the Wahweap. Pollock (1999) also suggested that estuarine conditions may have developed during Wahweap deposition.

More recently, Simpson *et al*. (2008), Hilbert-Wolf *et al*. (2009) and Tindall *et al.* (2010) have undertaken studies on the capping sandstone. The work of Simpson *et al*. (2008) has documented the presence of aeolian stratification in the capping sandstone and postulated that climatic conditions at the time of deposition were more arid than previously recognised, while Hilbert-Wolf *et al*. (2009) have identified seismites in the capping sandstone and noted that they were most likely caused by seismic activity along local normal faults.

2.4. Methods

Ten stratigraphic sections of the Wahweap Formation were measured in localities across the study area (fig. 2.1.), with a focus on the eastern side of the Kaiparowits Plateau where there is almost continuous exposure of the lower, middle and upper members for ~50 km along Smoky Mountain road. One section was measured through the Masuk Formation at Sweetwater Creek. Layer thicknesses were measured and detailed sedimentological analysis of extensive units was performed, including palaeocurrent data collection, examination of lateral variation and facies analysis. Stratigraphic sections from the Markagunt, Paunsaugunt and Kaiparowits plateaus are shown in Figure 2.2. Photographs and UTM co-ordinates of stratigraphic sections are shown in Figure 2.3.

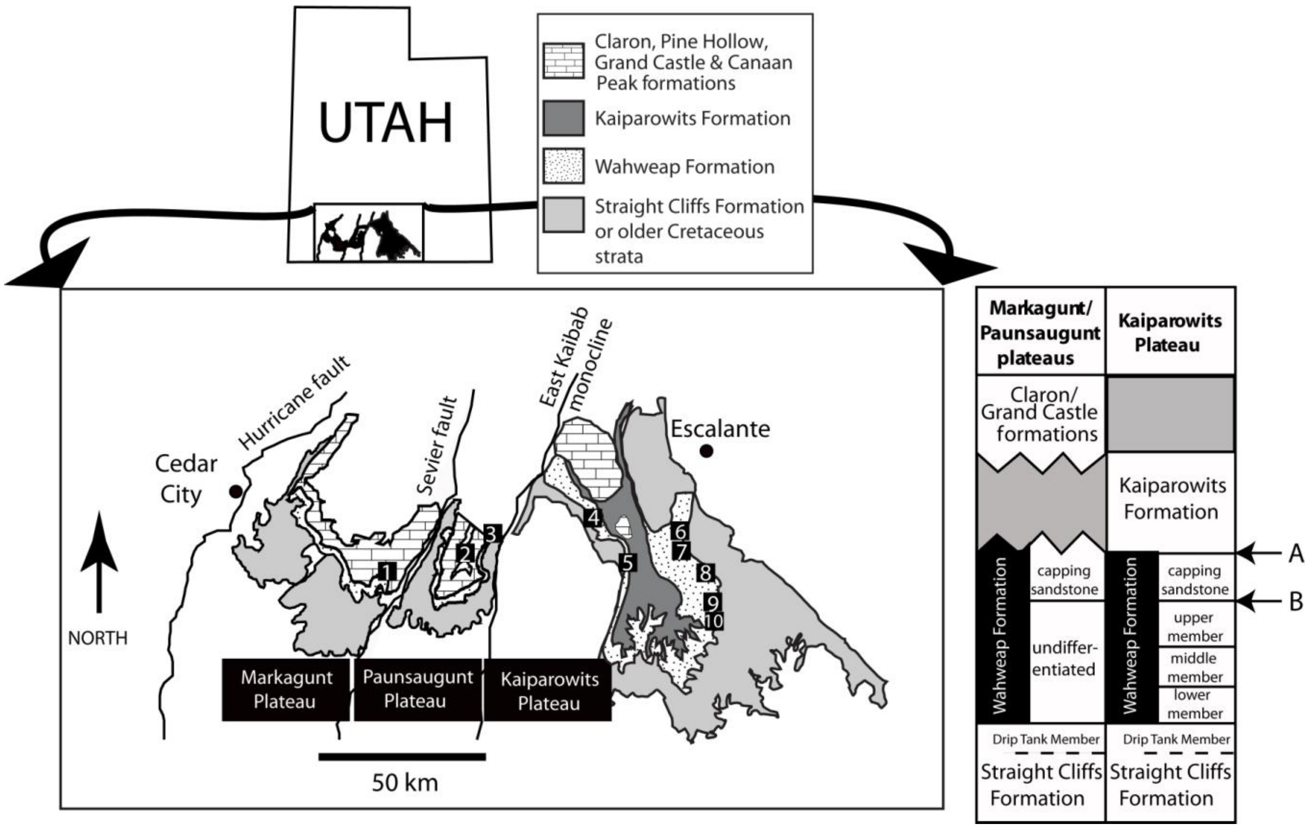


Figure 2.1. Location map of sections measured through the Wahweap Formation. Numbers refer to sections in Fig. 2.2.

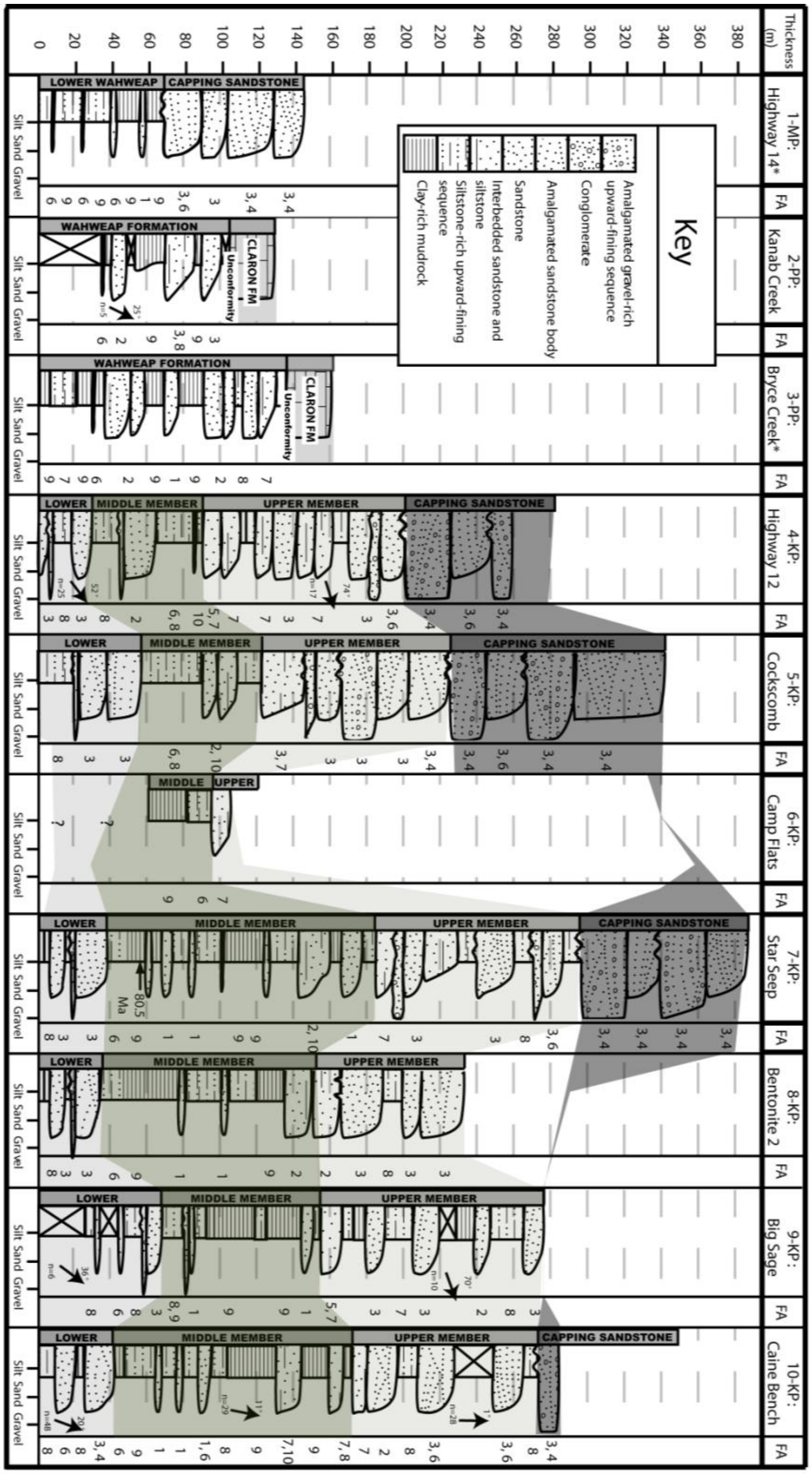


Figure 2.2. Stratigraphic sections through the Wahweap Formation. Section marked with an asterisk measured by Tilton (1991). Explanation of FAs in Table 2.2.

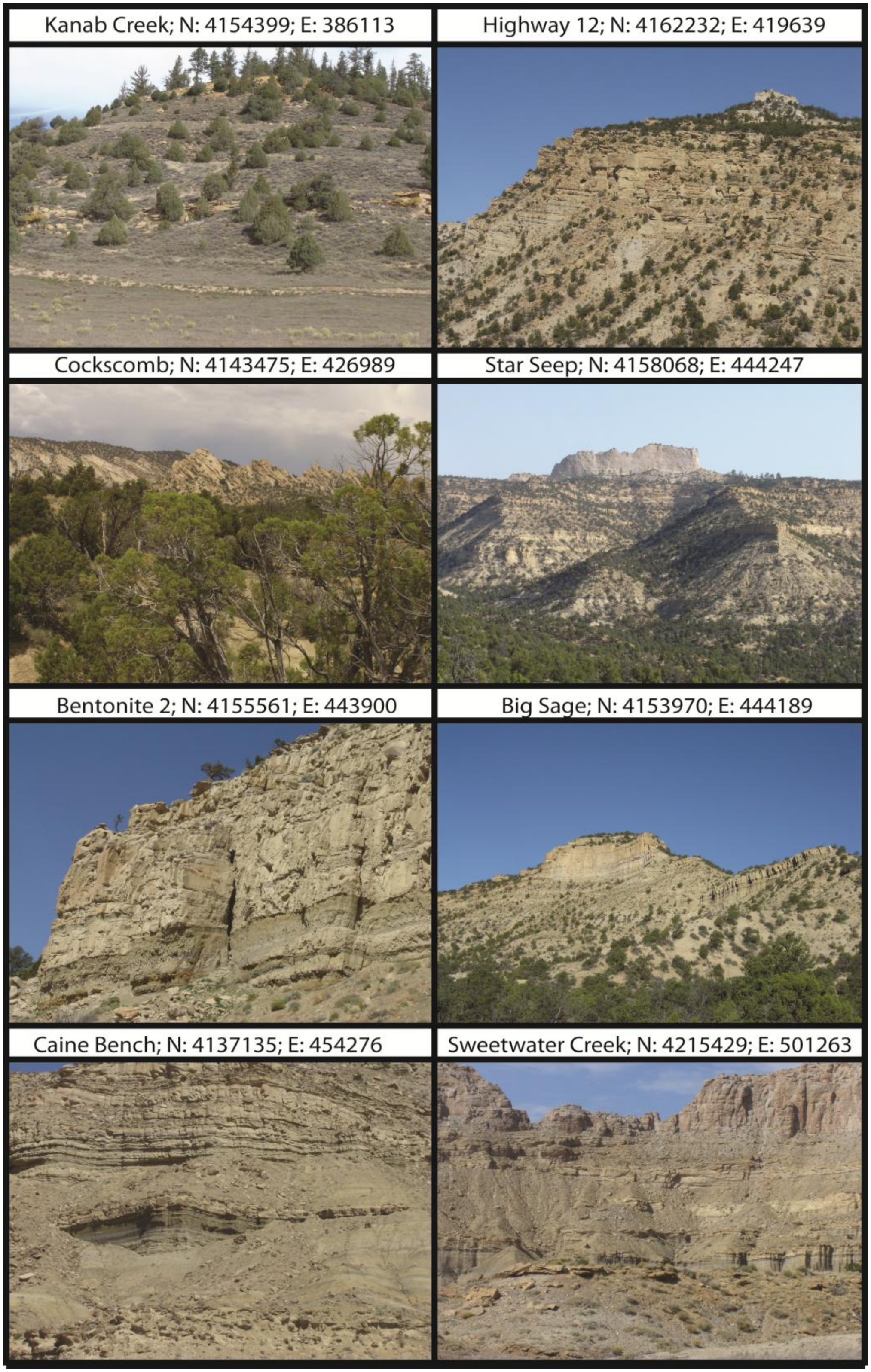


Figure 2.3. Photographs and UTM co-ordinates for stratigraphic sections of the Wahweap and Masuk formations

Localities were correlated based mainly on the position of laterally extensive sandstone bodies, several of which can be traced across the extent of the Wahweap Formation on the Kaiparowits Plateau, and sandstone petrology, which varies through the stratigraphy (Lawton *et al*. 2003). Additional correlation was possible from the position of a bentonite layer low in the middle member. Stratigraphic sections were then compared to reveal lateral and vertical changes in lithology.

2.5. Facies Architecture – Kaiparowits Plateau

The Wahweap Formation on the Kaiparowits Plateau was divided into 17 lithofacies (Table 2.1) and 10 facies associations (Table 2.2) based on grain-size, geometry and bounding surfaces.

Facies association 1: single-storey lenticular sandstones:

Description: FA 1 consists of thin sandstones that are not laterally extensive. Lenticular sandstones have a width:height ratio of less than 15 (e.g. Friend *et al*. 1979), and those in the Wahweap Formation have thicknesses between 1.5 and 7 m and widths of meters to tens of meters (for discrete channel elements). Facies Sh, St, Sp, Gmi, Gci and Sr are present. The basal surface of lenticular sandstones in the Wahweap Formation typically truncate underlying beds, with facies Gci or Gmi directly above the contact, fining upwards into facies Sh, St or Sp with rare instances of Sr. Upper contacts are sharp. Second order coset bounding surfaces (as defined by Miall 1985, 1996) occur within these bodies, and are typically 0.2 to 1 m apart, but internal truncation surfaces are not present. FA 1 is commonly found overlain and underlain by mudrock-dominated facies associations.

Table 2.1. Lithofacies of the Wahweap Formation

Lithofacies	**Grain-size/description**	**Colour**	**Features**
Gmi Matrix supported trough cross- bedded or massive intraformational conglomerate	*Clasts:* Intraformational siltstone and mudstone, shell, bone and wood, 2-200 mm, rounded to angular, poorly sorted. *Matrix:* Medium-coarse sand.	*Clasts:* Medium light grey (10Y 6/2) *Matrix:* Greyish orange pink (10R 8/2) to reddish brown (10R 5/4).	Occurs in scour fills, at the base of sandstone-dominated units, or overlies erosional surfaces within amalgamated sandstone bodies.
Gce Clast supported trough cross-bedded extraformational conglomerate	*Clasts:* Typically sub-rounded to well rounded chert or limestone, 2-100mm with rare wood or bone. *Matrix:* Medium-coarse sand.	*Matrix:* Pale yellow (5Y 8/3) to moderate orange pink (5Y 8/4)	Typically erode into mudstone or siltstone, and grade upward into sandstone. Forms units up to several metres thick.
Gci Clast supported intraformational conglomerate	*Clasts:* Intraformational siltstone and mudstone, fragmentary shell, bone and wood, 2-150 mm, rounded to sub-rounded. *Matrix:* Medium-coarse sand	*Clasts:* Medium light grey (10Y 6/2) *Matrix:* Pale yellow (5Y 8/3) to moderate orange pink (5Y 8/4)	Forms localised units overlying erosional surfaces within amalgamated sandstone bodies.
Gs Clast supported shell conglomerate	*Clasts:* Fragmentary to complete shells (gastropods and bivalves; 90%). Fragmentary vertebrate remains. Granule sized intraformational siltstone. *Matrix:* Medium-coarse sand.	*Matrix:* Pale yellow (5Y 8/3) to moderate orange pink (5Y 8/4)	Forms localised units, up to 1 m thick, which contain 90% shell material (dominantly fragmentary), as well as minor vertebrate bone and intraformational mudrock.
Gc Clast supported calcrete nodule conglomerate	*Clasts:* Entire or fragmentary calcrete nodules, 20-150 mm. *Matrix:* Medium sand	*Clasts:* White (8N) to very pale orange (10YR 8/3) or strong brown (5YR 4/6) *Matrix:* Pale yellow (5Y 8/3) to moderate orange pink (5Y 8/4)	Forms thin but laterally extensive units. Often overlies deeply incisive contacts within thick sandstone bodies.
Sh Horizontally laminated sandstone	Very fine to medium sand, minor silt	Very pale orange (10YR 8/3)	Mud drapes on cosets. Plant impressions on bedding plane.
Shl Horizontally laminated sandstone with parting lineation	Medium sand	Pale yellow (5Y 8/3)	Parting lineation.
Sr Rippled sandstone	Very fine to medium sand	Dusky red (5R 3/4) to pale yellow (5Y 8/3)	Localised facies, associated with Sh and Fg. Sr often grades upward into Fg.
St Trough cross-bedded sandstone	Fine to very coarse sand Rare granule to pebble sized clasts.	Pale yellow (5Y 7/4) to pale red (5R 6/2)	Often overlies Gc, Gm or Sm. Often has fragmentary plant or vertebrate remains.
Sp Planar cross-bedded sandstone	Medium sand	Pale yellow (5Y 8/3) to moderate orange pink (5Y 8/4), greyish orange pink (10R 8/2).	Occurs in association with St or Sm, or finely interbedded with Fg. Sometimes has wood with *Teredolites* boring or other plant material with sulphurous rims.
Sm Massive to weakly laminated sandstone	Fine to coarse sand Rare granule to cobble sized siltstone clasts	Pale yellow (5Y 7/4) to greyish orange pink (10R 8/2).	Often has erosive base and sometimes has well cemented sulphide nodules.
Fp Red/yellow pedogenic mudstone and siltstone	Clay, silt, minor sand	Light brown (5YR 5/8) to pale reddish brown (10R 5/4).	Oriented nodules, rhizocretions and wood fragments. Weakly developed horizontal laminations.
Fg Grey laminated claystone and siltstone	Clay, silt	Greyish red purple (5RP 4/2) to medium light grey (10Y 6/2).	Well developed horizontal laminations occur. Contains nodules (goethite?) and invertebrates (molluscs and crabs), vertebrates and plant matter. Often has gypsum veins.
Fc Carbonaceous mudstone	Clay, silt	Dark grey, weak red (5R 4/2) or dark yellowish brown (10YR 4/2)	Localised beds associated with Fg. High organic content, including wood fragments with sulphurous rims. Has gypsum veins and weakly developed horizontal laminations.
Fcr Calcrete nodule siltstone	Silt with minor sand	5Y 8/2 pale greenish yellow (matrix) 8N white (nodules)	Dense accumulations of calcium carbonate nodules up to 100 mm in diameter.
Fb Bentonitic claystone	Clay with minor silt, sand	Pale yellow (5Y 8/3) to blueish grey (5PB 5/1)	Phenocrysts of biotite and feldspar, concentrated near the base. Upper contact gradational with overlying Fg. Distinctive popcorn weathering due to clay content.
C Coal		Black	Occurs as isolated lenses, less than 0.5 m wide and 10 cm thick, in association with Fg and Fc.

Table 2.2. Facies Associations of the Wahweap Formation

Facies association	Lithofacies	Fossils	Interpretation	Occurrence
1.) Single-storey lenticular sandstone	Gci, Gmi Sh, St, Sp, Sr	Fragmentary bone, wood	Channel fill	middle upper
2.) Multi-storey lenticular sandstone	Gci, Gmi St, Sh, Sp	Shell conglomerate, fragmentary bone, wood	Amalgamated channel fill deposits	middle upper
3.) Major tabular sandstone body	Gci, Gmi, Gc, Gs St, Sp, Sh, Shl, Sr, Sm Fcr, C	Shell conglomerate, fragmentary bone, wood	Amalgamated channel fill deposits	lower upper capping sandstone
4.) Gravel bedform	Gce, Gci	Fragmentary vertebrate remains (extremely rare)	Braid bars	lower capping sandstone
5.) Low-angle heterolithic cross-strata	Sp Fg	Wood	Inclined heterolithic stratification	middle upper
6.) Minor tabular sandstone body	Sh, Sr	None	Crevasse channel	lower middle capping sandstone
7.) Lenticular interlaminated sandstone and mudrock	Sh, St Fg	Fragmentary bone, wood	Tidally-influenced channel	middle upper
8.) Inclined interbedded sandstone and mudrock	Sh Fg, Fp	Abundant fragmentary or complete vertebrate, invertebrate and plant remains	Proximal overbank deposits – levee, splay	lower middle upper
9.) Laterally extensive mudrock	Fg, Fc, Fp, Fb, C, Fcr	Fragmentary or complete vertebrate, invertebrate and plant remains	Distal overbank deposits – backswamp, pond, lake and soil	middle
10.) Lenticular mudrock	Sm Fg	None	Muddy channel	middle

Fossil content within FA 1 is limited to fragmentary vertebrate remains and coalified wood, as well as fragmentary to complete invertebrate remains. Complete isolated elements such as crocodile and dinosaur bones and teeth are also preserved. Fossils are most abundant in basal conglomerates and less commonly preserved in the sandy lithofacies of this FA.

Interpretation: These single-storey sandstone bodies are interpreted as the sedimentary fill of isolated channels (simple sandstone bodies of Friend *et al*. 1979). They represent small fluvial channels with facies Gmi, or Gci representing scour fill or channel lag deposits at the base, and facies Sh, St and Sp representing channel fill. This element is generally encased in overbank fine deposits and several single-storey lenticular sandstone bodies sometimes occur at the same stratigraphic level. Interpretation as channel fill is supported by the abundance of St, channel scours and the channel shape of the sandstone bodies (Holbrook 1996; Adams and Bhattacharya 2005), and the lack of deep erosive surfaces within the channel deposits together with the association of this FA with extensive mudrock units suggests aggradation as a dominant process (Wright and Marriott 1993).

Facies association 2: multi-storey lenticular sandstones:

Description: Multi-storey lenticular sandstones have width:height ratios of less than 15 and also contain internal truncation surfaces (third or fourth order surfaces as defined by Miall [1996]) with basal erosional relief of up to 1 m. These sandstone units are typically thickest towards the centre, where they attain thicknesses of 4 -12 m, and narrow towards their lateral margins, giving the unit an overall lenticular shape. Their lateral extent is typically on the order of tens of meters. Facies Gci, St or rarely Gmi typically occur at the base of FA2, where they overlie truncation surfaces, whereas St, Sh and Sp constitute the dominant facies within the body of FA 2. Cosets are typically of the order of 0.1 to 1 m in thickness. Fossil content and preservation in FA 2 is similar to FA 1.

Interpretation: Multi-storey lenticular sandstones are interpreted as the amalgamated deposits (e.g. Shanley and McCabe 1995; Mack and Madoff 2005) of a series of channel systems. Truncation surfaces are interpreted to represent periods of erosion and overlying successions represent the deposits of new channels. This is supported by the presence of intraformational conglomerates that commonly overlie truncation surfaces. Where facies Gci occurs above erosional surfaces it is interpreted as forming channel lag deposits. Facies Gmi is interpreted to result from small-scale sediment-gravity-flow deposits which occur within existing channel systems. Facies St, Sh and Sp constitute channel fill.

Facies association 3: major tabular sandstone bodies:

Description: FA 3 has width:height ratios greater than 15, and are traceable for significant distances laterally (up to several kms). Facies present (in order of abundance) are St, Sp, Gci, Gmi, Gc Sh, Shl, Sr, Fcr, C, Gs and Sm. Tabular sandstone bodies have undulating bases overlain by lens-shaped units of facies Gci, Gmi, Gc or Sm, that vary between 0.3 and 1.5 m thick. Clasts within the conglomerates are most commonly intraformational mudrock, although extraformational chert and quartzite, reworked calcrete nodules, fossil wood, fragmentary fossil material or complete elements such as teeth or shells are also present as clasts. Basal conglomerates typically grade into an upward-fining succession of sandstone which may be overlain by Fcr. Multiple upward-fining successions are commonly present within FA 3, which are typically separated by undulating contacts overlain by conglomerate. Thickness limits for these upward-fining cycles are between 2 and 8 m. The upper contact of FA 3 is usually sharp, except when the element is overlain by ripple-laminated sandstones and mudrocks, in which case the contact tends to be gradational. Inclined stratification, consisting of facies Sp with large, low angle cross-beds that have a moderate to high degree of bioturbation, and overlie an undulating base, also occur in this type of sandstone body, but are limited to the lower and upper member.

The most common form of fossil preservation in FA 3 is as fragmentary vertebrate, invertebrate and plant material in conglomeratic lithofacies. In some cases, concentrations of fragmentary fossil material are locally present in productive horizons within intraformational conglomerates that extend for up to 4 m. Isolated complete elements such as teeth, vertebrate skeletal elements (commonly vertebrae) and fish scales are also present. Additionally, coalified or silicified wood, in some cases logs at least 2 m long, can be found in FA 3. Shell conglomerates, in which bivalve and gastropod shell material makes up ~90 % of the clasts are rare, but locally present in FA 3 in both the lower and upper members. These shell conglomerates can be up to 0.5 m thick and 2-3 m wide.

Interpretation: FA 3 represents multi-storey channel complexes (as defined by Friend *et al*. 1979; Mack and Madoff 2005), with basal channel lag deposits, as well as intraformational rip-up clasts of underlying flood-basin deposits and wood, shells or bone fragments. Undulating contacts overlain by conglomerates are interpreted as erosive surfaces with channel lag deposits. These re-activation surfaces were formed by channel migration and avulsion, leading to the formation of amalgamated channel sandstones by a process that has been documented in other settings (e.g. Holbrook 1996; Komatsubara 2004; Adams and Bhattacharya 2005). The sedimentary structures present indicate that lower flow regime conditions were dominant, although parting lineation, while rare, is present, showing that deposition sometimes occurred in the upper flow regime. Upward-fining successions which terminate with Fcr record channel abandonment and pedogenesis in channel sediments similar to fluvial aggradational cycles of Atchley *et al*. (2004). Inclined stratification, interpreted to have formed by lateral accretion in point bars, indicate sinuous channel systems. Channel size was calculated from the size of point bar deposits (Fig. 2.4.; Table 2.3.) following the methodology of Allen (1965), and suggests a dominance of small to medium sized river channels**.** Facies Gs indicates the presence of freshwater clam communities in channels.

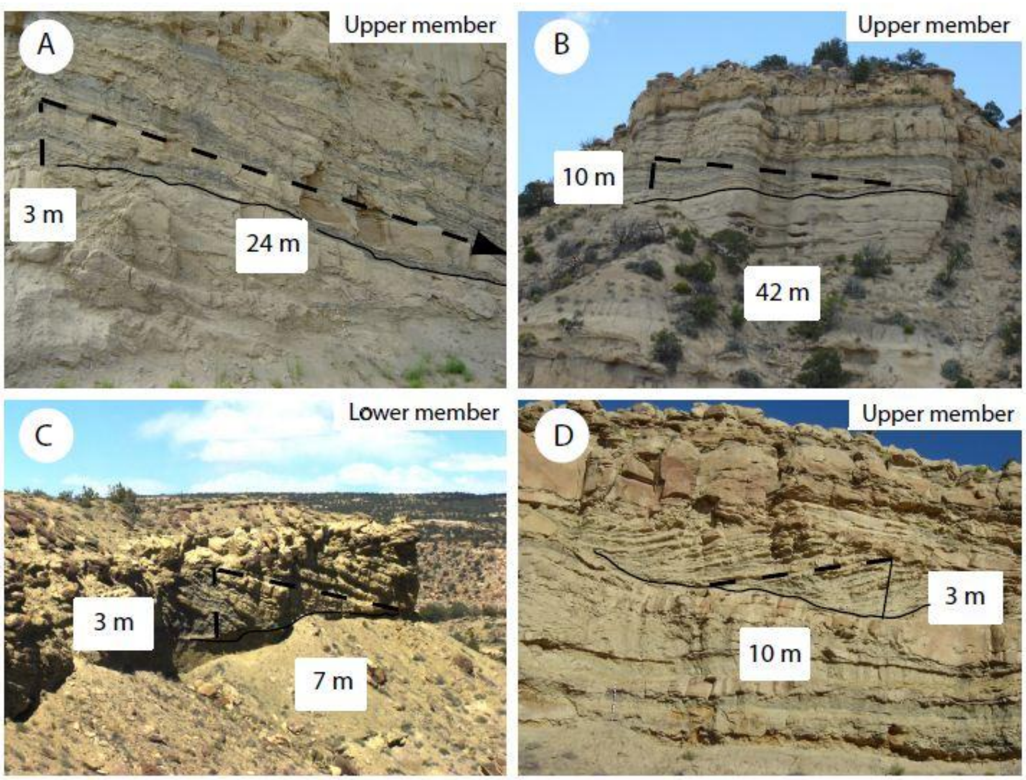


Figure 2.4. Lateral accretion sets of the Wahweap Formation used to calculate channel size

Table 2.3. Calculations of channel size in the Wahweap Formation

Channel location and member	Bankful depth (D)	Bankful width (W)	W/D ratio	Cross-sectional area (m^2)
A: H12 upper	3 m	24 m	8	72
B: CF upper 1	7 m	22 m	3.1	154
C: CF upper 2	10 m	42 m	4.2	420
D: CB lower	3 m	7 m	2.3	21
E: CB middle	0.7 m	6 m	8.6	4.2
F: PK upper	3 m	10 m	3.3	30
Average	**4.5 m**	**18.5 m**	**4.1**	**83.3**

Facies association 4: gravel bedforms:

Description: Gravel bedforms are Gce and Gci dominated features with sharp to undulating lower and sharp upper contacts. They are generally 1.5 to 3 m wide and 0.7 to 1.5 m high. They contain internal first and second order surfaces, and are always associated with laterally extensive major tabular sandstones (FA3). They are elongate features with palaeocurrent direction from trough-cross beds indicating flow directions parallel to their elongation. Individual trough cross-bed foresets in gravel bedforms can be up to 1.2 m high. They sometimes contain fragmentary silicified wood, but otherwise have a relatively low fossil content. FA 4 is most abundant towards the upper portion of the formation, where they may form thick, vertically stacked successions.

Interpretation: Gravel bedforms were deposited as prograding bars within channel deposits (sensu Miall 1985). They are typically elongate and oriented parallel to the palaeocurrent direction and are therefore likely to have formed by downstream accretion. Sharp lower contacts and sheet-like geometries characteristic of FA 4 are similar to features described from other braid-bar dominated systems (Cant and Walker 1978; Allen 1983; Johnson and Pierce 1990). They are typically stacked vertically and separated by third order erosive surfaces, suggesting deposition in high-energy channels.

Facies association 5: low-angle heterolithic cross-strata:

Description: FA 5 occurs at the contact between the middle and upper members and consist of facies Sp and Fg, which are interlaminated at a centimeter scale and inclined relative to the basal surface of the element. FA 5 is localised (less than 50 m wide), with sharp to undulatory upper contacts and lower contacts which truncate underlying elements, and commonly preserve plant matter, fragmentary invertebrate shells and wood with borings of the ichnogenus *Teredolites.* A moderate degree of bioturbation is preserved and typically concentrated in the lower part of the element. Cross stratification consists of large (about 1.5 m between second order surfaces), low angle beds. Within the element, Sp units are 5-30 cm thick and capped with drapes of Fg, which vary from 2-20 cm thick. Both Sp and Fg most commonly occur as continuous features across the length of the foreset or are more rarely restricted to discontinuous, 10-30 cm long, lens-shaped features. FA 5 generally has an upward-fining trend due to a decrease in the sand:mud ratio towards the top, although both sand- and silt- sized particles are present directly below the upper surface, and the succession never consists entirely of mudrock. FA 5 is typically overlain by channel sandstones.

Interpretation: This element is interpreted as inclined heterolithic stratification (IHS), formed by lateral accretion in meandering rivers. Thomas *et al*. (1987) have documented the possible depositional environments for inclined heterolithic stratification, which vary from fully fluvial to tide-influenced estuarine settings. The different depositional settings can be differentiated by such features as sand:mud proportions and cyclicity of interbedded sand and mud, trace fossil and fossil content and overall coarsening or fining trends. In the Wahweap Formation, the regularity of sand-mud couplets together with the presence of the trace fossil *Teredolites*, which is a brackish-water form (Gingras *et al*. 2004), as well as the restriction of FA 5 to a single stratigraphic level, suggests some degree of tidal influence in sediment deposition.

Facies association 6: minor tabular sandstone bodies:

Description: Minor tabular sandstone bodies are elements which have a width:height ratio of more than 15, but are not traceable for any significant distance. They are much thinner than laterally extensive sandstone bodies (FA 3) and do not have any internal erosive surfaces. The dominant facies types in this facies association are Sr and Sh, which have sharp contacts with underlying elements, and an upward-fining trend is typical within this element. Localised tabular sandstones sometimes have sharp upper contacts, but more commonly fine upwards gradationally into overbank fine deposits such as inclined interbedded sandstone and mudrock (FA 8). Vertebrate and invertebrate fossils are rare, but plant material (wood and leaf impressions) is present, as well as root casts and bioturbation. FA6 is typically moderately to intensely bioturbated, and both horizontal and vertical movement traces are common. FA 6 is commonly found overlying or adjacent to channel sandstones.

Interpretation: The abundant ripple and planar lamination in FA 6 suggests shallow flows (e.g. Reading 1996) that may most likely be found in crevasse channels and splays in the overbank subenvironment. Crevasse channels can be distinguished from fluvial channels by their finer grain-size and thinness (Amorosi *et al*. 2008) and in the Wahweap Formation they have an association with overbank deposits. Roberts *et al*. (1994) described splays from the Palaeogene of the Bighorn Basin with similar tabular geometry, sedimentary structures and sharp to gradational contacts. Bioturbation and plant material are also commonly preserved in splays (Smith *et al*. 1989), consistent with the interpretation given above.

Facies association 7: lenticular interlaminated sandstone and mudrock:

Description: FA 7 occurs at the middle-upper member contact and consists of facies Fg and Sh or St, interbedded at a centimeter scale. The sand:mud ratio within this element can vary from 1.5 to 0.6, with the majority of sandstone occurring in sheets. Sandstone lenses of carbon-draped Sh or St are less common and occur between layers of Fg. There is typically an overall upward-fining trend, with the succession becoming more mud-rich towards the top and possibly consisting entirely of mudrock directly below the upper contact. Flaser bedding is common and lenticular bedding is locally present. Lower contacts of FA 7 are wavy, and truncate underlying units, while upper contacts generally grade into mudstone or are truncated by sandstone units.

Interpretation: FA 7 is interpreted as tidally influenced channel deposits. Wavy lower surfaces are interpreted as erosional contacts with underlying elements, while sandstone lenses with carbon drapes, and finely interbedded, cyclical sandstone and mudstone indicate changes of carrying capacity in the depositional environment that may be attributed to tidally influenced channels (e.g. Reineck, 1972; Dalrymple *et al*. 1992; Gastaldo *et al*. 1995; Pontén and Plink-

Björklund 2007). FA 7 is restricted to the same stratigraphic level as FA 5, which suggests that they formed a part of the same tidally-influenced channel systems.

Facies association 8: inclined interbedded sandstone and mudrock:

Description: FA 8 consists of facies Fg, Fp and Sh interbedded at a centimeter to meter scale. Typically FA 8 contains successions with a 0.2 to 1 m thick unit of Sh at the base, fining up into Fg and finally Fp. Together the fine-grained components (Fg and Fp) attain thicknesses of 0.5 to 5 m, with the respective proportion of Fg to Fp fairly variable but more commonly dominated by Fg. This facies association has a sharp basal contact and fines upward into a gradational contact with overlying fine-grained elements. It occurs in association with minor tabular sandstone bodies and adjacent to channel elements, and typically has a lateral extent of tens of meters. Variable bedding orientation can be noted between channel elements and inclined interbedded sandstone and mudrock. Most commonly, bedding is subhorizontal in channels and gently inclined away from the channel in this facies association. Vertebrate, invertebrate and plant remains are abundant and typically fragmentary.

Interpretation: FA 8 is interpreted as proximal flood-basin deposits, such as levees or other overbank deposits. The upward-fining nature of this element represents progressively lower-energy conditions as overbank deposition initiated by high water volumes reduced with decreasing discharge, and the gently inclined beds and association with channel deposits are consistent with an interpretation as levee deposits (Fielding *et al*. 1993; Brierley *et al*. 1997). In situations where overbank deposition on levees alternated with sediment accumulation on floodplains and pedogenesis (as evidenced by facies Fp), thick (up to 5 m) units of FA 8 developed. The fragmentary nature of fossils preserved in this element suggests that they represent accumulations built up during quiet water conditions in adjacent channels, and underwent extended periods of subaerial exposure during which they were fragmented by animal activity. Thicker sandstone units within FA 8 may represent crevasse splays deposited on top of the levee deposits (Fielding *et al*. [1993] have made similar interpretations in other fluvial deposits).

Facies association 9: laterally extensive mudrock:

Description: FA 9 is very common and consists of lithofacies Fg, Fc, Fp, C and Fb. Lower contacts are sharp or gradational, while upper contacts can be sharp or undulatory. Facies Fg is the dominant component, with facies Fp and Fc either forming thin, laterally persistant horizons or else isolated lenses that are typically localised. Facies C occurs as thin stringers within Fg or Fc and is commonly associated with other identifiable plant material. Facies Fb forms thin beds that have a distinctive popcorn texture. These beds of Fb are either localised, contain euhedral crystals of feldspar, and have concave-up bases, or else have a tabular geometry and can be traced for tens or hundreds of meters.

The typical succession of beds observed in FA 9 consists of interbedded Fg and Fc in the basal part, with Fg forming a greater proportion of the succession. Fg units can vary between 0.5 and 1.5 m, whereas Fc units have not been observed to exceed 0.5 m. Fp commonly occurs at the top of the succession, separated from the underlying Fg or Fc with a gradational contact. Rhizocretions, nodules and bioturbation, as well as remnants of soil structures such as peds and cutans are preserved in facies Fp.

Fossil content in FA 9 is abundant and includes invertebrate, plant and vertebrate fossils. Vertebrate fossils from overbank fines can be fragmentary or contain complete, disarticulated skeletal elements.

Interpretation: FA 9 represents overbank deposition on a waterlogged floodplain with reducing conditions as indicated by the drab colour (Table 2.1.), which is similar to poorly drained floodplains of the Willwood Formation described by Kraus (1997). The invertebrate fossil content also supports a subaqueous environment. Facies Fc is considered to represent subaqueous deposition in ponds or lakes on the floodplain in which abundant organic material, especially leaves, wood, invertebrate shells and vertebrate bone, accumulated and was preserved in an anoxic environment, while coal stringers indicate organic content that has been preserved and compressed. Facies Fg, which typically has weakly to well-developed lamination, as well as invertebrate shells and vertebrate bone, most likely represents subaqueous sedimentation. This is further supported by the lack of pedogenic characteristics in Fg deposits. Facies Fp, which typically overlies Fg or Fc with gradational contacts, is interpreted to represent palaeosol accumulations that developed in response to extended subaerial exposure of the floodplain sediments under conditions of low net sedimentation. Rhizocretions, bioturbation and nodule development as well as soil structures (slickensides, cutans and peds) indicate the development of soil horizons, and the red colour indicates that oxidation occurred, suggesting that the sediment was well drained (Kraus 1997; Therrien 2005), in contrast to the underlying lithofacies. The transition from waterlogged Fg and Fc to well-drained Fp is repeated in cycles of overbank deposition in the Wahweap Formation and is possibly governed by proximity to major channels, with a waterlogged floodbasin occurring proximal to major channels and well-drained floodbasin with pedogenesis occurring distally (e.g. Reading 1996; Kraus 1997, McCarthy and Plint 2003). Facies Fb occurs where ash from volcanic eruptions settled on the floodplain and was preserved on floodplains. A similar mechanism is described for ash preservation in the overlying Kaiparowits Formation (Roberts *et al*. 2005). Localised beds of Fb with a lenticular geometry are from ash accumulations in palaeo-depressions on the floodplain which were subsequently covered by epiclastic floodplain sedimentation. The greatest concentrations of euhedral crystals occur in such beds. By contrast, tabular horizons of Fb are less ‘pure’ and most likely represent ash-rich sediment mixed with a minor component of epiclastic sediment.

Facies association 10: lenticular mudrock:

Description: Lenticular mudrocks occur in the upper part of the middle member. This element typically has a concave-up base which truncates underlying units, and has a lateral extent of approximately 4-8 m and a thickness of up to 3 m. Width:height ratio is less than 15. It is dominated by Fg, which varies from massive to possessing crude horizontal stratification. Minor, lens-shaped bodies of Sm, about 20 cm thick and up to 80 cm wide occur encased within the grey mudrock, but these sandstone lenses make up less than 10 % of the overall volume of the facies association. The upper contacts of lenticular mudrocks are sharp and they are typically overlain by channel sandstones.

FA 10 is exceptional in that the grey mudrock in this element was found to have very little fossil content. Carbonised plant remains were the only organic constituents recovered from this element.

Interpretation: Lenticular mudrocks are interpreted as the fill of muddy channels, similar to those described by Ghosh *et al*. (2006). The lower contact represents an incisive channel base which eroded into underlying overbank deposits and the lenticular shape, size and lack of internal erosive surfaces indicate single-storey channels. Muddy channel fill has been described from suspended-load dominated stream systems in fluvial systems (Gibling *et al*. 1998; Ghosh *et al*. 2006) and is considered to develop in response to a decrease in river energy brought on by avulsion**.** In the Wahweap Formation, muddy channel fill is most likely preserved as the result of low-energy stream action.

2.5.1. Alluvial architecture of the lower member

Facies associations present in the lower member are inclined interbedded sandstone and mudrock, minor tabular sandstones, major tabular sandstones and gravel bedforms. The lower-middle member contact is marked by a 5-10 m thick laterally extensive major tabular sandstone body with several unique characteristics. Facies Gc and Gs occur more abundantly in this horizon than anywhere else in the formation, while facies Fcr occurs in association with this sandstone unit. The lower member is found to be thinner (20-40 m) and more sandstone-rich in the northern part of the Kaiparowits Plateau and thicker (up to 60 m) and more mudrock-rich in the southern part of the plateau. The lower contact of the lower member is locally sharp, where an extensive mudrock unit overlies the Drip Tank Sandstone or else gradational, where a gradual decrease in the sand:mud ratio is present and extended sandstone bodies become less abundant upwards in the stratigraphic succession. The lower-middle member contact is sharp.

2.5.2. Alluvial architecture of the middle member

The dominant facies association in the middle member is laterally extensive mudrock. Single and multi-storey lenticular sandstones, minor tabular sandstones, lenticular interlaminated sandstone and mudrock, low-angle heterolithic cross-strata, inclined interbedded sandstone and mudrock, bioturbated mudrock and lenticular mudrock also occur. Extensive horizons of facies Fc, and minor C occur in association with Fg in the middle member. Lenticular interlaminated sandstone and mudrock, lenticular mudrock and low-angle heterolithic cross-strata only occur in the upper part of the middle member (Fig. 2.5.). The middle-upper member contact is typically gradational and based on an increase in the sand:mud ratio, or the first appearance of major tabular sandstone.

2.5.3. Alluvial architecture of the upper member

The upper member contains major tabular sandstones, single and multi-storey lenticular sandstones, low-angle heterolithic cross-strata, lenticular interlaminated sandstone and mudrock and inclined interbedded sandstone and mudrock (Fig. 2.6.; Fig. 2.7.). Mudrock-rich or heterolithic elements are more common at the base of the upper member while laterally extensive major tabular sandstone units (Fig. 2.8.) occur towards the top. The upper-capping sandstone member contact is sharp to erosive, in some cases with several meters of incision, and is marked by an increase in grain-size as well as the appearance of extraformational conglomerate and gravel bedforms. The degree of sandstone amalgamation also increases from the upper member to the capping sandstone.

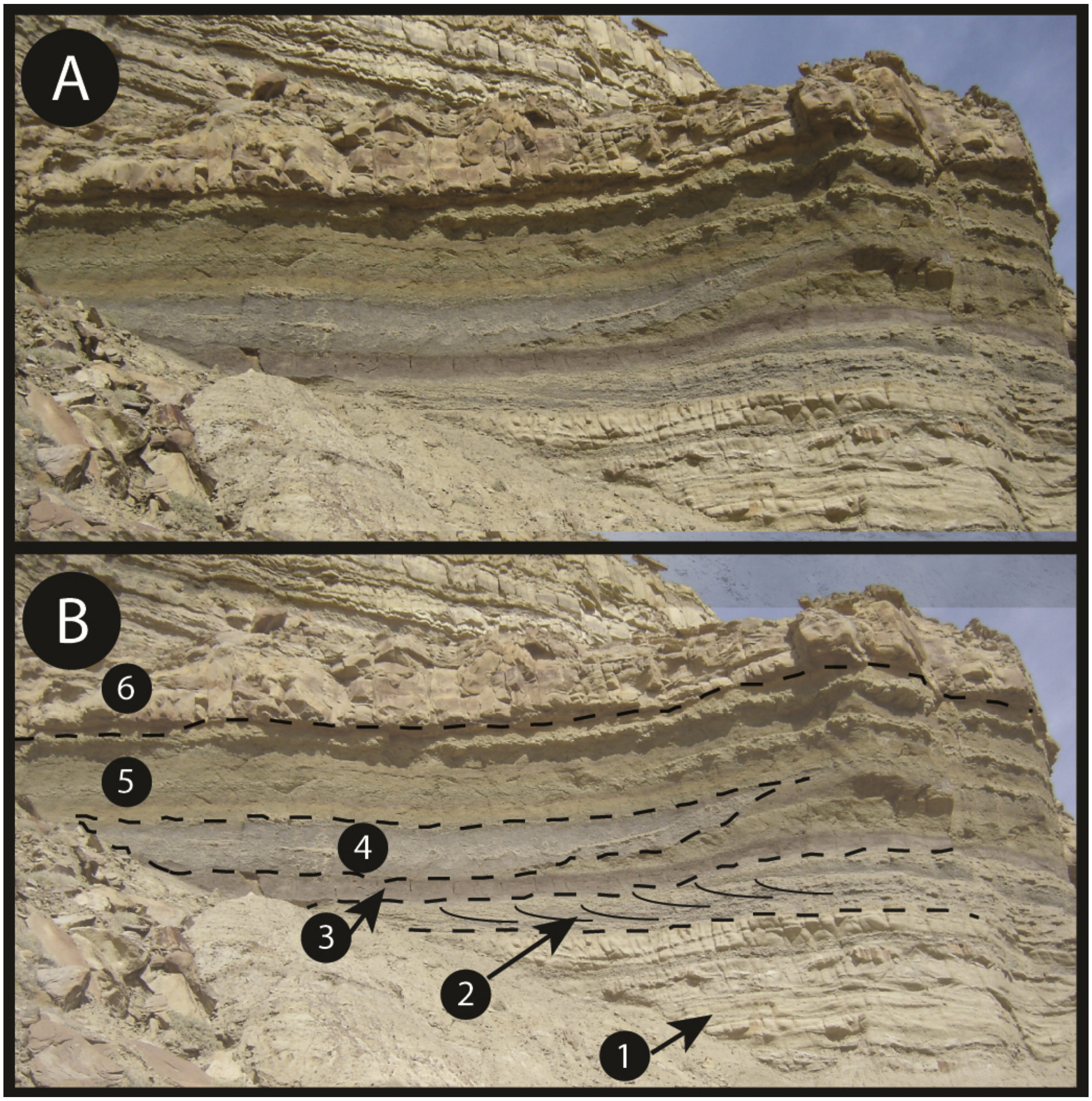


Figure 2.5. A) Photograph and B) Interpretation of facies and facies associations of the upper part of the middle member. 1, lenticular interlaminated sandstone and mudrock; 2, low-angle heterolithic cross-strata; 3, carbonaceous mudrock (facies Fc); 4, lenticular mudrock; 5, inclined interbedded sandstone and mudrock; 6, minor tabular sandstone body.

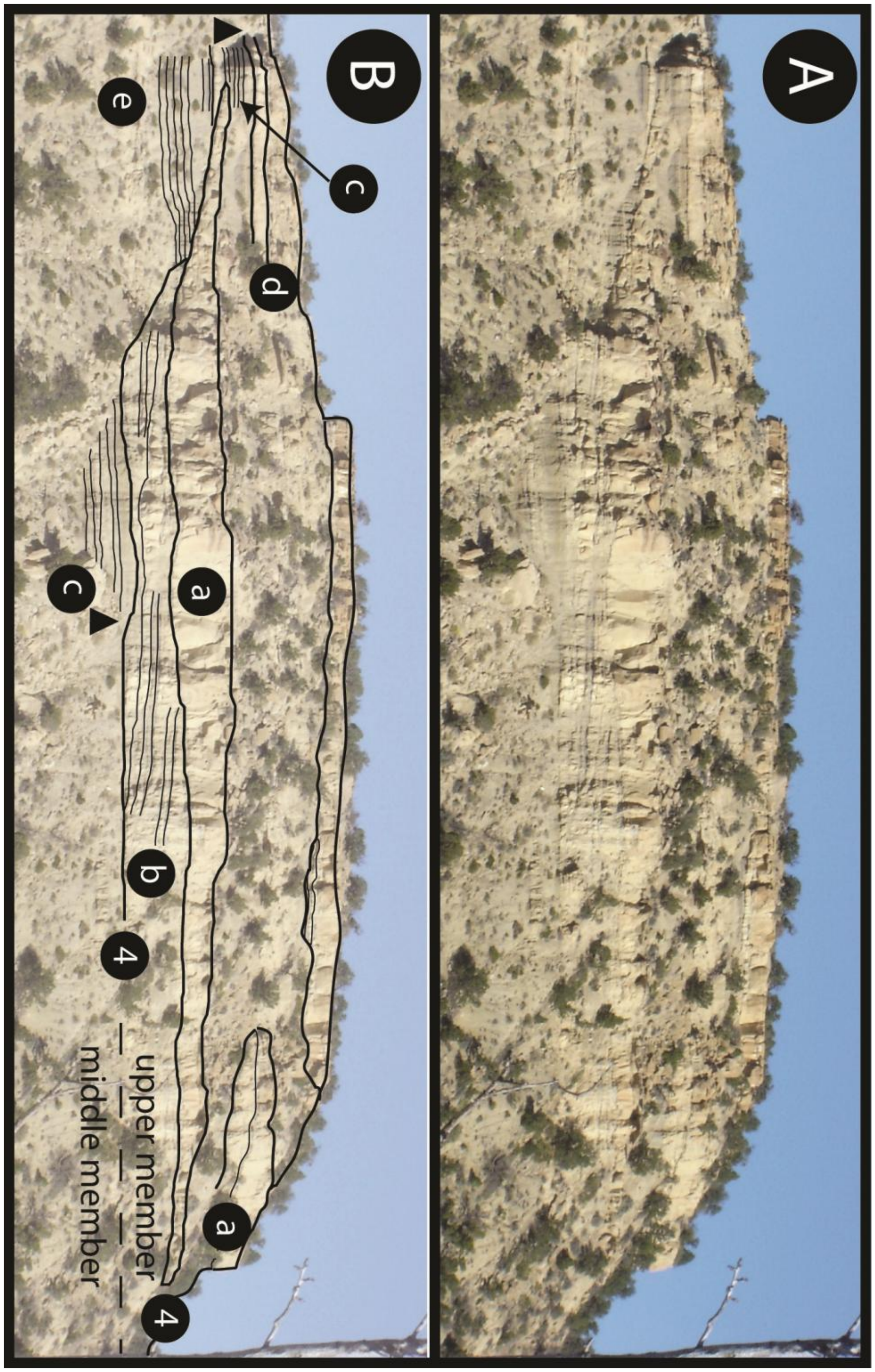


Figure 2.6. A) Photograph and B) Interpretation of common facies associations in the Wahweap Formation. a, multistory lenticular sandstone; b, low-angle heterolithic cross-strata; c, inclined interbedded sandstone and mudrock; d, minor tabular sandstone body; e, laterally extensive mudrock; 4, fourth-order bounding surface; black triangles indicate upward-fining successions.

Figure 2.7. A) Photograph and B) Interpretation of facies associations of the upper member. a, multistory lenticular sandstone; b, inclined interbedded sandstone and mudrock; c, low-angle heterolithic cross-strata; d, laterally extensive mudrock (middle member); e, laterally extensive mudrock (middle member).

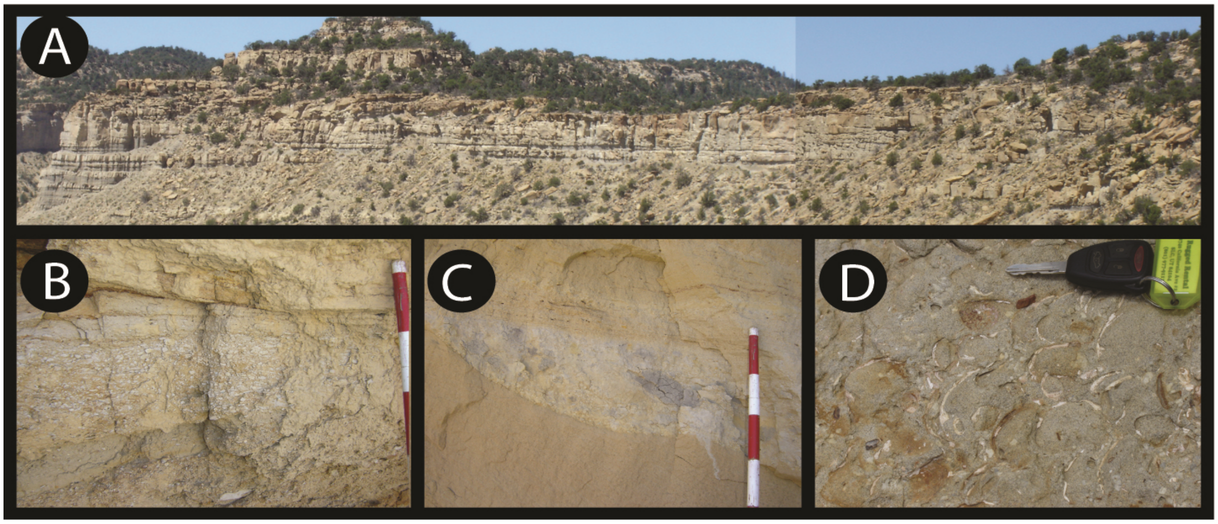


Figure 2.8. Photomosaic (A) of major tabular sandstone (FA3). Common lithofacies in FA3 include (B) intraformational clast-supported conglomerate (Gci); (C) intraformational matrix-supported massive conglomerate (Gmi) overlying a fourth-order surface; and (D) Clast-supported shell conglomerate (Gs).

2.5.4. Alluvial architecture of the capping sandstone member

The facies associations which occur in the capping sandstone are major tabular sandstone bodies, gravel bedforms and minor tabular sandstone bodies. The capping sandstone typically crops out as a thick conglomerate-sandstone unit with numerous conspicuous third- to fifth-order bounding surfaces (as defined by Miall, 1985). On the Kaiparowits Plateau, the top of the capping sandstone is truncated by erosion in most of the areas, but measurement of the complete unit where it is conformably overlain by the Kaiparowits Formation indicates a thickness of 60 – 90 m.

The contact between the capping sandstone and the Kaiparowits Formation is either sharp and concordant (although the presence of root traces suggests that there may have been a lapse in deposition), or else gradational, where a number of minor sandstone bodies similar to those of the capping sandstone persist in the lower part of the Kaiparowits Formation.

2.6. Sandstone composition and palaeocurrents

Point counting (Appendix A) reveals that sandstone composition varies through the lower, middle and upper members. In the lower member, 16 sandstone thin sections were analysed and found to have a composition variable from arkosic to litharenitic and sublitharenitic (Fig. 2.9.). Palaeocurrent direction in the lower member, from trough cross-bedding, has a mean of 33^{o} (n=119).

In the middle member, 5 thin sections reveal a litharenitic-sublitharenitic composition. Palaeocurrent direction in the middle member has a mean of 17.4^{o} (n=46). In the upper member, 22 thin sections have a litharenitic-sublitharenitic composition. Palaeocurrent direction has a mean of 52^{o} (n=93).

2.7. Palaeoenvironment

2.7.1. Palaeoenvironmental interpretations – lower member

The lower member preserves siltstone-dominated upward-fining successions which are interpreted to have formed mostly by overbank deposition in the proximal floodbasin subenvironment. The presence of abundant grey laminated mudrock in inclined interbedded sandstone and mudrock as well as coalified wood, invertebrate shell conglomerate and carbonaceous mudrock all signify a wet palaeoenvironment (e.g. Kraus 1997), while siltstone with calcrete nodules, and calcrete-nodule conglomerates may indicate seasonal aridity. The occurrence of lateral accretion deposits as well as the low sand:mud ratio are consistent with Eaton’s (1991) interpretation of a meandering-river depositional environment. Additionally, the high proportion of overbank elements and general lack

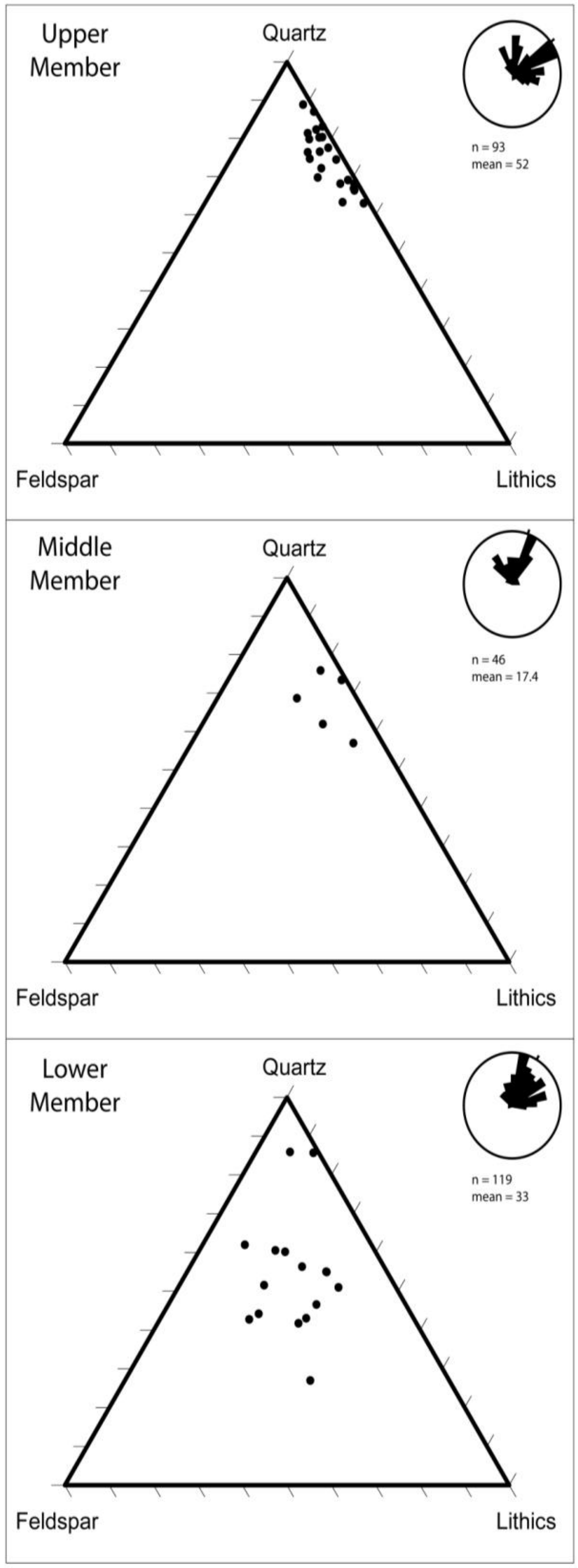


Figure 2.9. Sandstone composition and palaeocurrent for the lower, middle and upper members.

of features indicating subaerial exposure (such as desiccation cracks) are consistent with an interpretation of river systems that frequently burst their banks and developed waterlogged floodplains. Single-storey channel systems cut into overbank deposits suggesting that floodplains were cohesive and there was avulsion caused by aggradation within the channel system (Törnqvist 1994; Slingerland and Smith 2004). This feature, coupled with rhizocretions and abundant plant matter preserved in the overbank environment (distant from channels), suggests densely vegetated floodplains. The absence of continuous coal seams within the Wahweap Formation is more likely related to rapid sediment accumulation rates than to an absence of carbonaceous or organic-rich material. A similar reason for lack of coal seams is postulated for the overlying Kaiparowits Formation (Roberts 2007).

The major tabular sandstone at the top of the lower member (Fig. 2.10.) preserves intraformational conglomerates and contains third-order erosive surfaces that separate individual upward-fining cycles. This multi-storey sandstone body represents a shift from aggradation to amalgamation as the dominant depositional style, possibly indicative of a reduction in accommodation space (Wright and Marriott 1993; Komatsubara 2004).

2.7.2. Palaeoenvironmental interpretations - middle member

The abundance of grey laminated mudrock and carbonaceous mudrock in the laterally extensive mudrock facies association, together with single-storey lenticular channels and inclined interbedded sandstone and mudrock, suggests that depositional systems in the middle member were aggradational. Distal and proximal floodbasin deposits represent approximately 80 % of the middle member. Drab mudrock colours and the preservation of aquatic

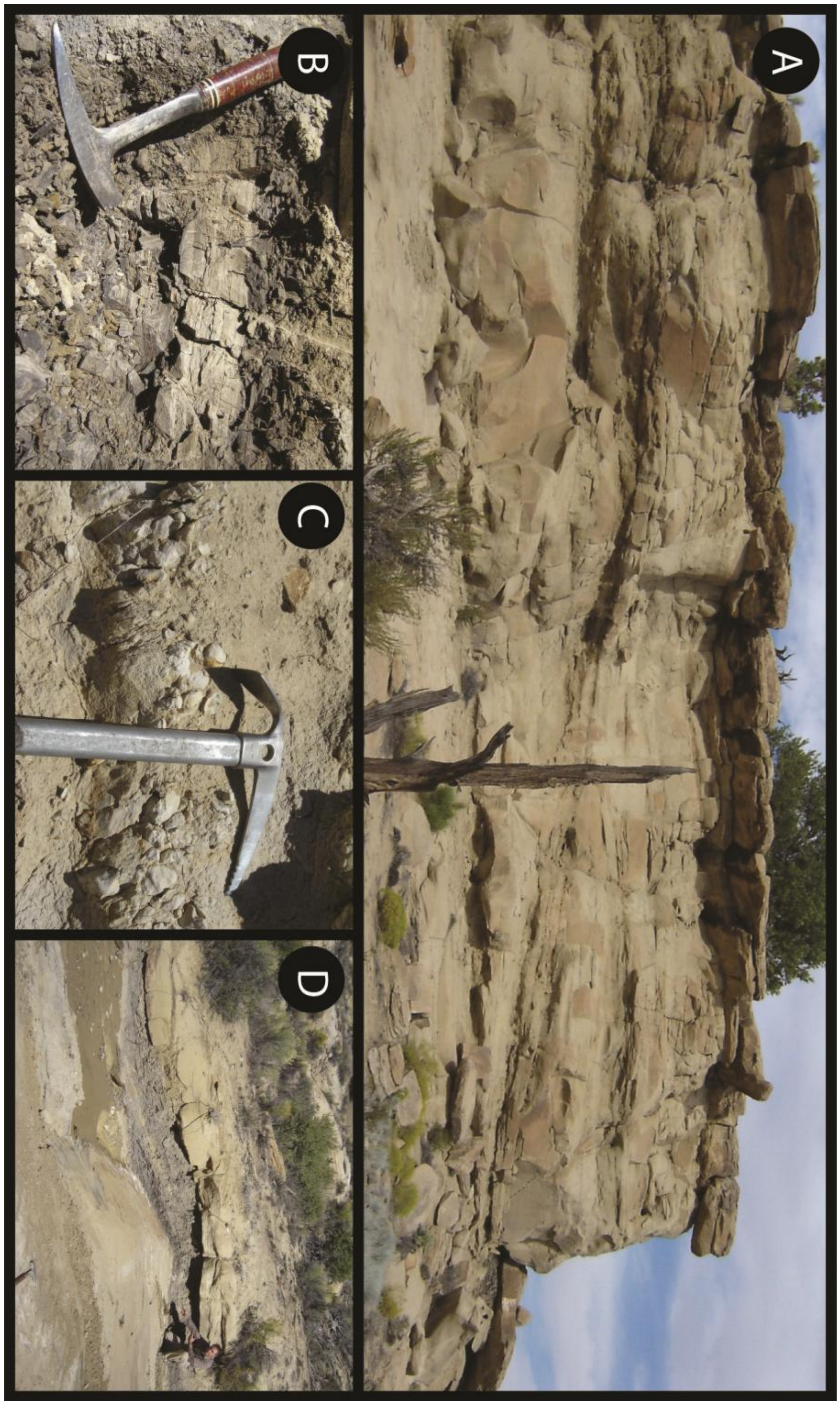


Figure 2.10. Features of the lower member: A) Major tabular sandstone (FA3) caps the lower member. B) horizontally laminated siltstone (Fg) from FA 9 (overbank deposits). C) clast-supported calcrete nodule conglomerate (Gc) from FA3. D) Inclined interbedded sandstone and mudrock (FA9).

invertebrates and plant matter in fine-grained deposits indicate a wet environment with a waterlogged and vegetated floodplain and numerous ponds and swamps (Kraus 1997). Mud-filled channels suggest an extremely low-energy environment, possibly with stagnant water conditions (Ghosh *et al*. 2006), while the presence of lenticular interlaminated sandstone and mudrock in the upper part of the middle member suggests a degree of tidal influence. The low sand:mud ratio and isolated lenticular nature of channels, as well as the lack of amalgamation surfaces are characteristic of rapid sedimentation by meandering or anastomosing rivers in a high-accommodation space setting (Wright and Marriott 1993; Adams and Bhattacharya 2005). Aggradation is the dominant fluvial style in such settings (Makaske 2001; Slingerland and Smith 2004) and leads to abundant overbank deposition and channel avulsion. In mud-rich systems such as this, with isolated single-storey sandstone bodies, Kraus and Wells (1999) have proposed that channel avulsion is an important process and rapid sedimentation inhibits pedogenesis in the floodbasin. This appears to be the case in the middle member of the Wahweap Formation on the Kaiparowits Plateau, where only incipient pedogenesis affected the fine-grained lithologies.

2.7.3. Palaeoenvironmental interpretations - upper member

Facies associations indicative of tidal influence (inclined heterolithic stratification and tidal couplets) occur at the base of the upper member (Fig. 2.11.), which, along with the presence of the brackish-water trace fossil *Teredolites* (Gingras *et al*. 2004), abundant mud drapes on foresets and flaser bedding suggests that marine incursion and possibly the development of estuarine conditions occurred during deposition of the upper member. The increase of sand:mud ratio and the presence of at least three laterally extensive major tabular sandstone bodies towards the top of the upper member, as well as an increase in the degree of amalgamation of sandstone units suggests a reduction in accommodation space at the time of

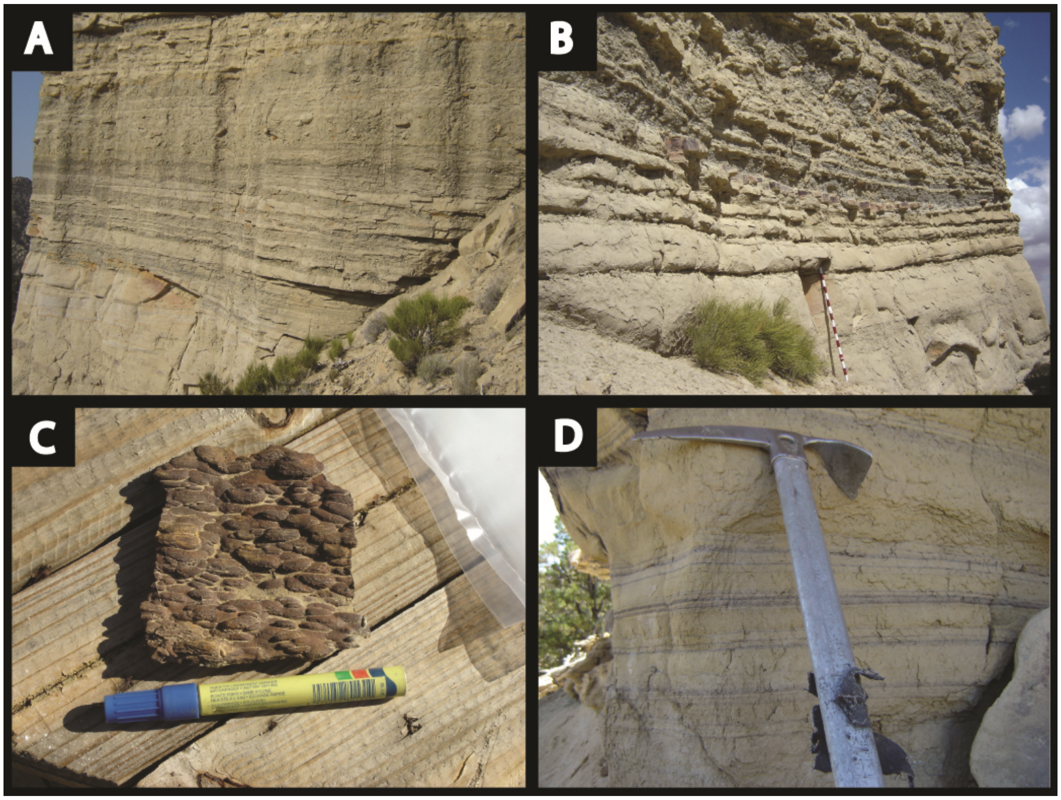


Figure 2.11. Features indicative of tidal influence at the top of the middle and bottom of the upper members: A) lenticular interlaminated sandstone and mudrock (FA8) interpreted as tidal couplets; B) heterolithic stratification; C) trace fossil *Teredolites* collected from the upper member; D) mud-draped laminae in upper member sandstone.

upper member deposition. Additionally, the absence of any tidal indicators in the upper part of the upper member suggest that tidal influence had diminished during its deposition.

2.7.4. Palaeoenvironmental interpretations - capping sandstone member

Alluvial architecture in the capping sandstone member records a drastic shift from the lower three members. The capping sandstone is sandstone-dominated with numerous amalgamation surfaces and both intra- and extra- formational conglomerates, suggestive of a low-accommodation space depositional setting. Previous studies (Eaton 1991; Pollock 1999) found the capping sandstone to have been deposited in a braided-river environment, in keeping with the facies (gravel bedforms, major tabular sandstone) noted in this study. The capping sandstone also preserves a shift in palaeocurrent direction (Lawton *et al*. 2003; Jinnah *et al*. 2009a), which corresponds to the change in depositional environment.

2.8. Conclusion

Facies analysis in the Wahweap Formation on the Kaiparowits Plateau reveals a meandering-anastomosed river depositional setting for the lower, middle and upper members, and a braided river setting for the capping sandstone, in-keeping with previous workers' assessment of the formation. Although conditions were generally wet, there are indications of seasonal aridity, especially in the lower member. The depositional system was tidally influenced during deposition of the upper middle and lower upper members, possibly with the development of fully estuarine conditions. Tidal influence diminished as deposition of the upper member progressed, and the depositional style changed to a braided system at the initiation of capping sandstone deposition.

3. ^{40}Ar-^{39}Ar Age Dating and Correlation of the Wahweap Formation

3.1. Introduction

Formations in the Western Interior Basin are generally dated by a combination of biostratigraphy and radiometric dating of ash beds. Biostratigraphically, invertebrates, especially molluscs have been especially useful for correlation, and Western Interior Ammonite zones are widely used to correlate marine strata (Gill and Cobban, 1973). For continental deposits, mammal fossils have been used to define “North American Land Mammal Ages” (NALMAs) which have also been used for correlation (Lillegraven and McKenna, 1986). NALMAs have since been modified to incorporate other elements of terrestrial faunas and are thus now referred to as “Land Vertebrate Ages” (e.g. Sullivan and Lucas, 2006).

Where ash beds are preserved in the Western Interior Basin (e.g. in strata deposited close to the Elkhorn Mountain Volcanic centre in the northern part of the basin), they have proved useful for obtaining high precision radiometric dates that can be used to further refine and test palaeoecological hypotheses.

3.2. Age bracket for the Wahweap Formation

Age constraints for the Wahweap Formation are currently based on a combination of mammalian and palynomorph biostratigraphy (Eaton, 1991, 2006; Lawton *et al*., 2003). The most significant work to this effect is based on the multituberculate mammal zonation by Eaton (2002), which indicates an early Campanian age (Eaton, 2002, 2006). A dated ash bed from ~80 m above the base of the overlying Kaiparowits Formation yields an age of 75.96 ± 0.15 Ma (Roberts *et al*., 2005). Based on calculations of sediment accumulation rates, Roberts *et al*. (2005) suggest an age of ~76.1 Ma for the Wahweap-Kaiparowits contact. Moreover, both Peterson (1969) and Roberts (2007) suggest a relatively conformable contact between the two formations. Thus, an upper age bracket of ~76.1 Ma is assumed for the top of the Wahweap Formation.

Age constraint for the base of the Wahweap Formation is more contentious due to a lack of age constraining taxa in the base of the Wahweap Formation and a paucity of fossils and palynomorphs from the underlying Drip Tank Member of the Straight Cliffs Formation (Eaton, 2006). Palynomorphs in the John Henry Member of the Straight Cliffs Formation suggest an age somewhere between Santonian and earliest Campanian for the Wahweap-Straight Cliffs contact.

3.3. Methods

Bentonites CF05-B and SS07-B are located within Grand Staircase-Escalante National Monument on the Kaiparowits Plateau. The bentonites are located along the Smoky Mountain Road between Camp Flats and Star Seep. CF05-B occurs approximately 300-500 m west of the road, towards the top of a small bench below the steep slope leading up to the prominent plateau. CF05-B crops out in the middle member (Fig. 3.1.), approximately 40 m above the base of the Wahweap Formation (Fig. 3.2.), where it has a sharp contact with the underlying Drip Tank Sandstone. CF05-B occurs at the following GPS coordinates: UTM 12 S 0429987/4172415, as a 20 cm-thick, pale yellow-green unit

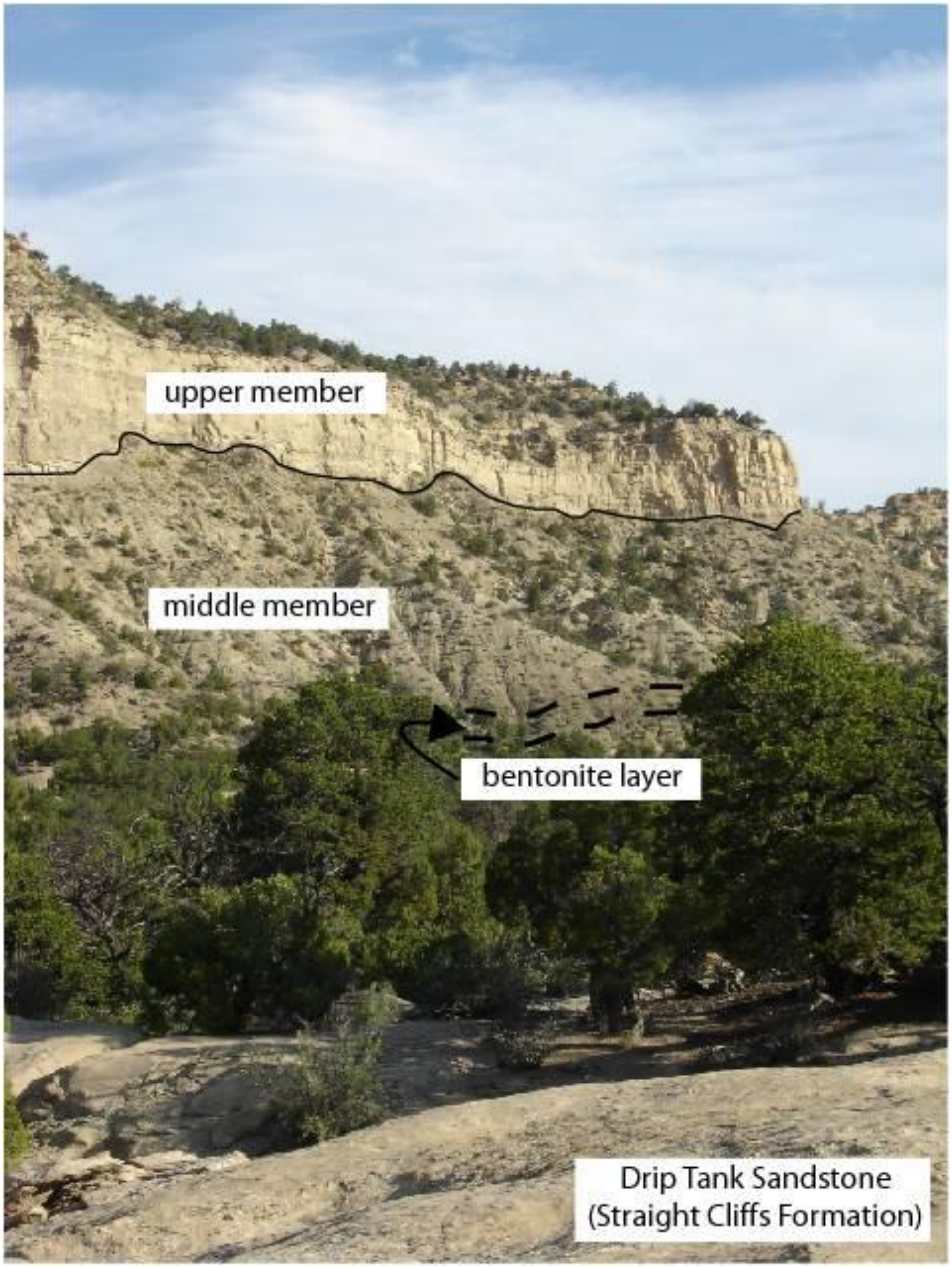


Figure 3.1. Position of bentonite CF05B stratigraphically in the Wahweap Formation on Smoky Mountain Road

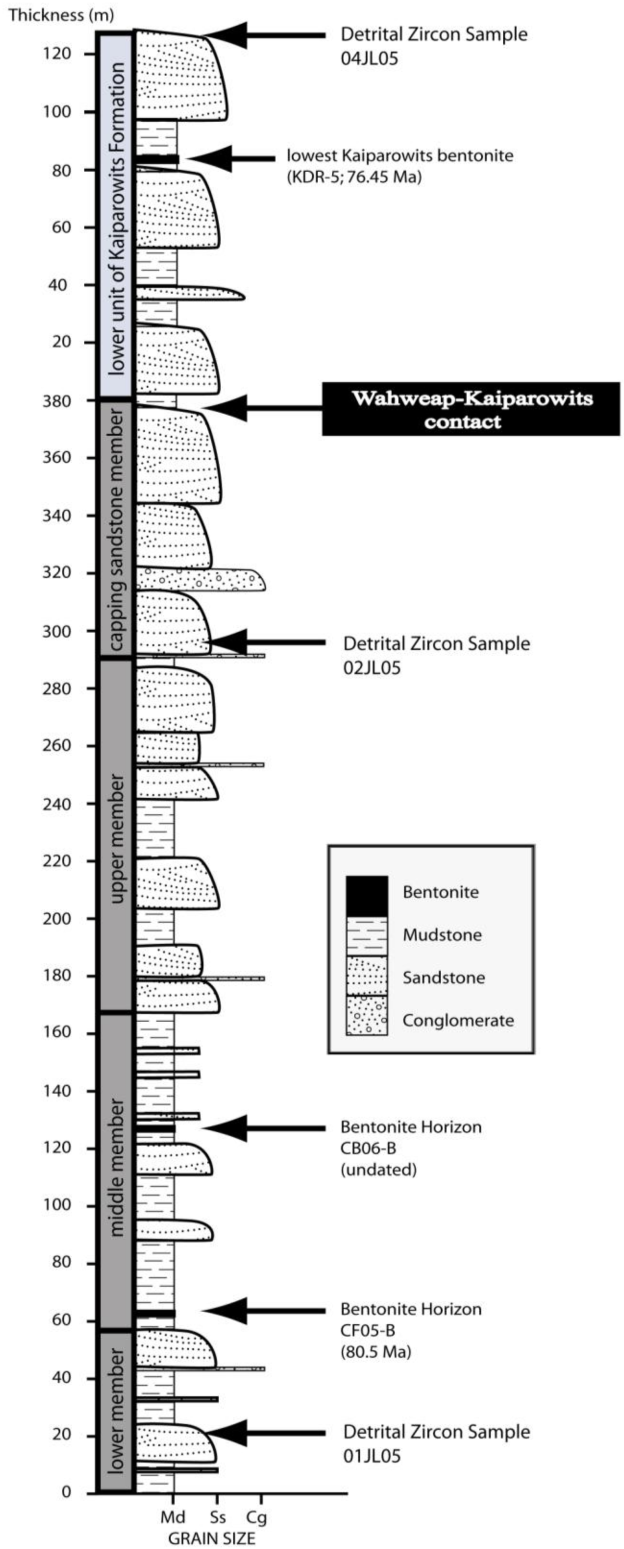


Figure 3.2. Position of bentonite CF05B in a composite stratigraphic section of the Wahweap Formation. Positions of detrital zircon samples from Larsen (2007) also shown.

with surficial popcorn weathering that lies sharply above a dark grey-green overbank mudstone, and is gradationally overlain by another similar mudstone. The bentonite can be traced laterally to the south for several hundred meters before it is erosionally truncated by a medium grained (fluvial) sandstone body.

SS07-B occurs south of CF05-B, at GPS coordinates UTM 12 S 0443826/4158544, and is exposed in the sloped bank of a dry wash about 300 m from the road.

Approximately 20 kg of bentonite from samples CF05-B and SS07-B was collected in the field. The samples were disaggregated using warm tap water and repeated agitation, and then sieved through 20, 40, 60, 100 mesh screens. Feldspar was separated from the remaining coarse fraction using a Franz isodynamic separator and then sanidine was isolated from plagioclase by density separations using heavy liquids.

^{40}Ar-^{39}Ar dating was performed at Berkeley Geochronology Center (see Jinnah *et al.*, 2009a, which reports the age of CF05-B, for details of the procedure). Sanidine from the Fish Canyon Tuff of Colorado was used as a mineral standard, with a reference age of 28.02 Ma (Renne *et al*., 1998). The reference age of the Fish Canyon Tuff was updated by Kuiper *et al.* (2008) to 28.201 Ma, and all dates that have used this a standard (including CF05-B and SS07-B) have therefore also required revision.

3.4. Ages of bentonites CF05B and SS07B

Five sanidine grains provided an eruptive age in CF05-B, and two sanidine grains provided an eruptive age in SS07-B.

The five acceptable sanidine analyses for CF05-B (80.1 ± 0.3 Ma) are shown in Fig. 3.3. accompanied by an age-probability density diagram summed from the individual ages. The two sanidine analyses for SS07-B (79.9 ± 0.3 Ma) are shown in Fig. 3.4.

Since the revision of the Fish Canyon tuff standard age from 28.02 to 28.2 Ma by Kuiper *et al.* (2008), this new standard has been widely applied to new Ar^{40}-Ar^{39} age dating and has been used to recalibrate legacy ages. Using the Ar/Ar legacy age converter Excel Macro freely available on the EARTHTIME website, the legacy age of 80.1 Ma published by Jinnah *et al*. (2009a), now becomes 80.6 Ma, and the legacy age of 79.9 Ma changes to 80.4 Ma, giving an average age of 80.5 Ma for the lower part of the middle member.

3.5. Average undecompacted sediment accumulation rate for the Wahweap Formation

The new average radiometric age of 80.5 Ma and maximum age constraints from detrital zircon populations (82 ± 2 Ma for the lower member and 77 ± 2 Ma for the capping sandstone) fit within constraints afforded by ages from both the base of the

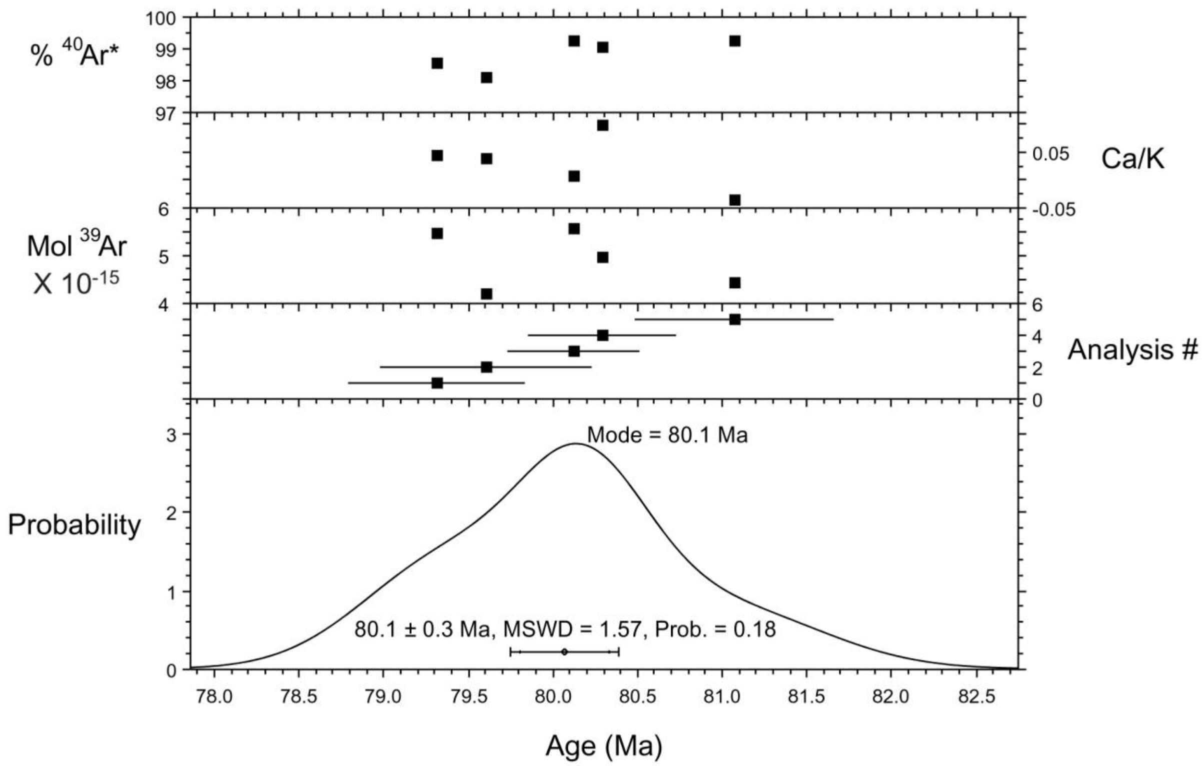


Figure 3.3. Age-probability spectrum for sanidine from sample CF05-B before correction to the new Fish River tuff standard. MSWD: Mean squared weighted deviate, prob: probability

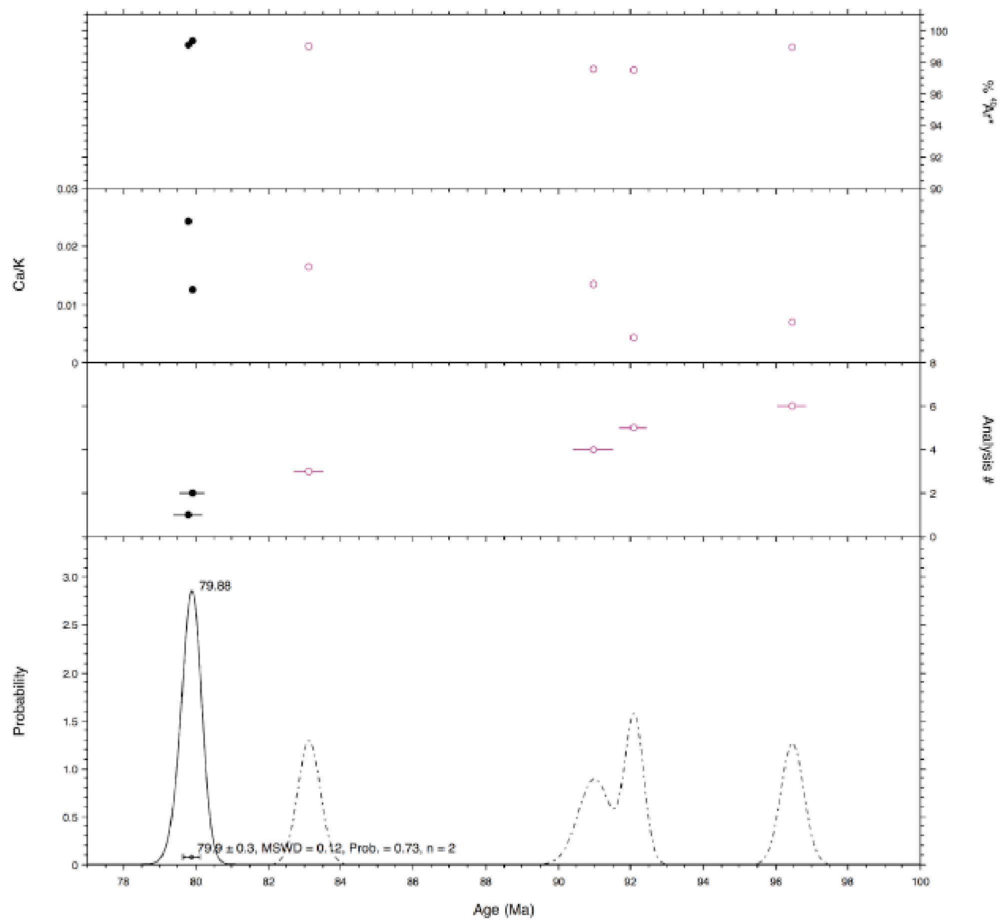


Figure 3.4. Age probability spectrum for sanidine from sample SS07-B before correction to the new Fish Canyon Tuff standard. MSWD: Mean squared weighted deviate, prob: probability

overlying Kaiparowits Formation (^{40}Ar-^{39}Ar age of 76.45 ± 0.14 Ma from a bentonite; Roberts et al., in press) and underlying Straight Cliffs Formation, biostratigraphically dated to latest Santonian or possibly earliest Campanian (Eaton, 1991, 2006; Lawton *et al*., 2003).

A preliminary, undecompacted sediment accumulation rate for the Wahweap Formation was determined by calculating the total rock thickness between the bentonite horizon, which is located 40 m from the base of the Wahweap Formation and the lowest dated bentonite in the overlying Kaiparowits Formation (76.45 Ma; ~80 m from the Wahweap top contact), and dividing that thickness (420 m in study area) by the difference between the ages of the two dated bentonites (4.05 Ma). This calculation indicates an undecompacted sediment accumulation rate of 10.4 cm/ka for the Wahweap Formation (plus the basal 80 m of the Kaiparowits Formation), which is significantly less than that of the overlying Kaiparowits Formation (~41 cm/ka; Roberts *et al*., 2005). This method does not account for unconformities that may exist in the Wahweap Formation, and is therefore prone to error, but is useful as a rough estimate of the age of the upper and lower contacts for comparison and correlation.

Utilizing the sediment accumulation rate, 10.4 cm/ka, suggests that the age for the base of the Wahweap Formation (40 m below the 80.6 Ma bentonite) is approximately 81.0 Ma. Therefore, deposition of the Wahweap Formation, in the study area likely commenced at 81 Ma and ended somewhere between ~77.5 and 76.1 Ma. However, the apparently gradational contact (Peterson, 1969; Roberts, 2007) between the Wahweap and Kaiparowits formations supports an age closer to 76.1 Ma (Roberts *et al*., 2005) for the top of the Wahweap Formation, which would mean that the formation fits wholly within the middle Campanian.

Further age constraint is provided by the youngest detrital zircons found in the lower (01JL05) and capping sandstone members (02JL05), which provide a maximum limit to the age of deposition (Larsen, 2007; Jinnah *et al*., 2009a). The youngest zircon found in the lower member has an age of 82 ± 1 Ma, which provides a maximum age limit slightly older than, but in keeping with the age of the base of the Wahweap calculated above. In the capping sandstone, the youngest zircon has an age of 77 ± 2 Ma which provides additional support for the overall middle Campanian age bracket for Wahweap deposition.

3.6. Regional stratigraphic correlation

Utilizing the new radiometric age constraints for the Wahweap Formation, improved regional correlations can now be made with greater confidence. Continental deposits coeval to the Wahweap Formation are exposed across the Western Interior Basin in Big Bend National Park in Texas, the San Juan Basin in New Mexico, and the Henry Basin, the Markagunt and Paunsaugunt plateaus, and the Book Cliffs, all in Utah (Fig. 3.5.). Of particular interest is the correlation between strata in the Kaiparowits, Markagunt and Paunsaugunt plateau with the Book Cliffs and the Henry Basin. Wahweap equivalent strata on the Markagunt and Paunsaugunt plateaus, are considered to be roughly equivalent (temporally) to deposits on the Kaiparowits Plateau (Lawton *et al*., 2003), however more precise correlation between these areas and equivalent deposits to the west in Cedar Canyon and elsewhere, requires further investigation. The Wahweap Formation correlates with the lower to middle Castlegate Sandstone in the Book Cliffs (Fig. 3.6.); however no marked hiatus occurs in the Wahweap Formation between ~80.7 -79.0 Ma, as in the Book Cliffs (see McLaurin and Steel, 2007 and references therein). As demonstrated by Eaton (1990), the Masuk Formation in the Henry Basin likely represents the eastern, downstream facies equivalent to the lower, middle and upper members of the Wahweap Formation, while the Tarantula Mesa Sandstone is presumed to be equivalent to the capping sandstone member (Eaton, 1990). The Wahweap Formation correlates with the Trail and Rusty members of the Ericson Sandstone Formation in the Rock Springs Uplift, Wyoming (Gill *et al*., 1970; Martinsen *et al*., 1999).

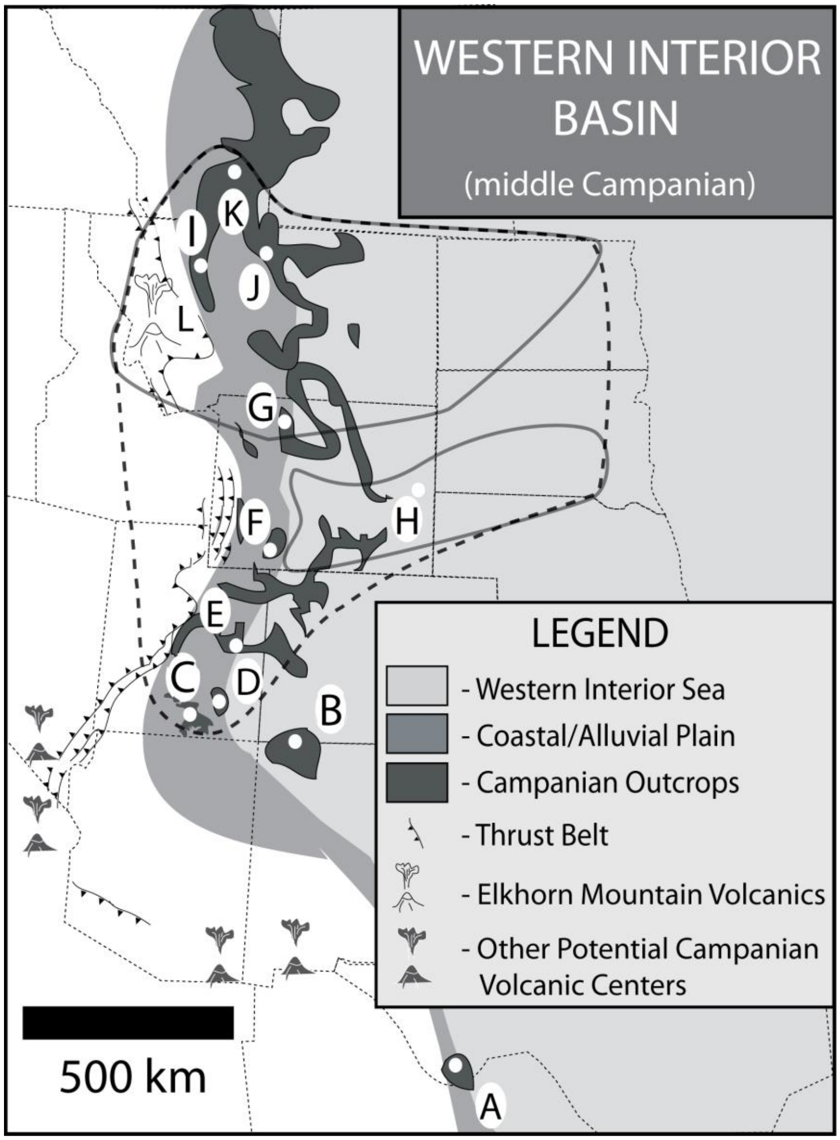


Figure 3.5. Map shows the approximate position of the Western Interior Seaway during the middle Campanian as well as the location of modern exposures of Campanian sedimentary strata. Letters in map refer to localities in Fig. 3.6. EMV: Elkhorn Mountain Volcanics, 1 (solid line): known distribution of Ardmore Bentonite (from Gill and Cobban, 1973; Thomas et al., 1990), 2 (dashed line): Proposed extent of Ardmore Bentonite during Campanian. Modified from Roberts et al. (2005).

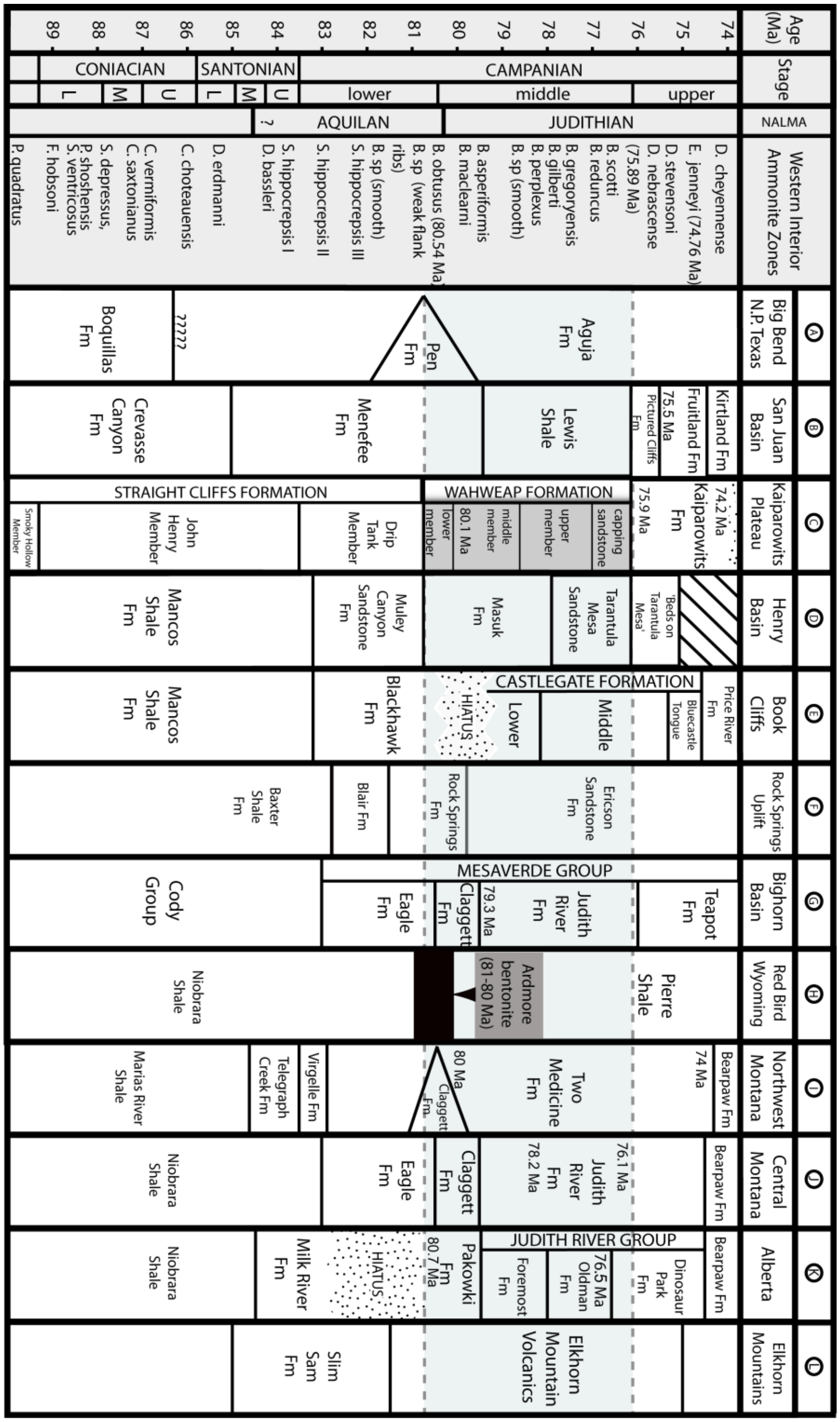


Figure 3.6. Correlation of Wahweap equivalent strata in the Western Interior Basin (data from Robinson et al., 1968; Gill et al., 1970; Gill and Cobban, 1973; Fouch et al., 1983; Lawton, 1986; Goodwin and Deino, 1989; Eaton, 1990, 1991, 2006; Rowe et al., 1992; Eberth and Hamblin, 1993; Elder and Kirkland, 1993; Rogers et al., 1993; Hicks et al., 1995, 1999; Martinsen et al., 1999; Fassett, 2000; Fitzsimmons and Johnson, 2000; Lageson et al., 2001; Sankey, 2001; Buillit et al., 2002; Payenberg et al., 2002; Lawton et al., 2003; Sullivan and Lucas, 2003; Roberts et al., 2005; McLaurin and Steel, 2007).

This study also permits improved regional correlation between a number of important vertebrate fossil-bearing continental successions (Fig. 3.6.). Most significantly with respect to mammalian and dinosaurian faunas, the Wahweap Formation can now be more precisely correlated with the lower Judith River Formation in Wyoming and central Montana and the middle portion of the Two Medicine Formation in western Montana. The lower and middle members of the Wahweap Formation are found to be correlative with the Pakowki and Foremost formations in Alberta, and younger than the Milk River Formation based on its radiometric age (Payenberg *et al*., 2002). The uppermost portion of the upper and the capping sandstone members of the Wahweap Formation correlate to the Oldman Formation and possibly with the very base of the Dinosaur Park Formation. To the south, the only significant continental vertebrate bearing succession coeval with the Wahweap Formation is the lower portion of the Aguja Formation in Big Bend National Park in Texas.

3.7. Possible sources for Wahweap Formation ash beds

The Cordilleran arc in California and western Nevada was the closest contemporaneous volcanic centre during deposition of the Wahweap Formation and is a likely source for ash beds. The eastern-most Cordilleran arc-related batholiths crop out approximately 450 km from the Kaiparowits Plateau, but palinspastic restoration of Tertiary extension places elements of the arc approximately 250 km closer to the study area (Wernicke *et al*., 1988), making this a likely source. Most of the Cretaceous extrusive volcanic rocks presumably associated with the Cordilleran Arc in southern-central California and eastern Nevada have been eroded during the Tertiary, leaving a limited record for correlation with ash beds in the Kaiparowits Basin. However, contemporaneous plutonic rocks such as the granite of the Sierra Nevada (Evernden and Kistler, 1970; Barth *et al*., 2004), as well as interbedded volcanic/volcaniclastic sedimentary complexes (Dunne *et al*., 1998) are still preserved indicating that igneous activity occurred in the arc throughout the Mesozoic. It is also important to recognise that a variety of other volcanic centres on the periphery of the Western Interior Basin were active during the Campanian and must also be considered as possible sources of ash in southern Utah.

A particularly important Campanian volcanic centre is the Elkhorn Mountain Volcanics, which were syn-tectonically emplaced between the Sapphire and Lombard thrust plates in west-central Montana during the early Campanian. They are widely regarded as the source for a large proportion of Campanian bentonites in the northern Western Interior Basin. In particular, the Elkhorn Mountain Volcanics have been identified as the source for the regionally extensive Ardmore bentonite, which represents a thick, succession of several closely spaced, Campanian ash beds preserved in alluvial to marine strata of the Western Interior Basin, extending across Alberta, Montana, Wyoming, North Dakota, South Dakota and Nebraska (Spivey, 1940; Gill and Cobban, 1973; Hicks *et al*., 1995, 1999; Payenberg *et al*., 2002). The age of the Ardmore bentonite (~81.0-80.04 Ma; see Hicks *et al*., 1995, 1999) overlaps with the ages reported here (80.4 and 80.6 Ma) for bentonites from the Wahweap Formation. Moreover, Smith (1960) recognised the ash-flow tuff field produced by the Elkhorn Mountain Volcanics as one of the most voluminous ever recorded.
Given the close temporal relationship between the ages of Wahweap Formation bentonites and the voluminous and explosive Elkhorn Mountain Volcanics, they must be considered a

possible source for air-fall volcanic deposits preserved in the Wahweap Formation. Although the Kaiparowits Basin is located ~900 km to the southwest of the Elkhorn Mountain Volcanics, the thin, discontinuous geometry and extremely fine phenocryst sizes of bentonites from the Wahweap Formation (and Kaiparowits Formation; Roberts *et al*., 2005) are certainly consistent with far-travelled ash falls. Additionally, models of large-scale Late Cretaceous wind patterns (e.g. Parrish and Curtis, 1982; Kump and Slingerland, 1999), coupled with isopach maps showing distribution patterns of the Ardmore bentonites (Gill and Cobban, 1973), support south-eastward transport in western North America.

Igneous activity in eastern Sonora (Mexico) began by 90 Ma and continued for 35 million years, based on U-Pb ages from volcanic rocks of the Tarahumara Formation (McDowell *et al*., 2001), making this a possible candidate for ash in the southern part of the Western Interior Basin during the Campanian. Rocks of the Tarahumara Formation include volcanic breccias and agglomerates consistent with the kind of explosive volcanism required to produce distal ash-fall fields.

Mapped Upper Cretaceous volcanic centres documented in eastern California, Arizona, New Mexico (the Hidalgo Formation) and Texas are ruled out as a source area as they are dated as late Campanian and Maastrichtian (Hayes, 1970; Drewes, 1978; Dickinson *et al*., 1989; Young *et al*., 2000; Breyer *et al*., 2007). The true source(s) of ash-fall units in the Kaiparowits Basin remains uncertain. While the Cordilleran arc would have been the closest contemporaneous volcanic centre and is perhaps the expected source area, the unprecedented explosive nature of the Elkhorn Mountain Volcanics requires that they also be considered. Such methods as geochemical fingerprinting of bentonites, which has recently been conducted in the contemporaneous Two Medicine Formation in Montana, holds promise for eventually determining the true source of air-fall volcanics preserved in the Kaiparowits Basin (Foreman *et al*., 2004).

3.8. Implications of radiometric dates for North American Land Mammal Ages (NALMAs)

The middle Campanian age demonstrated for the Wahweap Formation indicates that the original North American land mammal "age" of Aquilan assigned to the Wahweap Formation needs to be reconsidered. Cifelli (1990) and Eaton (2002) originally interpreted mammal specimens recovered from the Wahweap Formation to indicate an Aquilan age although they did note significant differences with the type-Aquilan fauna in the Milk River Formation of southern Alberta (Lillegraven and McKenna, 1986). More recently Eaton (2006) recognised that revised age assessments of the Milk River Formation by Payenberg *et al*. (2002) and others indicate that the Wahweap faunas are younger than those of the Milk River Formation, and Eaton (2006) suggested that the Aquilan extends into the early Campanian. In fact, radiometric dating of the Milk River Formation place it firmly within the Santonian and earliest Campanian (84.5 – 83.5 Ma; Leahy and Lerbekmo, 1995; Payenberg *et al*., 2002), significantly older than the Wahweap faunas (Kielan-Jaworowska *et al*., 2004). This implies that the age of the Aquilan NALMA is older than previously thought.

The data presented here confirm that the Wahweap Formation is significantly younger than the type-Aquilan Milk River Formation. Moreover, the majority of the Wahweap Formation is clearly coeval with the type-Judithian fauna from the Judith River Formation (Fig. 3.6.) which suggests that the Aquilan age assignment for the Wahweap Formation may need further consideration. This is significant because several mammal taxa found in the Wahweap Formation have been regarded as Aquilan (see Eaton, 2002; Kielan-Jaworowska *et al*., 2004).

Although North American land mammal “ages” were intended to be defined by faunal content rather than absolute age (Lillegraven and McKenna, 1986), it is notable that the Wahweap Formation overlaps temporally with the type-Judithian Judith River Formation, even though it has greater affinities to the type-Aquilan fauna (Eaton and Cifelli, 1988; Eaton, 2002; Kielan-Jaworowska *et al*., 2004). Several mammal taxa used to date the Wahweap Formation as Aquilan (*Cimolodon electus, C. similis* and *Cimexomys* cf. *antiquus*; Eaton, 2002) have also been recovered from other unambiguous Judithian aged deposits in the Western Interior Basin (the Kaiparowits, Fruitland/Kirtland and Foremost/Oldman formations respectively (see Table 2.25 in Kielan-Jaworowska *et al*., 2004). Thus there are difficulties in assigning the Wahweap Formation to either the Aquilan or the Judithian NALMA, suggesting that it may be appropriate to redefine the indicator taxa for one or both “ages”, or the NALMA boundary, or possibly construct a new NALMA between the Aquilan and Judithian. Kirkland and DeBlieux (2010), who have studied a suite of vertebrate faunas from the Wahweap Formation have suggested that a new Wahweapian age be constructed between the Aquilan and Judithian. However, the construction of a new

NALMA based on a fauna from the southern part of the WIB is risky due to faunal variations with latitude (Rowe *et al*., 1992; Lehman *et al*., 2006) and can therefore only be considered after further investigation of faunas across the WIB.

3.9. Conclusions

Radiometric dating reveals that the Wahweap Formation is slightly younger than was predicted by biostratigraphy. Deposition took place between approximately 81 and 76 Ma, which suggests that revision of the Aquilan age assessment is required. The creation of a new Wahweapian age assignment is possible but further investigation is required to ascertain if the Wahweap Formation's unique fauna is due to temporal or geographic distinctiveness.

4. Sequence Stratigraphy

4.1. Introduction

A wealth of stratigraphic and sedimentological studies on the Cretaceous Western Interior Basin and elsewhere has yielded important data and new models for unravelling and understanding the concepts and intricacies of sequence stratigraphy in foreland basins (Shanley and McCabe 1995; Van Wagoner 1995; Rogers 1998; Martinsen *et al*. 1999; Willis 2000; Miall and Arush 2001; Adams and Bhattacharya 2005; McLaurin and Steel 2007). Much of our present understanding of sea level cyclicity in the Western Interior Basin is based on the study of open marine and shoreline sequences from Alberta to Texas, but focused in Montana, Wyoming, Colorado and Kansas (Weimer 1960; Kauffman 1969, 1977; Gill and Cobban 1973; Kauffman and Caldwell 1993). These studies led to the recognition of four major transgressive-regressive events in the Western Interior Seaway during the Late Cretaceous (Kauffman 1969, 1977; Weimer 1986). Successions deposited during these events, from oldest to youngest, are referred to as the Greenhorn, Niobrara, Claggett and Bearpaw cyclothems. More recently, numerous workers have focused on the alluvial response to base level change in an attempt to place marginal marine to fully continental deposits into sequence-based frameworks. These studies have typically focused on the relationship between base level fluctuation and changes in fluvial stacking patterns, accommodation space and the identification of sequence boundaries and other diagnostic stratigraphic surfaces (e.g., transgressive surfaces, maximum flooding surfaces) (Shanley and McCabe 1991, 1995; Wright and Marriott 1993; Olsen *et al*. 1995; Van Wagoner 1995; Little 1997; Rogers 1998

Martinsen *et al*. 1999; McLaurin and Steel 2000, 2007; Willis 2000; Miall and Arush 2001; Lawton *et al*. 2003; Atchley *et al*. 2004; Ambrose and Ayers 2007 and many others).

Changes in base level that are large enough to affect fluvial systems have two major controls: eustasy and tectonism. The identification of eustatic control involves tying a sedimentary succession to changes in sea level of known age, whereas tectonic control is signified by localised change that typically varies across structural boundaries (Heller *et al*. 1988; Shanley and McCabe 1995; Lawton *et al*. 2003; Horton *et al*. 2004). The interplay between the two controlling mechanisms is complex (Mancini and Tew 1993; Dennison 1994; Emery and Myers 1996) and poor age constraints and stratal correlation commonly complicate studies of non-marine sequence stratigraphy.

Additionally, other factors such as changes in climate and sediment supply during deposition may cause an overprint that further complicates interpretation of changes in facies architecture (Blum and Törnqvist 2000; Feldman *et al*. 2005). Climate controls factors such as run-off in river systems, rate of sediment supply and vegetation density, which in turn affect river morphology (Blum and Törnqvist 2000; Weissmann *et al*. 2002; Feldman *et al*. 2005).

The identification of non-marine sequence boundaries is somewhat contentious. Originally it was thought that non-marine sequence boundaries could be identified by a transition in fluvial architecture, from isolated, single-storey channels with abundant preservation of overbank mudrock (high accommodation space deposits) of the transgressive system below the sequence boundary, to amalgamated multi-storey sandstone dominated units (low accommodation space deposits) of the lowstand system above the sequence boundary (Shanley and McCabe 1995). It has also been considered that a change in fluvial style from meandering to braided river systems occurs across the sequence boundary, although work by Adams and Bhattacharya (2005) has demonstrated that this is not always the case. Studies of the Straight Cliffs Formation in southern Utah have also shown that the latest highstand deposits directly below the sequence boundary are deposited at a time when the generation of accommodation space due to transgression has ceased and therefore an increasing (but still minor) degree of amalgamation is expected (Shanley and McCabe 1995). In this context, amalgamated sandstone units are predicted to occur both above (e.g. Shanley and McCabe 1995; Miall and Arush 2001; Lawton *et al*. 2003; Adams and Bhattacharya 2005) and, less commonly, below (e.g. Little 1997; Lawton and Christensen 2005) sequence boundaries, depending on whether they are interpreted as progradational systems which fill remnant accommodation space after maximum transgression or else lowstand deposits formed under low accommodation space triggered by a drop in base level. This necessitates a reliance on other data, such as palaeocurrent orientations, sandstone petrography and provenance, which are now considered critical tools for identifying sequence boundaries in fully alluvial systems (Fitzsimmons and Johnson 2000; Lawton *et al*. 2003). Additionally, evidence of a major basal disconformity is considered to be a key feature for a continental sequence (e.g. Roca and Nadon 2007).

4.2. Tectonism and eustasy in the Campanian Western Interior Basin: Claggett and Bearpaw cyclothems

Previous work in the Campanian part of the northern Western Interior Basin has identified an extensive 80 Ma sequence boundary in the Pierre Shale of Wyoming, Montana and the Dakotas (Van Wagoner *et al*. 1990), as well as in updip alluvial strata of the Two Medicine Formation in western Montana (Rogers 1998) that preserve well-dated bentonite horizons that have been used to date the maximum transgression of the Claggett sea at 79.6 Ma (Rogers *et al*. 1993). In the southern Western Interior Basin, the La Ventana Tongue of the Cliff House Sandstone in the San Juan Basin (New Mexico) is the southern-most occurrence of strata deposited during the Claggett transgression (Palmer and Scott 1984). The great extent of Claggett-associated deposits across the Western Interior Basin, together with the Claggett transgression occurring at a period of elevated global sea level (Haq *et al*. 1987; Ogg and Ogg 2008; Fig. 4.1.), have led to its recognition as a eustatic event with significant influence on fluvial systems on the western margin of the Western Interior Seaway (Fouch *et al*. 1983; Palmer and Scott 1984; Rogers *et al*. 1993).

Alternatively, Molenaar and Rice (1988) and Kristinik and DeJarnett (1995) have proposed that tectonism during the Campanian overprinted any eustatic effect on fluvial systems on the western margin of the Western Interior Seaway, and have regarded contemporaneous Campanian deposits as recording tectonically driven base level shifts. The absence of extensive sequence boundaries that can be traced along strike, together with the poor correlation of bounding surfaces and stacking patterns in synchronous deposits are cited as evidence of tectonically driven control of fluvial systems.

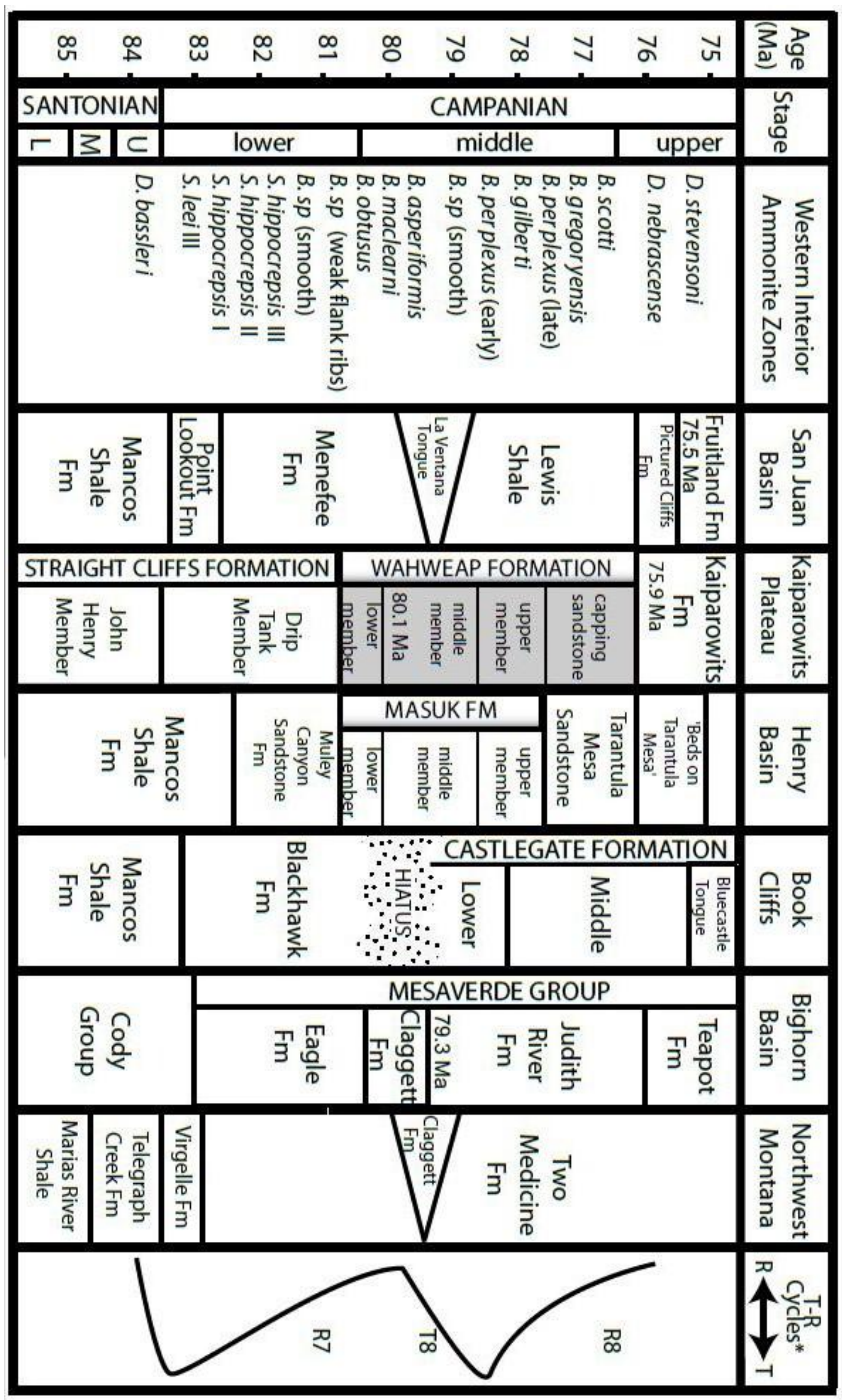


Figure 4.1. Correlation diagram of the Wahweap Formation with strata from the San Juan Basin (New Mexico), Henry Basin (southeastern Utah), Book Cliffs (central Utah), Bighorn Basin (Wyoming), and northwest Montana. Timescale and transgressive-regressive cycles are after Gradstein et al. (2004) as updated by Ogg and Ogg (2008). Correlation is after Jinnah et al. (2009a). R7, T8, and R8 refer to transgressive-regressive cycles of Kauffman and Caldwell (1993).

4.3. Previous sequence stratigraphic work

Aside from detailed biostratigraphic work (Eaton 1991, 2002) and sedimentological studies focusing on the capping sandstone member (Pollock, 1999; Simpson *et al*. 2008; Hilbert-Wolf *et al*. 2009; W. Simpson *et al*. 2010), most other studies relating to the Wahweap Formation have been broadly focused on the overall sequence stratigraphy of Upper Cretaceous strata in the Kaiparowits Plateau. Shanley and McCabe (1991, 1995) identified a sequence boundary at the base of the Drip Tank Member of the Straight Cliffs Formation, although they did not include the Wahweap Formation in their analysis. According to their work, the Drip Tank Member, an amalgamated sheet sandstone, represents a low-stand progradational sheet. In the sequence stratigraphic model of Little (1997), which includes the Straight Cliffs, Wahweap, Kaiparowits and Canaan Peak formations, sequence boundaries are placed above, rather than below amalgamated sheet sandstones. As such, Little (1997) placed sequence boundaries at the Drip Tank-Wahweap contact and at the Wahweap-Kaiparowits contact (i.e. at the top of the Drip Tank and capping sandstone members). According to Little (1997), sequences are tectonically controlled and initiated by subsidence which causes high accommodation space and the deposition of laterally restricted fine-grained sandstone interbedded with mudrock at the base of the sequence. Tectonic quiescence, together with continuous sedimentation after the initial tectonic subsidence leads to fill of accommodation space and eventually results in the deposition of amalgamated sheet sandstones at the top of tectonically driven sequences in Little's model. The next phase of tectonic subsidence then initiates deposition of the overlying sequence.

Lawton *et al*. (2003) also placed the Wahweap Formation into a sequence stratigraphic framework, more similar in principle to that proposed by Shanley and McCabe (1995) for the Straight Cliffs Formation. In this instance, the sequence boundary is placed below the capping sandstone, at the upper member-capping sandstone contact within the Wahweap Formation. The change in accommodation space in this case is the result of a tectonically driven drop in base level at the time that the capping sandstone was being deposited. As evidence for this, Lawton *et al*. (2003) cited a change in sandstone composition, and thus source area, from the Mogollan highlands and Cordilleran magmatic arc in the south and southwest for the lower three members of the Wahweap Formation, to the Sevier belt in the west and northwest for the capping sandstone. They were also the first to recognise a shift in palaeocurrent indicators from northward trending in the lower three members to eastward trending in the capping sandstone. Later work on detrital zircons and source areas by Larsen (2007), Dickinson and Gehrels (2008) and Jinnah *et al*. (2009a) confirmed that the lower three members had a southerly source while the capping sandstone had a different easterly source. The shift in source area is best explained by tectonic rejuvenation in the Sevier belt, accompanied by tectonic base level drop and the initiation of a new sequence with the capping sandstone at its base. Additionally, thinning of the capping sandstone from the Markagunt Plateau to the Kaiparowits Plateau has been used to infer normal movement along the East Kaibab monocline during the Campanian, resulting in basin partitioning and greater accommodation space in the Markagunt and Paunsaugunt regions, as well as a transition from a foredeep to a wedgetop depocentre as the capping sandstone was being deposited (Lawton *et al*. 2003).

4.4. Aims

In this chapter, data presented in the previous two chapters on member contacts, shifts in alluvial architecture, palaeocurrent, sandstone composition and member age constraints is used to refine the sequence stratigraphic framework of the Wahweap Formation. This is combined with palaeoenvironmental data from the down-dip Masuk and Tarantula Mesa formations to elucidate allocyclic controls during deposition, and place the Wahweap Formation into the context of Western Interior Seaway cyclothems.

4.5. Changes in base level

Through examination of the lower and upper contacts of the capping sandstone, placement of the sequence boundary, which has previously been indicated to lie at both the base and top of the unit (Little, 1997; Lawton *et al*. 2003) can be determined. In this study, the presence of erosional relief at the base of the capping sandstone, together with the conformable and in some areas gradational upper contact with the Kaiparowits Formation, both refute the notion that a sequence boundary exists at the top of the capping sandstone and support the alternative scenario that places a sequence boundary at the base of the capping sandstone member. A more rigorous test and confirmation of this hypothesis comes from analysis of the down-dip correlative Tarantula Mesa Sandstone in the Henry Mountains Basin, to the east of the Kaiparowits Plateau (Peterson *et al*. 1980; Eaton 1990). The Tarantula Mesa Formation was coevally deposited in a fully continental environment, yet overlies a thick succession of marginal-marine and tidally influenced fluvial deposits. The preservation of fully alluvial correlative strata in the down-dip direction (shoreward) strongly suggests that the capping sandstone member is not synchronous with base-level highstand, as suggested by the Little (1997) model, but rather that it was deposited immediately following base level fall. Thus, the capping sandstone records a period of fluvial incision, basinward shoreline movement and a low accommodation space facies tract dominated by highly amalgamated fluvial channel systems and intense cannibalization of overbank deposits.

The model of Lawton *et al*. (2003) places sequence boundaries at the base of the Drip Tank and capping sandstone, based on their amalgamated nature. The Drip Tank sequence boundary, across which there is no change in flow direction or sandstone composition, is regarded as eustatic in origin, while the capping sandstone sequence boundary, which records a shift in flow direction, sandstone composition and source area (Lawton *et al*. 2003; Jinnah *et al*. 2009a) is inferred to originate from tectonic activity.

New radiometric age constraint places deposition of the Wahweap Formation between 80.6 and 76.1 Ma (Jinnah *et al*. 2009a) and allows the lower three members to be better understood within a sequence stratigraphic context, and a clearer distinction between tectonic and eustatic control to be drawn. For instance, the Drip Tank Member of the Straight Cliffs Formation is younger than previously thought and was deposited in the Early Campanian (Eaton 2006; Jinnah *et al*. 2009a). Additionally, Lawton and Christensen (2005) place the Drip Tank sequence boundary at the top of the amalgamated sandstone unit rather than the base. This suggests that the Drip Tank sequence boundary is actually correlative with the eustatic ~80 Ma sequence boundary that has been recognised across the Western Interior Basin (Van Wagoner *et al*. 1990; Rogers 1998) and in global sea level curves (Haq *et al*. 1987). The development of this sequence boundary was followed by the Claggett transgression, with a highstand at 79.6 Ma in the northern portion of the Western Interior Basin (Rogers *et al*. 1993). In the Wahweap Formation, the age of the middle-upper member contact is constrained by radiometric age dates of 80.1 Ma in the lower middle member and 75.46 Ma in the basal part of the overlying Kaiparowits Formation (Roberts *et al*. 2005; Jinnah *et al*. 2009a). Thus, it is suggested that the middle-upper member contact in the Wahweap Formation correlates with maximum transgression of the Claggett cyclothem. Evidence supporting this hypothesis includes the presence of tidal sedimentary structures and overall fining and waterlogged conditions that developed above this boundary in the middle member as a result of rapid generation of accommodation space in the basin.

The uppermost part of the Drip Tank Sandstone and the lower member of the Wahweap Formation therefore fall into the amalgamated fluvial facies tract (Shanley and McCabe, 1995), while the middle member, which represents an increase in accommodation space, falls into the isolated fluvial facies tract. The top of the middle member, together with the entire upper member fall into the highstand facies tract, where there is evidence of tidal influence and marine incursion in the lower part, followed by an increase in amalgamation caused by reduced accommodation space due to prograding fluvial systems in the upper part (Fig. 4.2.). Therefore, the uppermost part of the Drip Tank sandstone, together with the lower, middle and upper members of the Wahweap Formation, form a complete non-marine sequence underlain and overlain by sequence boundaries.

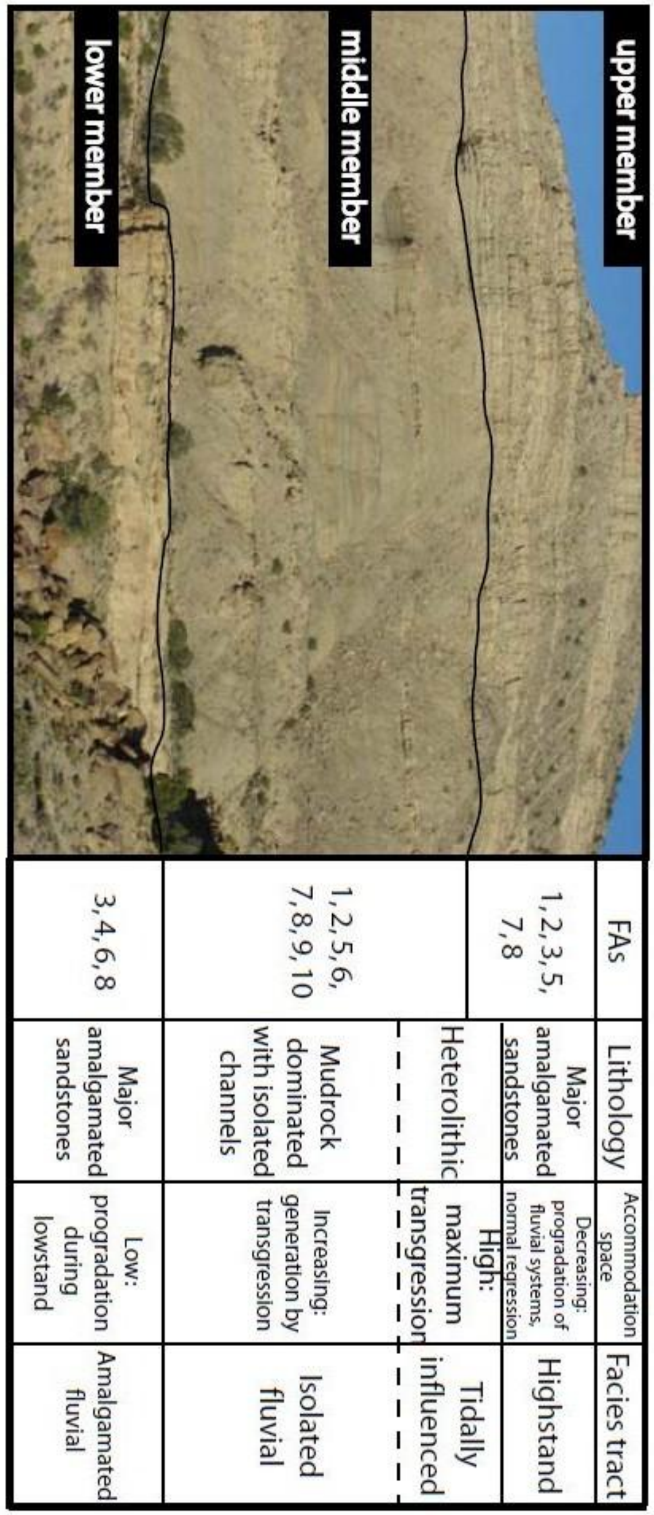


Figure 4.2. Photograph of the lower, middle, and upper members of the Wahweap Formation showing interpreted sequence stratigraphic context

4.6. Base level controls

One major consideration in alluvial sequence stratigraphy is the role of climate in alluvial architecture, including stacking patterns, sand:mud ratio and sediment load. Climatic control is an extremely challenging factor to account for (Blum and Törnqvist 2000; Feldman *et al*. 2005), but its relative influence can be largely ruled out or at least understood if one can demonstrate that climate does not change or vary significantly across a sequence boundary. This is the case in the central Western Interior Basin during deposition of the Wahweap and Kaiparowits formations in the middle-late Campanian. Detailed facies analysis presented here for the Wahweap Formation and by Roberts (2007) for the overlying Kaiparowits Formation indicate very similar climatic and palaeoenvironmental regimes between the two units. If this is accurate, then variations in sediment supply should have been relatively constant during this time, and thus, changes in alluvial architecture that we see taking place through the Wahweap Formation are most parsimoniously explained in terms of eustatic and tectonic factors, with climate playing a very limited role.

This assumption permits the interpretation that eustasy is the primary control on the Drip Tank sequence boundary, and that the rapid upsection decrease in sand:mud ratio in the middle member, followed by slowing and eventual increase in sand:mud ratio in the upper member are all eustatically controlled features. Therefore, the sequence that includes the Drip Tank Sandstone and the lower, middle and upper members of the Wahweap Formation formed with eustasy as the dominant allocyclic control.

In contrast, the overlying sequence, which has the capping sandstone at its base, displays a number of features which indicate that tectonic activity was the dominant allocyclic control. Evidence for this includes the abrupt change in palaeocurrent direction, source area, sandstone composition and palaeoenvironment (Lawton *et al*. 2003; Jinnah *et al*. 2009a) above the upper member – capping sandstone contact. These features are actually indicative of a reduction in accommodation space resultant from a tectonically-driven base level change. Further evidence for tectonic rather than eustatic or climatic control comes from the poor correlation of the capping sandstone sequence boundary with contemporaneous deposits of the Book Cliffs in central Utah, where tectonism is the major driving force (Olsen *et al*. 1995; Miall and Arush 2001; Adams and Bhattacharya 2005). Although the capping sandstone and the middle part of the Castlegate Sandstone were deposited at the same time from a source area in the Sevier belt, sequence boundaries between these two successions do not correlate, suggesting that both were under tectonic control. Additionally, evidence of tectonism in the Sevier thrust belt during the mid-Campanian (Horton *et al*. 2004; DeCelles and Coogan 2006) further supports tectonically-driven base level change in the capping sandstone.

Overall, a combination of factors thus suggests that eustasy was the driving allocyclic control in the sequence that includes the lower, middle and upper members of the Wahweap Formation, and a shift to tectonic control occurred at the initiation of capping sandstone deposition.

4.7. Conclusions

The Drip Tank Sandstone and lower, middle and upper members of the Wahweap Formation constitute a 3rd-order stratigraphic sequence in which eustasy was the dominant allocyclic control. By contrast, the capping sandstone and Kaiparowits Formation constitute a stratigraphic sequence in which tectonism was the dominant control. Analysis of the alluvial architecture in the Wahweap Formation supports the non-marine sequence stratigraphic model of Shanley and McCabe (1995) and Lawton *et al*. (2003) in which amalgamated fluvial, isolated fluvial, tidally influenced and highstand progradation systems tracts are defined. This analysis also shows that there was eustatic influence in the middle part of the Western Interior Basin during the Campanian.

5. Fluvial Architecture on the Markagunt and Paunsaugunt Plateaus and Basin Development

5.1. Introduction

Dickinson (1974) used the term 'retroarc basin' to denote continental lowlands and epicontinental seas that develop as linear features which extend parallel to the boundaries between cratons in the hinterland and magmatic arcs and thrust belts in the foreland. Since then, retroarc foreland systems have been identified as elongate depressions that develop on continental crust between thrust belts and cratons in response to compressional events related to subduction and orogenesis (Allen *et al*., 1986; Ingersoll, 1988). Foreland systems develop where thrust loads are isostatically compensated for by flexural subsidence resulting in basin development. Basin fill can be several kilometres thick (Dickinson, 1974; Jordan, 1981) and continental, marginal marine or fully marine depositional settings can develop. DeCelles and Giles (1996) divided foreland systems into four depozones: the wedgetop, foredeep, forebulge and backbulge depozones (Fig. 5.1.), based on differing accommodation space caused by variable flexural subsidence.

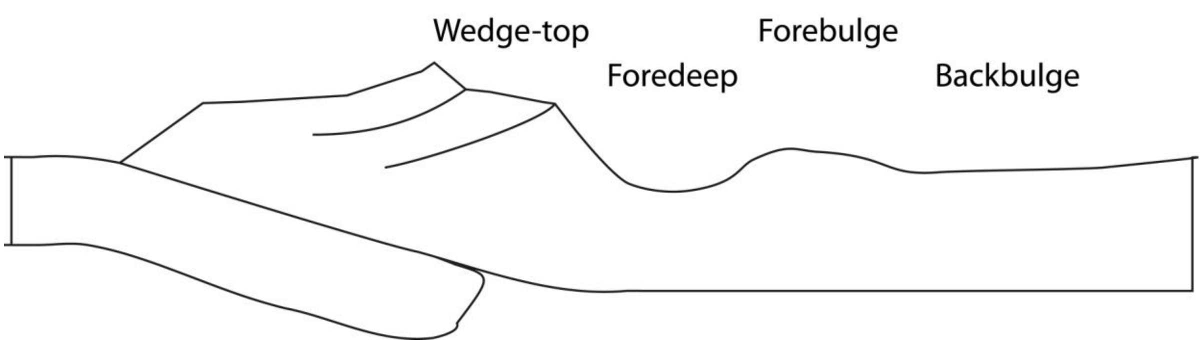


Figure 5.1. Diagram showing depozones of a foreland basin (after DeCelles and Giles, 1996).

The wedgetop depozone is a fault-bounded region at the front of the orogenic belt, often characterised by the presence of piggy-back basins (Ori and Friend, 1984) and coarse-grained clastic wedges (DeCelles and Giles, 1996). Sediment deposited in a wedge-top setting is typically deformed soon after deposition before deep burial (Flemings and Jordan, 1990; Lawton *et al.*, 1993; DeCelles and Giles, 1996).

The foredeep depozone (or the foreland basin) is the zone of greatest crustal subsidence, and sediment deposited within it can be sourced either from the thrust belt or craton by transverse flow, or from uplifts along strike in the foreland basin by longitudinal flow (Cowan, 1993; DeCelles and Giles, 1996; Lawton *et al*, 2003).

The forebulge is a zone of positive relief that forms in flexural response to subsidence in the foredeep (Flemings and Jordan, 1990; DeCelles and Giles, 1996). It is typically expected to be a zone of non-deposition represented by an unconformity in the stratigraphic record (Tankard, 1986; Flemings and Jordan, 1990). It may also be the site of section condensation or carbonate platform development in marine systems (DeCelles and Giles, 1996). Development of a forebulge zone in the underfilled stage of a foreland system depends on the relationship between flexural uplift and dynamic subsidence. Where flexural uplift in the forebulge region is greater than dynamic subsidence, unconformities can be expected, whereas if dynamic subsidence is greater than flexural uplift, condensed sections are expected (Catuneanu, 2004; Catuneanu, 2006).

The backbulge lies between the forebulge and the craton and is typically shallower than the foredeep (Plint *et al.*, 1993). Sediment accumulations in the backbulge are typically distal from their source area (the orogenic belt) and thus fine-grained (DeCelles and Giles, 1996).

Sediment accumulation within foreland systems is controlled by a number of factors including tectonic events in the thrust belt, eustatic sea level changes where marine conditions have developed or the basin is open to the sea (Dennison, 1994; Catuneanu, 2004), as well as climate, thrust belt lithologies and the style of subduction. Thus numerous workers (e.g. Ori and Friend, 1984; Cant and Stockmal, 1989; Yingling and Heller, 1992; Robinson and Slingerland, 1998; Houston *et al*., 2000; Liu *et al.*, 2005) have studied foreland basin fill with the aim of reconstructing basin history.

5.2. Aims

In this section, stratigraphic thicknesses of the Wahweap Formation in different sub-basins are used to re-evaluate the current basin development model, in which the capping sandstone is presumed to have been deposited in a wedgetop depocentre after basin partitioning occurred between the Markagunt and Paunsaugunt plateaus, and the Kaiparowits Plateau. Fluvial architecture and unit continuity across structural divides are used to interpret accommodation space and depositional setting in the foreland system.

5.3. The Wahweap Formation west of the East Kaibab monocline

On the Markagunt Plateau along highway 89, the thickness of the lower, mudrock-rich part of the Wahweap has been measured at approximately 165 m by Tilton (1991). The sequence is dominated by red, yellow and grey mudrock, in which non-pedogenic and pedogenic units can be distinguished. Non-pedogenic mudrock has a colour of 5PB 5/1 blueish grey or 10Y 8/1 blueish white and may have a massive sedimentary texture or else contain weakly developed planar bedding. Pedogenic mudrock has a variable colour (e.g. 5RP 4/2 greyish purple, 5R 4/4 weak red, 10YR 4/2 dark yellowish orange, 5R 8/8 red) and contains slickenside structures, root mottles (5Y 6/6 yellow) and horizontal stratification. Non-pedogenic and pedogenic mudrock typically occur cyclically through the lower part of the Wahweap Formation, with a non-pedogenic unit between 0.4 m and 2 m thick at the base of the cycle and a pedogenic unit of between 1 m and 2 m at the top. The cyclical interbedded nature of non-pedogenic and pedogenic mudrock therefore suggests that floodplain sequences on the Markagunt Plateau are cumulative palaeosols.

Sandstone bodies have a lenticular geometry with erosive bases and sharp or gradational upper contacts. They lack internal erosive surfaces and can thus be classified as single-storey, and are typically less than 2 m thick and up to 20 m wide. Trough cross-stratification (TCS), ripple lamination and planar bedding all occur within channel sandstones. The largest TCS foresets are approximately 0.6 m thick. Although lateral accretion macroform structures are not obvious, the overall fluvial architecture is most consistent with low-gradient channel systems with stable banks (i.e., meandering to anastomosing style streams). On the Markagunt and Paunsaugunt plateaus, there are no heterolithic stratification sets, or any other sedimentary structures indicative of tidal influence within sandstone bodies.

The capping sandstone unit on the Markagunt Plateau is approximately 70 m thick (Pollock, 1999). The fluvial architecture within the capping sandstone is more amalgamated than in the lower part of the Wahweap Formation, and there is minimal preservation of overbank fines. Gravel bedforms and major and minor tabular sandstone bodies form the bulk of the capping sandstone. Intraformational and extraformational conglomerates lenses occur throughout. Overall, these features indicate a high-gradient, most likely braided fluvial system.

The Wahweap Formation on the Paunsaugunt Plateau is similar to that on the Markagunt Plateau. The lower part has a thickness of approximately 100 m and consists of a mudrock-dominated succession with a high degree of pedogenesis, including slickensides, root mottling and horizonization. Thin sandstone bodies occur throughout the lower part of the succession, characterised by both lenticular and sheet-like geometries. Like the Markagunt Plateau, sandstone bodies on the Paunsaugunt Plateau lack internal erosive surfaces (although their basal surfaces may be erosive). Upper surfaces of sandstone bodies are sharp or gradational.

On the Paunsaugunt Plateau, outcrop of the capping sandstone is poor: in many places the entire unit has been eroded and the Claron Formation unconformably overlies the lower part of the Wahweap Formation. Where the capping sandstone is preserved, it has been interpreted as palaeovalley fill to account for its inconsistent thickness (Pollock, 1999).

Between the eastern edge of the Paunsaugunt Plateau at the Paunsaugunt fault and the western edge of the Kaiparowits Plateau at the East Kaibab-monocline, there is a moderate amount of exposure of the lower, mudrock-dominated part of the Wahweap Formation along several hills. These outcrops (best observed at Bristlecone Point) constitute pedogenic mudrock and siltstones that closely resemble the lower Wahweap Formation on the Paunsaugunt Plateau. In this locality the lower part of the Wahweap Formation is overlain unconformably by the Claron Formation.

5.4. The Wahweap Formation on the Kaiparowits Plateau

Compared to the relatively thin exposures on the Markagunt and Paunsaugunt plateaus, the Wahweap Formation attains its maximum thickness of ~400 m on the Kaiparowits Plateau. Several well-developed, laterally extensive amalgamated sandstone bodies occur on the Kaiparowits Plateau and form the basis for subdividing the Wahweap Formation into the lower, middle, upper and capping sandstone members. Additionally, the Wahweap Formation on the Kaiparowits Plateau is more sand-rich close to the East Kaibab-monocline in the west, and more mud-rich towards the east. In contrast to the Markagunt and Paunsaugunt plateaus, vertical stacking patterns of a number of stratigraphic sections on the Kaiparowits Plateau show distinct cyclicity in alluvial architecture. For example, laterally extensive amalgamated sandstone bodies occur in the lower, upper and capping sandstone members whereas aggradational mudrock occurs in the lower part of the lower member and the entire middle member. Mudrock units typically display features indicative of subaqueous deposition at their base and there is a transition to features more consistent with subaerial exposure and incipient pedogenesis towards the top. Minor occurrence of calcrete nodules in mudrocks (and both calcrete and mudrock fragments as rip-up clasts in intraformational conglomerates) indicate that the environment of deposition was seasonally arid. Sandstone bodies (both lenticular and sheet geometry) are either single- or multi-storey and are typically larger than on the Markagunt or Paunsaugunt plateaus. On the Kaiparowits Plateau large sandstone bodies may extend for tens to hundreds of meters (or more) and be tens of meters thick. Several facies associations indicative of tidal influence have been documented in the middle and upper members.

The lower three members of the Wahweap Formation on the Kaiparowits Plateau attain a thickness of between 200 m in the north and 280 m in the south. The capping sandstone is at its thickest to the north of the Kaiparowits Plateau, where it attains a thickness of more than 140 m. Across most of the southern part of the Kaiparowits Plateau the capping sandstone is undergoing modern-day erosion, but where the complete unit is seen to occur, its thickness is found to be between 60 m (close to Death Ridge) and 90 m (at Star Seep on Smoky Mountain Road; Fig. 5.2.).

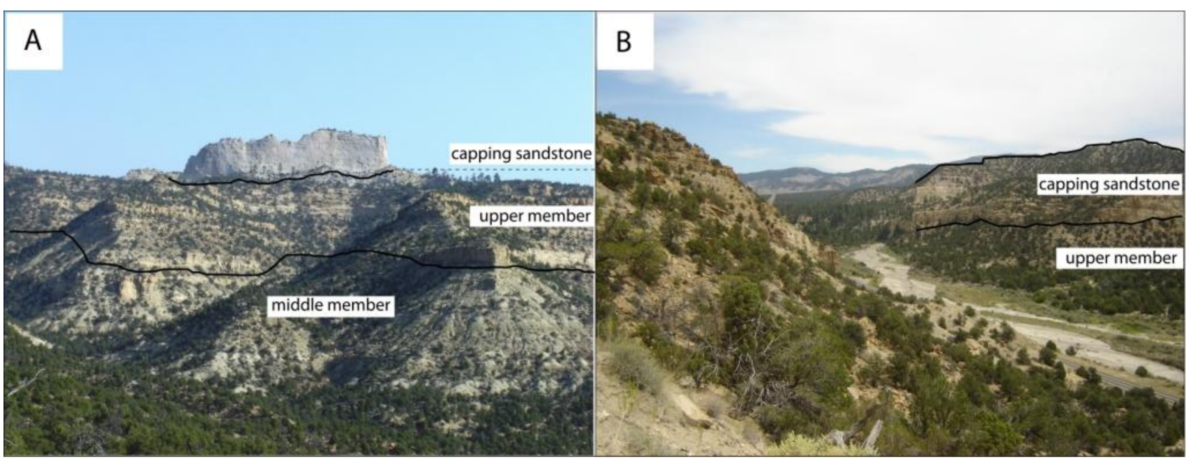


Figure 5.2. Stratigraphy of the Wahweap Formation on the Kaiparowits Plateau. A: Star Seep, southern Kaiparowits Plateau. B: Upper Valley along Utah highway 12, northern Kaiparowits Plateau

5.5. The Masuk and Tarantula Mesa formations in the Henry Mountains region

The Masuk Formation in the Henry Mountains region is regarded as equivalent to the lower part of the Wahweap Formation, whereas the Tarantula Mesa Formation is equivalent to the capping sandstone (Eaton, 1990). Like the lower part of the Wahweap on the Kaiparowits Plateau, the Masuk Formation is divided into a lower, middle and upper member with virtually identical distinguishing characteristics. Together, the three members of the Masuk Formation attain a thickness of 194 m. Overall, the Masuk Formation contains more mudrock than the Wahweap Formation and in this succession mudrock typically possesses well-developed laminations and a drab colour. There is no evidence of subaerial exposure, pedogenesis or seasonal aridity. Additionally, unlike the Wahweap Formation, there is a dominance of lenticular over tabular sandstone body geometry in the sandstone-rich lower and upper members. Sandstone bodies vary from meters to hundreds of meters in lateral extent. The Tarantula Mesa Formation, which is 100 m to 125 m thick as measured by Smith (1983), is similar to the capping sandstone member in both architecture and thickness, but the capping sandstone generally contains beds of both intraformational and extraformational conglomerate which are not well developed in the Tarantula Mesa Formation.

5.6. Environmental variations based on fluvial architecture

Distinct differences in sedimentology are recognised between the lower part of the Wahweap Formation on the Markagunt and Paunsaugunt plateaus and the Kaiparowits Plateau and Henry Mountains. The rivers which deposited the lower part of the Wahweap Formation on the Markagunt and Paunsaugunt plateaus had small channels (as evidenced by the size of lenticular sandstones) with confined flow.

There was little change in fluvial conditions throughout deposition of the succession on the Markagunt and Paunsaugunt plateaus, resulting in a fairly uniform lithostratigraphy. The lack of such features as heterolithic stratification and brackish water trace fossils suggest a fully fluvial environment.

On the Kaiparowits Plateau, larger channels with more interconnectivity developed, resulting in the deposition of amalgamated sheet sandstones. Overbank deposition occurred abundantly, especially in the lower and middle members. Heterolithic stratification with regular sand-mud couplets at the top of the middle and base of the upper member record a marine incursion and the development of estuarine conditions in the Kaiparowits Plateau.

In the more distal Masuk Formation, large channels persisted. Floodplain sediment accumulations which bear the remains of aquatic invertebrates and have no evidence of pedogenesis suggest a constantly waterlogged floodplain with large standing bodies of water in a low-lying coastal plain. Although heterolithic stratification is abundant throughout the section, the invertebrate fossils reported by Eaton (1990) from the Masuk Formation suggest a dominantly freshwater setting with periodic incursions of brackish water.

In terms of fossil preservation throughout the stratigraphic section, several trends are notable. On the Markagunt and Paunsaugunt plateaus, fossils are most commonly preserved as individual elements from a variety of taxa within fossiliferous horizons (microsites)(Eaton, 1999). No sites with associated or articulated material are known within the Wahweap

Formation in these areas. On the Kaiparowits Plateau, where the Wahweap Formation is very well exposed and has been more extensively prospected, fossil microsites which yield a higher number of taxa represented by isolated or fragmentary remains are most common in the sandstone-rich lower and upper members (Cifelli, 1990; Eaton, 1991). Sites with associated or articulated remains are much less common than microsites (DeBlieux *et al*. 2009); the mudrock-rich middle member is more likely to contain such sites. Jinnah *et al*. (2009b) document a site with associated hadrosaur remains from the middle member.

These trends in fossil preservation are related to the environment of deposition and ultimately controlled by the accommodation space available at time of deposition. Large amalgamated channel bodies of the lower and upper members typically contain much material which has been extensively reworked, resulting in preservation of isolated elements, whereas rapid floodplain sedimentation in the middle member (and parts of the lower member) was more conducive to preserving associated fossil material.

5.7. Interpretations of accommodation space

Interpretations of accommodation space in the Markagunt, Paunsaugunt, Kaiparowits and Henry Mountains basins are made based on alluvial architecture and total stratigraphic thickness (Table 5.1.). On the Markagunt and Paunsaugunt plateaus, the isolated nature of channels in the lower part of the Wahweap Formation, and their entombment in thick pedogenic mudrock sequences, coupled with their lack of significant internal erosional discontinuities suggests that deposition took place in a relatively high accommodation space setting. Cumulative palaeosols, such as those described here from the Markagunt and

Table 5.1. Summary of thickness (T) and accommodation space (AS) conditions across the extent of the Wahweap, Masuk and Tarantula Mesa formations. *Measured by Tilton (1991). **Measured by Smith (1983).

Markagunt Plateau			***Paunsaugunt Plateau***			***Kaiparowits Plateau***			***Henry Mountains***		
Wahweap unit	**T**	**AS**	**Wahweap unit**	**T**	**AS**	**Formation/ Member**	**T**	**AS**	**Formation/Member**	**T**	**AS**
capping sandstone	*70 m**	low	*capping sandstone*	*75 m**	low	*Wahweap/capping sandstone*	*90 m*	low	*Tarantula Mesa Formation*	*125 m***	low
lower part	*260 m*	high	*lower part*	*100 m*	high	*Wahweap/upper*	*280 m*	high, decreasing	*Masuk/upper*	*194 m*	high, decreasing
						Wahweap/middle		high, increasing	*Masuk/middle*		high, increasing
						Wahweap/lower		low, increasing	*Masuk/lower*		low, increasing

Paunsaugunt plateaus, indicate conditions where sedimentation occurs at a constant rate (Kraus, 1999). On the Paunsaugunt Plateau, the preserved thickness of the succession is only 100 m, but the alluvial architecture suggests a high accommodation space setting. Since the lower part of the Wahweap Formation is mostly overlain unconformably by the Palaeocene-Eocene Claron Formation in this area, the thickness is likely accounted for by erosion between the Cretaceous and the Palaeocene. Santucci and Kirkland (2010) suggest that this erosion is related to uplift along the Paunsaugunt fault during the latest-Cretaceous to Early Tertiary. The capping sandstone, which contains numerous erosive surfaces, intraformational conglomerate lenses and very little preserved overbank material, was deposited under conditions of much less accommodation space than the lower part of the Wahweap Formation.

On the Kaiparowits Plateau, accommodation space varied during deposition of the lower, middle and upper members. Deposition of the lower member took place in a low accommodation space setting, and the middle member was deposited in a high accommodation space setting. The basal part of the upper member was deposited under high accommodation space settings, although accommodation space gradually declined as the member was deposited and the amalgamated sandstones at the top of the upper member suggest that deposition at that time occurred in a low accommodation space setting. Alluvial architecture of the capping sandstone on the Kaiparowits Plateau is mostly similar to that on the Markagunt and Paunsaugunt plateaus, and indicates low accommodation space.

In the Henry Mountains, the lower, middle and upper members of the Masuk Formation closely mirror the Wahweap Formation in terms of the position of amalgamated sandstone units and these members were deposited under similar accommodation space settings to their equivalents on the Kaiparowits Plateau.

The Tarantula Mesa Formation, like the capping sandstone, has numerous erosive surfaces and a lack of overbank material, and is similarly interpreted to have been deposited in a low accommodation space setting.

5.8. Sequence controls

There is a change in fluvial architecture and sandstone:mudstone ratio in the lower part of the Wahweap Formation across the East Kaibab monocline/ Paunsaugunt fault. To the west, mudrock-dominated strata with a high degree of pedogenesis and a mostly uniform vertical stacking pattern occur, whereas to the east, the Wahweap and Masuk formations display an overall upward-coarsening pattern. On the Markagunt and Paunsaugunt plateaus the dominant architectural style of isolated lenticular sandstone bodies encased in mudrock suggests that high-accommodation space conditions prevailed throughout deposition whereas on the Kaiparowits Plateau and in the Masuk Formation the variation in fluvial architecture suggests that there was fluctuation between high- and low-accommodation space in these regions. The poor correlation of high- and low- accommodation space deposits across the East Kaibab monocline suggests that different mechanisms were responsible for generating and influencing accommodation space in these discrete regions during deposition of the lower part of the Wahweap Formation. Because the western deposits were proximal to the Sevier fold-and-thrust belt and far from the contemporaneous shoreline it is most likely that tectonic forces were the controlling mechanism for accommodation space generation in this part of the basin.

For the deposits of the Kaiparowits Plateau and Henry Mountains, which were proximal to the contemporaneous shoreline, eustatic variations in the Western Interior Seaway would have also had an influence on sedimentation. As such, the Markagunt and Paunsaugunt plateaus host the deposits of proximal river systems (sensu Catuneanu, 2006) whereas the Kaiparowits Plateau and Henry Mountains Basin host distal deposits.

By the time of capping sandstone and Tarantula Mesa deposition, eastward migration of the fold-and-thrust belt, together with a regional drop in base level eliminated eustatic influence as a factor in deposition and alluvial architecture. The deposition of terrestrial strata (Tarantula Mesa Formation) across the former shoreline, with little evidence of tidal influence, in the Henry Mountains Basin suggests that the Western Interior Seaway had retreated significantly by this time.

5.9. Basin development

Within the foreland system, four depocentres, the wedgetop, foredeep (foreland basin), forebulge and backbulge have been identified (DeCelles and Giles, 1996; Catuneanu, 2004). Each of these depocentres is characterised by different patterns of lateral and vertical stratigraphic architecture, as well as varying accommodation space exemplified in the concept of reciprocal stratigraphies, where high accommodation space deposits in the foredeep are contemporaneous with low accommodation space deposits in the forebulge and vice versa (Heller *et al*., 1988; Catuneanu, 2004). The most proximal depocentre, the wedgetop, is characterised by piggyback basins that develop on active thrust sheets (Ori and Friend, 1984), coarse clastic wedges, discontinuous facies and syndepositional deformation of sediments by tectonic activity. The foredeep is typically the zone of maximum subsidence and rapid generation of accommodation space due to crustal flexure as a response to tectonic loading in the adjacent thrust belt. Periods of intense thrust activity therefore drive near concurrent accommodation space generation in the foredeep. In the forebulge, deposition is controlled by the underfilled vs. overfilled nature of the basin and the rate of sediment supply (Burbank *et al*. 1989; Catuneanu, 2004).

Studies of the sedimentary record that focus on identifying different depocentres and the transitions between these depocentres have important implications for understanding foreland basin evolution and controls on the sedimentary record and can also aid in local and regional correlation by explaining proximal-distal facies variations (Ori and Friend, 1984; Lawton and Trexler, 1991; DeCelles, 2004). The alluvial architecture of the lower part of the Wahweap Formation on the Markagunt and Paunsaugunt plateaus suggests that deposition took place in a relatively high accommodation space setting with no discernible changes in facies architecture within this interval. However the maximum preserved thickness of this unit is about 165 m along highway 89, less than the lower, middle and upper members on the Kaiparowits Plateau (280 m), where the alluvial architecture suggests that accommodation space was more variable. It is therefore likely that there was deep incision along the lower Wahweap-capping sandstone contact on the Markagunt and Paunsaugunt plateaus.

In the study of Lawton *et al*. (2003), the capping sandstone is interpreted to form in a wedgetop depocentre. Evidence cited for this includes the drastic change in alluvial architecture (and hence accommodation space), sandstone petrology (from quartzolithic /

subfeldspathic to quartzose; Pollock, 1999) and the inconsistent thickness of the capping sandstone across the Paunsaugunt and Kaiparowits plateaus.

The capping sandstone is overlain unconformably by the Palaeocene-Eocene Claron Formation on the Markagunt and Paunsaugunt plateaus. In many places, the capping sandstone is not present at all and the Claron Formation lies directly on the lower part of the Wahweap Formation. In some places the entire Wahweap Formation is missing and the Claron overlies the Drip Tank Sandstone. On the southern Kaiparowits Plateau, the capping sandstone is undergoing modern day erosion across most of its extent but the total thickness has been measured at between 60 and 90 m. The consistency of both the thickness and the fluvial architecture of the capping sandstone and equivalent Tarantula Mesa Formation across large structural features such as the Sevier and Paunsaugunt faults and the East Kaibab monocline, which separate distinct proximal-distal facies in the lower part of the Wahweap Formation, shows that the capping sandstone forms an extensive conglomeratic sheet that was deposited across the boundaries of several structural basins. The presence of seismites close to the East Kaibab monocline (Hilbert-Wolf *et al*. 2009; Tindall *et al.*, 2010) shows that some syndepositional tectonic activity occurred along this structure, but the similar thickness of the capping sandstone on either side of the structure, together with the remarkable continuity of lithofacies, indicates that the Markagunt, Paunsaugunt and Kaiparowits basins were not fully partitioned at the time of deposition, and syndepositional fault movement did not disrupt sedimentation patterns in the wedgetop.

Alternatively, the continuity of the capping sandstone across the Sevier fault and East Kaibab monocline could indicate that deposition took place in a foredeep rather than wedgetop setting, during a time of low-accommodation space in the foredeep. There has been debate about the exact conditions required for progradation of coarse-grained units into the distal parts of a foreland basin. In the two-phase model of Heller *et al*. (1988), gravel deposits accumulate in the proximal reaches of the foredeep syntectonically, when uplift of thrust sheets and simultaneous subsidence in the foredeep provide a source area and abundant accommodation space respectively. Progradation of coarse-grained sediment into the distal reaches of the foredeep occurs during periods of thrust quiescence when accommodation space in the proximal foredeep has been reduced and transverse drainages are set up. Burbank *et al*. (1989) later recognised that other factors such as rate of sediment supply also affect the distance that coarse-grained units prograde.

The great distance of progradation of coarse-grained detritus in the capping sandstone-Tarantula Mesa system seems to imply that deposition took place during a phase of thrust-belt quiescence and low accommodation space in the foredeep. However, numerous indicators of syndepositional tectonism in the Late Campanian (DeCelles and Coogan, 2006; Hilbert-Wolf *et al.* 2009; Simpson *et al.* 2009) suggest that active thrusting occurred during deposition of the capping sandstone. In the Sevier fold-and-thrust belt in central Utah (close to the source of the capping sandstone) thrusting on either the Paxton or Gunnison thrusts occurred throughout the Campanian (DeCelles and Coogan, 2006). This suggests that coarse-grained detritus prograded into the foredeep system syntectonically.

6. Palaeosol Geochemistry

6.1. Introduction

A number of techniques have been developed which use aspects of palaeosol chemistry and mineralogy as a proxy for the temperature and rainfall conditions at the time of soil formation. Both quantitative and qualitative approaches exist, although palaeosol geochemistry is best used together with other proxies, such as a palaeobotanical approach, where scope for this exists. The qualitative approach involves comparison with modern analogues such as the comprehensive list maintained by Soil Survey Staff (1999) or specific classification schemes focused on palaeosols (Mack *et al*., 1993). Quantitative approaches use indices such as the chemical index of alteration (CIA), gleization, base ratio, or isotope ratios as a proxy for rainfall or temperature.

6.1.1. Qualitative Methods

Qualitative methods for estimating the climatic conditions in which palaeosols formed involve comparing palaeosol properties to a set of predefined qualities in a soil classification scheme. The most comprehensive of these schemes is that of Soil Survey Staff (1999), which defines the following modern soil taxa: alfisols, andisols, aridisols, entisols, gelisols, histosols, inceptisols, mollisols, oxisols, spodosols, ultisols and vertisols. Due to difficulties in applying certain criteria, such as cation exchange capacity, bulk density, moisture content and base status, of modern soils to palaeosols, Mack *et al*. (1993) created a scheme specifically for palaeosols while retaining some of the modern soil nomenclature. They refer to argillisols, calcisols, gypsisols, protosols, histosols, gleysols, oxisols, spodosols and vertisols with slightly modified definitions to facilitate field identification in the rock record.

The clay minerals formed in a soil horizon by pedogenic processes have also been used to infer palaeoprecipitation (e.g. Barshad, 1966; Fig. 6.1.). However, Singer (1980) demonstrated that the use of clay minerals for palaeoclimatic reconstructions is underlain by numerous assumptions, which are only all true in exceptional cases. These assumptions are:

1.) Clay minerals and climatic parameters are quantitatively related.

2.) Once formed, clay minerals are stable as long as climate is stable.

3.) Clay mineral assemblages are uniform throughout a weathering profile.

4.) Clay mineral assemblages are stable during burial.

5.) Clay minerals have a uniform sensitivity towards environmental change.

In soils or palaeosols that contain pedogenic carbonate (calcrete), the depth to the calcrete horizon from the top of the soil has been proposed as an estimator of precipitation (McFadden and Tinsley, 1985; Wright and Tucker, 1991). However, Royer (1999) found that although the presence and abundance of calcrete can be useful as an approximate indicator of precipitation, there is only a weak relationship between depth to the calcrete

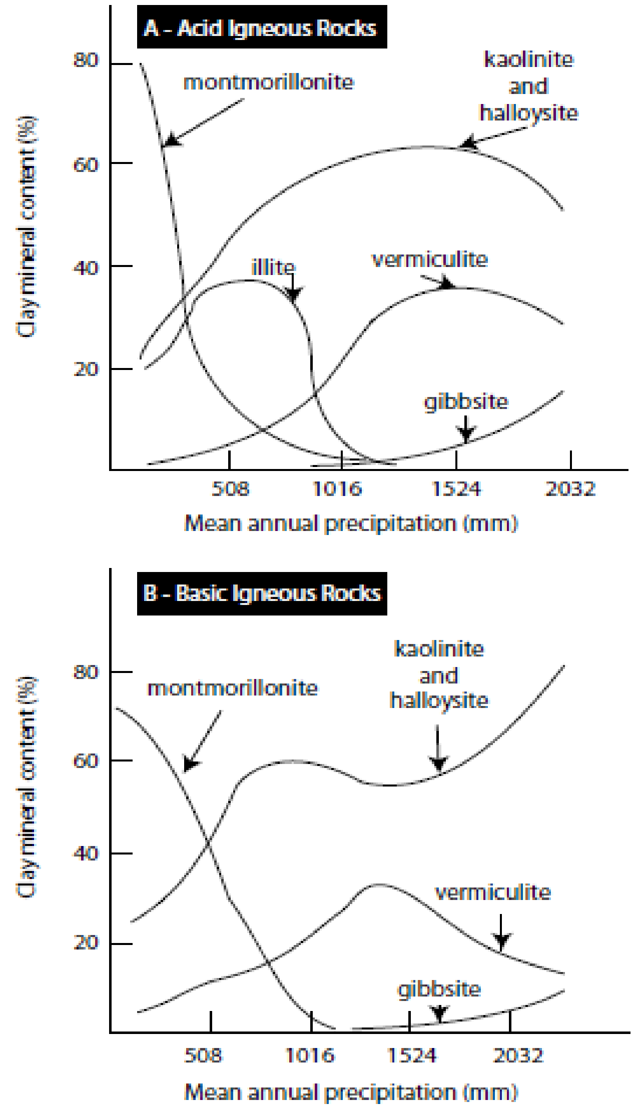


Figure 6.1. Barshad's (1966) diagram relating clay occurrence to precipitation and source rock. Winter precipitation is assumed.

horizon in a soil and the precipitation parameter under which it formed. Royer (1999) also found that presence of pedogenic carbonate correlates well with precipitation values of 760 mm or less, and are therefore useful to differentiate between hydric and arid or seasonally arid environments.

6.1.2. Quantitative Methods

Several quantitative methods have been suggested for estimating rainfall and temperature conditions during pedogenesis based on the chemistry of palaeosols. These methods are relatively new and at present require additional testing to determine the range of their applicability. They generally also require a large sample size to ensure validity of the calculated figures, but can also be used in other circumstances for simpler comparative work.

The chemical index of alteration (CIA=100 x [Al/{Al+Ca+K+Na}]; Nesbitt and Young, 1982) is used as an indicator of the degree of weathering which the soil underwent prior to burial. The underlying principle is to compare the amount of relatively insoluble Al to the amount of more soluble base ions. Some workers use the modified CIA-K, which excludes K, to account for later potassium metasomatism which would alter the overall result (Sheldon and Tabor, 2009; Goldberg and Humayun, 2010). The CIA-K can be related to mean annual precipitation (MAP) by use of the experimentally derived relationship:

$$\text{MAP (mm.yr}^{-1}\text{)} = 221.1e^{0.0197(\text{CIA-K})}$$

The gleization index (GI=FeO/Fe_2O_3; Sheldon and Tabor, 2009) compares the amount of Fe present as Fe^{2+} to Fe^{3+} during pedogenesis and uses this as an indication of fugacity in the soil. This is used to differentiate waterlogged and well-drained soils.

The clayeyness (Al/Si; Sheldon and Tabor, 2009) compares the amount of Al (presumed to be present in clay minerals) to Si (presumed to be present in silt- and sand- sized particles) to provide an estimate of the ratio of clays to larger particles. Although this is relatively imprecise, it has been found to have a highly significant relationship with temperature in inceptisols (protosols).

Salinization ([K+Na]/Al; Sheldon and Tabor, 2009) is used as an indicator of aridity, on the assumption that K and Na build up in salts in arid soils. The salinization index of a soils' Bw- or Bt- horizon is related to mean annual temperature (MAT) by the equation:

$$T\ (^{o}C) = 18.5S + 17.3$$

where S is salinization (Sheldon and Tabor, 2009). This equation is not applicable to hillslope, wetland or tropical soils.

Base loss ([Ca+Mg+Na+K]/Ti; Sheldon and Tabor, 2009) is used to define the extent of leaching in a palaeosol. The soluble base ions should be depleted relative to insoluble Ti during leaching at normal pH.

6.1.3. Palaeosol analysis in the Wahweap Formation

In this study, palaeosols of the Wahweap Formation from the Markagunt, Paunsaugunt and Kaiparowits plateaus are analysed in order to determine if there are variations in pedogenesis across the proximal-distal extent of the formation. Specifically, the idea that pedogenesis progressed to a greater degree in the upland (proximal) setting of the Markagunt and

Paunsaugunt plateaus, as compared to the coastal plain (distal) setting of the Kaiparowits Plateau will be tested. Such variations in pedogenesis, if they exist, would elucidate variable sedimentation patterns and have implications for basin development.

6.2. Methods

Seven palaeosols were selected from across the proximal-distal extent of the Wahweap Formation and across the stratigraphy as part of this study (Fig. 6.2.). On the Markagunt Plateau, the most continuous section of Wahweap Formation crops out along Highway 89 and is well exposed in Stout Canyon (SC) and in a Kampground of America (KOA) campsite. On the Paunsaugunt Plateau, a well-exposed section was located at Kanab Creek (KCB and KC). On the Kaiparowits Plateau exposed palaeosols were found at Big Sage (BSB and BS) and Caine Bench (CB) Each palaeosol was described in the field, and samples were collected from the top and bottom of the B horizon, or, in cases where pedogenic horizons were poorly developed, at the top and bottom of the unit. Whole rock X-ray fluorescence and X-ray diffraction was undertaken in order to determine the chemistry and mineralogy of each sample (Appendix B). The CIA-K, GI, clayeyness, salinization and base loss (Sheldon and Tabor, 2009) were then calculated (Table 6.1.) in order to evaluate similarities and differences in soil-forming processes across regions and through time.

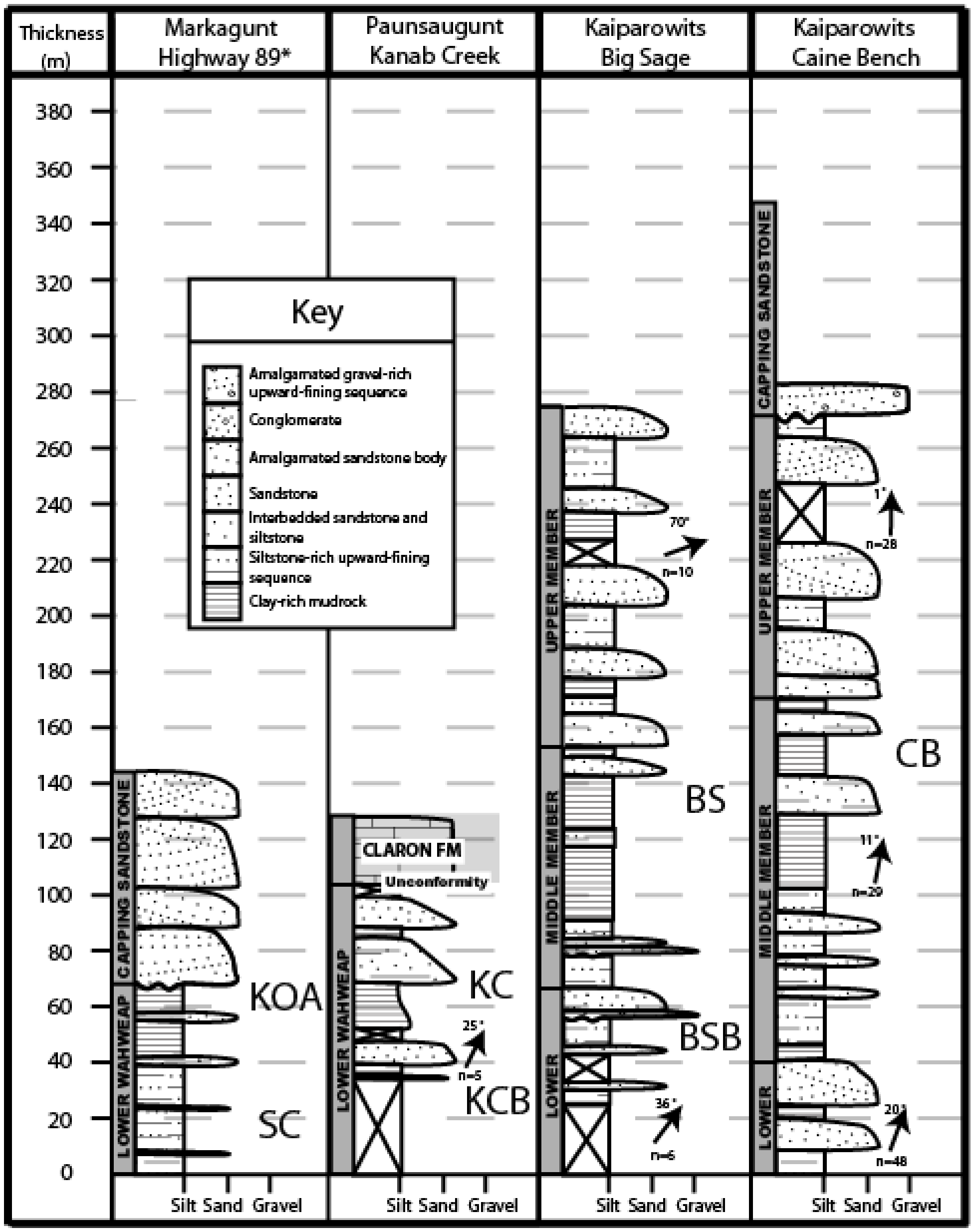


Figure 6.2. Stratigraphic positions of the seven palaeosols sampled in this study. KOA: Kampground of America, SC: Stout Canyon, KC: Kanab Creek, KCB: Kanab Creek Base, BS: Big Sage, BSB: Big Sage Base, CB: Caine Bench.

Table 6.1. Geochemical indicators for palaeosols of the Wahweap Formation. CIA-K: Chemical index of alteration excluding potassium. MAP: Mean annual precipitation. GI: Gleization index

Sample	Mineral content	CIA-K	MAP	GI	Clayeyness	Salinization	Base Loss
SC1	quartz, kaolinite, calcite	45.96	546.49	17.78	0.07	0.18	89.34
SC3	quartz, kaolinite, calcite	56.27	669.60	17.78	0.08	0.17	69.56
KOA1	quartz, kaolinite, calcite, illite	80.25	1073.86	17.60	0.10	0.12	21.66
KOA2	quartz, kaolinite, calcite	38.72	473.91	17.59	0.09	0.16	102.52
KCB3	quartz, kaolinite, calcite, illite	37.98	466.99	18.08	0.11	0.16	120.18
KCB2	quartz, kaolinite, calcite	62.76	760.91	17.95	0.11	0.18	63.12
KC3	quartz, kaolinite, dolomite	68.11	845.50	18.17	0.08	0.15	37.28
KC2	quartz, kaolinite, calcite, dolomite	35.28	442.81	17.99	0.10	0.16	129.06
BSB1	quartz, dolomite, illite, chlorite	37.81	465.49	17.85	0.09	0.16	126.86
BSB2	quartz, illite, chlorite	65.04	795.92	17.99	0.11	0.19	62.18
BS3	quartz, illite, chlorite	87.48	1238.33	17.78	0.11	0.13	24.62
BS5	quartz, kaolinite, chlorite	89.77	1295.39	18.11	0.10	0.19	20.71
CB1	quartz, illite, chlorite	85.88	1199.83	18.09	0.09	0.30	18.88
CB3	quartz, illite, chlorite	87.18	1231.15	17.78	0.10	0.26	19.09

6.3. Palaeosols of the Markagunt and Paunsaugunt plateaus

On the Markagunt and Paunsaugunt plateaus, the most common lithofacies in the Wahweap Formation is pedogenic mudrock. The bulk of floodplain deposits occur as compound palaeosols which are interspersed with isolated fluvial sandstone bodies. At Stout Canyon (SC) at the base of the Wahweap Formation on the Markagunt Plateau, pedogenic horizons within cumulative palaeosols are characterized by Munsell colours ranging from pale yellow (5Y 8/3), to weak red (5R 4/4) to bluish grey (5B 7/1). Individual palaeosols within the cumulative horizon are at least 40 cm thick and contain small (< 1 cm) nodular pedogenic carbonate in a 5 cm thick layer close to the palaeosol surface. Pedogenic horizonization, peds and root mottles are well developed. XRD shows that the palaeosol contains kaolinite. Palaeosol SC is classified as a calcisol.

On the east side of the Markagunt Plateau at the KOA campground on highway 89 (KOA), in the upper part of the lower Wahweap Formation, palaeosols are common and typically cumulative. Individual pedogenic horizons within the cumulative palaeosols are primarily very dark red (5R 3/6) in colour. The total thickness of individual pedogenic horizons within the cumulative palaeosol is unknown because the top of the palaeosol is not exposed, but the horizon is at least 1.2 m thick. Pedogenic carbonate is nodular and occurs in a 12-17 cm thick horizon that is at least 35 cm from the top of the palaeosol. XRD analysis shows that the palaeosol contains kaolinite. Palaeosol KOA is classified as a calcisol (Table 6.1).

At Kanab Creek at the base of the Wahweap Formation on the Paunsaugunt Plateau (KCB), pedogenic horizons within cumulative palaeosols have a colour of light red (5R 6/6), red (5R 8/8) or greyish orange (10YR 7/4). Individual palaeosols are approximately 70 cm thick with disseminated pedogenic carbonate that is concentrated towards the upper surface of the palaeosol. Pedogenic horizons, peds and root mottles are well developed. XRD shows that the palaeosol contains kaolinite (Table 6.1). Palaeosol KCB is classified as an argillic calcisol.

At Kanab Creek at the top of the lower part of the Wahweap Formation on the Paunsaugunt Plateau (KC), pedogenic horizons within cumulative palaeosols have a colour of dark yellowish orange (10YR 4/2) to moderate red (5R 5/4) and are approximately 85 cm thick. Disseminated pedogenic carbonate occurs, and is concentrated approximately 30 cm below the palaeosol surface. Pedogenic horizons, peds and root mottles are well developed. XRD shows that kaolinite is present. Palaeosol KC is classified as an argillic calcisol.

6.4. Palaeosols of the Kaiparowits Plateau

On the Kaiparowits Plateau, palaeosols are typically thin (~50 cm or less) and occur in association with more abundant non-pedogenic subaqueous floodplain mudrock.

At the Big Sage locality, in the lower member of the Wahweap Formation (BSB) on the Kaiparowits Plateau, pedogenic mudrock occurs in a 35 cm thick horizon and is olive brown (2.5Y 4/4). Charcoalified wood and other organic matter occurs in the unit. Pedogenic carbonate nodules occur close to the upper surface of the palaeosol. Pedogenic horizons are not developed, and poorly developed root mottling is evident. XRD analysis shows the presence of chlorite and illite. Palaeosol BSB is classified as a carbonaceous protosol.

At Big Sage, in the upper part of the middle member of the Wahweap Formation (BS) on the Kaiparowits Plateau, pedogenic horizons occur in association with non-pedogenic floodplain mudrocks. Pedogenic mudrock has a colour of olive yellow (5Y 6/8) and the horizon is 50 cm thick. Pedogenic carbonate and horizonization is not developed, but peds and root mottles are present. XRD shows the presence of illite, chlorite and kaolinite. Palaeosol BS is classified as a protosol.

At Caine Bench, no suitable pedogenic horizon was found in the lower member of the Wahweap Formation, although the presence of reworked carbonate nodules in amalgamated sandstones in the lower member suggests that pedogenesis did occur during deposition of the lower member.

In the upper part of the middle member of the Wahweap Formation at Caine Bench (CB) on the Kaiparowits Plateau, pedogenic mudrock occurs in association with non-pedogenic subaqueous floodplain mudrock. The pedogenic unit has a colour of pale yellow (5Y 8/3) and is approximately 30 cm thick. Peds and root mottles are well developed. XRD indicates that illite and chlorite are present. Palaeosol CB is classified as an argillisol.

6.5. Variations in pedogenesis

On the Markagunt and Paunsaugunt plateaus, the sediment which accumulated on river floodplains was well-drained, vegetated, and underwent pedogenesis to a significant degree, resulting in the development of features such as root mottling and soil horizonization. Kraus (1999) documented the relationship between sedimentation rate and pedogenesis. Cumulative soil horizons such as those described here from the Markagunt and Paunsaugunt plateaus are typical of scenarios where floodplains receive constant sedimentation and there is little erosion. Pedogenesis in these conditions is interrupted by sedimentation on the floodplain resulting in interbedded pedogenic and non-pedogenic horizons. In the lower part of the Wahweap Formation on the Markagunt and Paunsaugunt plateaus, palaeosols are poorly to moderately leached (as indicated by the base loss ratio), have calcrete horizons close to the upper surface, and have low to moderate CIA-K values, all indicative of general aridity at the time of pedogenesis (e.g. Wright and Tucker, 1991; Royer, 1999; Metzger and Retallack, 2010; Srivastava, 2010).

In the upper part of the Wahweap Formation in this area (below the capping sandstone), palaeosols are more leached closer to their upper surfaces than they are at depth, and pedogenic carbonate is generally more disseminated and deeper in the soil profile. These features, together with moderately higher CIA-K ratios than lower in the section, indicate wetter conditions.

On the Kaiparowits Plateau, sediment accumulations on floodplains underwent incipient pedogenesis, as evidenced by the presence of pedogenic carbonate, slickensides and root mottling, but well-defined soil horizons did not develop. Soil development was likely interrupted by rapid and continuous overbank deposition or else prohibited by large standing bodies of water developing on low-lying floodplains – as a result palaeosol horizons are isolated and associated with thick subaqueous mudrock successions rather than forming part of cumulative palaeosols.

In the lower member of the Wahweap Formation on the Kaiparowits Plateau, palaeosol horizons have calcrete close to the upper surface, as well as moderate leaching, generally indicative of aridity. Aridity in the Kaiparowits region during deposition of the lower member was most likely seasonal due to the association of these palaeosols with thick subaqueous mudrock accumulations.

In the upper part of the middle member of the Wahweap Formation, palaeosols have no pedogenic carbonate, and base loss and CIA-K ratios indicate a high degree of leaching, all indicative of wet conditions (e.g. Sheldon and Tabor, 2009; Goldberg and Humayun, 2010). These palaeosols are also associated with thick, subaqueous mudrocks, suggesting a lower degree of seasonality than during lower member deposition.

Overall, several trends are notable. More intense pedogenesis in the Markagunt and Paunsaugunt plateaus may indicate that floodplain sedimentation rates in the upland system were slower than in the coastal plain (Kaiparowits Plateau). This would have been a result of the occurrence of relatively small channels with low discharge in the upland setting, where flooding would have been a minor process and pedogenesis was largely unhindered. By contrast, in the lower reaches, larger channels with higher discharge would have been more prone to flooding and thus accumulated thick subaqueous mudrocks on their floodplains, with subordinate pedogenic horizons. Proximity to large channels would therefore have been an important control on pedogenesis in the Wahweap Formation (similar to the findings of McCarthy and Plint [2003] in the Dunvegan Formation). Additionally, pedogenic processes would have operated for a longer time in the upland system, leading to the development of more mature soils than in the distal basin.

During deposition of the lower part of the Wahweap Formation, there was general aridity in the more proximal (upland) parts of the depositional system, and seasonal aridity in the distal parts. During deposition of the upper part of the Wahweap Formation, wetter conditions prevailed, with the wettest conditions occurring in the distal reaches (Kaiparowits Plateau) of the system during deposition of the middle and upper members. This suggests that climatic conditions became increasingly wet as the Wahweap Formation was deposited. The suggestion of a seasonally arid but overall wet palaeoclimate from palaeosol analysis is consistent with the results of the facies analysis presented previously, where abundant subaqueous mudrock with a notable lack of features such as mudcracks or armoured mud balls, indicates wet conditions. It is also consistent with reconstructions of the overall regional climate as presented by Fricke *et al*. (2010) for the central Western Interior Basin.

Finally, as deposition of the capping sandstone began, a switch to a more arid palaeoclimatic regime occurred (e.g. Simpson *et al.*, 2008; W. Simpson *et al*., 2010), accompanied by tectonism and possibly uplift along the East-Kaibab monocline (Tindall *et al*., 2010).

6.6. Conclusion

Variations in pedogenesis suggest that climate became wetter as deposition of the lower part of the Wahweap Formation progressed. This trend is consistent across the east-west extent of the Wahweap Formation; however, a more quantitative estimate of climate change would require a study with a greater sample number. During deposition of the upper part of the Wahweap Formation, variations occurred in pedogenesis from the west (where it was more arid) to the east (where it was more humid), possibly driven by topography. This would also explain the isolated (as opposed to cumulative) nature of Kaiparowits Plateau palaeosols and their association with thick subaqueous swamp deposits.

The exact cause of increased moisture is unknown, but it is intriguing that this occurs in correlation with transgression during deposition of the isolated fluvial facies tract on the Kaiparowits Plateau. Whether or not there is a causal link between these two features is unclear, but presents an interesting question for future work.

7. Discussion and Conclusion

7.1. Depositional environment of the Wahweap Formation

Combined facies analysis and palaeosol geochemistry show that climatic conditions during the deposition of the lower member of the Wahweap Formation were wet with seasonal aridity. As deposition progressed, wetter conditions prevailed and there is less evidence of a pronounced arid season in the middle and upper members. This was followed by more arid conditions during deposition of the capping sandstone, where work by Simpson *et al*. (2008) and W. Simpson *et al.* (2010) reveals the presence of aeolian stratification and cryptobiotic soil textures indicative of aridity.

Many workers (e.g. Lehman, 1987; Sankey, 2001 and references therein) have suggested that there were strong palaeoclimatic differences between the northern and southern Western Interior Basin in the latest-Campanian to Maastrichtian. In comparison to coeval northern deposits such as the Two Medicine and Judith River formations, where there is some evidence for greater aridity, such as drought-related mass mortality (Rogers, 1990) and minor ephemeral stream flow (Eberth, 1990), the Wahweap Formation does represent a wetter palaeoenvironment. However, in light of the seasonally arid environment postulated herein for the Wahweap Formation, especially in the Markagunt and Paunsaugunt regions, the degree of climatic variation between the middle and northern Western Interior during the middle Campanian could be as modest as the intensity of the arid season.

7.2. Reassessment of stratigraphy

A synthesis of the stratigraphic data available from the Markagunt, Paunsaugunt and Kaiparowits plateaus, as well as the Henry Mountains syncline allows for a re-evaluation of the stratigraphic scheme for these areas. A number of questions can be addressed: 1.) is there enough lithological similarity between the Wahweap Formation on the Markagunt and Paunsaugunt plateaus and the type area on the Kaiparowits Plateau to justify placement in the same formation?; 2.) does the capping sandstone have enough lithological similarity to the Tarantula Mesa Formation to justify placement in the same formation? and 3.) What is the most appropriate way to subdivide the Wahweap Formation into members?

Because formations are defined on lithological similarity, it can be argued that the Wahweap Formation on the Markagunt and Paunsaugunt plateaus is not properly classified as the same formation as the Wahweap Formation on the Kaiparowits Plateau. West of the East Kaibab monocline, the succession is dominated by palaeosols and lenticular sandstones, whereas east of the East Kaibab monocline the formation is dominated by drab non-pedogenic mudrocks and tabular sandstones. Additionally, three of the four informal members that are used on the Kaiparowits Plateau (lower, middle and upper) are not valid to the west. Although usage of the term “Wahweap Formation” to describe these rocks is somewhat entrenched, it is not entirely appropriate and stratigraphic revision and the erection of a new formation seems appropriate.

With regard to the synonymy of the capping sandstone member and the Tarantula Mesa Formation, the lithological similarity between these two units permits them to be placed in the same formation. In this case, the name “Tarantula Mesa Formation” has precedence, and can be applied to all “capping sandstone” outcrop from the Markagunt to Kaiparowits Plateau, which would then fall into a separate formation from the Wahweap Formation on the Kaiparowits Plateau.

The four informal members that are defined on the Kaiparowits Plateau are not applicable to the succession on the Markagunt and Paunsaugunt plateaus. However, they are useful due to the fine resolution they provide in subdividing the succession on the Kaiparowits Plateau, where the bulk of new fossil prospecting is focused, and a clear subdivision scheme is needed. On the Kaiparowits Plateau, the contacts between members are frequently used as markers to place fossil localities into stratigraphic sections. It is noteworthy that member boundaries have variable positions in stratigraphy across the Kaiparowits Plateau, and therefore are most likely diachronous, which has implications for correlating fossil sites across regions. Thus, some caution needs to be applied when using members to correlate fossil sites.

8. References

Adams, M.M., and Bhattacharya, J.P. 2005. No change in fluvial style across a sequence boundary, Cretaceous Blackhawk and Castlegate formations of central Utah, U.S.A. Journal of Sedimentary Research 75:1038-1051.

Allen, J.P., and Fielding, C.R. 2007. Sedimentology and stratigraphic architecture of the Late Permian Betts Creek Beds, Queensland, Australia. Sedimentary Geology 202: 5-34.

Allen, J.R.L. 1965. The sedimentation and palaeogeography of the Old Red Sandstone of Anglesey, North Wales. Yorkshire Geological Society, Proceedings 35: 139-185.

Allen, J.R.L. 1983. Studies in fluviatile sedimentation: bars, bar-complexes and sandstone sheets (low sinuosity braided streams) in the Brownstones (L Devonian), Welsh borders. Sedimentary Geology 33: 237-293.

Allen, P. A., Homewood, P., and Williams, G. D. 1986. Foreland basins: An introduction. In: Allen, P.A. and Homewood, P. (eds.). *Foreland Basins*. Special Publication 8. International Association of Sedimentologists. Blackwell Scientific Publications, London. 3-12.

Ambrose, W.A., and Ayers, W.B. Jr. 2007. Geologic controls on transgressive-regressive cycles in the upper Pictured Cliffs Sandstone and coal geometry in the lower Fruitland

Formation, northern San Juan Basin, New Mexico and Colorado. American Association of Petroleum Geologists, Bulletin 91: 1099-1122.

Amorosi, A., Pavesi, M., Lucchi, M.R., Sarti, G., and Piccin, A. 2008. Climatic signature of cyclic fluvial architecture from the Quaternary of the central Po Plain, Italy. Sedimentary Geology 209: 58-68.

Armstrong, R.L. 1968. Sevier orogenic belt in Nevada and Utah. Geological Society of America, Bulletin 79: 429-458.

Atchley, S.C., Nordt, L.C., and Dworkin, S.I. 2004. Eustatic control on alluvial sequence stratigraphy: a possible example from the Cretaceous-Tertiary transition of the Tornillo Basin, Big Bend National Park, west Texas, U.S.A. Journal of Sedimentary Research 74: 391-404.

Barshad, L. 1966. The effect of a variation in precipitation on the nature of clay mineral formation in soils from acid and basic igneous rocks. Proceedings of the International Clay Conference, Jerusalem 1: 167-173.

Barth, A.P., Wooden, J.L., Jacobson, C.E., Probst, K. 2004. U-Pb geochronology and geochemistry of the McCoy Mountains Formation, southeastern California: a Cretaceous retroarc foreland basin. Geological Society of America, Bulletin 116: 142-153.

Bilodeau, W.L. 1986. The Mesozoic Mogollon highlands, Arizona: an Early Cretaceous rift shoulder. Journal of Geology 94: 724-735.

Blakey, R.C. 1997. North American Paleogeographic Maps: Late Cretaceous (85 Ma). Available at cpgeosystems.com. Accessed January 2011.

Blakey, R.C. 2008. Pennsylvanian-Jurassic sedimentary basins of the Colorado Plateau and southern Rocky Mountains: *in* Miall, A.D. (ed.). The Sedimentary Basins of the United States and Canada. Sedimentary Basins of the World 5: 245-296.

Blum, M.D., and Tőrnqvist, T.E. 2000. Fluvial responses to climate and sea level: a review: Sedimentology 47 (supplement 1): 2-48.

Bowers, W. E. 1990. Geologic maps of Bryce Canyon National Park and vicinity, southwestern Utah. U.S. Geological Survey Miscellaneous Investigation Series Map I-2108.

Breyer, J.A., Busbey, A.B., Hanson, R.E., Befus, K.E., Griffin, W.R., Hargrove, U.S., Bergman, S.C. 2007. Evidence for Late Cretaceous volcanism in Trans-Pecos Texas. Journal of Geology 115: 243-251.

Bridge, J.S. 2003. Rivers and floodplains. Blackwell, Malden MA. 491 p.

Brierley, G.J., Ferguson, R.J., and Woolfe, K.J. 1997. What is a fluvial levee? Sedimentary Geology 114: 1-9.

Brinkman, D.B., Ryan, M.J., and Eberth, D.A. 1998. The paleogeographic and stratigraphic distribution of ceratopsids (Ornithischia) in the Upper Judith River Group of Western Canada. Palaios 13:160-169.

Buillit, N., Lallier-Vergés, E., Pradier, B. Nicolas, G. 2002. Coal petrographic genetic units in deltaic-plain deposits of the Campanian Mesa Verde Group (New Mexico, USA). International Journal of Coal Geology 51: 93-110.

Burbank, D.W., Beck, R.A., Raynolds, R.G.H., Hobbs, R., and Tahirkheli, R.A.K. 1989. Thrusting and gravel progradation in foreland basins: a test of post-thrusting gravel dispersal. Geology 16:1143-1146.

Burchfiel, B.C., and Davis, G.A. 1975. Nature and controls of Cordilleran orogenesis, western United States: Extensions of an earlier synthesis. American Journal of Science 275: 363-396.

Cant, D.J., and Stockmal, G.S. 1989. The Alberta foreland basin: relationship between stratigraphy and Cordilleran terrane-accretion events. Canadian Journal of Earth Sciences 26: 1964-1975.

Cant, D.J., and Walker, R.G. 1978. Fluvial processes and facies sequences in the sandy braided South Saskatchewan River, Canada. Sedimentology 25: 625-648.

Catuneanu, O. 2004. Retroarc foreland systems – evolution through time. Journal of African Earth Sciences 38: 225-242.

Catuneanu, O. 2006. Principles of Sequence Stratigraphy. Elsevier, Amsterdam. 375 p.

Catuneanu, O., Sweet, A.R., and Miall, A.D. 2000. Reciprocal stratigraphy of the Campanian-Paleocene Western Interior of North America. Sedimentary Geology 134:235-255.

Cifelli, R.L. 1990. Cretaceous mammals from southern Utah. II. Marsupials and marsupial like mammals from the Wahweap Formation (early Campanian). Journal of Vertebrate Paleontology 10: 320-331.

Cowan, E.J. 1993. Longitudinal fluvial drainage patterns within a foreland basin fill: Permo-Triassic Sydney Basin, Australia. Sedimentary Geology 85: 557-577.

Cross, T.A. 1986. Tectonic controls of foreland basin subsidence and Laramide style deformation, western United States. In: Allen, P.A. and Homewood, P. (eds.). *Foreland Basins*. Special Publication 8. International Association of Sedimentologists. Blackwell Scientific Publications, London p. 13-39.

Cross, T.A., and Pilger, R.H. 1982. Controls of subduction geometry, locations of magmatic arcs, and tectonics of arcs and backarc regions. Geological Society of America Bulletin 93: 545-562.

Dalrymple, R.W., Makino,Y., Zaitlin, B.A. 1991. Temporal and spatial patterns of rhythmite deposition on mud flats in the macrotidal, Cobequid Bay–Salmon River estuary, Bay of Fundy, Canada. In: Smith, D.G., Reinson, G.E., Zaitlin, B.A. and Rahmani, R.A. (eds.). Clastic Tidal Sedimentology. Canadian Society of Petroleum Geologists Memoir, 16, p. 137–160.

Dalrymple, R.W., Zaitlin, B.A., and Boyd, R. 1992. Estuarine facies models: conceptual basis and stratigraphic implications: Journal of Sedimentary Petrology 62: 1130-1146.

De Blieux, D. D., Kirkland, J.I., Eaton, J.G., Titus, A.L., Getty, M.A., Sampson, S.D., and Hayden, M.C. 2009. Vertebrate paleontology of the Lower-Middle Campanian Wahweap Formation in Grand Staircase-Escalante National Monument (GSENM). Advances in Western Interior Late Cretaceous Paleontology and Geology, Abstracts with Program, p. 17.

DeCelles, P.G. 2004. Late Jurassic to Eocene evolution of the Cordilleran thrust belt and foreland basin system, western U.S.A. American Journal of Science 304:105-168.

DeCelles, P.G., and Coogan, J.C. 2006. Regional structure and kinematic history of the Sevier fold-and-thrust belt, central Utah. Geological Society of America Bulletin 118: 841-864.

DeCelles, P.G., and Currie, B.S. 1996. Long-term sediment accumulation in the Middle Jurassic-early Eocene Cordilleran retroarc foreland-basin system. Geology 24: 591-594.

DeCelles, P.G., and Giles, K.N. 1996. Foreland basin systems. Basin Research 8: 105-123.

Dennison, J.M. 1994, Assessing the roles of tectonism and eustasy. In: Dennison, J.M., and Ettensohn, F.R. (eds.). Tectonic and Eustatic Controls on Sedimentary Cycles: SEPM, Concepts in Sedimentology and Paleontology 4, p. 1-4.

Dickinson, W.R. 1974. Plate tectonics and sedimentation. Special Publication of SEPM 22:1-27.

Dickinson, W.R., and Gehrels, G.E. 2008. Sediment delivery to the Cordilleran foreland basin: insights from U-Pb ages from detrital zircons in Upper Jurassic and Cretaceous strata of the Colorado Plateau. American Journal of Science 308: 1041-1082.

Dickinson, W.R., Klute, M.A., Hayes, M.J., Janecke, S.U., Lundin, E.R., McKittrik, M.A., and Olivares, M.D. 1988. Paleogeographic and paleotectonic setting of Laramide sedimentary basins in the central Rocky Mountain region. Geological Society of America, Bulletin 100: 1023-1039.

Dickinson, W.R., Fiorillo, A.R., Hall, D.L., Monreal, R., Potochnik, A.R., Swift, P.N., 1989. Cretaceous strata of southern Arizona. In: Jenney, J.P., and Reynolds, S.J. (eds.). Geologic Evolution of Arizona. Geological Society Digest 17: 447-461.

Doelling, H.H., 1997. Interim geologic map of the Smoky Mountain 30`x60` quadrangle, Kane and San Juan Counties, Utah and Coconino County, Arizona. Utah Geological Survey Open File Report 359.

Doelling, H.H., and Willis, G.C., 1999. Interim geologic map of the Escalante and parts of the Loa and Hite crossing 30` x 60` quadrangles, Garfield and Kane counties, Utah. Utah Geological Survey Open File Report. 368. 22p.

Drewes, H., 1978. The Cordilleran orogenic belt between Nevada and Chihuahua. Geological Society of America, Bulletin 89: 641-647.

Dunne, G.C., Garvey, T.P., Oborne, M., Schneidereit, D., Fritsche, A.E., Walker, J.D. 1998. Geology of the Inyo Mountains Volcanic Complex: implications for Jurassic paleogeography of the Sierran magmatic arc in eastern California. Geological Society of America, Bulletin 110: 1376-1397.

Dyman, T. S., Merewether, E. A., Molenaar, C. M., Cobban, W. A., Obradovich, J. D., Weimer, R. J., and Bryant, W. A., 1994. Stratigraphic transects for cretaceous rocks, Rocky Mountain and Great Plains regions. In: Caputo, M. V., Peterson, J. A., and Franczyk, K. J.

(eds.). Mesozoic systems of the Rocky Mountain region, USA, Rocky Mountain Section, Society for Sedimentary Geology, p. 365–391.

Eaton, J.G. 1990. Stratigraphic revision of Campanian (Upper Cretaceous) rocks in the Henry Basin, Utah. The Mountain Geologist 27: 27-38.

Eaton, J.G. 1991. Biostratigraphic framework for the Upper Cretaceous rocks of the Kaiparowits Plateau, southern Utah. In: Eaton, J.G. and Nations, J.D. (eds.). Stratigraphy, Depositional Environments, and Sedimentary Tectonics of the Western Margin, Cretaceous Western Interior Seaway, Geological Society of America Special Paper 260:47-63.

Eaton, J.G. 1999. Vertebrate paleontology of the Paunsaugunt Plateau, Upper Cretaceous, southwestern Utah. In: Gillette, D.D. (ed.). Vertebrate Paleontology in Utah: Utah Geological Survey Miscellaneous Publication 99-1: 335-339.

Eaton, J.G. 2002. Multituberculate mammals from the Wahweap (Campanian, Aquilan) and Kaiparowits (Campanian, Judithian) formations, within and near Grand Staircase-Escalante National Monument, southern Utah. Utah Geological Survey Miscellaneous Publication 02-4, 1-66.

Eaton, J.G. 2006 Santonian (Late Cretaceous) mammals from the John Henry Member of the Straight Cliffs Formation, Grand Staircase-Escalante National Monument, Utah. Journal of Vertebrate Paleontology 26: 446-460.

Eaton, J.G., and Cifelli, R.L. 1988. Preliminary report on Late Cretaceous mammals of the Kaiparowits Plateau, southern Utah. Contributions to Geology, University of Wyoming 26: 45-55.

Eaton, J.G., Laurin, J., Kirkland, J.I., Tibert, N.E., Leckie, R.M., Sageman, B.B., Goldstrand, P.M., Moore, D.W. Straub, A.W., Cobban, W.A., and Dalebout, J.D. 2001. Cretaceous and Early Tertiary geology of Cedar and Parowan canyons, western Markagunt Plateau, Utah. Utah Geological Association Publication 30: 337-362.

Eberth, D.A. 1990. Stratigraphy and sedimentology of vertebrate microfossil sites in the uppermost Judith River Formation (Campanian), Dinosaur Provincial Park, Alberta, Canada. Palaeogeography, Palaeoclimatology, Palaeoecology 78: 1-36.

Eberth, D.A., and Hamblin, A.P. 1993. Tectonic, stratigraphic and sedimentologic significance of a regional discontinuity in the upper Judith River Group (Belly River wedge) of southern Alberta, Saskatchewan, and northern Montana. Canadian Journal of Earth Science 30: 174-200.

Elder, W.P., and Kirkland, J. 1993. Cretaceous paleogeography of the Colorado Plateau and adjacent areas. In: Morales, M. (ed.). Aspects of Mesozoic Geology and Paleontology of the Colorado Plateau. Museum of Northern Arizona Bulletin, 59: 129-152.

Emery, D., and Myers, K., 1996, Sequence Stratigraphy. Blackwell Science, Oxford, U.K. 297 p.

Evernden, J.F., and Kistler, R.W. 1970. Chronology of emplacement of Mesozoic batholithic complexes in California and western Nevada: U.S. Geological Survey, Professional Paper 623, 42 p.

Fassett, J.E. 2000. Chapter Q. In: Kirschbaum, M.A., Roberts, L.N.R. and Biewick, L.R.H (eds.). Geologic Assessment of Coal in the Colorado Plateau: Arizona, Colorado, New Mexico and Utah. United States Geological Survey Professional Paper CD-ROM 1625-B, Q1-117.

Feldman, H.R., Franseen, E.K., Joeckel, R.M., and Heckel, P.H. 2005. Impact of longer-term modest climate shifts on architecture of high-frequency sequences (cyclothems), Pennsylvanian of midcontinent U.S.A. Journal of Sedimentary Research 75: 350-368.

Fielding, C.R. 2006. Upper flow regime sheets, lenses and scour fills: extending the range of architectural elements for fluvial sandstone bodies. Sedimentary Geology 190: 227-240.

Fielding, C.R., Falkner, A.J., and Scott, S.G. 1993. Fluvial response to foreland basin overfilling; the Late Permian Rangal Coal Measures in the Bowen Basin, Queensland, Australia. Sedimentary Geology 85: 475-497.

Fitzsimmons, R., and Johnson, S. 2000. Forced regressions: recognition, architecture and genesis in the Campanian of the Bighorn Basin, Wyoming. In: Hunt, D. and Gawthorpe, R.L. (Eds.). Sedimentary Responses to Forced Regressions. Geological Society, London, Special Publications 172: 113-139.

Flemings, P. B. and Jordan, T. E. 1990. Stratigraphic modelling of foreland basins: interpreting thrust deformation and lithospheric rheology. Geology 18: 430–434.

Fouch, T.D., Lawton, T.F., Nichols, J.D., Cashion, W.B., and Cobban, W.A. 1983. Patterns and timing of synorogenic sedimentation in Upper Cretaceous rocks of central and northwestern Utah. In: Reynolds, M.W. and Dolly, E.D. (eds.). Mesozoic Paleogeography of the West-central United States. Society of Economic Paleontologists and Mineralogists, Rocky Mountain Section 305-336.

Foreman, B.Z., Rogers, R., Wirth, K., and Hanna, R. 2004. Geochemical correlation and characterization of Two Medicine Formation bentonites, northwestern Montana. Abstracts. Journal of Vertebrate Paleontology 24 (3), 59.

Fricke, H.C., Foreman, B.Z., and Sewall, J.O. 2010. Integrated climate model-oxygen isotope evidence for a North American monsoon during the Late Cretaceous. Earth and Planetary Science Letters 289: 11-21.

Friend, PF., Slater, M.J., and Williams, R.C. 1979. Vertical and lateral building of river sandstone bodies, Ebro Basin, Spain: Geological Society of London, Journal 146: 39-46.

Gastaldo, R.A., Allen, G.P., and Huc, A.Y. 1995. The tidal character of fluvial sediments of the modern Mahakam River delta, Kalimantan, Indonesia. In: Flemming, B.W., and Bartholoma, A. (eds.). Tidal Signatures in Modern and Ancient Sediments: International Association of Sedimentologists, Special Publication 24 p. 171-181.

Ghosh, P., Sarkar, S., and Maulik, P. 2006. Sedimentology of a muddy alluvial deposit: Triassic Denwa Formation, India: Sedimentary Geology 191: 3-36.

Gilbling, M.R. 2006. Width and thickness of fluvial channel bodies and valley fills in the geological record: a literature compilation and classification. Journal of Sedimentary Research 76: 731-770.

Gibling, M.R., Nanson, G.G., and Maroulis, J.C. 1998. Anastomosing river sedimentation in the Channel Country of central Australia. Sedimentology 45: 699-719.

Gilbert, H., Velasco, A.A., and Zandt, G. 2007. Preservation of Proterozoic terrane boundaries within the Colorado Plateau and implications for its tectonic evolution. Earth and Planetary Science Letters 258: 237-248.

Gill, J.R., and Cobban, W.A. 1973. Stratigraphy and geologic history of the Montana Group and equivalent rocks, Montana, Wyoming, and North and South Dakota. Geological Survey Professional Paper 776: 1-37.

Gill, J.R., Merewether, E.A., Cobban, W.A. 1970. Stratigraphy and nomenclature of some Upper Cretaceous and Lower Tertiary rocks in south-central Wyoming. United States Geological Survey Professional Paper 667: 1-53.

Gingras, M.K., Maceachern, J.A., and Pickerill, R.K. 2004. Modern perspectives on the *Teredolites* ichnofacies: observations from Willipa Bay, Washington. Palaios 19: 79-88.

Goldberg, K. and Humayun, M. 2010. The applicability of the Chemical Index of Alteration as a paleoclimatic indicator: an example from the Permian of the Paraná Basin, Brazil. Palaeogeography, Palaeoclimatology, Palaeoecology 293: 175-183.

Goldstrand, P.M. 1992. Evolution of Late Cretaceous and early Tertiary basins of southwest Utah based on clastic petrology. Journal of Sedimentary Petrology 62: 495-507.

Goldstrand, P.M. 1994. Tectonic development of Upper Cretaceous to Eocene strata of southwestern Utah. Geological Society of America, Bulletin 106: 145–154.

Goodwin, M.B., Deino, A.L. 1989. The first radiometric ages from the Judith River Formation (Upper Cretaceous), Hill County, Montana. Canadian Journal of Earth Science 26: 1384-1391

Gradstein, F.M., Ogg, J.G., and Smith, A.G. 2004. A Geologic Time Scale 2004. Cambridge University Press, Cambridge 589p.

Gregory, H.E., and Moore, R.C. 1931. The Kaiparowits Region. The United States Geological Survey Professional Paper 164: 1-161.

Hancock, J.M., and Kauffman, E.G. 1979. The great transgressions of the Late Cretaceous. Journal of the Geological Society of London 136: 175-186.

Haq, B.U., Hardenbol, J., and Vail, P.R. 1987. Chronology of fluctuating sea levels since the Triassic: Science 235: 1156-1166.

Hayes, P.T. 1970. Cretaceous paleogeography of southeastern Arizona and adjacent areas. US Geological Survey Professional Paper 658-B: 1-42.

Heller, P.L., Angevine, C.L., Winslow, N.S., and Paola, C. 1988. Two-phase stratigraphic model of foreland-basin sequences. Geology 16: 501-504.

Heller, P.L., and Paola, C. 1989. The paradox of Lower Cretaceous gravels and the initiation of thrusting in the Sevier orogenic belt, United States Western Interior. Geological Society of America, Bulletin 101: 864-875.

Heller, P.L., and Paola, C. 1996. Downstream changes in alluvial architecture: An exploration of controls on channel-stacking patterns. Journal of Sedimentary Research 66: 297-306.

Hicks, J.F., Obradovich, J.D., and Tauxe, L. 1995. A new calibration point for the Late Cretaceous time scale: The 40Ar/39Ar isotopic age of the C33/C33n geomagnetic reversal from the Judith River Formation (Upper Cretaceous), Elk Basin, Wyoming, USA. Journal of Geology 103: 243-256.

Hicks, J.F., Obradovich, J.D., and Tauxe, L. 1999. Magnetostratigraphy, isotopic age calibration, and intercontinental correlation of the Red Bird section of the Pierre Shale, Niobrara County, Wyoming, USA. Cretaceous Research 20: 1-27.

Hilbert-Wolf, H.L., Simpson, E.L., Simpson, W.S., Tindall, S.E., and Wizevich, M.C. 2009. Insights into syndepositional fault movements in a foreland basin; trends in seismites of the Upper Cretaceous, Wahweap Formation, Kaiparowits Basin, Utah, U.S.A. Basin Research 21: 856-871.

Hintze, L. 1997. Geologic Highway Map of Utah. Brigham Young University Geology Studies Special Publication 3.

Holbrook, J.M. 1996. Complex fluvial response to low gradients at maximum regression: a genetic link between smooth sequence boundary morphology and architecture of overlying sheet sandstone: Journal of Sedimentary Research 66: 713-722.

Horton, B.K., Constenius, K.N., and DeCelles, P.G. 2004. Tectonic control on coarse-grained foreland-basin sequences: an example from the Cordilleran foreland basin, Utah. Geology 32: 637-640.

Houston, W.S., Huntoon, J.E., and Kamola, D.L., 2000. Modeling of Cretaceous foreland-basin parasequences, Utah, with implications for timing of Sevier thrusting. Geology 28: 267–270.

Huntoon, P.W. 1993. Influence of inherited Precambrian basement structure on the localization and form of Laramide monoclines, Grand Canyon, Arizona. In Schmidt, C.J., Chase, R.B., and Erslev, E.A., (eds.). Laramide Basement Deformation in the Rocky Mountain Foreland of the Western United States. Geological Society of America Special Paper 280, 243-256.

Ingersol, R.V. 1988. Tectonics of sedimentary basins. Geological Society of America, Bulletin 100: 1704-1719.

Jinnah, Z.A., and Roberts, E.M. 2011. Facies associations, paleoenvironment and base-level changes in the Upper Cretaceous Wahweap Formation, Utah, U.S.A. Journal of Sedimentary Research 81: 266-283.

Jinnah, Z.A., Roberts, E.M., Deino, A.L., Larsen, J.S., Link, P.K., and Fanning, C.M. 2009a. New ^{40}Ar-^{39}Ar and detrital zircon U-Pb ages for the Upper Cretaceous Wahweap and Kaiparowits formations on the Kaiparowits Plateau, Utah: implications for regional correlation, provenance and biostratigraphy. Cretaceous Research 30: 287-299.

Jinnah, Z.A., Getty, M.A., and Gates, T.A. 2009b. Taphonomy of a hadrosaur bonebed from the Upper Cretaceous Wahweap Formation, Grand Staircase-Escalante National Monument, Utah. Advances in Western Interior Late Cretaceous Paleontology and Geology, Abstracts with Program, p. 28.

Johnson, E.A., and Pierce, F.W. 1990. Variations in fluvial deposition on an alluvial plain: an example from the Tongue River Member of the Fort Union Formation (Paleocene), southeastern Powder River Basin, Wyoming, U.S.A.: Sedimentary Geology 69: 21-36.

Jordan, T. E. 1981. Thrust loads and foreland basin evolution, Cretaceous, western United States. American Association of Petroleum Geologists Bulletin 65: 2506–2520.

Kauffman, E.G. 1969. Cretaceous marine cycles of the Western Interior. Mountain Geologist 6: 227-245.

Kauffman, E.G. 1977. Geological and biological overview: Western Interior Cretaceous Basin. Mountain Geologist 14: 75-99.

Kauffman, E.G., and Caldwell, W.G.E. 1993. The Western Interior Basin in space and time. In: Caldwell, W.G.E. and Kauffman, E.G. (eds.). Evolution of the Western Interior Basin. Geological Association of Canada Special Paper 39: 1-30.

Kielan-Jaworowska, Z., Cifelli, R.L. and Luo, Z-X. 2004. Mammals From the Age of Dinosaurs: Origins, Evolution, and Structure. Columbia University Press, New York, 630 p.

Kirkland, J. and De Blieux, D. 2005. Dinosaur remains from the lower to middle Campanian Wahweap Formation at Grand Staircase-Escalante National Monument, southern Utah. Journal of Vertebrate Paleontology, Abstracts of posters 25(3), 78A.

Kirkland, J. and De Blieux, D. 2010. New basal centrosaurine ceratopsian skulls from the Wahweap Formation (Middle Campanian), Grand Staircase–Escalante National Monument, southern Utah, In: Ryan, M.J., Chinnery-Allgeier, B.J. and Eberth, D.A. (eds.). New Perspectives on Horned Dinosaurs: The Royal Tyrrell Museum Ceratopsian Symposium. Bloomington, Indiana University Press, p. 117–140

Komatsubara, J. 2004. Fluvial architecture and sequence stratigraphy of the Eocene to Oligocene Iwaki Formation, northeast Japan: channel fills related to sea-level change. Sedimentary Geology 168: 109-123.

Kraus, M.J. 1997. Lower Eocene alluvial paleosols: pedogenic development, stratigraphic relationships, and paleosol/landscape associations. Palaeogeography, Palaeoclimatology, Palaeoecology 129: 387-406.

Kraus, M.J. 1999. Paleosols in clastic sedimentary rocks: their geologic applications. Earth Science Reviews 47: 41-70.

Kraus, M.J. and Wells, T.M. 1999. Recognizing avulsion deposits in the ancient stratigraphic record, In: Smith, N.D., and Rogers, J., (eds.). Fluvial Sedimentology VI. International Association of Sedimentologists, Special Publication, 28: 251-268.

Kristinik, L.F., and DeJarnett, B.B. 1995. Lateral variability of sequence stratigraphic framework in the Campanian and Lower Maastrichtian of the Western Interior Seaway, In: Van Wagoner, J.C. and Bertram, G.T. (eds.). Sequence Stratigraphy of Foreland Basin Deposits: Outcrop and Subsurface examples from the Cretaceous of North America. American Association of Petroleum Geologists, Memoir 64: 11-25.

Kuiper, K. F., Deino, A., Hilgen, F.J., Krijgsman, W. Renne, P.R. and Wijbrans, J.R. 2008. Synchronizing rock clocks of Earth history. Science 320: 500-504.

Kump, L.R. and Slingerland, R.L. 1999. Circulation and stratification of the early Turonian Western Interior Seaway: sensitivity to a variety of sources. In: Barrera, E., and Johnson, C.C.

(Eds.). Evolution of the Cretaceous Ocean-Climate System. Geological Society of America, Special paper 332.

Lageson, D.R., Schmitt, J.G., Horton, B.K., Kalakay, T.J. and Burton, B.R. 2001. Influence of Late Cretaceous magmatism on the Sevier orogenic wedge, western Montana. Geology 29: 723-726.

Larsen, J.S. 2007. Facies and Provenance of the Pine Hollow Formation: Implications for Sevier foreland basin evolution and the Paleocene climate of southern Utah [Unpublished M.Sc. Thesis]: Idaho State University, Pocatello, 83 p.

Lawton, T.F. 1986. Fluvial systems of the Upper Cretaceous Mesaverde Formation, Central Utah: A record of transition from thin-skinned to thick-skinned deformation in the foreland region. In: Peterson, J.A. (ed.). Paleotectonics and Sedimentation in the Rocky Mountain Region. American Association of Petroleum Geologists Memoir 41: 423-442.

Lawton, T.F. 2008. Laramide sedimentary basins. In: Miall, A.D. (ed.). The Sedimentary Basins of the United States and Canada. Sedimentary Basins of the World 5: 429-450.

Lawton, T.F., and Trexler, J.H. 1991. Piggyback basin in the Sevier orogenic belt, Utah: implications for development of the thrust wedge. Geology 19: 827-830.

Lawton, T.F., and Christensen, A. 2005. Sequence boundaries in terrestrial foreland-basin strata: do they lie above or below the amalgamated fluvial facies tract? (abstract). American Association of Petroleum Geologists, Annual Convention, Technical Program.

Lawton, T.F., Pollock, S.L., and Robinson, R.A.J. 2003. Integrating sandstone petrology and nonmarine sequence stratigraphy: application to the Late Cretaceous fluvial systems of southwestern Utah, U.S.A. Journal of Sedimentary Research 73: 389-406.

Leahy, G.D. and Lerbekmo, J.F. 1995. Macrofossil magnetobiostratigraphy for the upper Santonian – lower Campanian interval in the Western Interior of North America: comparisons with European stage boundaries and planktonic foraminiferal zonal boundaries. Canadian Journal of Earth Sciences 32: 247-260.

Leeder, M.R. 1999. Sedimentology and Sedimentary Basins: from Turbulence to Tectonics. Oxford, Blackwell Science, 592p.

Lehman, T.M. 1987. Late Maastrichtian paleoenvironments and dinosaur biogeography in the western interior of North America. Palaeogeography, Palaeoclimatology, Palaeoecology 60: 189–217.

Lehman, T.M. 1997. Late Campanian dinosaur biogeography in the western interior of North America. In: Wolberg, D.L., Stump, E., Rosenberg, G.D. (eds.). Dinofest International. Academy of Natural Sciences, Philadelphia. 223–240.

Lehman, T.M., McDowell, F.W., and Connelly, J.N. 2006. First isotopic (U-Pb) age for the Late Cretaceous Alamosaurus vertebrate fauna of west Texas, and its significance as a link between two faunal provinces. Journal of Vertebrate Paleontology 26: 922-928.

Lillegraven, J.A., and McKenna, M.C. 1986. Fossil mammals from the “Mesaverde” Formation (Late Cretaceous, Judithian) of the Bighorn and Wind River Basins, Wyoming, with definitions of Late Cretaceous North American Land Mammal “Ages”. American Museum Novitates 2840: 1-68.

Little, W.W. 1995. The influence of tectonics and eustasy on alluvial architecture, Middle Coniacian through Campanian strata of the Kaiparowits Basin, Utah [unpublished Ph.D. thesis]: University of Colorado, Boulder, 328 p.

Little, W.W. 1997. Tectonic and eustatic controls on cyclical fluvial patterns, Upper Cretaceous strata of the Kaiparowits Basin, Utah. In: Hill, L.M. (ed.) Learning from the Land; Grand Staircase-Escalante National Monument Science Symposium Proceedings. 489-504.

Liu, S.F., Nummedal, D., Yin, P.G., and Luo, H.J. 2005. Linkage of Sevier thrusting episodes and Late Cretaceous foreland basin megasequences across southern Wyoming (USA). Basin Research, 17: 487-506.

Mack, G.H. and Madoff, R.D. 2005. A test of models of fluvial architecture and palaeosol development: Camp Rice Formation (Upper Pliocene-Lower Pleistocene), southern Rio-Grande rift, New Mexico, USA. Sedimentology 52: 191-211.

Mack, G.H., James, W.C., and Monger, H.C. 1993. Classification of paleosols. Geological Society of America Bulletin, 105: 129-136.

Makaske, B. 2001. Anastomosing rivers: a review of their classification, origin and sedimentary products: Earth-Science Reviews 53: 149-196.

Mancini, E.A., and Tew, B.H. 1993. Eustasy versus subsidence: Lower Paleocene depositional sequences from southern Alabama, eastern Gulf Coastal Plain. Geological Society of America Bulletin 105: 3-17.

Martinsen, O.J., Ryseth, A., Helland-Hansen, W., flesche, H., Torkildsen, G., and Idil, S. 1999. Stratigraphic base level and fluvial architecture: Ericson Sandstone (Campanian), Rock Springs Uplift, SW Wyoming, USA. Sedimentology 46: 235-259.

McCarthy, P.J. and Plint, A.G. 2003. Spatial variability of palaeosols across Cretaceous interfluves in the Dunvegan Formation, NE British Columbia, Canada: palaeohydrological, palaeogeomorphological and stratigraphic implications. Sedimentology 50: 1187-1220.

McDowell, F.W., Roldán-Quintana, J., and Connelly, J.N. 2001. Duration of Late Cretaceous-early Tertiary magmatism in east-central Sonora, Mexico. Geological Society of America, Bulletin 113: 521-531.

McFadden, L. D., and Tinsley, J. C. 1985. Rate and depth of pedogenic-carbonate accumulation in soils: Formulation and testing of a compartment model, In: Weide, D. L., (ed.). Soils and Quaternary geology of the southwestern United States. Geological Society of America Special Paper 203: 23–41.

McLaurin, B.T., and Steel, R.J. 2000. Fourth order nonmarine to marine sequences, middle Castlegate Formation, Book Cliffs, Utah. Geology 28: 359-362.

McLaurin, B.T., and Steel, R.J. 2007. Architecture and origin of an amalgamated fluvial sheet sand, lower Castlegate Formation, Book Cliffs, Utah. Sedimentary Geology 197: 291-311.

Metzger, C.A., and Retallack, G.J. 2010. Paleosol record of Neogene climate change in the Australian outback. Australian Journal of Earth Sciences 57: 871-885.

Miall, A.D. 1977. A review of the braided river depositional environment. Earth Science Reviews 13: 1-62.

Miall, A.D. 1978. Lithofacies types and vertical profile models in braided river deposits: a summary. In: A.D. Miall (ed.). Fluvial Sedimentology. Canadian Society of Petrology and Geology Memoirs, 5: 597-604.

Miall, A.D. 1985. Architectural element analysis: a new method of facies analysis applied to fluvial deposits. Earth-Science Reviews 22: 261-308.

Miall, A.D. 1996. The Geology of Fluvial Deposits; Sedimentary Facies, Basin Analysis, and Petroleum Geology. Heidelberg, Springer-Verlag, 582 p.

Miall, A.D., and Arush, M. 2001. The Castlegate Sandstone of the Book Cliffs, Utah: sequence stratigraphy, paleogeography, and tectonic controls. Journal of Sedimentary Research 71: 537-548.

Miall, A.D., Catuneanu, O., Vakarelov, B.K., and Post, R. 2008. The Western Interior Basin. In: Miall, A.D. (ed.). The Sedimentary Basins of the United States and Canada. Sedimentary Basins of the World 5: 329-362.

Molenaar, C.M., and Rice, D.D. 1988. Cretaceous rocks of the Western Interior Basin. In: Sloss, L.L., (ed.). Sedimentary Cover – North American Craton, US: The Geological Society of America, The Geology of North America, vol. D-2: 77-82.

Moore, D.W., and Straub, A.W. 2001. Correlation of Upper Cretaceous and Paleogene (?) rocks beneath the Claron Formation, Crow Creek, western Markagunt Plateau, southwest Utah. Utah Geological Association Publication 30: 75-95.

Nesbitt, H.W., and Young, G.M. 1982. Early Proterozoic climates and plate motions inferred from major element chemistry of lutites. Nature 299: 715–717.

Nichols, D.J. 1997. Palynology and ages of some Upper Cretaceous formations in the Markagunt and northwestern Kaiparowits plateaus, southwestern Utah. United States Geological Survey Bulletin 2153-E: 80-95.

Ogg, J., and Ogg, G. 2008. Late Cretaceous (65-100 Ma time slice). Update to Ogg et al. (2004). Available at www.stratigraphy .org. Accessed 12 April 2010.

Olsen, T., Steel, R., Høgseth, K., Skar, T., and Røe, S., 1995. Sequential architecture in a fluvial succession: sequence stratigraphy in the Upper Cretaceous Mesaverde Group, Price Canyon, Utah. Journal of Sedimentary Research B65: 265-280.

Ori, G.G. and Friend, P.F. 1984. Sedimentary basins formed and carried on active thrust sheets. Geology 12: 475-478.

Palmer, J.J., and Scott, A.J. 1984. Stacked shoreline and shelf sandstone of the La Ventana Tongue (Campanian), northwestern New Mexico. American Association of Petroleum Geologists, Bulletin 68: 74-91.

Parrish, J.T., and Curtis, R.L. 1982. Atmospheric circulation, upwelling, and organic-rich rocks in the Mesozoic and Cenozoic eras. Palaeogeography, Palaeoclimatology, Palaeoecology 40: 31-66.

Payenberg, T.H.D., Braman, D.R., Davis, D.W., and Miall, A.D., 2002. Litho- and chronostratigraphic relationships of the Santonian-Campanian Milk River Formation in southern Alberta and Eagle Formation in Montana utilising stratigraphy, U-Pb geochronology, and palynology. Canadian Journal of Earth Sciences 39: 1553-1577.

Peterson, F., 1969. Cretaceous sedimentation and tectonism in the southeastern Kaiparowits region, Utah. US Geological Survey Open File Report 69-202: 1-259

Peterson, F. and Kirk, A.R. 1977. Correlation of Cretaceous rocks in the San Juan, Black Mesa, Kaiparowits and Henry basins, southern Colorado Plateau. New Mexico Geological Society Field Guide 28:167-178.

Peterson, F., Ryder, R.T., and Law, B.E. 1980. Stratigraphy, sedimentology and regional relationships of the Cretaceous system in the Henry Mountains region, Utah. In: Picard, M.D.

(ed.). Henry Mountain Symposium: Utah Geological Association, Special Publication 8: 151-170.

Plint, A. G. 1991. High-frequency relative sea-level oscillations in Upper cretaceous shelf clastics of the Alberta foreland basin: possible evidence for a glacio-eustatic control?, in Macdonald, D. I. M. (ed.). Sedimentation, tectonics and eustasy: sea-level changes at active margins, International Association of Sedimentologists Special Publication 12: 409–428.

Plint, A. G., and Kreitner, M. A. 2007. Extensive, thin sequence spanning Cretaceous foredeep suggest high-frequency eustatic control: Late Cenomanian, Western Canada foreland basin. Geology 35: 735–738.

Plint, A. G., Hart, B. S., and Donaldson, W. S. 1993. Lithospheric flexure as a control on stratal geometry and facies distribution in Upper Cretaceous rocks of the Alberta foreland basin. Basin Research 5: 69–77.

Pollock, S.L. 1999. Provenance, geometry, lithofacies and age of the Upper Cretaceous Wahweap Formation, Cordilleran foreland basin, southern Utah [unpublished M.Sc. thesis]: New Mexico State University, Las Cruces, 117 p.

Pontēn, A., and Plink-Björklund, P. 2007. Depositional environments in an extensive tide-influenced delta plain, Middle Devonian Gauja Formation, Devonian Baltic Basin. Sedimentology 54: 969-1006.

Reading, H.G. 1996. Sedimentary Environments; Processes, Facies and Stratigraphy: Oxford, U.K., Blackwell Science, 688 p.

Reineck, H.E. 1972. Recognition of Ancient Sedimentary Environments: SEPM, Tulsa, 439 p.

Renne, R.R., Swisher, C.C., Deino, A.L., Kaarner, D.B., Owens, T.L., and DePaolo, D.J. 1998. Intercalibration of standards, absolute ages and uncertainties in 40Ar/39Ar dating. Chemical Geology 145: 117-152.

Roberts, E.M. 2007. Facies architecture and depositional environments of the Upper Cretaceous Kaiparowits Formation, southern Utah: Sedimentary Geology 197: 207-233.

Roberts, E.M., and Hendrix, M.S. 2000. Taphonomy of a petrified forest in the Two Medicine Formation (Campanian) northwest Montana: implications for palinspastic restoration of the Boulder batholith and Elkhorn Mountains Volcanics. Palaios 15: 476-482.

Roberts, E.M., Deino, A.L., and Chan, M.A. 2005. $^{40}Ar/^{39}Ar$ age of the Kaiparowits Formation, southern Utah, and correlation of contemporaneous Campanian strata and vertebrate faunas along the margin of the Western Interior Basin. Cretaceous Research 26: 307-318.

Roberts, S.B., Stanton, R.W., and Flores, R.M. 1994. A debris-flow deposit in alluvial coal-bearing facies, Big-horn Basin, Wyoming, USA: evidence for catastrophic termination of a mire: International Journal of Coal Geology 25: 213-241.

Robinson, R.A.J., and Slingerland, R.L. 1998. Grain-size trends, basin subsidence and sediment supply in the Campanian Castlegate Sandstone and equivalent conglomerates of central Utah. Basin Research 10: 109-127.

Roca, X., and Nadon, G.C. 2007. Tectonic control on the sequence stratigraphy of nonmarine retroarc foreland basin fills: insights from the Upper Jurassic of central Utah. Journal of Sedimentary Research 77: 239-255.

Rogers, R.R. 1990. Taphonomy of three dinosaur bone beds in the Upper Cretaceous Two Medicine Formation: evidence for drought-related mortality. Palaios 5: 394-413.

Rogers, R.R. 1998. Sequence analysis of the Upper Cretaceous Two Medicine and Judith River formations, Montana: nonmarine response to the Claggett and Bearpaw marine cycles. Journal of Sedimentary Research 68: 615-631.

Rogers, R.R., Swisher, C.C., and Horner, J.R. 1993. $^{40}Ar/^{39}Ar$ age and correlation of the nonmarine Two Medicine Formation (Upper Cretaceous), northwestern Montana, U.S.A. Canadian Journal of Earth Sciences 30: 1066-1075.

Rowe, T., Cifelli, R.L., Lehman, T.M., and Weil, A. 1992. The Campanian Terlingua local fauna, with a summary of other vertebrates from the Aguja Formation, Trans-Pecos, Texas. Journal of Vertebrate Paleontology 12: 472-493.

Rowley, P.D., Williams, V.S., Vice, G.S., Maxwell, D.J., Hacker, D.B., Snee, L.W., and Mackin, J.H. 2006. Interim Geologic Map of the Cedar City 30`x60` Quadrangle, Iron and Washington Counties, Utah. Utah Geological Survey Open File Report 476 DM.

Royer, D.L. 1999. Depth to pedogenic carbonate horizon as a paleoprecipitation indicator? Geology 27: 1123–1126.

Royse, F. Jr. 1993. Case of the phantom foredeep: Early Cretaceous in west-central Utah. Geology 21: 133–136.

Sable, E.G., and Hereford, R. 2004. Geologic Map of the Kanab 30’ x 60’ quadrangle, Utah and Arizona. United States Geological Survey, Geologic Investigations Series I-2655.

Sampson, S.D., Loewen, M.A., Gates, T.A., Zanno, L.E., and Kirkland, J.I. 2002. New evidence of dinosaurs and other vertebrates from the Upper Cretaceous Wahweap and

Kaiparowits formations, Grand Staircase-Escalante National Monument, southern Utah. Geological Society of America, Rocky Mountain Section, 54th Annual Meeting, Abstracts with Programs 34: 5.

Sampson, S.D., Loewen, M.A., Farke, A.A., Roberts, E.M., Forster, C.A., Smith, J.A., and Titus, A.L. 2010. New horned dinosaurs from Utah provide evidence of dinosaur endemism. Public Library of Science One 5: 1-12.

Sankey, J.T. 2001. Late Campanian southern dinosaurs, Aguja Formation, Big Bend, Texas. Journal of Paleontology 75: 208-215.

Santucci, V.L., and Kirkland, J.I. 2010. An Overview of National Park Service Paleontological Resources from the Parks and Monuments in Utah. In: D.A. Sprinkel, T.C. Chidsey, Jr., and P.B. Anderson (eds.). Geology of Utah's Parks and Monuments 2010 Utah Geological Association Publication 28 (3rd edition).

Shanley, K.W. and McCabe, P.J. 1991. Predicting facies architecture through sequence stratigraphy – an example from the Kaiparowits Plateau, Utah. Geology 19: 742-745.

Shanley, K.W. and McCabe, P.J. 1993. Alluvial architecture in a sequence stratigraphic framework: a case history from the Upper Cretaceous of southern Utah, USA. Special Publication of the International Association of Sedimentologists 15: 21-56.

Shanley, K.W., and McCabe, P.J. 1995. Sequence stratigraphy of Turonian-Santonian strata, Kaiparowits Plateau, southern Utah, U.S.A.: implications for regional correlation and foreland basin evolution. In: Van Wagoner, J.C., and Bertram, G.T. (eds.). Sequence Stratigraphy of Foreland Basin Deposits: Outcrop and Subsurface examples from the Cretaceous of North America. American Association of Petroleum Geologists Memoir 64: 103-136.

Sheldon, N.D., and Tabor, N.J. 2009. Quantitative paleoenvironmental and palaeoclimatic construction using paleosols. Earth Science Reviews 95: 1-52.

Simpson, E.L., Hilbert-Wolf, H.L., Simpson, W.S., Tindall, S.E., Bernard, J.J., Jenesky, T.A., and Wizevich, M.C. 2008. The interaction of aeolian and fluvial processes during deposition of the Upper Cretaceous capping sandstone member, Wahweap Formation, Kaiparowits Basin, Utah, U.S.A. Palaeogeography, Palaeoclimatology, Palaeoecology 270: 19-28.

Simpson, E.L., Wizevich, M.C., Hilbert-Wolf, H.L., Tindall, S.E., Bernard, J.J., and Simpson, W.S. 2009. An Upper Cretaceous sag pond deposit: implications for recognition of local seismicity and surface rupture along the Kaibab monocline, Utah. Geology 37: 967-970.

Simpson, E.L., Hilbert-Wolf, H.L., Wizevich, M.C., Tindall, S.E., Fasinski, B.R., Storm, L.P., and Needle, M.D. 2010. Predatory digging behaviour by dinosaurs. Geology 38: 699-702.

Simpson, W.S., Simpson, E.L., Wizevich, M.C., Malenda, H.F., Hilbert-Wolf, H.L., and Tindall, S.E. 2010. A preserved Late Cretaceous biological soil crust in the capping sandstone member, Wahweap Formation, Grand Staircase-Escalante National Monument, Utah: paleoclimatic implications. Sedimentary Geology 230: 139-145.

Singer, A. 1980. The palaeoclimatic interpretation of clay minerals in soils and weathering profiles. Earth Science Reviews 15: 303-326.

Slingerland, R., and Smith, N.D. 2004. River avulsions and their deposits: Annual Review of Earth and Planetary Sciences 32: 257-285.

Smith, C. 1983. Geology, depositional environments and coal resources of the Mt. Pennell 2 NW Quadrangle, Garfield County, Utah. Brigham Young University Geological Studies 30: 145-169.

Smith, N.D., Cross, T.A., Dufficy, J.P, and Clough, S.R. 1989. Anatomy of an avulsion: Sedimentology 36: 1-23.

Smith, R.L. 1960. Ash flows. Geological Society of America, Bulletin 71: 795-842.

Soil Survey Staff. 1999. Soil Taxonomy: A Basic System of Soil Classification for Making and Interpreting Soil Surveys (2nd edition). United States Department of Agriculture Natural Resource Conservation Service, Agriculture Handbook 436, 871 p.

Spivey, R.C. 1940. Bentonite in southwestern South Dakota. South Dakota Geological Survey, Report of Investigations 36: 1-56.

Srivastava, P., Rajak, M.K., Sinha, R., Pal, D.K., and Bhattacharyya, T. 2010. A high-resolution micromorphological record of the Late Quaternary paleosols from GangaeYamuna interfluve: stratigraphic and paleoclimatic implications. Quaternary International 227: 127-142.

Sullivan, R.M., and Lucas, S.G. 2003. The Kirtlandian, a new Land-Vertebrate "Age" for the Late Cretaceous of Western North America. In: Lucas, S. G., Berglof, W., and Ulmer-Scholle, D. (Eds.). Geology of the Zuni Plateau, New Mexico Geological Society Guidebook, 54: 369-377.

Sullivan R.M., and Lucas, S.G. 2006. The Kirtlandian land-vertebrate ‘‘age’’ – faunal composition, temporal position and biostratigraphic correlation in the nonmarine Upper Cretaceous of western North America. New Mexico Museum of Natural History Science Bulletin 35: 7–29.

Tankard, A. J.1986. Depositional response to foreland deformation in the Carboniferous of eastern Kentucky. Bulletin of the American Association of Petroleum Geologists 70: 853–868.

Therrien, F. 2005. Palaeoenvironments of the latest Cretaceous (Maastrichtian) dinosaurs of Romania: insights from fluvial deposits and paleosols of the Transylvanian and Hateg basins: Palaeogeography, Palaeoclimatology, Palaeoecology 218: 15-56.

Thomas, R.G., Smith, D.G., Wood, J.M., Visser, J., Calverley-Range, E.A., and Koster, E.H. 1987. Inclined heterolithic stratification – terminology, description, interpretation and significance. Sedimentary Geology 53: 123-179.

Thomas, R.G., Eberth, D.A., Deino, A.L., and Robinson, D. 1990. Composition, radioisotopic ages, and potential significance of an altered volcanic ash (bentonite) from the Upper Cretaceous Judith River Formation, Dinosaur Provincial Park, southern Alberta, Canada. Cretaceous Research 11: 125-162.

Tilton, T. L. 1991. Upper Cretaceous Stratigraphy of the Southern Paunsaugunt Plateau, Kane Country, Utah. Ph. D. dissertation, University of Utah, Salt Lake City, 162 p.

Tindall, S.E., and Davis, G.H. 1999. Monocline development by oblique-slip fault-propagation folding: the East Kaibab monocline, Colorado Plateau, Utah. Journal of Structural Geology 21: 1303-1320.

Tindall, S.E., Storm, L.P., Jenesky, T.A., and Simpson, E.L. 2010. Growth faults in the Kaiparowits Basin, Utah, pinpoint initial Laramide deformation in the western Colorado Plateau. Lithosphere 2: 221-231.

Titus, A.L., Powell, J.D., Roberts, E.M., Sampson, S.D., Pollock, S.L., Kirkland, J.I., and Albright, L.B. 2005. Late Cretaceous stratigraphy, depositional environments and macrovertebrate paleontology of the Kaiparowits Plateau, Grand Staircase-Escalante National Monument, Utah. In: Pederson, J., Dehler, C.M. (Eds.). Interior Western United States. Geological Society of America Field Guide 6: 101-128.

Titus, A.L., Knell, M.J., Wiersma, J.P., and Getty, M.A. 2008. First report of the hyper-giant Cretaceous crocodylian Deinosuchus from Utah. Geological Society of America Abstracts with Programs 40 (1): 58.

Törnqvist, T.E. 1994. Middle and late Holocene avulsion history of the River Rhine (Rhine-Meuse delta, Netherlands). Geology 22: 711-714.

Tristán-González, M., Aguirre-Díaz, G.J., Labarthe-Hernández, G.L., Torres-Hernández, J.R., and Bellon, H. 2009. Post-Laramide and pre-Basin and Range deformation and implications for Paleogene (55–25 Ma) volcanism in central Mexico: A geological basis for a volcano-tectonic stress model. Tectonophysics 471: 136-152.

Van Wagoner, J.C. 1995. Sequence stratigraphy and marine to nonmarine facies architecture of foreland basin strata, Book Cliffs, Utah, U.S.A. In: Van Wagoner, J.C., Bertram, G.T. (eds.). Sequence Stratigraphy of Foreland Basin Deposits: Outcrop and Subsurface examples from the Cretaceous of North America: American Association of Petroleum Geologists Memoir 64: 137-223.

Van Wagoner, J.C., Mitchum, R.M., Jr., Campion, K.M., and Rahmanian, V.D. 1990, Siliciclastic Sequence Stratigraphy in Well Logs, Core and Outcrops: Concepts for High-Resolution Correlation of Time and Facies. American Association of Petroleum Geologists, Methods in Exploration Series no. 7: 1-55.

Weimer, R.J. 1960. Upper Cretaceous stratigraphy, Rocky Mountain area: American Association of Petroleum Geologists, Bulletin 44: 1-20.

Weimer, R.J. 1986. Relationship of unconformities, tectonics, and sea level changes in the Cretaceous of the Western Interior, United States. In: Peterson, J.A. (ed.). Paleotectonics and Sedimentation in the Rocky Mountain Region, United States: American Association of Petroleum Geologists Memoir 41: 397-422.

Weissmann, G.S., Mount, J.F., and Fogg, G.E. 2002. Glacially driven cycles of accumulation space and sequence stratigraphy of a stream-dominated alluvial fan, San Joaquin Valley, California, U.S.A. Journal of Sedimentary Research 72: 240-251.

Wernicke, B., Axen, G.J., and Snow, J.K. 1988. Basin and range extensional tectonics at the latitude of Las Vegas, Nevada. Geological Society of America, Bulletin 100: 1738-1757.

White, T, Furlong, K & Arthur, M. 2002. Forebulge migration in the Cretaceous Western Interior basin of the central United States. Basin Research 14: 43-54.

Willis, A. 2000. Tectonic control of nested sequence architecture in the Sego Sandstone, Nelson Formation and Upper Castlegate Sandstone (Upper Cretaceous), Sevier Foreland Basin, Utah, USA. Sedimentary Geology 136: 277-317.

Willis, B.J., Bhattacharya, J.B., Gabel, S.L., and White, C.D. 1999. Architecture of a tide-influenced delta in the Frontier Formation of central Wyoming, U.S.A. Sedimentology 46: 667-688.

Wright, V.P., and Marriott, S.B. 1993. The sequence stratigraphy of fluvial depositional systems: the role of floodplain sediment storage. Sedimentary Geology 86: 203-210.

Wright, V.P. and Tucker, M.E. 1991. Calcretes: an introduction. In: Wright, V.P. (ed.). Calcretes. Reprint Series 2, International Association of Sedimentologists. Blackwell Science, Oxford; 1–22.

Yan, Z., Wang, Z.Q., Wang, T., Yan, Q.R., Xiao, W.J., and Li, J.L. 2006. Provenance and tectonic setting of clastic deposits in the Devonian Xicheng Basin, Qinling orogen, central China. Journal of Sedimentary Research 76: 557–574.

Yingling, V.L., and Heller, P.L. 1992. Timing and record of foreland sedimentation during the initiation of the Sevier orogenic belt in central Utah Basin Research 4: 279-290.

Yoshida, S., Willis, A., and Miall, A.D. 1996. Tectonic control of nested sequence architecture in the Castlegate Sandstone (Upper Cretaceous), Book Cliffs, Utah. Journal of Sedimentary Research 66: 737–748.

Young, J.R., McMillan, N.J., Lawton, T.F., Esser, R.P. 2000. Volcanology, geochemistry, and structural geology of the Upper Cretaceous Hidalgo Formation, southwestern New Mexico. In: Lawton, T. F., McMillan, N. J., McLemore, V. T. (eds.). Southwest Passage – A Trip through the Phanerozoic, New Mexico Geological Society Guidebook 51: 149-156.

Zanno, L.E. 2005. Dinosaur diversity and biogeographical implications of the Kaiparowits Formation (Late Campanian), Grand Staircase-Escalante National Monument, southern Utah. Geological Society of America, Abstracts with Programs 37: 115.

APPENDIX A:

Compositional data for sandstone thin sections

Sample	Qtz count	FS count	Sed lith count	Matrix count	Cement count	Total count	Qtz % norm	Sed lith % norm	FS % norm	Matrix %	Member	FA
190606CC1	122	9	11	9	149	350	85.915	7.746	6.338	2.571429	lower	major tabular sandstone
190606CC2	144	2	22	0	170	353	85.714	13.095	1.190	0	lower	major tabular sandstone
190606CC4	201	14	35	54	70	400	80.400	14.000	5.600	13.5	upper	major tabular sandstone
190606CC5	213	20	46	23	36	350	76.344	16.487	7.168	6.571429	upper	major tabular sandstone
200606CC2	166	4	80	75	3	352	66.400	32.000	1.600	21.30682	upper	major tabular sandstone
090606H12 1	82	95	127	42	8	362	26.974	41.776	31.250	11.60221	lower	major tabular sandstone
110606H12 6	123	53	88	20	65	350	46.591	33.333	20.076	5.714286	lower	major tabular sandstone
110606H12 2	187	6	40	66	28	340	80.258	17.167	2.575	19.41176	upper	major tabular sandstone
210606H12 2	176	4	83	76	19	366	66.920	31.559	1.521	20.76503	upper	major tabular sandstone
210606B2 1	125	31	71	25	93	352	55.066	31.278	13.656	7.102273	lower	minor tabular sandstone
210606B2 2	65	37	24	110	114	369	51.587	19.048	29.365	29.8103	lower	minor tabular sandstone
210606B2 3	68	59	32	9	170	350	42.767	20.126	37.107	2.571429	lower	major tabular sandstone
170606B2 1	80	62	39	25	147	353	44.199	21.547	34.254	7.082153	lower	major tabular sandstone
210606B2 4	114	64	87	26	50	350	43.019	32.830	24.151	7.428571	lower	major tabular sandstone
210606B2 5	107	35	48	85	60	357	56.316	25.263	18.421	23.80952	lower	inclined interbedded sandstone and mudrock
210606B2 7	85	5	22	159	18	350	75.893	19.643	4.464	45.42857	middle	minor tabular sandstone
210606B2 8	96	17	42	67	115	353	61.935	27.097	10.968	18.98017	middle	inclined interbedded sandstone and mudrock
210606B2 9	61	7	39	132	94	350	57.009	36.449	6.542	37.71429	middle	minor tabular sandstone
210606B2 10	167	2	76	72	56	379	68.163	31.020	0.816	18.99736	upper	major tabular sandstone
210606B2 11	174	16	85	44	24	351	63.273	30.909	5.818	12.53561	upper	major tabular sandstone
210606B2 13	172	4	38	68	58	350	80.374	17.757	1.869	19.42857	upper	major tabular sandstone
210606B2 14	182	5	34	41	72	351	82.353	15.385	2.262	11.68091	upper	major tabular sandstone
210606B2 MF	134	3	43	37	97	346	74.444	23.889	1.667	10.69364	upper	major tabular sandstone
210606B216 1	158	10	30	26	128	356	79.798	15.152	5.051	7.303371	upper	single-storey lenticular sandstone

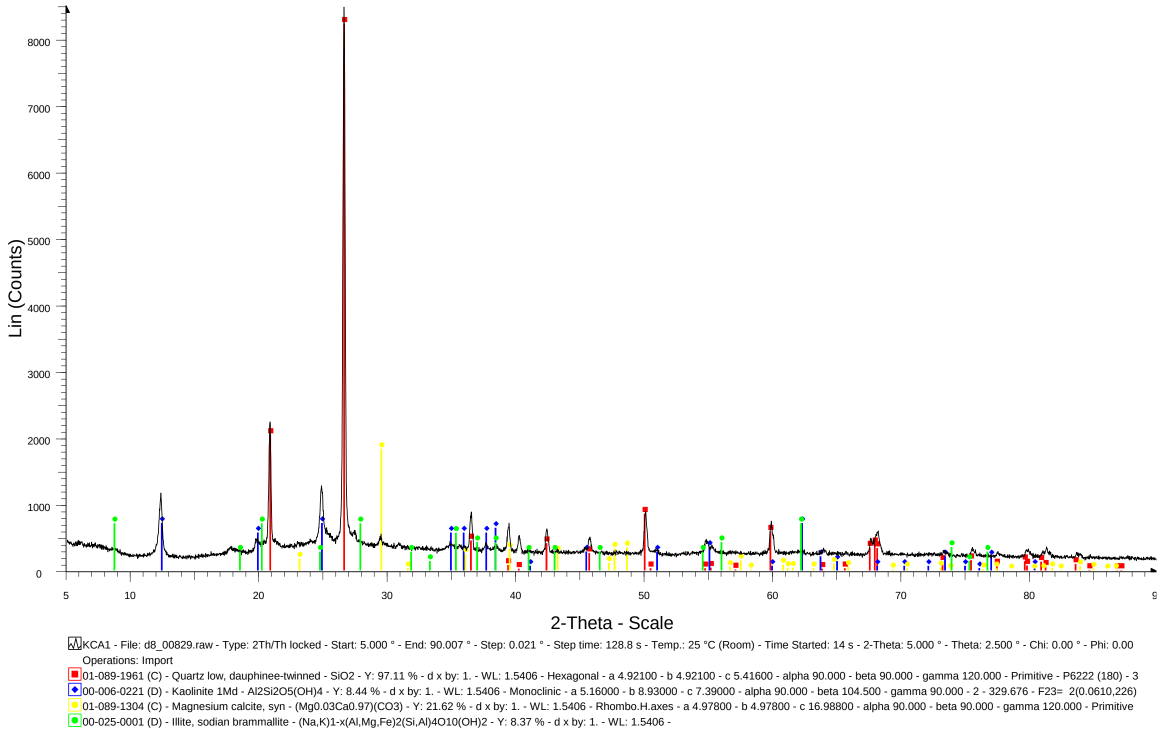
210606B216 2	139	8	57	76	56	357	68.137	27.941	3.922	21.28852	upper	single-storey lenticular sandstone
210606B216 5	215	1	31	7	95	352	87.045	12.551	0.405	1.988636	upper	major tabular sandstone
210606B216 6	183	4	19	3	112	351	88.835	9.223	1.942	0.854701	upper	major tabular sandstone
040606BS 3	117	29	67	59	54	344	54.930	31.455	13.615	17.15116	lower	major tabular sandstone
230606BS 1	154	39	109	11	24	350	50.993	36.093	12.914	3.142857	lower	major tabular sandstone
170606BS 4	105	11	92	21	100	351	50.481	44.231	5.288	5.982906	lower	minor tabular sandstone
170606 BS 2	218	3	76	17	36	350	73.401	25.589	1.010	4.857143	middle	single-storey lenticular sandstone
040606BS 1	212	1	42	75	13	352	83.137	16.471	0.392	21.30682	upper	multi-storey lenticular sandstone
220606CB 10	152	56	43	69	21	365	60.558	17.131	22.311	18.90411	lower	single-storey lenticular sandstone
210606CB 11	145	49	47	9	90	350	60.166	19.502	20.332	2.571429	lower	major tabular sandstone
220606CB 12	88	56	67	62	70	350	41.706	31.754	26.540	17.71429	lower	major tabular sandstone
220606CB 14	111	51	17	83	81	367	62.011	9.497	28.492	22.6158	lower	minor tabular sandstone
220606CB 9	184	36	48	18	113	399	68.657	17.910	13.433	4.511278	middle	inclined interbedded sandstone and mudrock
220606CB 1	176	10	44	58	62	356	76.522	19.130	4.348	16.29213	upper	multi-storey lenticular sandstone
220606CB 2	168	17	40	30	80	345	74.667	17.778	7.556	8.695652	upper	major tabular sandstone
220606CB 3	196	23	62	56	6	351	69.751	22.064	8.185	15.95442	upper	major tabular sandstone
220606CB 4	148	3	84	61	53	350	62.979	35.745	1.277	17.42857	upper	major tabular sandstone
220606CB 5	161	4	68	12	124	336	69.099	29.185	1.717	3.571429	upper	single-storey lenticular sandstone
220606CB 6	140	8	24	26	156	366	81.395	13.953	4.651	7.103825	upper	major tabular sandstone
220606CB 7	159	4	42	113	29	352	77.561	20.488	1.951	32.10227	upper	major tabular sandstone
220606CB 8	140	12	42	59	59	312	72.165	21.649	6.186	18.91026	upper	major tabular sandstone

APPENDIX B:

XRF and XRD data for palaeosols

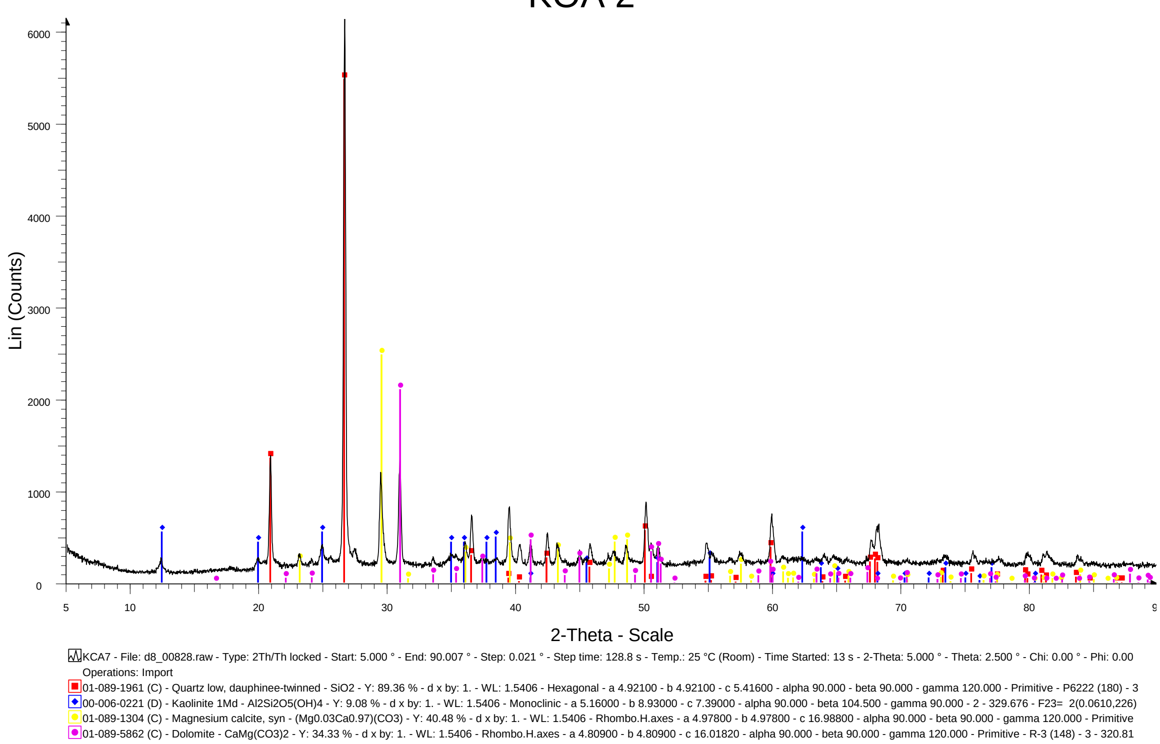
	SiO2	Al2O3	Fe2O3	FeO	MnO	MgO	CaO	Na2O	K2O	TiO2	P2O5	Cr2O3	NiO	TOTAL	LOI
SC1	79.2	8.94	0.14	1.12	0.01	2.41	5.71	0.08	1.34	0.46	0.28	0.01	0	99.71	9.04
SC3	78.06	10.95	0.11	0.88	0.01	2.74	4.59	0.1	1.55	0.56	0.2	0.01	0	99.77	9.21
KOA1	79.44	13.4	0.24	1.9	0.01	0.71	1.76	0.06	1.45	0.7	0.06	0.01	0	99.74	6.72
KOA2	71.68	10.78	0.23	1.82	0.02	3.14	9.31	0.08	1.46	0.59	0.16	0.01	0	99.28	13.37
KCB3	68.58	12.56	0.15	1.22	0.04	3.06	11.21	0.08	1.71	0.57	0.23	0.02	0	99.43	14.76
KCB2	71.46	13.36	0.54	4.36	0.02	2.99	4.27	0.1	2.09	0.64	0.19	0.01	0	100.03	17.87
KC3	80.89	11.15	0.17	1.39	0.01	1.21	2.79	0.09	1.43	0.6	0.15	0.01	0	99.9	7.01
KC2	63.56	10.3	0.96	7.77	0.1	3.59	10.33	0.07	1.4	0.52	0.23	0.02	0	98.86	15.82
BSB1	69.98	10.68	0.34	2.73	0.06	4.35	9.57	0.1	1.42	0.54	0.3	0.01	0	100.08	10.44
BSB2	70.96	13.52	0.54	4.37	0.03	3.1	3.87	0.14	2.2	0.64	0.17	0.01	0	99.55	11.11
BS3	76.27	14.24	0.45	3.6	0.01	1.35	0.94	0.2	1.46	0.65	0.05	0.01	0	99.23	8.93
BS5	79.06	12.85	0.4	3.26	0.01	1.27	0.48	0.36	1.65	0.71	0.06	0.02	0	100.13	7.31
CB1	80.09	12.24	0.29	2.36	0.01	0.7	0.32	0.87	2.1	0.73	0.04	0	0	99.75	0.25
CB3	78.57	13.53	0.33	2.64	0.01	0.87	0.37	0.8	2.02	0.76	0.03	0	0	99.93	7.28

KOA1



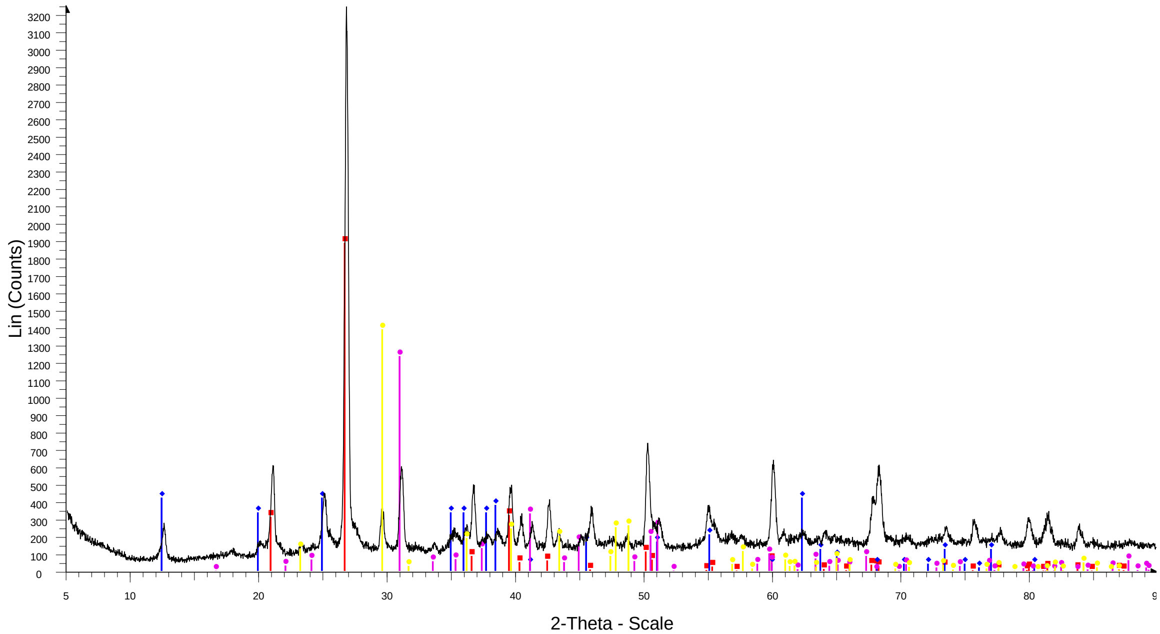
Lin (Counts)
2-Theta - Scale
KCA1 - File: d8_00829.raw - Type: 2Th/Th locked - Start: 5.000 ° - End: 90.007 ° - Step: 0.021 ° - Step time: 128.8 s - Temp.: 25 °C (Room) - Time Started: 14 s - 2-Theta: 5.000 ° - Theta: 2.500 ° - Chi: 0.00 ° - Phi: 0.00
Operations: Import
01-089-1961 (C) - Quartz low, dauphinee-twinned - SiO2 - Y: 97.11 % - d x by: 1. - WL: 1.5406 - Hexagonal - a 4.92100 - b 4.92100 - c 5.41600 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - P6222 (180) - 3
00-006-0221 (D) - Kaolinite 1Md - Al2Si2O5(OH)4 - Y: 8.44 % - d x by: 1. - WL: 1.5406 - Monoclinic - a 5.16000 - b 8.93000 - c 7.39000 - alpha 90.000 - beta 104.500 - gamma 90.000 - 2 - 329.676 - F23= 2(0.0610,226)
01-089-1304 (C) - Magnesium calcite, syn - (Mg0.03Ca0.97)(CO3) - Y: 21.62 % - d x by: 1. - WL: 1.5406 - Rhombo.H.axes - a 4.97800 - b 4.97800 - c 16.98800 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive
00-025-0001 (D) - Illite, sodian brammallite - (Na,K)1-x(Al,Mg,Fe)2(Si,Al)4O10(OH)2 - Y: 8.37 % - d x by: 1. - WL: 1.5406 -

KOA 2



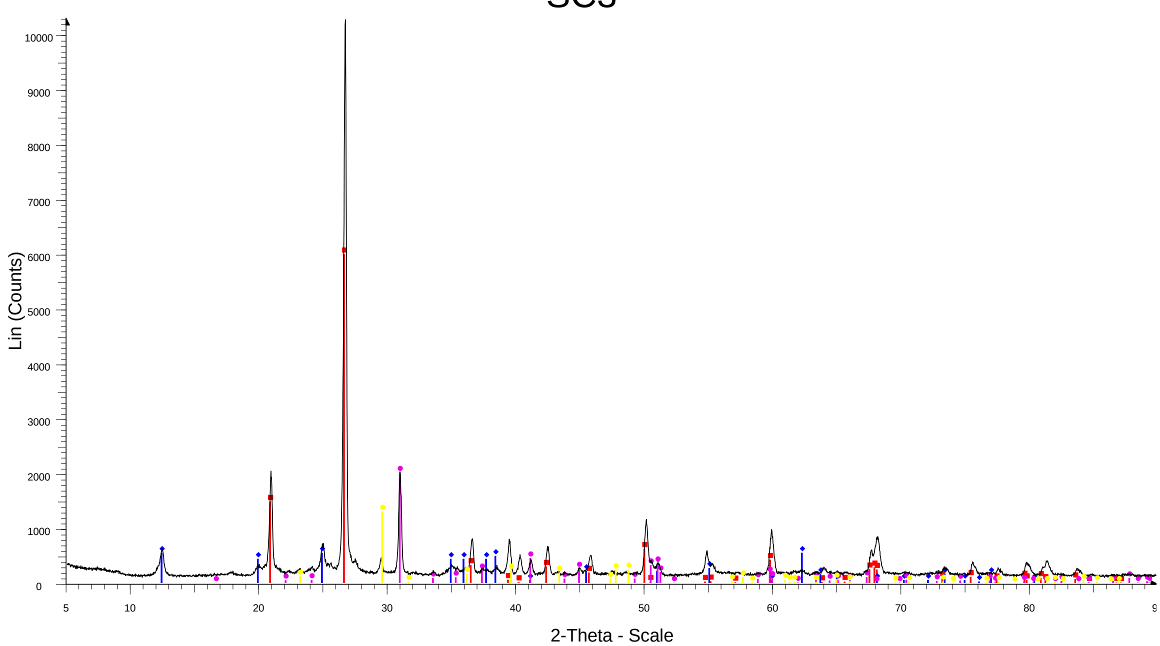
Lin (Counts)
2-Theta - Scale
KCA7 - File: d8_00828.raw - Type: 2Th/Th locked - Start: 5.000 ° - End: 90.007 ° - Step: 0.021 ° - Step time: 128.8 s - Temp.: 25 °C (Room) - Time Started: 13 s - 2-Theta: 5.000 ° - Theta: 2.500 ° - Chi: 0.00 ° - Phi: 0.00
Operations: Import
01-089-1961 (C) - Quartz low, dauphinee-twinned - SiO2 - Y: 89.36 % - d x by: 1. - WL: 1.5406 - Hexagonal - a 4.92100 - b 4.92100 - c 5.41600 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - P6222 (180) - 3
00-006-0221 (D) - Kaolinite 1Md - Al2Si2O5(OH)4 - Y: 9.08 % - d x by: 1. - WL: 1.5406 - Monoclinic - a 5.16000 - b 8.93000 - c 7.39000 - alpha 90.000 - beta 104.500 - gamma 90.000 - 2 - 329.676 - F23= 2(0.0610,226)
01-089-1304 (C) - Magnesium calcite, syn - (Mg0.03Ca0.97)(CO3) - Y: 40.48 % - d x by: 1. - WL: 1.5406 - Rhombo.H.axes - a 4.97800 - b 4.97800 - c 16.98800 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive
01-089-5862 (C) - Dolomite - CaMg(CO3)2 - Y: 34.33 % - d x by: 1. - WL: 1.5406 - Rhombo.H.axes - a 4.80900 - b 4.80900 - c 16.01820 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - R-3 (148) - 3 - 320.81

SC1



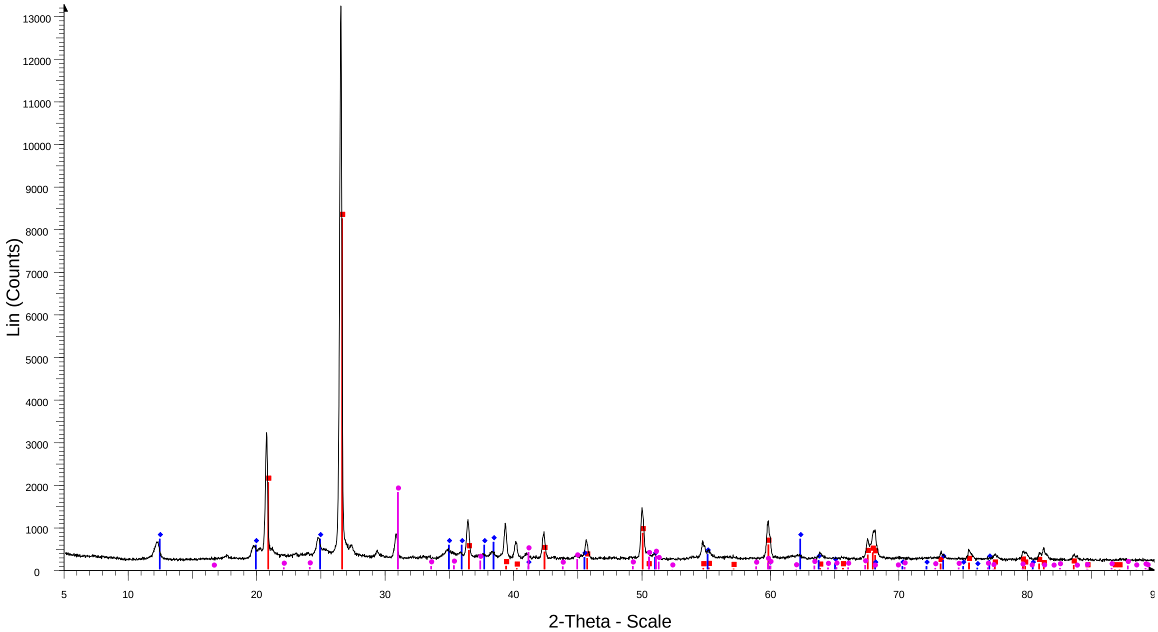
SC1 - File: d8_00826.raw - Type: 2Th/Th locked - Start: 5.000 ° - End: 90.007 ° - Step: 0.021 ° - Step time: 128.8 s - Temp.: 25 °C (Room) - Time Started: 14 s - 2-Theta: 5.000 ° - Theta: 2.500 ° - Chi: 0.00 ° - Phi: 0.00 °
Operations: Import
01-087-2096 (A) - Quartz - SiO2 - Y: 58.12 % - d x by: 1. - WL: 1.5406 - Hexagonal - a 4.91270 - b 4.91270 - c 5.40450 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - P3221 (154) - 3 - 112.961 - I/Ic PDF 4.1
00-006-0221 (D) - Kaolinite 1Md - Al2Si2O5(OH)4 - Y: 12.99 % - d x by: 1. - WL: 1.5406 - Monoclinic - a 5.16000 - b 8.93000 - c 7.39000 - alpha 90.000 - beta 104.500 - gamma 90.000 - 2 - 329.676 - F23= 2(0.0610,226
01-083-1530 (A) - Dolomite - CaMg(CO3)2 - Y: 38.05 % - d x by: 1. - WL: 1.5406 - Rhombo.H.axes - a 4.81200 - b 4.81200 - c 16.02000 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - R-3 (148) - 3 - 321.25
01-089-1305 (C) - Magnesium calcite, syn - (Mg0.06Ca0.94)(CO3) - Y: 42.82 % - d x by: 1. - WL: 1.5406 - Rhombo.H.axes - a 4.96300 - b 4.96300 - c 16.95700 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive

SC3



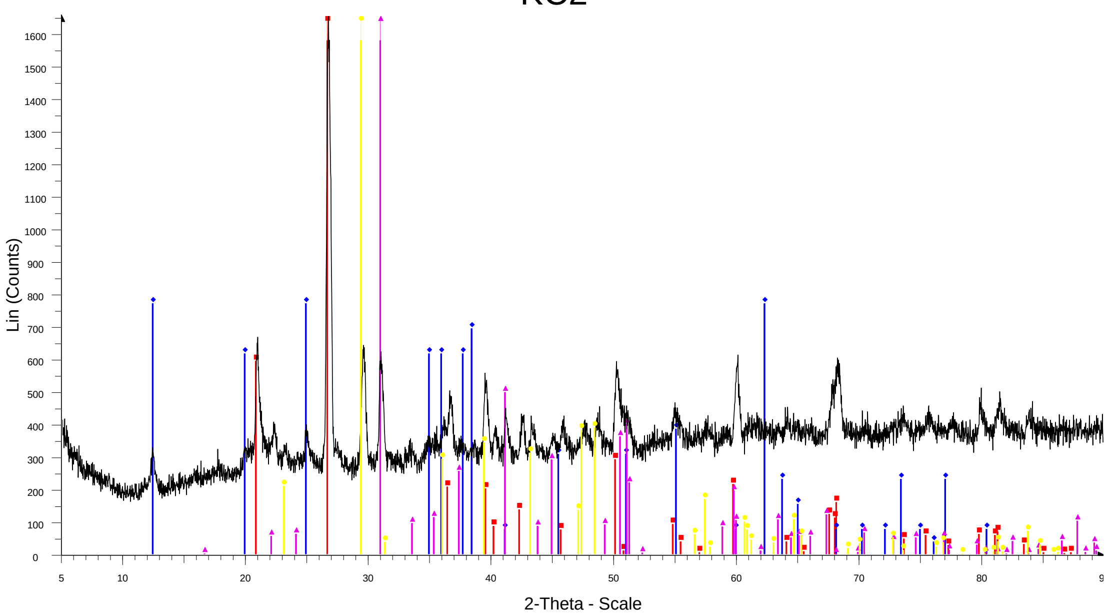
SC3 - File: d8_00825.raw - Type: 2Th/Th locked - Start: 5.000 ° - End: 90.007 ° - Step: 0.021 ° - Step time: 128.8 s - Temp.: 25 °C (Room) - Time Started: 15 s - 2-Theta: 5.000 ° - Theta: 2.500 ° - Chi: 0.00 ° - Phi: 0.00 °
Operations: Import
01-089-1961 (C) - Quartz low, dauphinee-twinned - SiO2 - Y: 58.41 % - d x by: 1. - WL: 1.5406 - Hexagonal - a 4.92100 - b 4.92100 - c 5.41600 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - P6222 (180) - 3
00-006-0221 (D) - Kaolinite 1Md - Al2Si2O5(OH)4 - Y: 5.36 % - d x by: 1. - WL: 1.5406 - Monoclinic - a 5.16000 - b 8.93000 - c 7.39000 - alpha 90.000 - beta 104.500 - gamma 90.000 - 2 - 329.676 - F23= 2(0.0610,226)
01-089-5862 (C) - Dolomite - CaMg(CO3)2 - Y: 19.65 % - d x by: 1. - WL: 1.5406 - Rhombo.H.axes - a 4.80900 - b 4.80900 - c 16.01820 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - R-3 (148) - 3 - 320.81
01-089-1305 (C) - Magnesium calcite, syn - (Mg0.06Ca0.94)(CO3) - Y: 12.74 % - d x by: 1. - WL: 1.5406 - Rhombo.H.axes - a 4.96300 - b 4.96300 - c 16.95700 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive

KC3



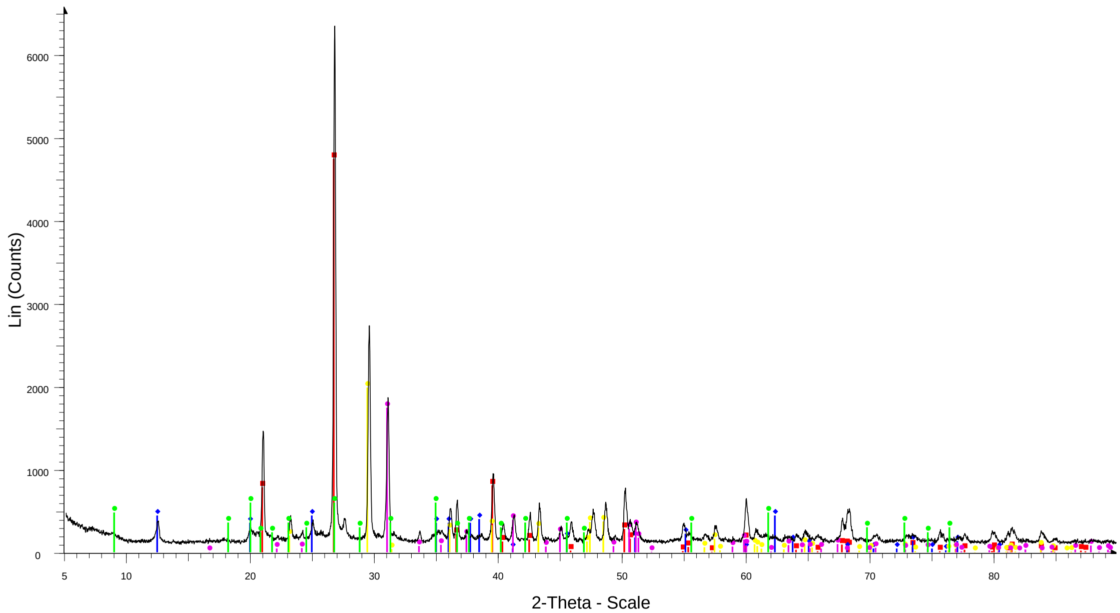
KC3 - File: d8_00821.raw - Type: 2Th/Th locked - Start: 5.000 ° - End: 90.007 ° - Step: 0.021 ° - Step time: 128.8 s - Temp.: 25 °C (Room) - Time Started: 14 s - 2-Theta: 5.000 ° - Theta: 2.500 ° - Chi: 0.00 ° - Phi: 0.00 °
Operations: Import
01-089-1961 (C) - Quartz low, dauphinee-twinned - SiO2 - Y: 62.33 % - d x by: 1. - WL: 1.5406 - Hexagonal - a 4.92100 - b 4.92100 - c 5.41600 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - P6222 (180) - 3
00-006-0221 (D) - Kaolinite 1Md - Al2Si2O5(OH)4 - Y: 5.43 % - d x by: 1. - WL: 1.5406 - Monoclinic - a 5.16000 - b 8.93000 - c 7.39000 - alpha 90.000 - beta 104.500 - gamma 90.000 - 2 - 329.676 - F23= 2(0.0610,226)
01-089-5862 (C) - Dolomite - CaMg(CO3)2 - Y: 13.71 % - d x by: 1. - WL: 1.5406 - Rhombo.H.axes - a 4.80900 - b 4.80900 - c 16.01820 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - R-3 (148) - 3 - 320.81

KC2



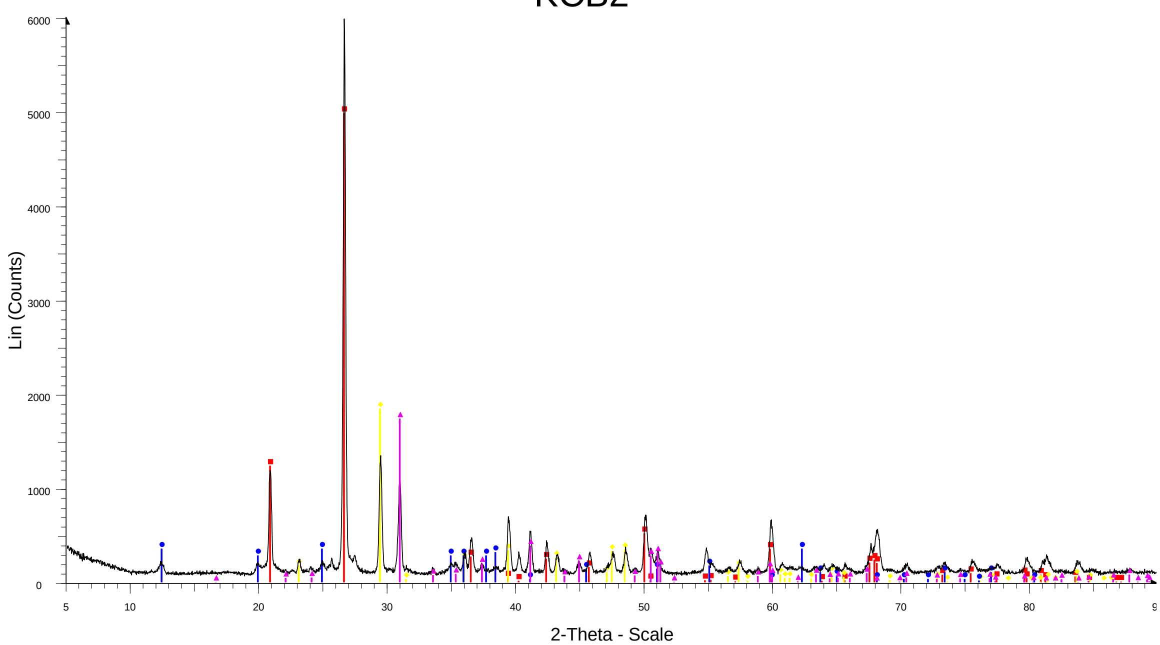
KC2 - File: d8_00823.raw - Type: 2Th/Th locked - Start: 5.000 ° - End: 90.007 ° - Step: 0.021 ° - Step time: 128.8 s - Temp.: 25 °C (Room) - Time Started: 14 s - 2-Theta: 5.000 ° - Theta: 2.500 ° - Chi: 0.00 ° - Phi: 0.00 °
Operations: Import
01-088-2487 (C) - Quartz low - SiO2 - Y: 169.24 % - d x by: 1. - WL: 1.5406 - Hexagonal - a 4.93000 - b 4.93000 - c 5.38500 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - P3221 (154) - 3 - 113.347 - I/Ic PD
00-006-0221 (D) - Kaolinite 1Md - Al2Si2O5(OH)4 - Y: 46.61 % - d x by: 1. - WL: 1.5406 - Monoclinic - a 5.16000 - b 8.93000 - c 7.39000 - alpha 90.000 - beta 104.500 - gamma 90.000 - 2 - 329.676 - F23= 2(0.0610,226
01-086-2339 (C) - Calcite - Ca(CO3) - Y: 126.45 % - d x by: 1. - WL: 1.5406 - Rhombo.H.axes - a 4.98400 - b 4.98400 - c 17.12100 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - R-3c (167) - 6 - 368.312 - I/
01-089-5862 (C) - Dolomite - CaMg(CO3)2 - Y: 134.26 % - d x by: 1. - WL: 1.5406 - Rhombo.H.axes - a 4.80900 - b 4.80900 - c 16.01820 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - R-3 (148) - 3 - 320.8

KCB3



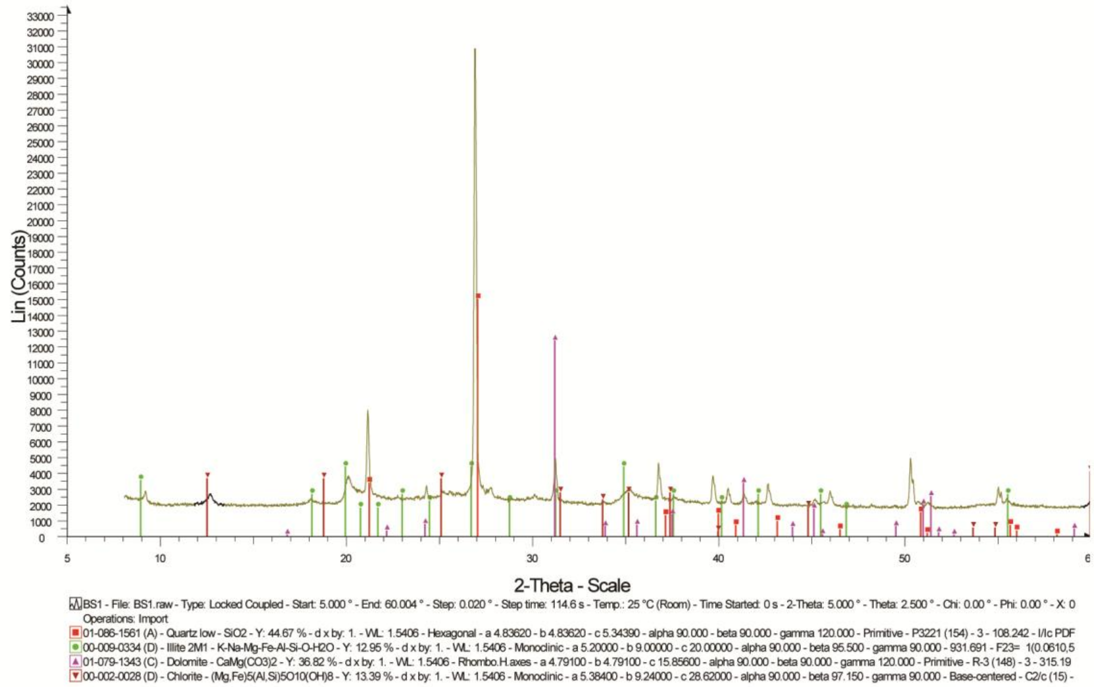
KCB3 - File: d8_00820.raw - Type: 2Th/Th locked - Start: 5.000 ° - End: 90.007 ° - Step: 0.021 ° - Step time: 128.8 s - Temp.: 25 °C (Room) - Time Started: 15 s - 2-Theta: 5.000 ° - Theta: 2.500 ° - Chi: 0.00 ° - Phi: 0.00
Operations: Import
01-087-2096 (A) - Quartz - SiO2 - Y: 74.79 % - d x by: 1. - WL: 1.5406 - Hexagonal - a 4.91270 - b 4.91270 - c 5.40450 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - P3221 (154) - 3 - 112.961 - I/Ic PDF 4.1
00-006-0221 (D) - Kaolinite 1Md - Al2Si2O5(OH)4 - Y: 6.97 % - d x by: 1. - WL: 1.5406 - Monoclinic - a 5.16000 - b 8.93000 - c 7.39000 - alpha 90.000 - beta 104.500 - gamma 90.000 - 2 - 329.676 - F23= 2(0.0610,226)
01-086-2339 (C) - Calcite - Ca(CO3) - Y: 31.31 % - d x by: 1. - WL: 1.5406 - Rhombo.H.axes - a 4.98400 - b 4.98400 - c 17.12100 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - R-3c (167) - 6 - 368.312 - I/Ic
01-089-5862 (C) - Dolomite - CaMg(CO3)2 - Y: 27.47 % - d x by: 1. - WL: 1.5406 - Rhombo.H.axes - a 4.80900 - b 4.80900 - c 16.01820 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - R-3 (148) - 3 - 320.81
00-009-0334 (D) - Illite 2M1 - K-Na-Mg-Fe-Al-Si-O-H2O - Y: 9.43 % - d x by: 1. - WL: 1.5406 - Monoclinic - a 5.20000 - b 9.00000 - c 20.00000 - alpha 90.000 - beta 95.500 - gamma 90.000 - 931.691 - F23= 1(0.0610,56

KCB2

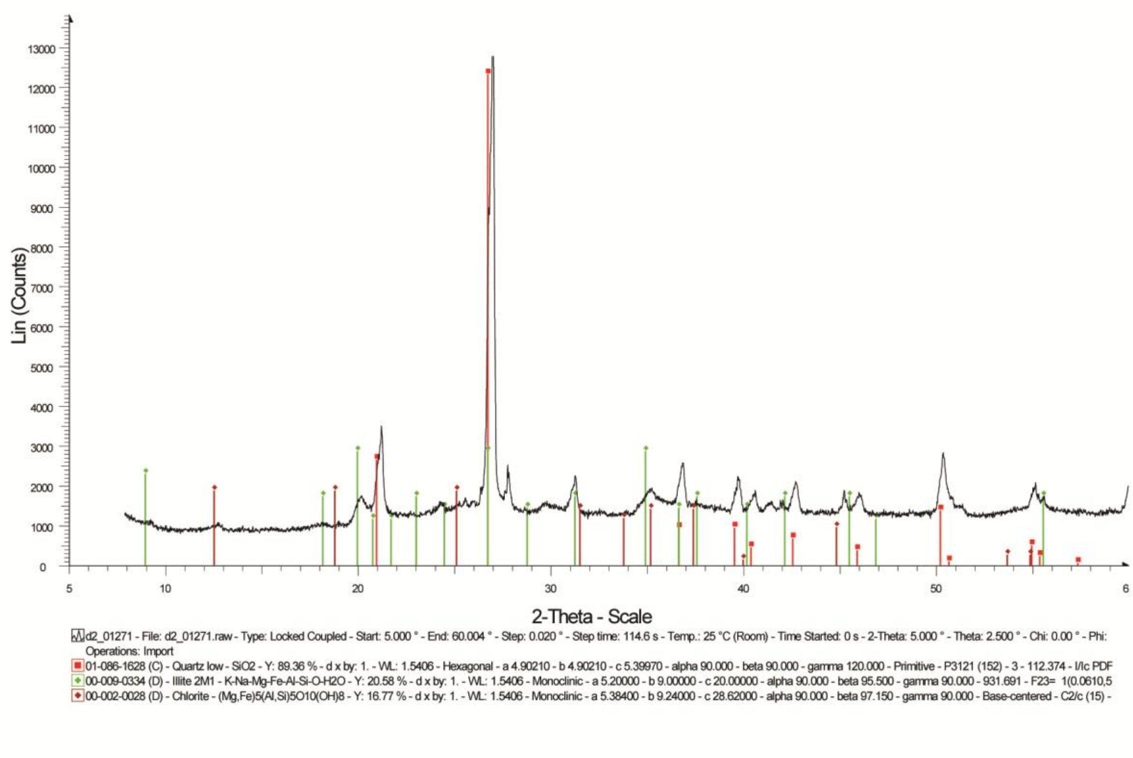


KCB2 - File: d8_00819.raw - Type: 2Th/Th locked - Start: 5.000 ° - End: 90.007 ° - Step: 0.021 ° - Step time: 128.8 s - Temp.: 25 °C (Room) - Time Started: 14 s - 2-Theta: 5.000 ° - Theta: 2.500 ° - Chi: 0.00 ° - Phi: 0.00
Operations: Import
01-089-1961 (C) - Quartz low, dauphinee-twinned - SiO2 - Y: 83.33 % - d x by: 1. - WL: 1.5406 - Hexagonal - a 4.92100 - b 4.92100 - c 5.41600 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - P6222 (180) - 3
01-083-1762 (A) - Calcite - Ca(CO3) - Y: 30.85 % - d x by: 1. - WL: 1.5406 - Rhombo.H.axes - a 4.98960 - b 4.98960 - c 17.06100 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - R-3c (167) - 6 - 367.847 - I/Ic
00-006-0221 (D) - Kaolinite 1Md - Al2Si2O5(OH)4 - Y: 5.95 % - d x by: 1. - WL: 1.5406 - Monoclinic - a 5.16000 - b 8.93000 - c 7.39000 - alpha 90.000 - beta 104.500 - gamma 90.000 - 2 - 329.676 - F23= 2(0.0610,226)
01-089-5862 (C) - Dolomite - CaMg(CO3)2 - Y: 29.07 % - d x by: 1. - WL: 1.5406 - Rhombo.H.axes - a 4.80900 - b 4.80900 - c 16.01820 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - R-3 (148) - 3 - 320.81

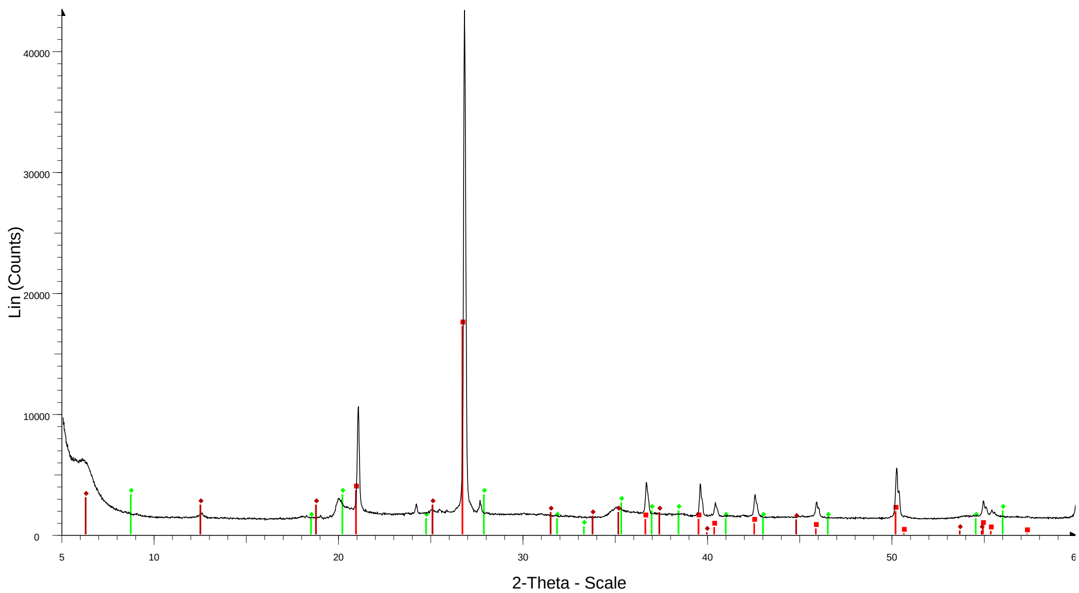
BSB 1


Lin (Counts)
2-Theta - Scale
BS1 - File: BS1.raw - Type: Locked Coupled - Start: 5.000 ° - End: 60.004 ° - Step: 0.020 ° - Step time: 114.6 s - Temp.: 25 °C (Room) - Time Started: 0 s - 2-Theta: 5.000 ° - Theta: 2.500 ° - Chi: 0.00 ° - Phi: 0.00 ° - X: 0
Operations: Import
01-086-1561 (A) - Quartz low - SiO2 - Y: 44.67 % - d x by: 1. - WL: 1.5406 - Hexagonal - a 4.83620 - b 4.83620 - c 5.34390 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - P3221 (154) - 3 - 108.242 - I/Ic PDF
00-009-0334 (D) - Illite 2M1 - K-Na-Mg-Fe-Al-Si-O-H2O - Y: 12.95 % - d x by: 1. - WL: 1.5406 - Monoclinic - a 5.20000 - b 9.00000 - c 20.00000 - alpha 90.000 - beta 95.500 - gamma 90.000 - 931.691 - F23= 1(0.0610,5
01-079-1343 (C) - Dolomite - CaMg(CO3)2 - Y: 36.82 % - d x by: 1. - WL: 1.5406 - Rhombo.H.axes - a 4.79100 - b 4.79100 - c 15.85600 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - R-3 (148) - 3 - 315.19
00-002-0028 (D) - Chlorite - (Mg,Fe)5(Al,Si)5O10(OH)8 - Y: 13.39 % - d x by: 1. - WL: 1.5406 - Monoclinic - a 5.38400 - b 9.24000 - c 28.62000 - alpha 90.000 - beta 97.150 - gamma 90.000 - Base-centered - C2/c (15) -

BSB 2

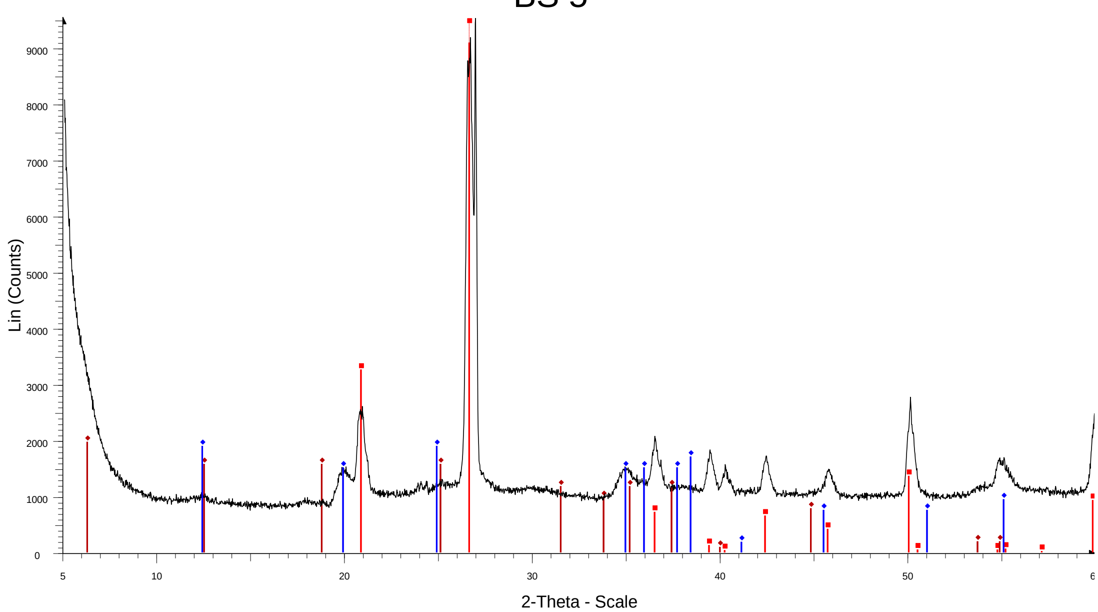

Lin (Counts)
2-Theta - Scale
d2_01271 - File: d2_01271.raw - Type: Locked Coupled - Start: 5.000 ° - End: 60.004 ° - Step: 0.020 ° - Step time: 114.6 s - Temp.: 25 °C (Room) - Time Started: 0 s - 2-Theta: 5.000 ° - Theta: 2.500 ° - Chi: 0.00 ° - Phi:
Operations: Import
01-086-1628 (C) - Quartz low - SiO2 - Y: 89.36 % - d x by: 1. - WL: 1.5406 - Hexagonal - a 4.90210 - b 4.90210 - c 5.39970 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - P3121 (152) - 3 - 112.374 - I/Ic PDF
00-009-0334 (D) - Illite 2M1 - K-Na-Mg-Fe-Al-Si-O-H2O - Y: 20.58 % - d x by: 1. - WL: 1.5406 - Monoclinic - a 5.20000 - b 9.00000 - c 20.00000 - alpha 90.000 - beta 95.500 - gamma 90.000 - 931.691 - F23= 1(0.0610,5
00-002-0028 (D) - Chlorite - (Mg,Fe)5(Al,Si)5O10(OH)8 - Y: 16.77 % - d x by: 1. - WL: 1.5406 - Monoclinic - a 5.38400 - b 9.24000 - c 28.62000 - alpha 90.000 - beta 97.150 - gamma 90.000 - Base-centered - C2/c (15) -

BS 3



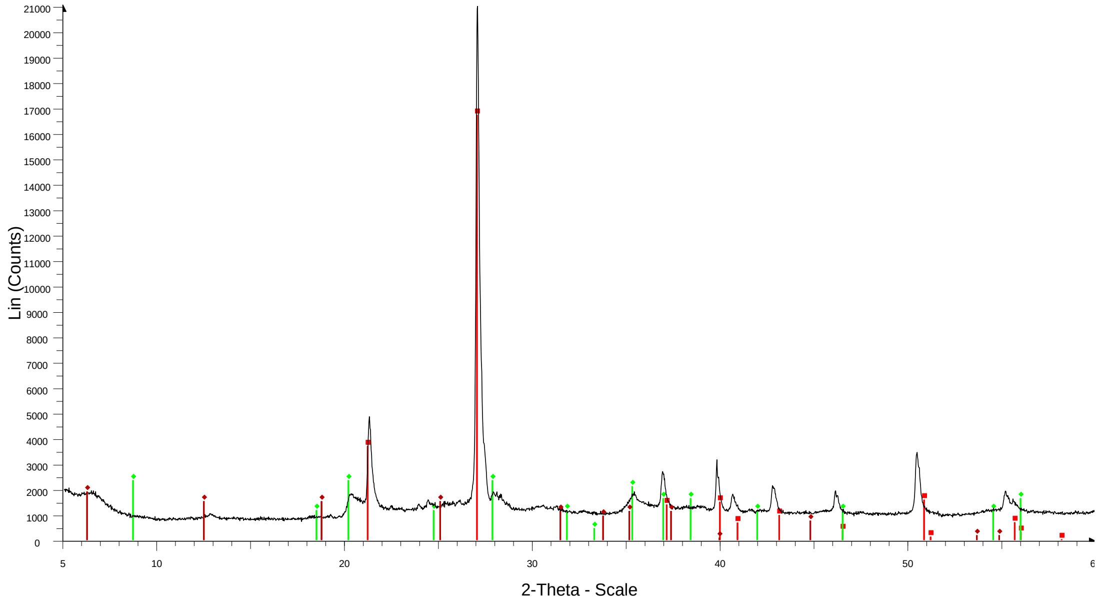
d2_01270 - File: d2_01270.raw - Type: Locked Coupled - Start: 5.000 ° - End: 60.004 ° - Step: 0.020 ° - Step time: 114.6 s - Temp.: 25 °C (Room) - Time Started: 0 s - 2-Theta: 5.000 ° - Theta: 2.500 ° - Chi: 0.00 ° - Phi:
Operations: Import
01-086-1629 (A) - Quartz low - SiO_2 - Y: 39.75 % - d x by: 1. - WL: 1.5406 - Hexagonal - a 4.90300 - b 4.90300 - c 5.39990 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - P3121 (152) - 3 - 112.419 - I/Ic PDF
00-025-0001 (D) - Illite, sodian brammallite - $(Na,K)_{1-x}(Al,Mg,Fe)_2(Si,Al)_4O_{10}(OH)_2$ - Y: 7.63 % - d x by: 1. - WL: 1.5406 -
00-002-0028 (D) - Chlorite - $(Mg,Fe)_5(Al,Si)_5O_{10}(OH)_8$ - Y: 7.06 % - d x by: 1. - WL: 1.5406 - Monoclinic - a 5.38400 - b 9.24000 - c 28.62000 - alpha 90.000 - beta 97.150 - gamma 90.000 - Base-centered - C2/c (15) -

BS 5



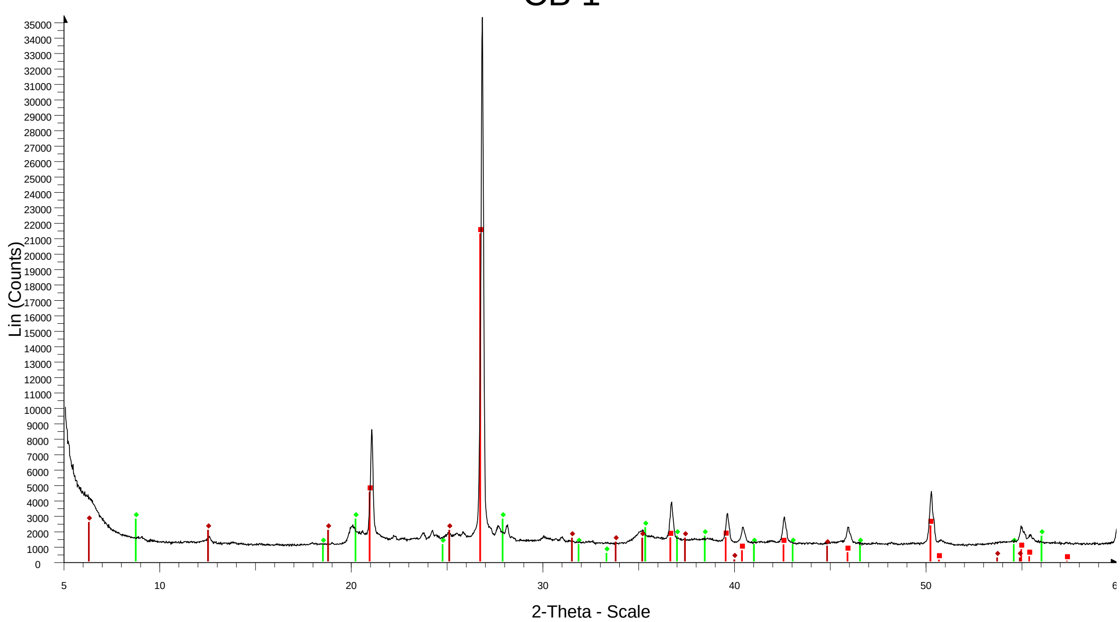
d2_01273 - File: d2_01273.raw - Type: Locked Coupled - Start: 5.000 ° - End: 60.004 ° - Step: 0.020 ° - Step time: 114.6 s - Temp.: 25 °C (Room) - Time Started: 0 s - 2-Theta: 5.000 ° - Theta: 2.500 ° - Chi: 0.00 ° - Phi:
Operations: Import
01-089-1961 (C) - Quartz low, dauphinee-twinned - SiO_2 - Y: 137.96 % - d x by: 1. - WL: 1.5406 - Hexagonal - a 4.92100 - b 4.92100 - c 5.41600 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - P6222 (180) -
00-002-0028 (D) - Chlorite - $(Mg,Fe)_5(Al,Si)_5O_{10}(OH)_8$ - Y: 20.72 % - d x by: 1. - WL: 1.5406 - Monoclinic - a 5.38400 - b 9.24000 - c 28.62000 - alpha 90.000 - beta 97.150 - gamma 90.000 - Base-centered - C2/c (15) -
00-006-0221 (D) - Kaolinite 1Md - $Al_2Si_2O_5(OH)_4$ - Y: 19.95 % - d x by: 1. - WL: 1.5406 - Monoclinic - a 5.16000 - b 8.93000 - c 7.39000 - alpha 90.000 - beta 104.500 - gamma 90.000 - 2 - 329.676 - F23= 2(0.0610,226

CB 3



d2_01269 - File: d2_01269.raw - Type: Locked Coupled - Start: 5.000 ° - End: 60.004 ° - Step: 0.020 ° - Step time: 114.6 s - Temp.: 25 °C (Room) - Time Started: 0 s - 2-Theta: 5.000 ° - Theta: 2.500 ° - Chi: 0.00 ° - Phi: Operations: Import
01-086-1561 (A) - Quartz low - SiO2 - Y: 79.62 % - d x by: 1. - WL: 1.5406 - Hexagonal - a 4.83620 - b 4.83620 - c 5.34390 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - P3221 (154) - 3 - 108.242 - I/Ic PDF
00-002-0042 (D) - Illite, sodian - (Na,K)Al2(Si3AlO10)(OH)2 - Y: 11.22 % - d x by: 1. - WL: 1.5406 -
00-002-0028 (D) - Chlorite - (Mg,Fe)5(Al,Si)5O10(OH)8 - Y: 9.11 % - d x by: 1. - WL: 1.5406 - Monoclinic - a 5.38400 - b 9.24000 - c 28.62000 - alpha 90.000 - beta 97.150 - gamma 90.000 - Base-centered - C2/c (15) -

CB 1



d2_01272 - File: d2_01272.raw - Type: Locked Coupled - Start: 5.000 ° - End: 60.004 ° - Step: 0.020 ° - Step time: 114.6 s - Temp.: 25 °C (Room) - Time Started: 0 s - 2-Theta: 5.000 ° - Theta: 2.500 ° - Chi: 0.00 ° - Phi: Operations: Import
01-086-1629 (A) - Quartz low - SiO2 - Y: 60.29 % - d x by: 1. - WL: 1.5406 - Hexagonal - a 4.90300 - b 4.90300 - c 5.39990 - alpha 90.000 - beta 90.000 - gamma 120.000 - Primitive - P3121 (152) - 3 - 112.419 - I/Ic PDF
00-025-0001 (D) - Illite, sodian brammallite - (Na,K)1-x(Al,Mg,Fe)2(Si,Al)4O10(OH)2 - Y: 7.87 % - d x by: 1. - WL: 1.5406 -
00-002-0028 (D) - Chlorite - (Mg,Fe)5(Al,Si)5O10(OH)8 - Y: 7.24 % - d x by: 1. - WL: 1.5406 - Monoclinic - a 5.38400 - b 9.24000 - c 28.62000 - alpha 90.000 - beta 97.150 - gamma 90.000 - Base-centered - C2/c (15) -