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# Re-examination of Changes in Fluvial Stacking Pattern Across the P-t Boundary in the Central Transantarctic Mountains, Antarctica

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# RE-EXAMINATION OF CHANGES IN FLUVIAL STACKING PATTERN ACROSS THE P-T BOUNDARY IN THE CENTRAL TRANSANTARCTIC MOUNTAINS, ANTARCTICA

by

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ABSTRACT

RE-EXAMINATION OF THE CHANGES IN FLUVIAL STACKING PATTERN ACROSS THE CENTRAL TRANSANTARCTIC MOUNTAINS, ANTARCTICA

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Under the Supervision of Professor John Isbell

A change in fluvial style and a change in the stacking pattern of fluvial channel sandstone bodies occur across the Buckley□Fremouw formational contact in the central Transantarctic Mountains in Antarctica. Strata in the Buckley Formation are characterized by thick floodplain deposits in the Middle to Upper Permian Buckley Formation; whereas, stacked interconnected sandstone bodies occur in the Triassic Fremouw Formation (Barrett et al., 1986; Isbell & Macdonald, 1991a, 1991b; Collinson et al., 1994; Isbell et al., 1997; 2005). Such changes in fluvial stacking patterns have been attributed to changes in the creation of accommodation within basins due to changes in relative sea level, changes in accommodation due to tectonism, and changes in sediment flux associated with loss of vegetation and increased erosion rates following the end-Permian mass extinction event. To explain the changes in the Buckley-Fremouw Formation in Antarctica, Isbell & Macdonald (1991a, 1991b) and Isbell et al. (1997) argued for changing tectonic conditions in the basin while Retallack et al. (2006) suggested the changes were associated with the P□T mass extinction event causing the loss of peat forming plants. This study found that the change in the accommodation across the PTB was a result of tectonism based on evidence of changing sandstone

composition, changing paleocurrent orientations, and changing fluvial stacking patterns between the Buckley Formation and the Fremouw Formation. This suggests differential subsidence in the Transantarctic foreland basin with an under-filled basin in the Late Permian changing to an over-filled basin in the Early Triassic.

To my parents, David and Judy
and my grandparents
Chuck and Heidi

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## LIST OF ABBREVIATIONS

**PTB** Permian–Triassic Boundary

**BGR** Beardmore Glacier Region

**CTM** Central Transantarctic Mountains

**EPME** End-Permian mass extinction

**FUS** Fining upward succession

**CGSF** Coarse-grained sandstone sheet facies association

**FGSF** Fine-grained sandstone sheet facies association

**st** Large-scale trough cross-stratification

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# Introduction

## 1.1 Problem and Significance

The purpose of this study is to investigate the environmental and tectonic conditions in Antarctica during deposition of the strata of the Permian Buckley and the Triassic Fremouw formations in the central Transantarctic Mountains (CTM), by investigating changes in fluvial stacking patterns. Stacking patterns are defined as the architecture of fluvial channels, and the density and interconnectedness of channel sandstone bodies relative to floodplain deposits in time and space (Leeder, 1978; Bridge & Leeder, 1979; Miall, 1983; Einsele, 1992; Bridge & Mackey, 1993). These fluvial strata record an important interval of geologic time and provide a unique view of conditions in Polar Gondwana during the late Paleozoic and early Mesozoic. Possible events that influenced fluvial stacking patterns in these strata include: 1) development and evolution of the foreland basin, 2) changing climatic conditions from an icehouse to a greenhouse world, 3) severe biotic and environmental crises associated with the end-Permian mass extinction, and 4) possible changes in sea level. By determining what the drivers were that resulted in changing fluvial stacking patterns in the Transantarctic Basin, a better understanding of polar Gondwana during this critical time in Earth's history can be identified. Four hypotheses have been proposed as the controlling mechanism for the changes in fluvial stacking patterns, which include: changes in eustatic sea level, changes in climate, plant extinction, and tectonism.

The Buckley and Fremouw formations were deposited in an evolving foreland basin within the late Paleozoic-early Mesozoic South Polar Circle (Figure 1) at a time when the climate was transitioning from the late Paleozoic ice age into a Late Permian-

Mesozoic Hothouse. During this time these strata experienced and recovered from the Phanerozoic's most severe mass extinction (the end Permian Mass extinction (EPME) and the ensuing Early Triassic recovery). The Transantarctic Basin was also evolving due to changing tectonic conditions along the Panthalassan margin of Antarctica and throughout Gondwana. Results of this study will help define fluvial stacking patterns in the Transantarctic Basin and aid in understanding the controls on fluvial stacking patterns and landscape evolution during this time of severe environmental change. This study will also provide insight on how the relationship between accommodation and sediment flux influences fluvial stacking patterns, which will aid in developing and testing models for Terrestrial Sequence Stratigraphy.

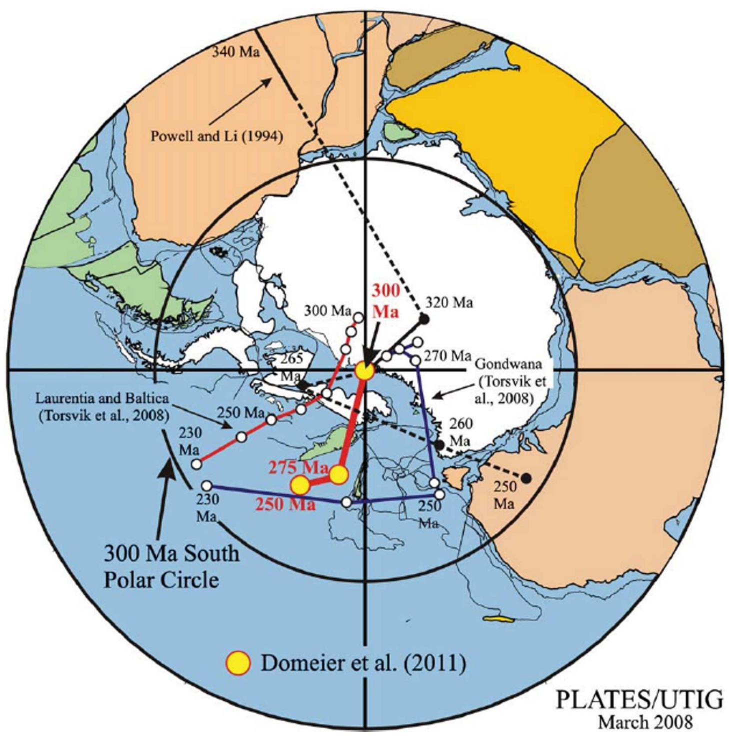



Figure 1. Late Permian Gondwanan plate reconstruction showing the position of the South Pole during the Permian and Earliest Triassic (Domeier et al., 2011; Isbell et al., 2012).

Changes in stacking patterns from the Late Permian through the Triassic have been observed throughout Gondwana including: South Africa, Antarctica, and Australia (Barrett et al., 1986; Collinson et al., 1994, 2006; Isbell et al., 1997; Ward et al., 2000; Catuneanu & Elango, 2001; Michaelsen, 2002). In Antarctica, a change in stacking patterns across the Permian-Triassic boundary (PTB) has been identified and is characterized by a change from laminated siltstones, carbonaceous shales, coals, and individual thin sandstone bodies isolated in thick floodplain deposits (low stacking pattern) in the Upper Permian Buckley Formation to thick amalgamated sandstone bodies consisting of stacked channels and bar forms with minor interstratified, non-carbonaceous, greenish gray mudstone (high stacking pattern) in the Lower Triassic Fremouw Formation (Collinson et al., 1994, 2006; Isbell et al., 1997). Changes in fluvial stacking patterns of channel deposits are traditionally linked to changes in fluvial style (e.g., a change from braided to meandering to anastomosed streams; Miall, 1996). More recently, changes in fluvial style have also been linked to changes in the rate of the creation of accommodation (e.g., subsidence, eustatic changes in sea level) (Flemings & Jordan, 1990; Frostick & Steel, 1993; Shanley & McCabe, 1994; DeCelles & Giles, 1996; Miall, 1996, 1997; Blum & Törnqvist, 2000; Clevis et al., 2004) relative to changes in avulsion rates, which are a proxy for sedimentation or sediment influx rates in fluvial environments (e.g., climate change, loss of vegetation; Schumm, 1977; Bridge & Leeder, 1979; McLoughlin et al., 1997; Cecil, 2003). The changes in fluvial stacking patterns across the PTB in Antarctica have been interpreted as lacustrine and meandering and/or braided river deposits with poorly drained floodplains in the Upper Permian Buckley Formation (Barrett et al., 1986; Isbell 1990, Isbell and Collinson, 1991; Isbell et

al., 1997) and braided river deposits with well-drained floodplains in the Lower Triassic Fremouw Formation (Barrett et al., 1986; Collinson et al., 1994, 2006).

A change in climate and landscape across the PTB has been proposed by Collinson (1997) and Retallack (1999) as a change from a cool humid climate in the Late Permian to a warmer climate with seasonal precipitation in the Early Triassic. These changes are represented by the occurrence of fossil leaves of *Glossopteris,* coal beds, and carbonaceous shales that represent humid climates within Permian strata, and an absence of coal, the occurrence of a *Dicoridium* fossil flora, and the occurrence of red-green-mottled claystones that represent drier climates as recorded in Triassic strata. Whether the change in climate and landscape are linked to the end-Permian mass extinction event that occurred at ~ 252 Ma (cf. Bowring & Erwin, 1998; Mundil et al., 2004) is unclear. If the Upper Permian Buckley and Lower Triassic Fremouw formational contact represents the PTB, then these strata contain a record of the EPME event from a high latitude perspective. As of yet, precise radiometric dating of these strata has not been successfully conducted. If, however, the Buckley-Fremouw contact formed due to the EPME, it suggests that the lithologic and facies changes observed across the formational contact possibly represent instantaneous changes to the landscape due to the extinction and loss of land plants and an associated increase in erosion rates (cf. Retallack & Krull, 1999; Ward et al., 2000; Michaelsen, 2002; Retallack 2005a; Retallack et al., 2005). However, McManus et al. (2002) and Collinson et al. (2006) raised the possibility that the Buckley Fremouw contact is diachronus across the CTM, which calls into question the validity of the hypothesis of a synchronous change in stacking patterns in the Transantarctic basin across the PTB. The basin evolution and the role of the EPME in

Antarctica are poorly understood. The EPME is the largest extinction in Earth's history, resulting in a loss of about 90% of marine species, about 70% of terrestrial vertebrate species (Erwin, 1994), and similar estimates (70-90%) have been made for terrestrial peat-forming plants (Retallack, 1995; Gastaldo et al., 2005). Both the marine and terrestrial realms were greatly affected with profound consequences for sedimentary environments, such as possible delayed recovery of coral reefs and peat swamps, the spread of stromatolites, claystone breccias, and the instantaneous development of braided stream deposits within terrestrial realms (Veevers et al., 1994; Retallack et al., 1996, 2005; Ward et al., 2000). The degree to which the EPME influenced extinction and environmental change in the Polar Regions has only recently been explored. Preliminary results suggest that the extinction event may not have been as severe near the poles (Miller & Collinson, 1994; Waterhouse & Shi, 2010; S.T. Hasiotis, personal communications, 2012). Many hypotheses have been proposed for causes of the EPME including, but not limited to: 1) draining of the continental shelves during exceptionally low sea level (Newell, 1973); 2) radiation from extraterrestrial sources (Visscher et al., 2004); 3) bolide impact (Retallack et al., 1998); 4) anoxia in stratified oceans (Wignall & Hallam, 1992; Wignall & Twitchett, 2002); 5) atmospheric changes caused by release of gases from the eruption of the Siberian Traps (Reichow et al., 2002); and 6) large methane gas releases (Berner, 2002; Retallack & Krull, 2006).

The focus of this study is to determine the fluvial stacking patterns of the Buckley and Fremouw formations, and to test possible causes for the abrupt shift in depositional environments across the Buckley-Fremouw formational contact in the central Transantarctic Mountains

This study examines well-exposed sections of Permian strata in the central Transantarctic Mountains at Mt. Achenar, Mt. Bowers, and Lewis Cliffs, and Triassic strata from Wahl Glacier and Gordon Valley (Figure 2). Previously reported data on the Permian-Triassic strata in the region is also used to supplement the findings of this study (Barrett, 1972; Barrett et al., 1986; Collinson & Isbell, 1986; Collinson, 1990; Isbell, 1990; Isbell et al., 1997; Isbell & Collinson, 1991; Flaig, 2005; Collinson et al., 2006). Results of this study will help determine what fluvial styles and stacking patterns (meandering, braided, anastomosed, and avulsion signatures) are contained in the strata, why fluvial stacking patterns change across the Buckley-Fremouw contact (e.g., tectonics, climate, sea-level change, plant extinction), and how the basin evolved during the Late Permian–Early Triassic.

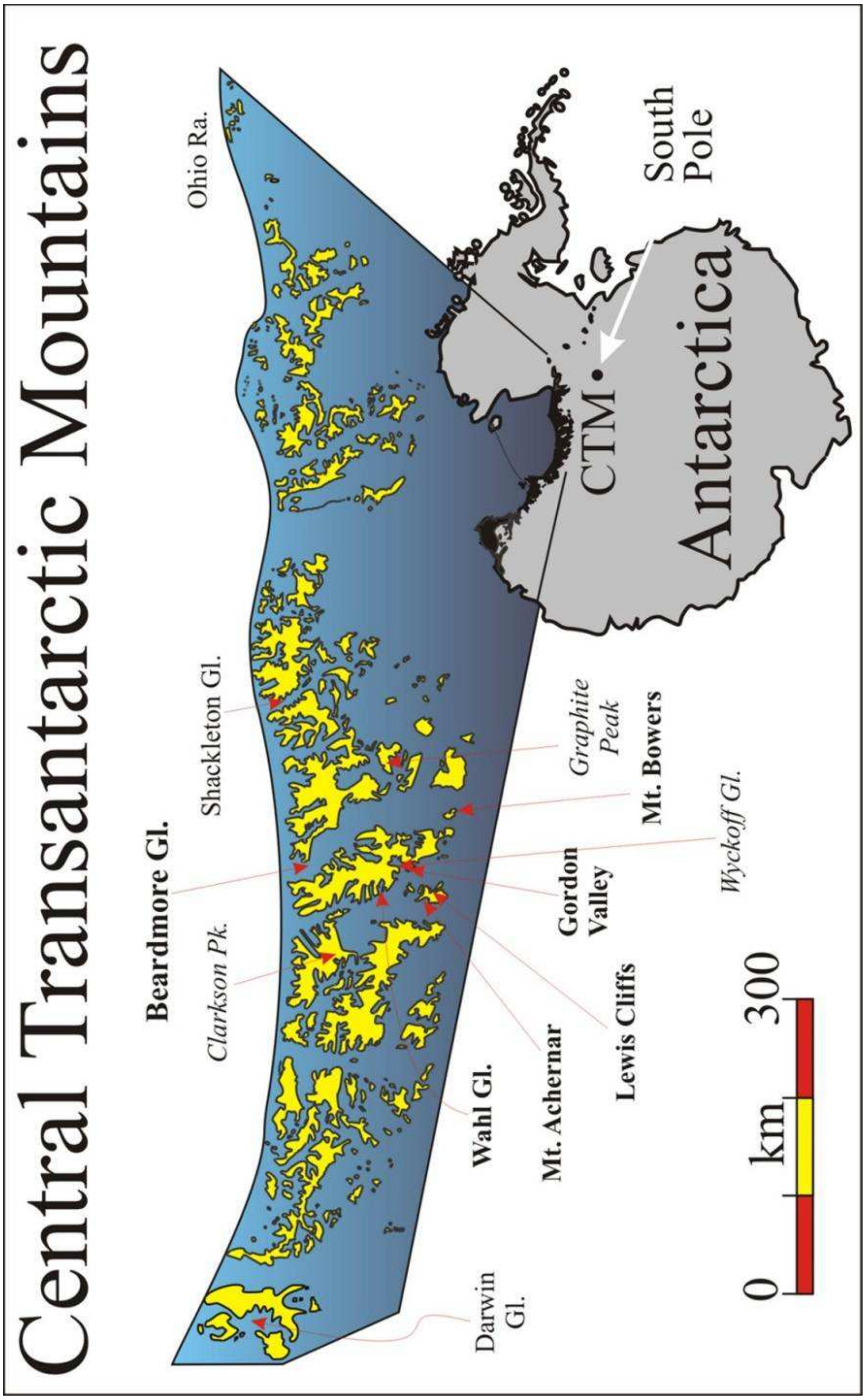



Figure 2. Study locations in the CTM. Late Permian strata occur at Mt. Bowers, Mt. Achenar, and Lewis Cliffs. Early Triassic strata occur at Wahl Glacier and Gordon Valley. Other locations in italics will be used from previous studies (modified from, Isbell, 2010).

## 1.2 Regional Geology

The Permian□Triassic Transantarctic (Beacon) basin extended along the East Antarctic cratonic margin from the Ellsworth Mountains to Victoria Land (Collinson, 1990; Isbell, 1990; Collinson et al., 1994; Isbell et al. 1997). Collinson et al. (1994) proposed that the basin evolved above a passive continental margin through four main stages. The first stage consisted of extension during the Carboniferous and Permian. The second stage consisted of the formation of a late Early Permian back-arc basin. The third stage was the formation of the Late Permian and Early Triassic foreland basin. Isbell (1990, 1999) and Isbell et al. (1997) provided evidence that suggests the Fairchild-Buckley formational contact marks the point at which the basin evolved into a foreland basin. The fourth stage consisted of extension and tholeiitic volcanism during the Jurassic. Isbell (1999) suggested that strata in the CTM were deposited within two intermontane or successor basins after post-orogenic uplift of the Ross terrain. In a regional context, the Transantarctic foreland basin developed as a part of the more extensive Pan-Gondwanian Mobile Belt that formed due to compression, collision, and subduction of the paleo-Pacific plate beneath the Gondwanan plate (Veevers et al., 1994; Catuneanu & Elango, 2001; Gastaldo et al., 2005). At that time, a series of related basins developed along the Panthalassan margin of Gondwana. These basins include: the Parana Basin (South America), the Karoo Basin (South Africa), the Transantarctic Basin (Antarctica), and the Sydney and Bowen Basin (Australia) (Collinson et al., 1994; Veevers et al., 1994; Isbell et al., 1997; Lopez-Gamundi, 1997; Fielding et al., 2001; Catuneanu, 2004).

The Transantarctic basin in the CTM is composed of a sequence of flat-lying, Devonian to Early Jurassic marine and continental deposits known as the Beacon Supergroup (cf. Barrett et al., 1986; Barrett, 1991; Collinson 1990). The Beacon Supergroup in the Beardmore Glacier region (BGR) consists of the Taylor Group and Victoria Group. The Taylor Group, in the BGR, consists of the Alexandra Formation, which is thought to be Devonian in age due to the correlation with other quartz arenites in southern Victoria Land and in the Ohio Range of Antarctica (Barrett et al., 1986). There is uncertainty, however, that the Alexandra Formation is truly Devonian due to the lack of fossils or radiometrically datable materials to support this age assignment (Barrett et al., 1986; Collinson, 1990). The Maya Erosion Surface, a regional unconformity, separates the Taylor Group from the Victoria Group (Barrett et al., 1986; Isbell, 1999). The Victoria Group in the BGR consists of six formations ranging from Permian to Triassic in age (Barrett et al., 1986; Collinson et al., 1994). A generalized sequence of the Victoria Group (Figure 3) begins with the glaciogenic latest Carboniferous (?)-Early Permian strata of the Pagoda Formation (Barrett et al., 1986; Isbell et al., 2008). The Pagoda Formation is overlain by the 60–140 m thick postglacial basinal shales, turbidites, and, prograding, subaqueous deltaic deposits of the Lower Permian Mackellar Formation (Barrett et al., 1986; Collinson et al., 1994; Miller & Collinson, 1994; Miller & Isbell, 2010). The Mackellar Formation is overlain by the 130–220 m thick fluvial-deltaic strata of the Fairchild Formation. Overlying the Fairchild Formation is the Buckley Formation, which is a 750 m thick succession of thin to thick sandstone sheets, coal, siltstone, and mudrock deposited in fluvial channel, levee, floodplain, mire, and lacustrine settings (Barrett et al., 1986; Isbell 1990; Collinson et al., 1994; Isbell et al.,

1997). The Buckley Formation is informally subdivided into two members. The lower member consists of quartzo-feldspatic sand and the upper member consists of volcaniclastic sand (Barrett et al., 1986; Collinson et al., 1994). An 180˚ reversal in paleocurrents also occurs between the lower and upper member of the Buckley Formation (Isbell, 1990, 1991). The Buckley Formation is overlain by the 650 m thick Triassic Fremouw Formation. The Fremouw Formation is informally subdivided into three members (lower, middle, and upper) (Barrett et al., 1986; Collinson et al., 1994). The lower member in the BGR consists of thick quartzose sandstone sheets and green-gray, non-carbonaceous siltstones (Barrett et al., 1986; Collinson et al., 1994). The middle member consists primarily of mudstone and thin sandstone sheets, and the upper member consists of volcanic sandstone, thin coal beds, logs, and *Dicoridium* (Barrett, 1972). The upper formation (Falla) is not noted in Figure 3, as it was not part of this study.

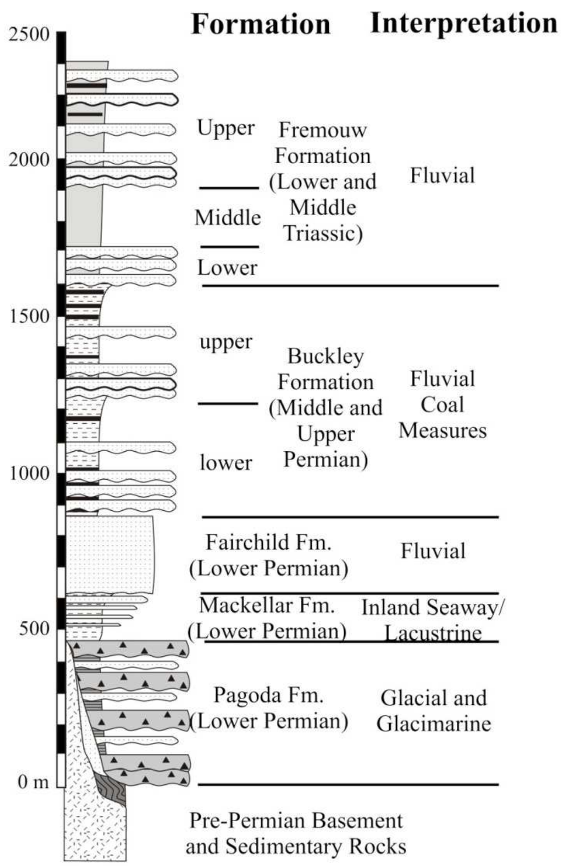



Figure 3. Generalized stratigraphic column of the Victoria Group, Beacon Supergroup, in CTM (After Isbell, 2012).

The Buckley-Fremouw contact has been proposed by Retallack et al. (1998) to represent the P-T boundary. The criteria used to determine the boundary was: a decrease in $\delta^{13}$ $C_{org}$; extinction of *Glossopteris* flora; the disappearance of coal in the upper Buckley; the appearance of the *Dicoridium* flora in the middle and upper Fremouw Formation; and the first appearance of the *Lystrosaurus* therapsid fauna in the lower Fremouw Formation. Isbell and Askin (1999) noted problems with this imprecise means of defining the location of the PT boundary. Furthermore, *Glossopteris* flora, along with roots, leaves, and wood have been found in the Shackleton Glacier region in the Fremouw Formation (McManus et al., 2002; Collinson et al., 2006). These findings suggest that the lower part of the Fremouw Formation in the Shackleton Glacier region maybe uppermost Permian (Collinson et al., 2006). Elsewhere, the *Glossopteris* flora occurs in Lower Triassic strata in Tasmania (Holmes, 1992), South Africa (Thomas, 1952; 1958), and India (Jacob, 1952; Pant and Pant, 1987).

### 1.3 Hypotheses for changing fluvial patterns and changes in fluvial stacking patterns

Fluvial architecture is thought to be controlled by three factors: accommodation, sediment supply, and hydrodynamics (Puigdefabregas, 1973). These controlling factors can be affected by sea level, tectonics (subsidence), climate, and vegetation, all of which could have an influence on fluvial patterns across the PTB in Antarctica (cf. Isbell et al., 1997, 2005; Retallack & Krull, 1999; Ward et al., 2000; McManus et al., 2002; Michaelsen, 2002; Retallack et al., 2005; Collinson et al., 2006). Characteristics of the

hypotheses for these controls will be used to help determine the cause for changing fluvial styles across the PTB in the CTM, Antarctica.

**Characteristics of accommodation and relative sea-level change**

If changes in sea level are the reason for a change in fluvial stacking patterns and fluvial style across the PTB, then the change in depositional environments should occur everywhere throughout the Transantarctic Basin that was influenced by eustatic changes in sea level. These changes should be diachronus as the rising sea progressively transgressed the land surface through time. Changes in accommodation in basins that are long distances from the ocean are driven by climate change and tectonism (Shanley & McCabe, 1994). Shanley and McCabe (Figure 4; 1994) suggested that during falling relative sea level, channel incision would be preserved as incised valley fills. During a lowstand and early transgression, amalgamated sand bodies fill the incised valley. This results in a high density channel body stacking pattern. As transgression proceeds, resulting from the shoreline migrating landward, strata records mixed marine-continental conditions and changes in accommodation as the rate of relative sea level rise increases. Shanley and McCabe (1994) suggested that such changes consist of a change from fluvial to tidally influenced fluvial deposits and a change in stacking pattern from overlapping, interconnected sandstone bodies (high density stacking pattern) to isolated sand bodies (low density channel body stacking pattern) due to an increase in the rate of the creation of accommodation space reflecting an increase in the rate of sea-level rise relative to the influx of clastics. Farther from the shoreline, the tidal signature might not be as pronounced and the deposits might simply show a change from high fluvial stacking

patterns to low fluvial stacking patterns. As the shoreline migrated seaward during the highstand stage, and a decrease in the rate of sea level rise to zero, multilateral sand bodies would be deposited due to a decrease in the rate of the creation of accommodation, resulting in a higher fluvial sandstone stacking pattern (Shanley & McCabe, 1994; Blum & Törnqvist, 2000; Blum & Aslan, 2006; Blum et al., 2012).

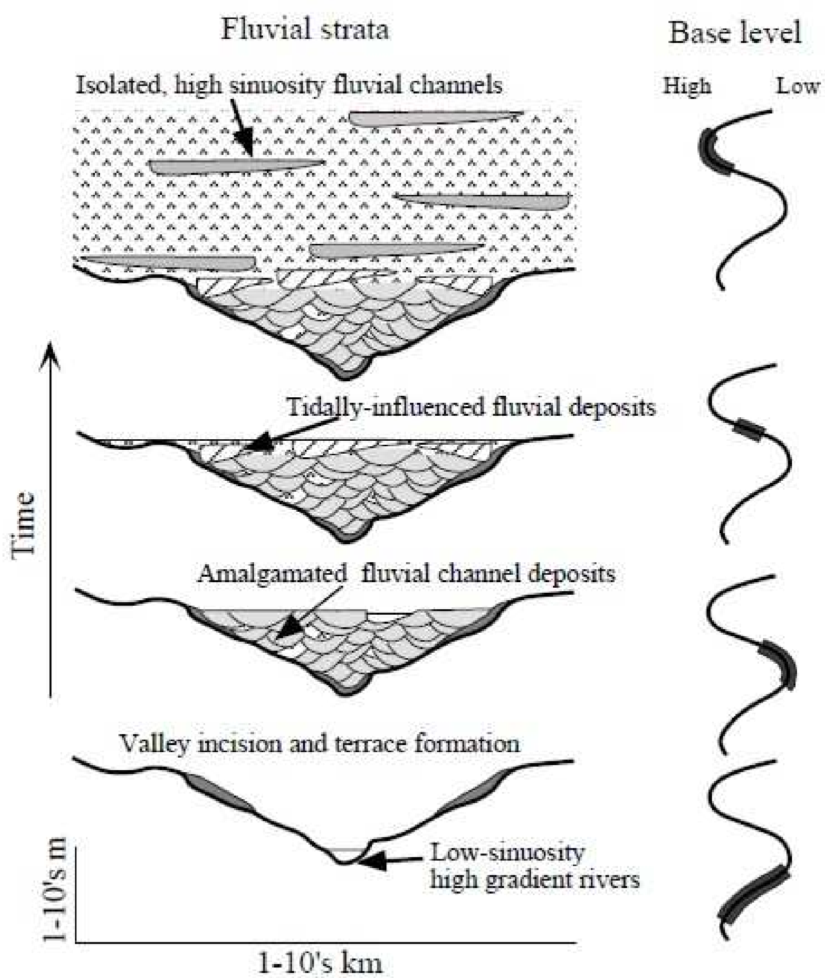



Figure 4. Diagram showing fluvial architecture as a function of base-level change (modified from Shanley & McCabe, 1993, 1994).

**Characteristics of tectonics**

There are many sedimentological characteristics that would suggest tectonics as the primary factor in changing fluvial style and stacking patterns. Subsidence due to tectonism is also a control on the development of accommodation. All basins experience differential subsidence. The drivers of subsidence can be thermal heating, sediment loading, or crustal thinning (Einsele, 1992; Leeder, 1999; Allen, 2005). In a foreland basin setting, such as the Transantarctic basin, the main driver of subsidence is tectonic loading due to emplacement of a fold and thrust belt (Flemings & Jordan, 1990; Jordan, 1995). Collinson et al. (1994) suggested that the fold and thrust belt for the Transantarctic basin is exposed in the Ellsworth and Pensacola Mountains because the present Transantarctic Mountains form an oblique cut across the trend of the late Paleozoic-early Mesozoic foreland basin. Once a foreland basin is formed, sediment loading becomes an important driver of continued subsidence (Flemings & Jordan, 1990; Jordan, 1995). Characteristics that may signal tectonism include a change in provenance (determined by changes in sandstone composition due to changing source terrains and erosional unroofing of different stratigraphic horizons in the fold and thrust belt), changes in regional lithofacies patterns due to changing locations of clastic influx and accommodation patterns, and changes in paleocurrent direction due to changes in subsidence (Isbell, 1990). Changing fluvial stacking patterns reflect changing accommodation and clastic flux into the basin (cf. Fleming & Jordan, 1990). Jordan (1995) defines a foreland basin as being either under-filled or over-filled depending on the topography of the basin and the amount of sediment supply vs. transport processes. An under-filled basin typically has a topographic trough between the orogenic belt and

the forebuldge (flexure uplift on the cratonic side of the basin) with paleocurrents flowing off the orogenic belt and forebuldge towards the basin axis (Figure 5). The drainage pattern in the basin axis of an under-filled basin is transverse to paleoflow coming off the orogenic belt and forebuldge. An over-filled basin has sedimentation rates exceeding accommodation so that a depositional trough doesn't form and the basin has an alluvial plain that extends off of the thrust belt and across the basin and forebulge with paleoflow solely coming off the orogenic belt and flowing out across the forebulge onto the craton (Figure 5). Models by Flemings and Jordan (1990) and Jordan (1995) predict that in an under-filled foreland basin there will be thrust-belt-derived, coarse-clastic facies that display high fluvial stacking patterns adjacent to the orogenic belt due to the greatest amount of sediment being derived from the adjacent high despite this also being the area of maximum subsidence, which occurs beneath the orogenic load (i.e., orogenic basin-margin facies). Further into the basin where subsidence is still high but where there is reduced clastic influx due to trapping of clastic in the proximal high subsidence zone, a topographic depression develops. Here, sparsely stacked channel body sandstone is isolated within thick basin-axis mudstones that produce low fluvial stacking patterns (i.e., Basin axis facies). However, locally, truck streams, which flow longitudinally down the topographic depression that marks the basins drainage axis, may produce a moderate to high stacking pattern. Finally, the strata deposited along the cratonic basin margin (i.e., Cratonic basin margin facies) would display moderately to highly stacked, forebulge-derived, coarse-clastic facies due to sediment influx being greater than the slow subsidence rates. In an under-filled basin, as the thrust sheet propagates cratonward and/or sediment loading redistributes the load, the prograding orogenic basin margin

displaces the basin axis and cratonic basin margin facies cratonward. In this scenario, the initial distal forebulge-derived facies would be overlain by the basin-axis facies and possibly even overlain by the proximal thrust-belt-derived coarse-clastic facies depending on the extent of progradation and the amount of clastics available to be eroded from the load and reworked within the basin. Paleocurrents should show flow from the thrust sheet as well as the forebulge flowing towards the basin axis. In the basin axis paleocurrents will flow perpendicular to flow coming off the thrust sheet and forebulge, and toward the lowest topographic region. During expansion of the basin due to continued tectonism or sediment loading, the location of axial drainage shift cratonward in front of a clastic wedge prograding off the orogenic margin. An over-filled basin occurs when sediment supply exceeds subsidence and infills the trough. In an over-filled basin the provenance and paleocurrents from proximal and distal clastic facies are from the thrust belt and facies should coarsen upward through the section.

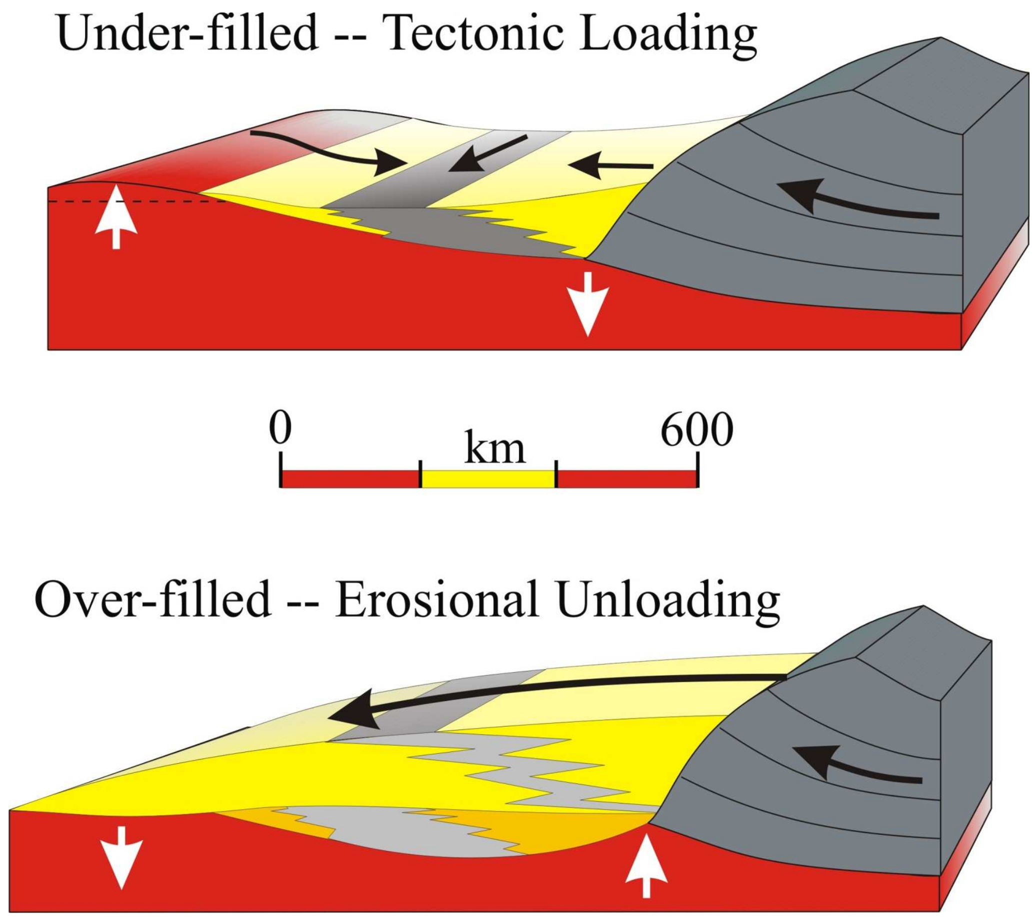



Figure 5. Diagram showing differences in paleocurrent directions and basin-fill of an under-filled and over-filled basin (Isbell, 2005). The white arrows represent subsidence and uplift and the black arrows represent paleoflow directions.

**Characteristics of climate induced stacking patterns**

Climatic changes are driven by changes in the amount of $CO_2$ and other greenhouse gases in the atmosphere, a change in plate location and/or configuration, uplift, and orbital parameters (Cecil, 2003; Church & Coe, 2003). If climate was the leading factor for changes in fluvial patterns one would expect to see a gradual shift in regional lithofacies due to changing precipitation patterns and changes in the volume of sediment influx into a basin due to changes in vegetation, and no significant change in paleocurrent orientations. The climate of a particular region is determined typically by precipitation, temperature, wind, and sunlight resulting in a type of seasonality (Cecil, 2003). In low latitudes wind direction and precipitation determine seasonality because temperature and sunlight typically are constant (Cecil, 2003). In high latitudes, however, variations in all the parameters control seasonality (Cecil, 2003). Seasonality of rainfall controls such parameters as weathering, pedogenesis, vegetative cover, erosion, and sediment yield for a catchment basin (Cecil & Dulong, 2003). Schumm (1977) noted that temperature is also an important factor influencing sediment yield, along with precipitation. The warmer the climate the more precipitation is needed to produce a significant amount of runoff; whereas, in a colder climate less precipitation is needed (Schumm, 1977). Langbein & Schumm (1958) and Schumm (1977) proposed a sediment yield curve (Figure 6), which suggests sediment yield reaches a maximum at about 31 cm/yr (12 inches/yr) of precipitation due to more runoff being available to transport sediment, but with greater amounts of precipitation (>31 cm/yr) vegetation becomes more abundant leading to less runoff and a decrease in sediment yield (Schumm, 1977). Newell et al. (1999) proposed a change in discharge regime due to a change in climate (subhumid to more arid) can increase sediment yield by reducing vegetation cover and a

change from a high-frequency and low-magnitude discharge regime to a low-frequency and high-magnitude discharge regime. So, under a more arid environment (low-frequency and high-magnitude discharge regime) channel size increases due to infrequent but large flood events that favor transport of more coarse-grained material (Newell et. al., 1999).

An increase in sedimentation rate, due to an increase in precipitation, would result in aggradation of the alluvial plain. An increase in channel-belt aggradation can result in an increase in avulsion frequency (Bridge & Leeder, 1979; Bridge & Mackey, 1993; Törnqvist, 1994; Bryant et al., 1995; Heller & Paola, 1996; Blum & Törnqvist, 2000; Mohrig et al., 2000; Törnqvist & Bridge, 2002; Stouthamer & Berendsen, 2007) by creating an alluvial ridge above the floodplain. Early models by Allen (1978), Leeder (1978), and Bridge & Leeder (1979) all emphasized avulsions as a control on the distribution of channel stacking pattern and all models predict an increase in 2D channel connectedness and density when subsidence and aggradation rates are low and vice versa (Mohrig et al., 2000). Avulsing systems that reoccupy abandoned channels in topographic lows should produce multistory channel sand bodies due to sedimentation rates exceeding accommodation (Blum & Törnqvist, 2000; Mohrig et al., 2000). In avulsing systems where accommodation is greater than sedimentation rate isolated ribbon channels will be encased in floodplain deposits (Blum & Törnqvist, 2000). Although the rate of clastic influx maybe driven in part by climate, accommodation is a function of tectonism (subsidence) and sea level change (Jervey, 1988; Catuneanu, 2006).

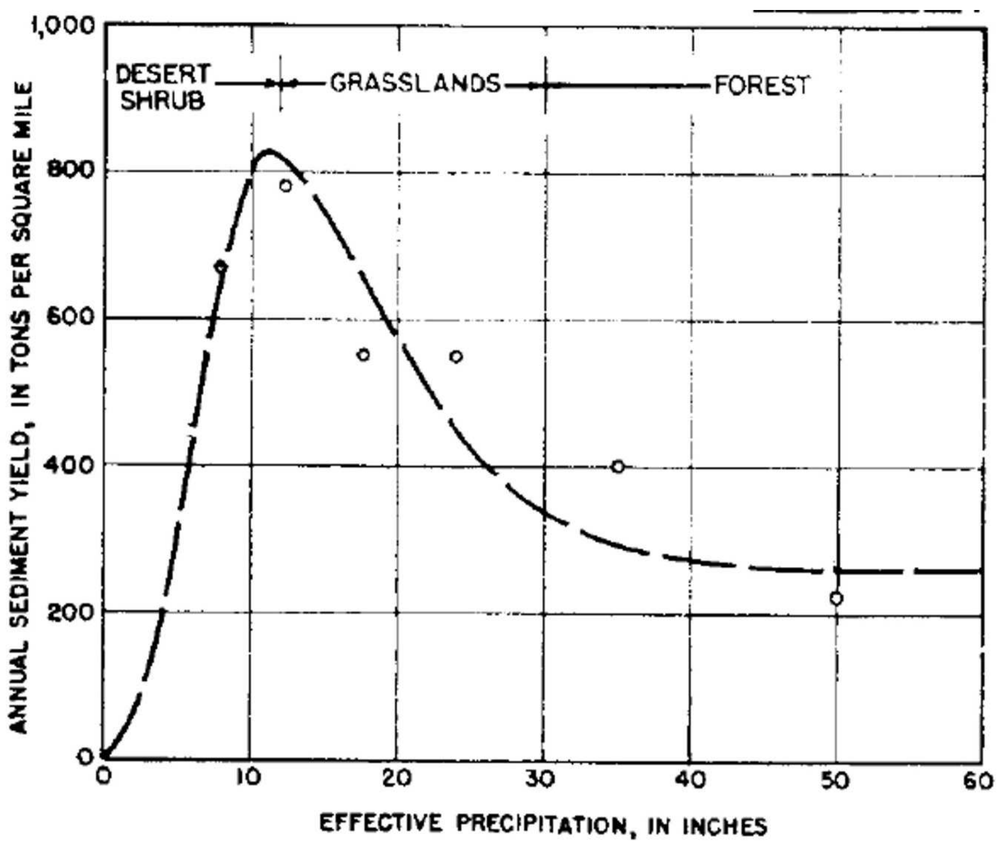



Figure 6. Diagram illustrating the relationship between effective precipitation, annual sediment yield, and vegetation (Langbein & Schumm, 1958; Schumm, 1977).

**Characteristics of plant extinction**

If plant extinction is the primary cause for changes in fluvial stacking patterns one would expect to see a change in flora to occur across the entire Transantarctic basin simultaneously as fluvial stacking patterns change. There should also be no change in sandstone composition, no change in paleoslope, little to no change in regional lithofacies, and a decrease in number and diversity of rooted plants across the PTB. Ward et al. (2000) and Michaelsen (2002) suggest that a global extinction of plants occurred at the PTB, resulting in a change in fluvial style (meandering to braided) and a change in stacking pattern (low to high) due to an increase in sediment yield (see chapter 3 for further discussion). Retallack et al. (2005) also proposed that plant extinction was the main cause for the change in fluvial style across the PTB in Antarctica. Retallack et al. (2005) based their conclusions on the change in fossil flora from Late Permian leaves (*Glossopteris*) and roots (*Vertebraria*) to middle Fremouw leaves (*Dicroidium*). Such a hypothesis for changing fluvial stacking patterns is dependent on the location of the PTB relative to the Buckley Fremouw contact, and that the contact is a time line rather than being diachronus as many formational boundaries are.

## 1.4 Methodology

Over an eight week period during the 2010□2011 Austral season a remote helicopter-supported field camp was established on the Bowden Neve in the BGR. A number of sites in the CTM were visited by a five person field team. Research for this thesis involved measuring numerous stratigraphic sections of sandstone bodies in the Permian Buckley Formation and the Triassic Fremouw Formation. The Permian sections

were measured at Mt. Achenar, Mt. Bowers, and Lewis Cliffs. Two sites on Mt. Achenar were visited on two separate occasions. The location of the first site visited at Mt. Achenar is 84□ 11.353’S 160□ 57.849’E. The location of the second site is 84□ 11.789’ S160□ 58.535’E. Mt. Bowers located at 85°00.122’S 164°09.606’E consisted of a 10 day remote camping trip. A total of 466 m of the Buckley Formation were measured by the field party for this thesis. A total of 8 vertical sections were measured and 1 lateral section was measured on the same sand body to determine any lateral changes in geometry, sedimentary structures, and grain size. Part of the lower Buckley Formation was measured at Lewis Cliffs (84°15.620’S 161°11.615’E) in 4 lateral stratigraphic columns of the same sand body. The Lower Triassic Fremouw Formation was measured at Wahl Glacier (84□5.028’S 165□58.535’E) and Gordon Valley (84□21.196’S 163□50.929’E). At Wahl Glacier about 7 m of sandstone was measured due to encroaching fog and the need for extraction from the site by helicopters before conditions deteriorated to the point where flying was impossible. Previous work by Barrett et al. (1986) and Flaig (2005) as well as photos of the outcrop were used to supplement the identification of fluvial stacking patterns at this site. At Gordon Valley 26.5 m of sandstone was measured in the upper Fremouw Formation. Sedimentologic data collected included: lithology, facies, sedimentary structures, deformational structures, grain sizes, contacts (sharp, gradational, or erosional), paleocurrents, and sandstone geometries and architecture. Detailed stratigraphic columns can be found in Appendix A.

Paleocurrent orientations were taken at Mt. Bowers, Mt. Achenar, Lewis Cliffs, and Gordon Valley. Paleocurrents were taken on ripples (sr), trough cross-stratification (st), foresets (f), and primary current lineation (pcl) where available. The magnetic

declination of the brunton was set to zero and corrections were made back at the lab. Measurements were corrected using the (NOAA National Geophysical Data Center) website (http://www.ngdc.noaa.gov/geomag-web/#declination). The paleocurrents were plotted on rose diagrams in 20˚ classes using the shareware GEOrient (version 9.5.0) by Rod Holcome of the Department of Earth Sciences, the University of Queensland, Australia. Only trough cross-stratification was used to determine paleo-flow trends for this study. Detailed paleocurrent orientation tables can be found in Appendix B.

The following chapters will define facies to determine the types of systems that are operating in the basin and their dynamics (Chapter 2) and define the fluvial stacking patterns and their controls (Chapter 3). Finally, these two sedimentologic tools will be used to address the depositional history of the basin across the PTB.

## 2. Description of Facies Associations

The facies associations for this study have been divided into three groups. The groups are defined by lithology and sedimentary structures. The first group is the coarse-grained sandstone facies association (Table 1), which includes medium- to coarse-grained sandstone sheets (see section 2.1). The second group is the fine- to medium-grained sandstone sheet facies association (Table 2; see section 2.2). The third group consists of the fine-grained facies association, which includes: siltstone, mudstone, and coal deposits (Table 3; see section 2.3).

Table 1. Coarse-grained Sandstone Facies Associations

| Facies | Grain size, contacts, thickness, width, and geometry | Sedimentary structures and fossils | Interpretation | Formation |
|---|---|---|---|---|
| Coarse-grained sandstone | • Medium- to coarse-grained sandstones.<br>• Erosional basal contacts.<br>• Up to 20 m thick.<br>• Extend laterally for 100’s m to km.<br>• Sheet bodies with a few (rare) isolated ribbons.<br>• Sharp or gradational upper contact with adjacent facies. | • Single storied or multistoried.<br>• Abundant trough and planar cross-beds.<br>• Occasional ripple cross-lamination.<br>• Rip-up clasts consisting of intraformational siltstone, mudrock, coal, and subrounded quartz pebbles.<br>• Reactivation surfaces common.<br>• Downstream accretion (dominant), vertical accretion, lateral accretion, and upstream accretion (not every sandstone body contains all of these accretionary deposits).<br>• Sand-filled abandoned channels.<br>• Rare to common transported fossil wood and plant debris.<br>• Paleocurrents show little variation. | • Braided stream channel deposits | • Buckley & Fremouw formations |

Table 2. Fine- to medium-Grained Sandstone Sheet Facies Association

| **Facies** | **Grain size, contacts, thickness, width, and geometry** | **Sedimentary structures and fossils** | **Interpretation** | **Formation** |
|---|---|---|---|---|
| Fine- to medium-grained sandstone sheets | □Fine- to medium-grained sandstone.<br>□Erosional or planar-sharp basal contacts. Some units have gradational bases.<br>□Up to 4 m thick and extends laterally up to 100's m across the outcrop.<br>□Sheets, may also be associated with ribbon bodies (small channels-form bodies).<br>-May occur as wings that thin laterally away from channel-sandstone bodies.<br>□Sharp or erosional upper contact with adjacent facies. | □Large scale trough and planar cross-beds.<br>□Small scale ripple cross-lamination (abundant).<br>□Horizontal lamination containing primary current lineation.<br>□Climbing ripples.<br>□Wavy and lenticular bedding.<br>□Massive bedding.<br>□Deformational structures (flames and convolute bedding).<br>□Paleocurrents vary at high angles with those contained in the Coarse-grained Sandstone Facies Associations. | Crevasse channels and splays, crevasse-splay complexes, and sheet floods | Buckley Formation |

Table 3. Fine-grained Facies Associations

| Facies | Grain size, contacts, thickness, width, and geometry | Sedimentary structures and fossils | Interpretation | Formation |
|---|---|---|---|---|
| Siltstone | Silt-sized grains.<br>Sharp to gradational basal contacts.<br>Sharp or gradational upper contact.<br>Up to 12 m thick and extend 10’s to 1000’s m across outcrop. | Massive or laminated<br>Laminated<br>Horizontal laminations (alternating silt and clay or silt and organic rich layers containing leaf compressions, impressions and carbonized remains).<br>Rare ripples.<br>Rare deformed sandy forset beds. | Massive<br>Floodplain deposit<br><br>Laminated<br>Lacustrine deposit | Buckley and Fremouw formations |
| Mudrock | Clay and silt sized grains.<br>Gradational or sharp basal contact.<br>Erosional or bioturbated (rooted horizon) upper contact.<br>Up to 50 m thick (up to1.5 m thick for this study) and extend 10’s to 100’s m across outcrop. | Carbonaceous in Buckley Formation (grey or black color).<br>Non-carbonaceous in Fremouw Formation (green, red, and brown).<br>Rooted horizons and fossil forests common in Buckley Formation. Rooted horizons also common in the Fremouw Formation. | Floodplain deposit and soil horizons. Poorly drained flood plains (carbonaceous) in the Buckley Formation while the Fremouw Formation mudrocks likely represent well-drained floodplains. | Buckley Formation Fremouw Formation |
| Coal | 0.1 m thick to several meters thick and extends 100’s of m to kms across the outcrop.<br>Sharp or gradational basal contact.<br>Erosional or sharp upper contact. | Interbedded with siltstone, mudrock, and fine-grained sand sheet facies. | Marsh and mire deposits. | Buckley Formation (lower and upper members) and upper Fremouw Formation |

## 2.1 Coarse-grained sandstone facies associations of the Buckley and Fremouw formations

*Buckley Formation:* Coarse-grained sandstone sheets in the lower and upper members of the Buckley Formation are similar and up to 20 m thick and extend laterally for 100's of m to km across the outcrop perpendicular to paleoflow directions taken from primary sedimentary structures. At Lewis Cliffs, the sheets thin dramatically over a few 10's of m to laterally extensive sandstone wings along the margins of the sheets. These wings and associated deposits are described below in section 2.2 with the fine-grained sandstone sheet facies associations. The coarse-grained sandstone sheets can occur isolated within finer-grained deposits or as amalgamated sheets displaying high fluvial stacking patterns (see chapter 3). The basal contact of individual sheets is erosional and generally of low relief, but in places, the base of some sheets can have several meters relief (Figure 7). The sandstone sheets typically cut into deposits of the fine-grained sandstone sheet facies associations (Figure 8; see section 2.2). The sheets are multistoried (Figure 9) and sometimes multilateral and consist of multiple upward-fining packets (FUS; Figure 10) and sandstone-filled abandoned channels. Rip-up clasts (Figure 11) are common directly above the basal erosional surface and above erosional surfaces at the base of each FUS, and consist of intraformational clasts of siltstone, mudstone, and coal. Paleocurrent orientation obtained from planar and trough cross-beds contained within the FUS and abandoned channels show little variation in flow directions (Figure 12). The upper bounding surface of the sheet is sharp to gradational with overlying fine-grained facies associations.

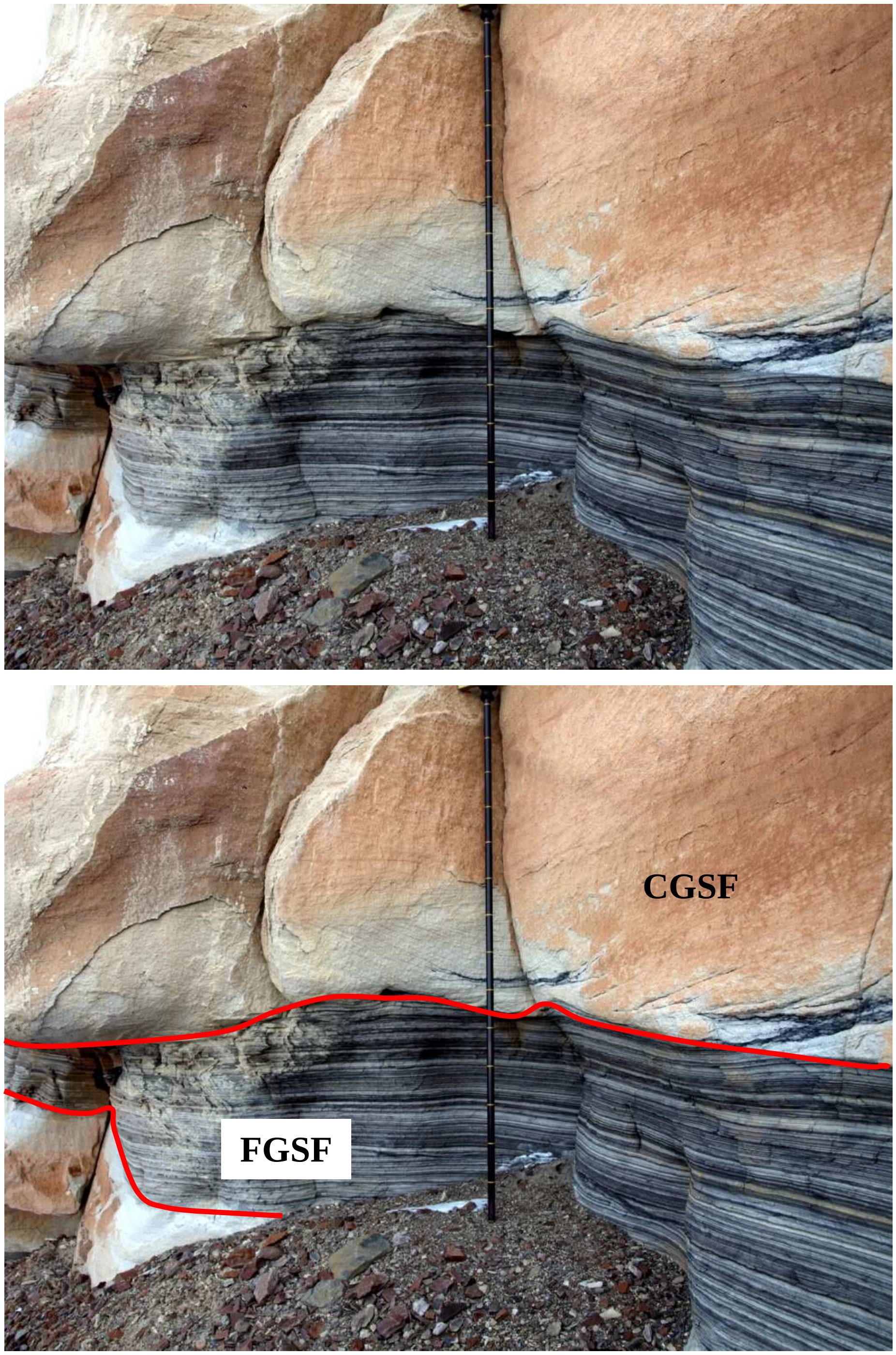



Figure 7. Erosional base of the coarse-grained sandstone facies (CGSF) truncating underlying laminated organic rich fine-grained sandstone facies (FGSF) at Lewis Cliffs. Jacob staff for scale.


Figure 8.  Coarse-grained sandstone facies (CGSF) erosionally overlying the fine-grained sandstone facies (FGSF) interbedded with siltstone at Lewis Cliffs. Person for scale.

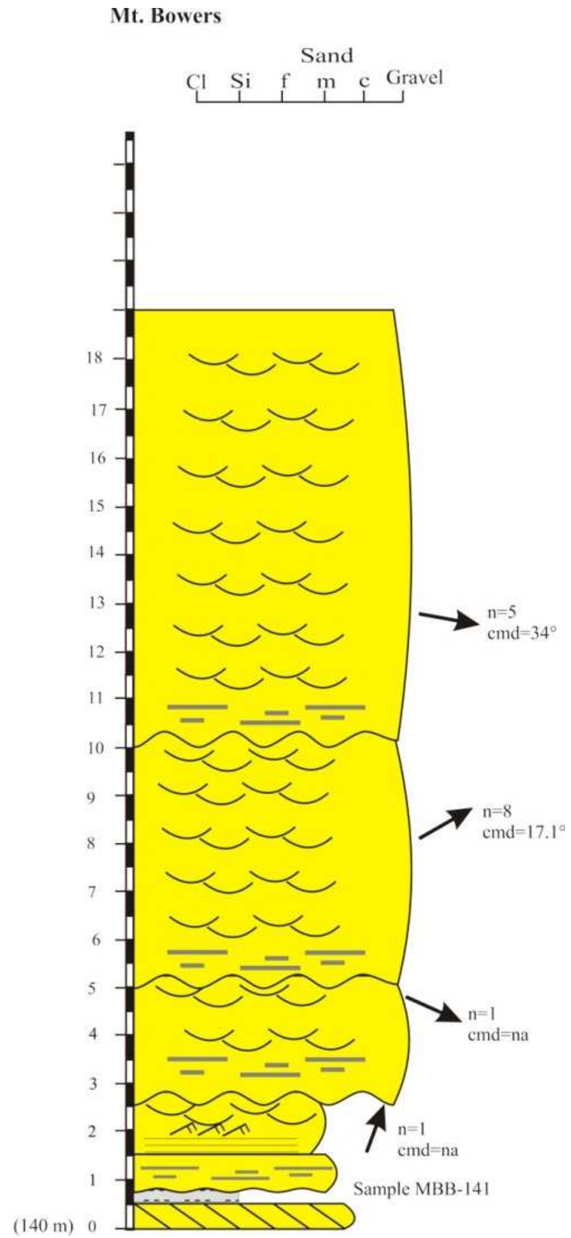



Figure 9. Stratigraphic column showing a multistoried coarse-grained sandstone facies association overlying a thin fine- to medium-grained sandstone sheet facies association at the base of the column at Mt. Bowers.


Figure 10. Multiple medium- to coarse-grained sheet sandstones in the upper Buckley Formation at Mt. Achenar. Each sandstone sheet contains multiple FUS successions (represented by blue arrows) that define macroforms. The third sandstone from the bottom of the photograph is ~ 14m thick.

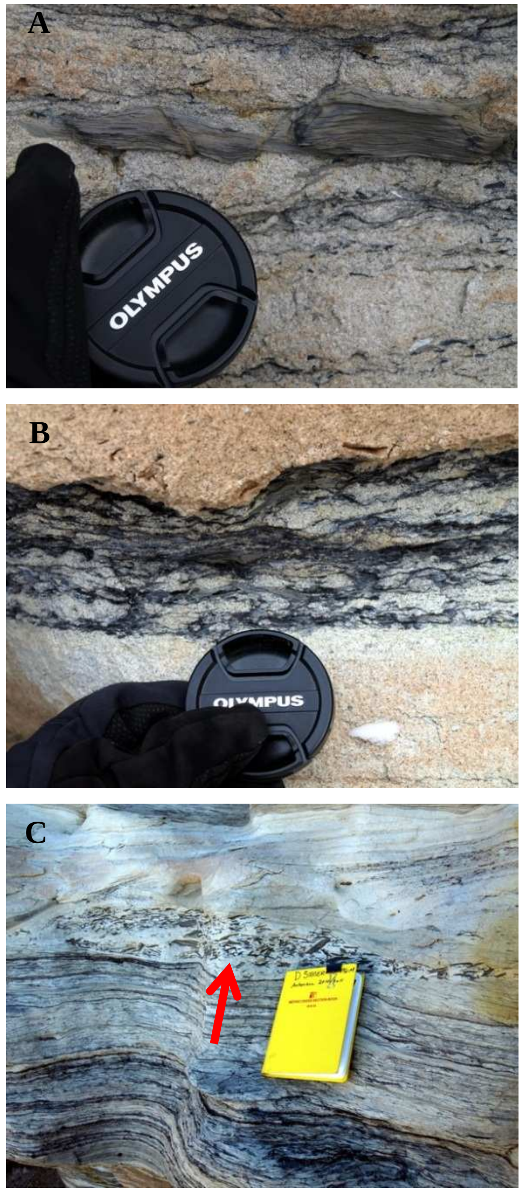



Figure 11. Rip-up clasts in the medium- to coarse-grained sandstone facies association from the Buckley Formation. The rip-up clasts of the Buckley Formation consists of siltstone (A), carbonaceous mudrock and coalified plant debris (B and C). Lens cap and field notebook for scale.

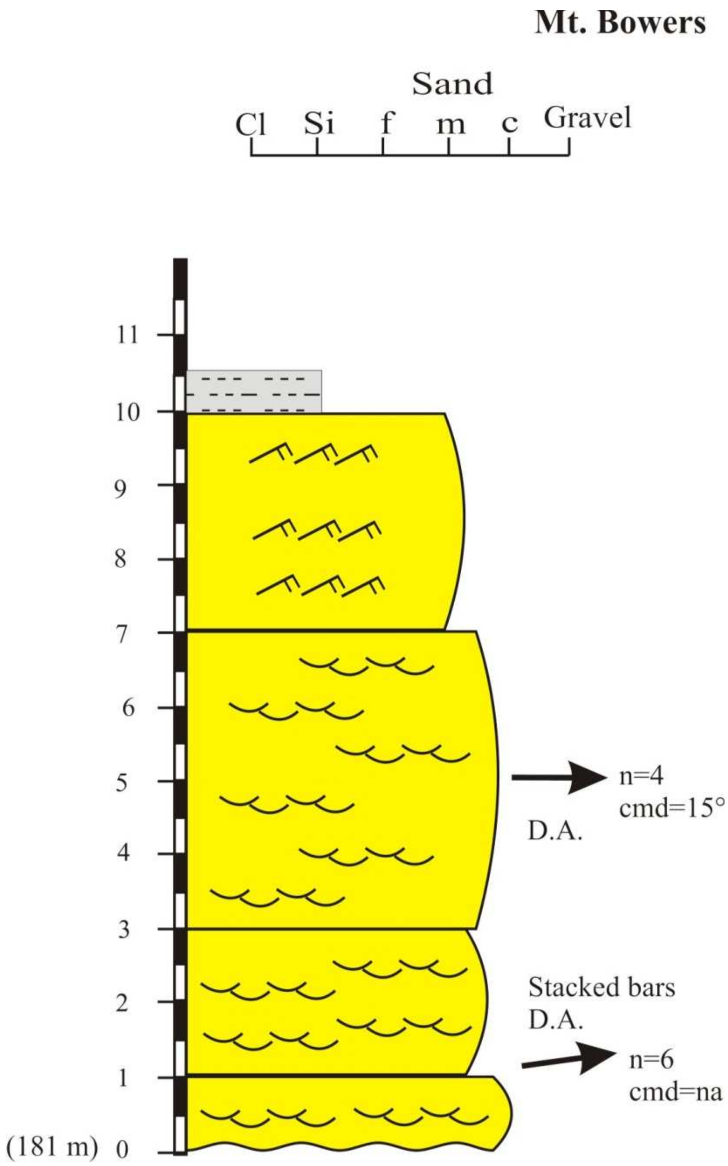



Figure 12. Stratigraphic column showing medium- to coarse-grained sandstone facies with little paleocurrent variation in this FUS in the lower Buckley Formation, Mt. Bowers. Black arrows represent azimuth directions, where north is towards the top of the page. n= the number of paleocurrents and cmd= the circular mean deviation.

Within an FUS, coarse-grained trough and planar cross-bedded (dunes) (Figure 13) sandstones rest on a basal erosion surface of low relief (cm to dm), and progressively, each packet fines upward to medium- and occasionally fine-grained ripple cross-laminated sandstone. Downstream accretion (Figure 14) and vertical accretion are

abundant within the packets, and reactivation surfaces (Figure 15) occur within individual cross-bed sets (erosional reactivation and some organic-rich fine-grained drapes). Each FUS is up to several meters thick and can be traced laterally across the outcrop for 10’s to 100’s of m. Individual packets are truncated by the basal erosional surface of the next overlying FUS package. Silicified wood, wood impressions and logs were found at Mt. Bowers and Mt. Achenar but are rare in the medium- to coarse-grained facies association (Figure 16).

Abandoned channels are scoop shaped and are a few to 10 m thick and are 10’s to a few hundred meters wide. The channels are filled with medium- to coarse-grained trough cross-bedded sandstone. These channels occur as both individual channel bodies (single storied) cut within the FUS and as stacked channels cutting into each other both vertically (multistoried) and laterally (multilateral). Channel fill consists of both lateral accretion and vertical accretion deposits. Vertical accretion, however, is possibly downstream accretion that is cut perpendicular to flow.

*Interpretation:* Medium- to coarse-grained sandstone sheets within the Buckley Formation contain characteristics that suggest deposition from sandy braided streams. The occurrence of laterally extensive multistoried, multilateral sand sheets with extensive wings are interpreted as levee deposits that extended out into the surrounding floodplain (see fine-grained facies associations for further description and discussion), which

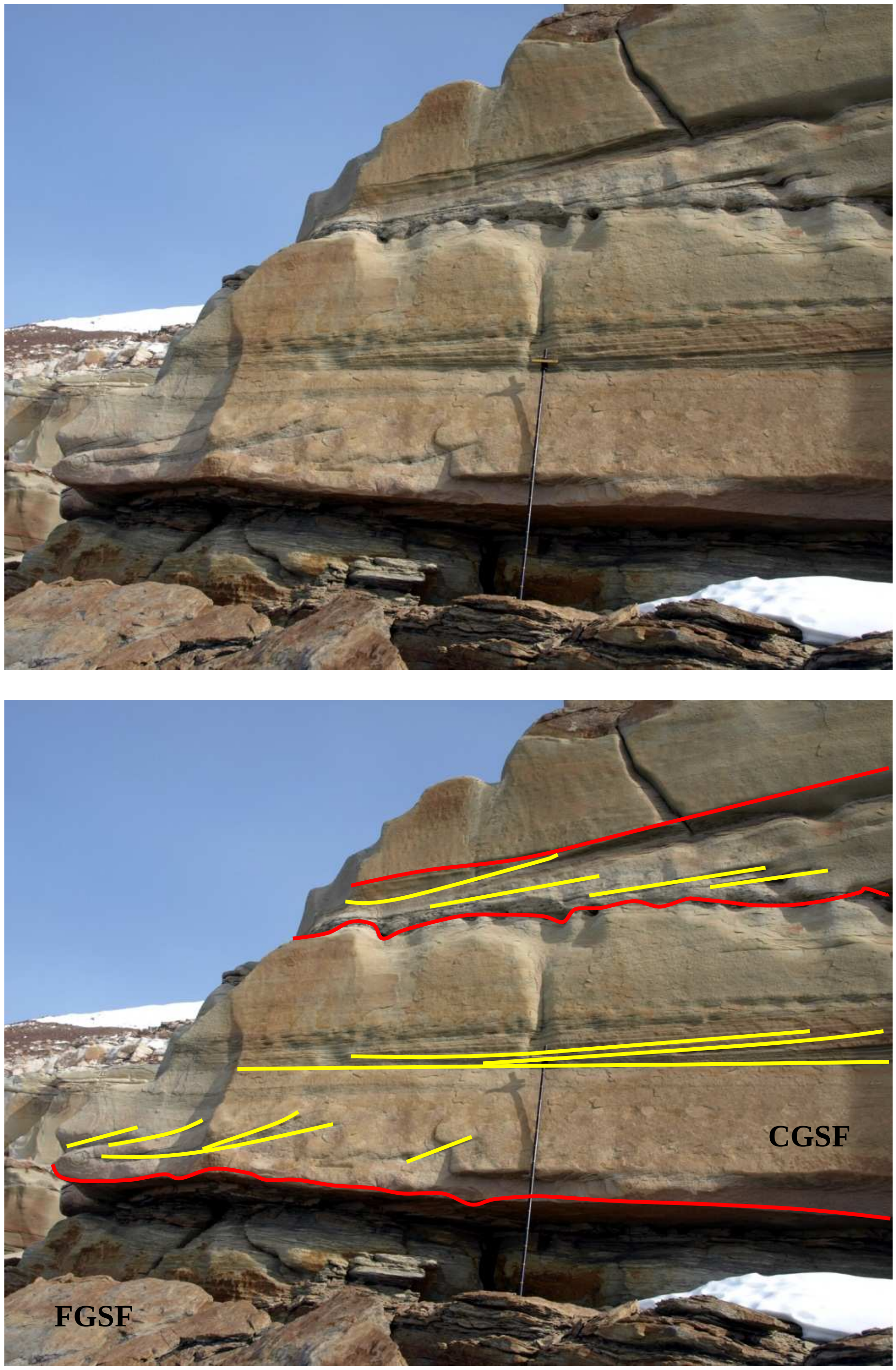



Figure 13. Multistoried medium- to coarse-grained sandstone facies showing large-scale cross-stratification and small-scale low angle cross-stratification truncating underlying fine-grained sandstone facies at Mt. Bowers. Jacob staff for scale.

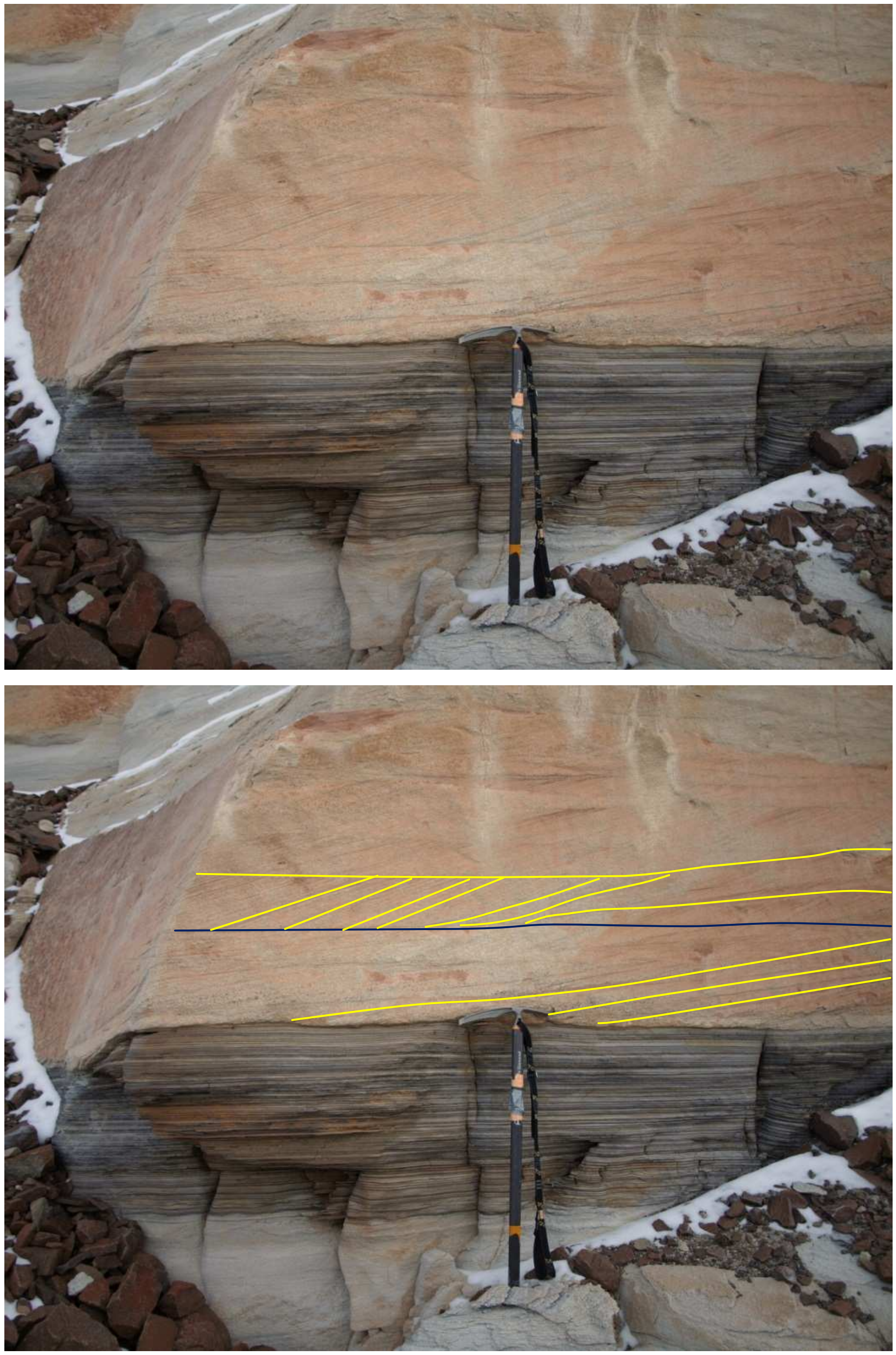


Figure 14. Down climbing ripples to possibly small dunes (yellow) overlain by nicely preserved downstream accretion at Lewis Cliffs. Ice axe for scale.

Figure 15. Multistoried medium- to coarse-grained sandstone facies with several erosional bounding surfaces that are the base of FUS successions within individual channels. The medium-to coarse-grained facies truncates underlying fine-grained sandstone facies at Mt. Bowers. Person for scale.

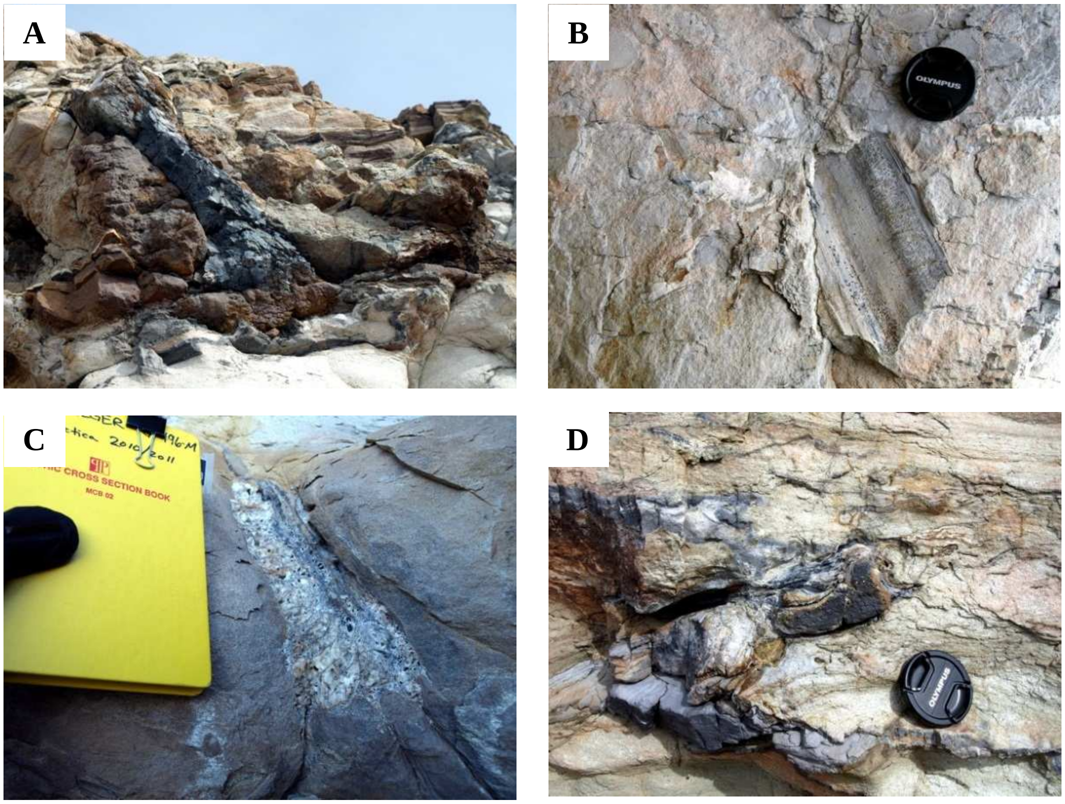



Figure 16. Medium- to coarse-grained sandstone facies containing (A) a preserved log, (B) wood impression with lens cap for scale, (C) Silicified wood with notebook for scale, and (D) petrified wood with lens cap for scale in the Buckley Formation.

suggests that these streams were laterally mobile on their alluvial surfaces (cf., Friend, 1983). Low paleocurrent dispersion suggests deposition by bars oriented parallel to the trend of channels within a relatively straight channel system (Miall, 2003).

The FUS contain abundant downstream accretionary surfaces. Such surfaces indicate deposition along the front of migrating macroforms or bars (cf., Bridge, 1986; Miall, 2003). Each package likely represents the migration of a single longitudinal bar. Reactivation surfaces could represent a halt in sedimentation following individual floods especially where fine-grained or organic-rich, fine-grained drapes line the accretionary

surfaces of these macroforms (cf., Miall, 1985; Nichols & Fisher, 2007; Bridge & Demicco, 2008).

The sand-filled abandoned channels indicate that coarse-grained sediment (sand) was transported down the channels during the abandonment phase. This happens under conditions of low-angle channel abandonment within braided streams, which allows for continued, but diminishing, discharge as a channel segment is gradually abandoned (Allen, 1964; Jackson, 1978; Miall, 1985; Bridge, 2006; Nichols & Fisher, 2007). Abandoned channels are ribbon shaped representing stable channel switching rather than lateral migration of channels (cf., Friend et al., 1979, 1983). The internal architecture consists of lateral accretion indicating the channels were filled by bank attached bars, vertical accretion representing gradual abandonment of channels, and downstream accretion indicates the migration of bars downstream (cf., Bridge, 1986; Bristow, 1987; Miall, 2003).

The thinning and interfingering with fine-grained sandstone sheets is consistent with deposition in sandy braided stream systems within aggrading river settings (cf., Miall, 1985, 2003; Bridge 1984, 1985; Bristow et al., 1999).

*Fremouw Formation:* The medium- to very coarse-grained sandstone sheets occur primarily in the lower and upper members of the Fremouw Formation. These sheets are up to 20 m thick and extend laterally for 100's of m to km across the outcrop. The sheets display high fluvial stacking patterns (see chapter 3). There is a difference, however, in the fluvial stacking patterns in the lower, middle, and upper members of the Fremouw Formation. For example, the lower Fremouw Formation at Wahl Glacier lacks

fine-grained facies associations (floodplain) between individual sand sheets. At Gordon Valley, the upper member of the Fremouw Formation has some fine-grained facies associations between individual sand sheets, whereas the middle member, which is exposed at both sites, consist of several hundred meters of fine-grained facies association with few coarse-grained sandstone sheets. The basal contact of individual sheets is generally planar with low relief. The bases of individual sandstone sheets, however, are erosional and cut typically into fine-grained sandstone sheets and floodplain facies associations (Figure 17). Gravel is common directly above the erosional surface at the base of the sandstone sheets and consists of subrounded quartz pebbles and intraformational clasts of siltstone and mudstone (Figures 17 and 18). The upper bounding surface of the sheet is sharp with the overlying fine-grained sand sheet facies associations.

Figure 17. Coarse-grained sandstone facies with large siltstone rip-up clasts (black arrows) truncating underlying mudrock facies (~1 m thick) of the lower Fremouw Formation at Wahl Glacier.

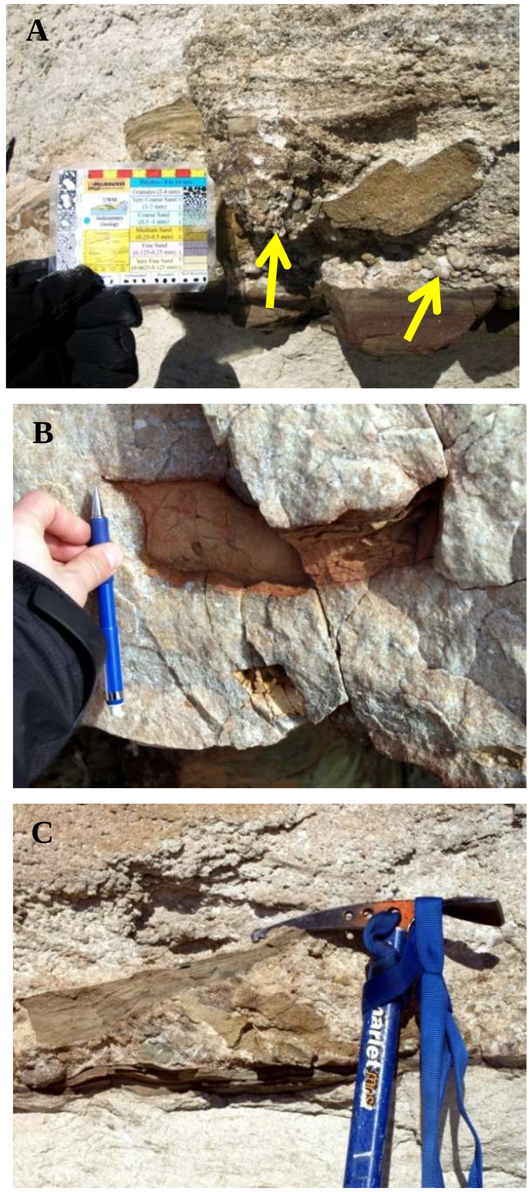



Figure 18. Rip-up clasts and gravel contained in the coarse-grained sandstone sheet facies association in the Fremouw Formation consisting of subrounded to subangular quartz pebbles with a grain size comparator card (10 cm long) for scale (yellow arrows; A) and non-carbonaceous siltstone and mudrock (B and C) with a pencil and ice axe for scale respectively.

The coarse-grained sandstone sheets are multistoried and multilateral packages containing upstream and downstream accretion (Figure 19). Downstream accretion is dominant throughout the sheets, and in some places the downstream accretion surfaces are contained between accretion surfaces that dip at low angles in the upstream direction. If traced laterally in the direction of paleoflow, however, some of the upstream accretion surfaces reach a peak or crest and then descend in a downstream direction (in the direction of paleoflow indicators) beyond the crest. Downstream accretion occurs between the higher-order upstream accretion surfaces. Therefore, the upstream accretion surfaces define macroforms (bars). Some of the upstream accretion surfaces have the appearance of climbing macroforms, especially in the upper member of the Fremouw Formation.

The sheets contain multiple stacked macroforms (multistoried; Figure 20). The macroforms may consist of a fining-upward package bounded by either basal erosion surfaces or by horizontal or upstream accretion surfaces, or may consist of a package of sandstone that shows little vertical variability in grain size (Figure 21). Within these multistoried sand sheets, very coarse-grained trough and planar cross-bedded (dunes) sandstone rests on the planar to low relief (cm to dm) erosional surface near the base of the packet. Packages progressively fine upward (FUS) to medium-grained, smaller, trough cross-bedded sandstone. Downstream accretion is abundant and some erosional reactivation surfaces occur on cross-beds. In some places, foresets of the macroforms become folded and consist of meter-scale convolute beds (Figure 22). Each macroform or story is up to several meters thick and can be traced laterally across the outcrop for

Figure 19. Coarse-grained sandstone sheet facies association in the upper Fremouw Formation at Gordon Valley. Photo shows downstream accretion and cross-bed sets (yellow lines) and upstream accretion (blue line) with a hammer for scale (Photograph courtesy of John Isbell).

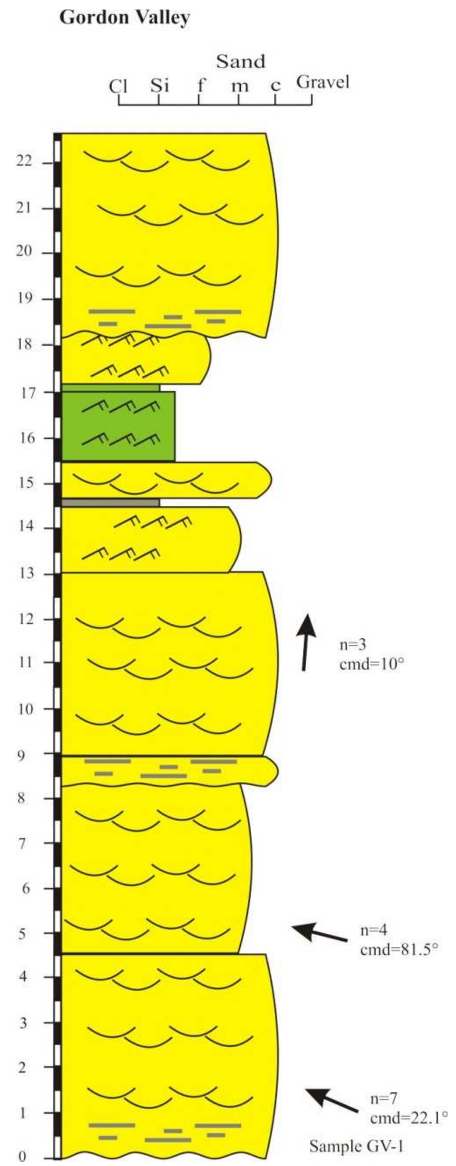



Figure 20. Stratigraphic column from Gordon Valley showing multistoried coarse-grained sandstone sheet facies association. Paleocurrents show little variation in each FUS.

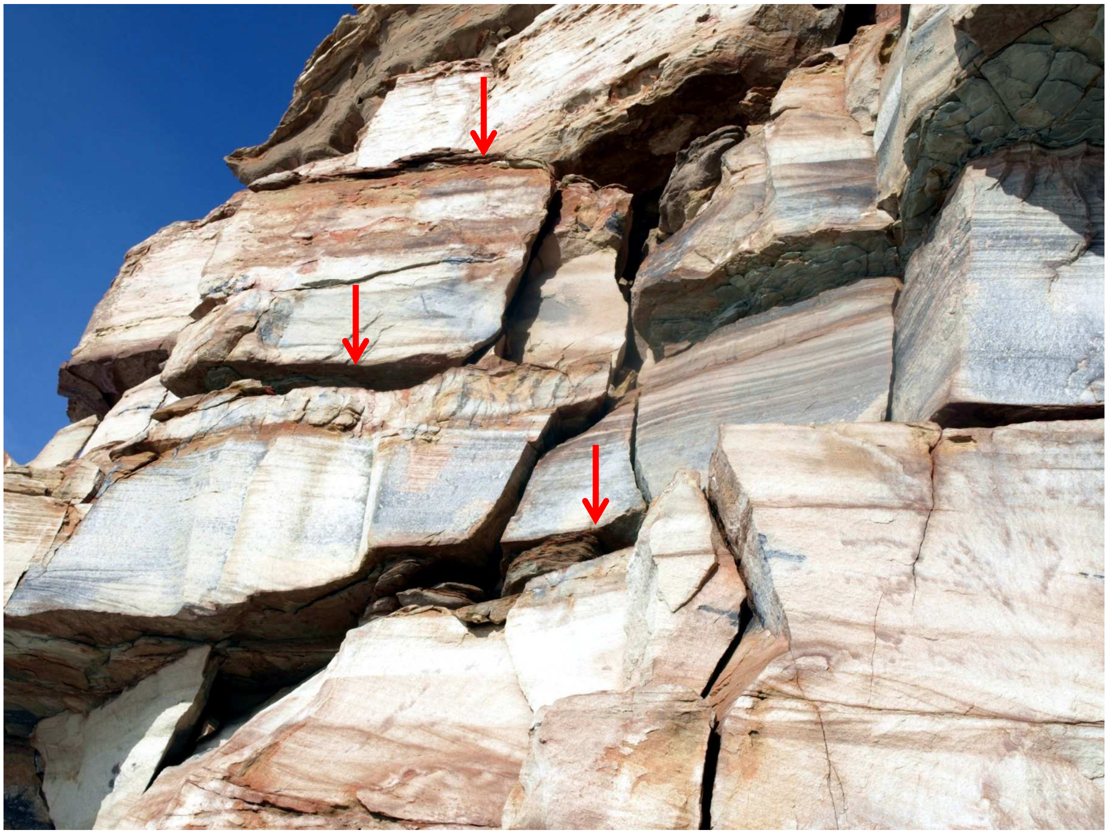


Figure 21. Fining upward coarse-grained sandstone facies showing bounding surfaces (red arrows) of individual macroforms in the lower Fremouw Formation, Wahl Glacier. Bottom FUS ~ 1 m thick.

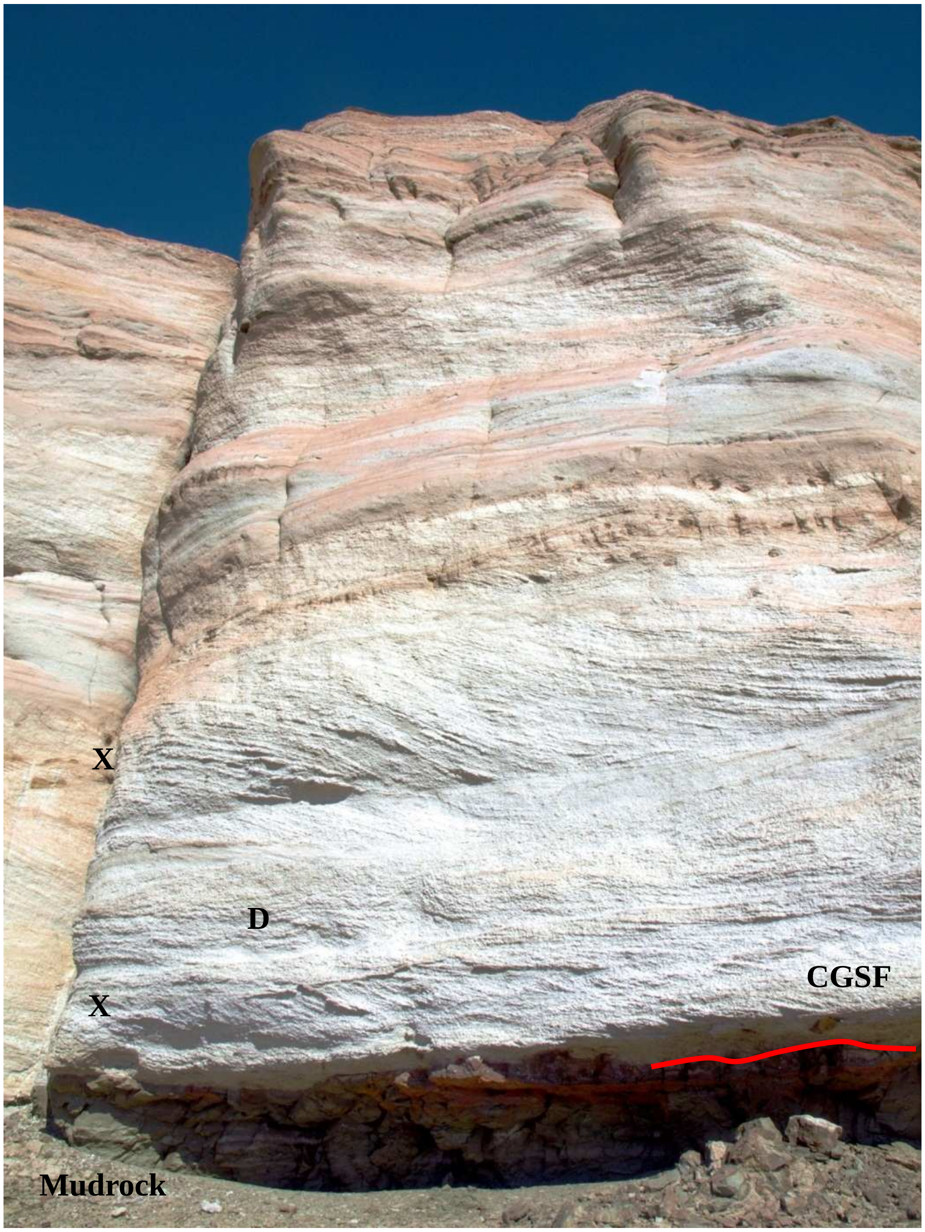



Figure 22. Coarse-grained sandstone facies (CGSF) erosionally overlying mudrock in the upper Fremouw Formation at Gordon Valley. Photo includes (X) large-scale cross-stratification and (D) deformed cross-beds. Sandstone ~12 m thick.

10's to 100's of m. The abandoned channels are both sheet like and ribbon shaped and are a few meters to 10's of m thick. The channels are filled with medium- to very coarse-grained trough cross-bedded sandstone (Figures 22-24). These channels occur as stacked channels cutting into each other vertically. Channel fill consists of both lateral accretion,

vertical accretion, and possibly downstream accretion cut perpendicular to flow (Figure 23). Paleocurrent orientations obtained from planar and trough cross-beds within the macroforms and abandoned channels show little variation in flow direction (Figure 20).

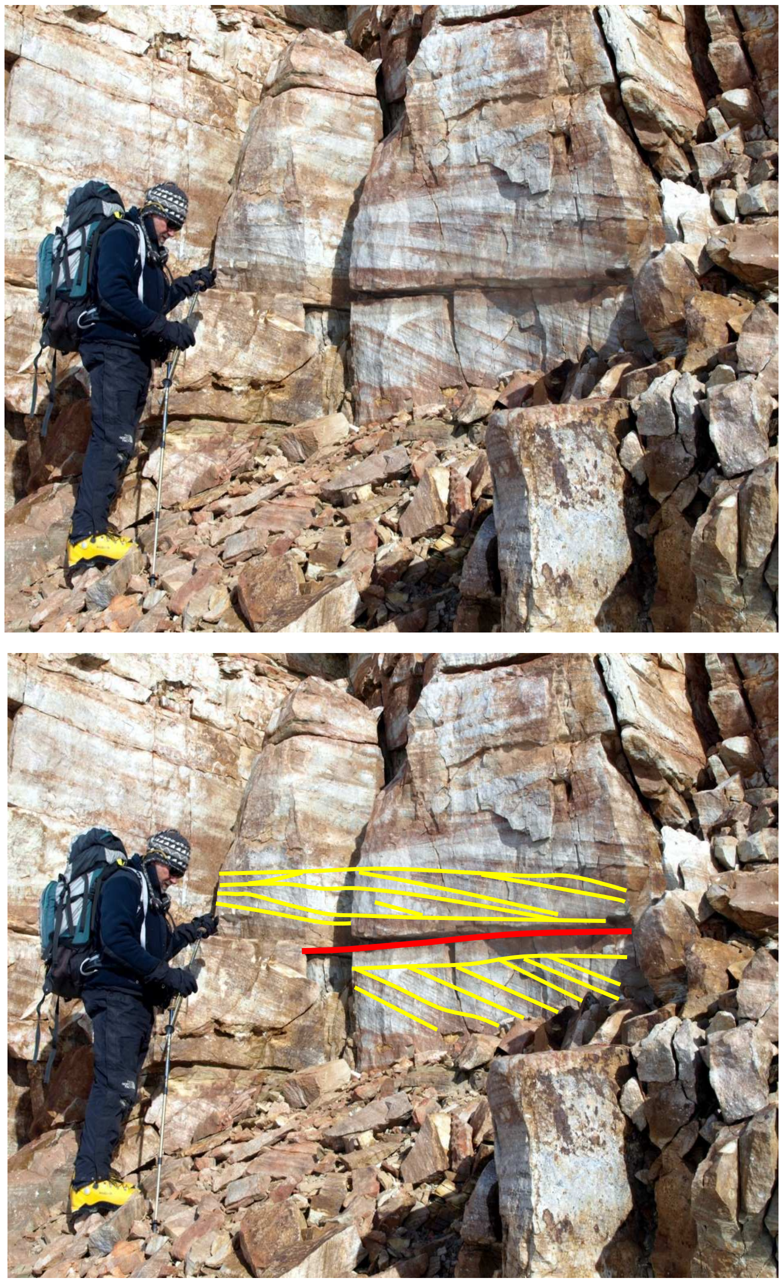


Figure 23. Two downstream accreting macroforms (yellow) separated by a bounding surface (red) at Wahl Glacier. Person for scale.

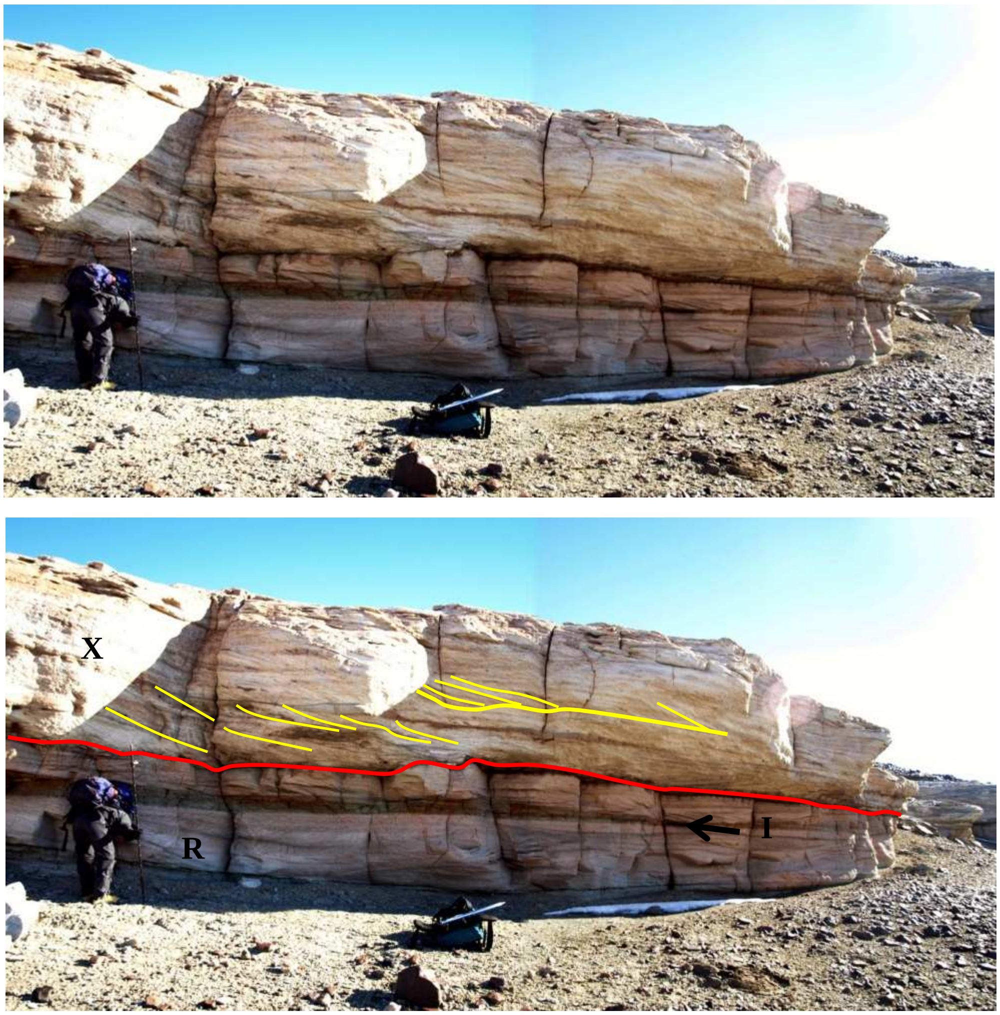



Figure 24. Coarse-grained sandstone facies (red line) truncating fine- to medium-grained sandstone facies interbedded with very fine sand and siltstone (I; black arrow) that thins in the direction of flow at Gordon Valley. Photo shows (X) large-scale cross-stratification and (R) small-scale cross-stratification. Person for scale.

*Interpretation:* Medium- to coarse-grained sandstone sheet facies associations within the Fremouw Formation contain characteristics that suggest deposition from sandy braided streams. The occurrence of multistoried and multilateral sheet sandstones suggests that these streams may have migrated laterally on their alluvial surface within the channel belt (cf., Friend, 1983). Low paleocurrent dispersion suggests deposition by bars oriented parallel to the trend of channels within a relatively straight channel system (Miall, 2003).

The packages of multistoried macroforms contain abundant downstream accretion separated by upstream accretionary surfaces. The upstream accretionary surfaces indicate the bars are overtaking bars located downstream, and the bars are climbing up the back of the downstream bars. The downstream accretion represents deposition along the front of migrating macroforms or bars (cf., Bridge, 1986; Miall, 2003). In some places foresets become folded and convolute suggesting rapid sedimentation rates and collapse of the bar fronts. Each package likely represents the migration of a single longitudinal bar. Reactivation surfaces could represent a halt in sedimentation following individual floods especially where fine-grained facies associations drape the accretionary surfaces of these macroforms (cf., Williams & Rust, 1969; Collinson, 1970; Reineck & Singh, 1980; Miall, 1996; Reading, 1996).

The sandstone-filled abandoned channels represent low angle abandonment. Some abandoned channels are sheet-like with flat-bottomed, scoop-shaped scours at the end of the channel bodies representing lateral migration of channels and some are ribbon-shaped suggesting stable channels with no migration (cf., Friend, 1983). The internal architecture consists of lateral accretion indicating the channels were filled by bank

attached bars, vertical accretion representing gradual abandonment of channels, and downstream accretion indicating the downstream migration of the bars (cf., Bridge, 1986; Bristow, 1987; Miall, 2003).

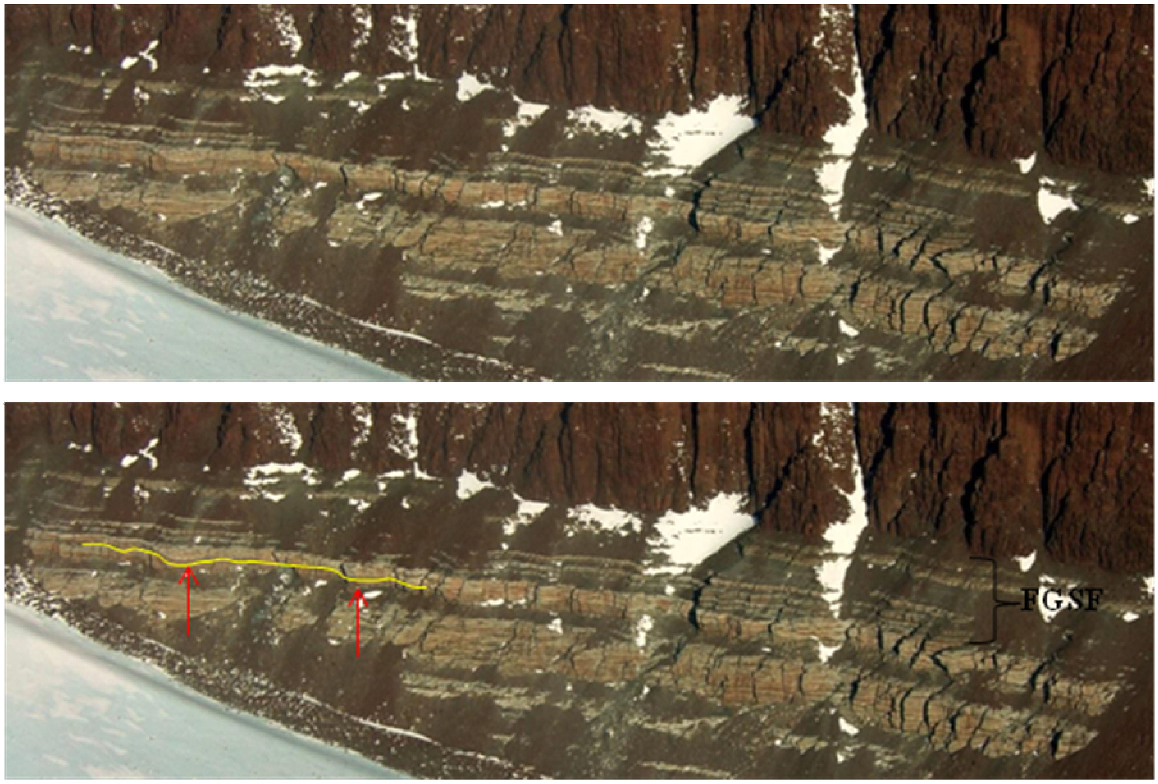
The presence of these multistoried, laterally extensive sandstone sheets lacking significant fine-grained deposits with multistoried macroforms is consistent with deposition in sandy braided stream systems where little floodplain sediment is preserved (cf., Allen, 1965, 1978; Bridge & Leeder, 1979; Bridge, 1985; Miall, 2003). The multistoried macroforms exhibiting downstream accretion, upstream accretion, and vertical accretion are consistent with the buildup and progradation of bars contained within a much larger primary channel.

Differences in the abundance of coarse-grained sandstone sheets between the three informal members of the Fremouw Formation will be discussed in Chapter 3.

### 2.2 Fine- to medium-grained sandstone sheet facies association

The fine- to medium-grained sandstone sheet facies associations are as much as 4 m thick and extend laterally for 100's of m across the outcrop in the direction of paleoflow (Figure 25). The contact at the bases of these sheets consists of both low relief (cm to dm) erosional, planar-sharp, and gradational contacts. Erosional contacts, with associated intraformational rip-up clasts, typically cut into siltstone, mudrock, and coal deposits (see 2.3 for further description and discussion). These surfaces typically grade laterally into planar-sharp contacts in the direction of paleoflow.

A




B

Figure 25. Photograph of the lower Buckley Formation at Lewis Cliffs. A) Photograph and line drawn (yellow) shows the FGSF extending across the outcrop for 10’s of meters. The yellow line outlines ribbon shaped sandstone bodies (red arrows) with wings extending as sheet like bodies for 100’s of meters. B) Close up of the ribbon shaped sandstone body (underlined in yellow) from photograph A. This particular deposit is ~ 15 m thick and is the channel sandstone deposits that the wings are attached to. (Photographs courtesy of John Isbell). The stratigraphic section is approximately 125 m thick.

This facies association is abundant throughout the Buckley Formation. Similar sandstone sheets also occur in the Fremouw Formation. In the Fremouw Formation, this facies association consists of fine- to coarse-grained sandstone and only occurs in the

middle and upper members of the formation in the study area. However, just to the south of the study area, at Graphite Peak and in the Shackleton Glacier region, the lower member of the Fremouw Formation also contains strata of this facies association. In both the Buckley and the Fremouw formations, fine- to medium-grained sandstone sheet facies associations thin and fine in the direction of paleoflow (obtained from cross-stratification). Distally, the sheets pinch out into strata of the fine-grained facies associations (see 2.3 for description). Within the sheets, in the direction of paleoflow, sedimentary structures progressively change from large-scale trough and planar cross-bedding (dunes), to ripple cross-lamination (Figure 26), to interstratified ripple cross-laminations, and horizontal laminations (Figure 26) containing primary current lineation (PCL: plane bed). Climbing ripple-laminations are common within the distal portions of these sandstone sheets (Figure 27) as does wavy and lenticular bedding (Figures 28 and 29). At Lewis Cliffs where the margins of coarse-grained facies associations (see section 2.1) are exposed, the channel-like bodies thin over a few 10's of m into sandstone wings that extend as sheet-like bodies for 100's of m away from the thicker sandstone bodies (Figure 25). Small channels, up to 10 m wide and 1-2 m deep, occur within these wings.

The sheets are typically interbedded with siltstone and coal-bearing mudrock deposits (Figure 30). The sheets also commonly occur directly above coal seams and occur as stacked amalgamated sheets. Abrupt changes in grain sizes and paleocurrent orientations (obtained from small trough cross-beds) occur between packages of amalgamated sheets (cf. Isbell, 1990). Paleocurrent orientations derived from cross-stratification contained within the individual packages show variation in flow direction at high angles to the coarse-grained sandstone facies associations (channels) and

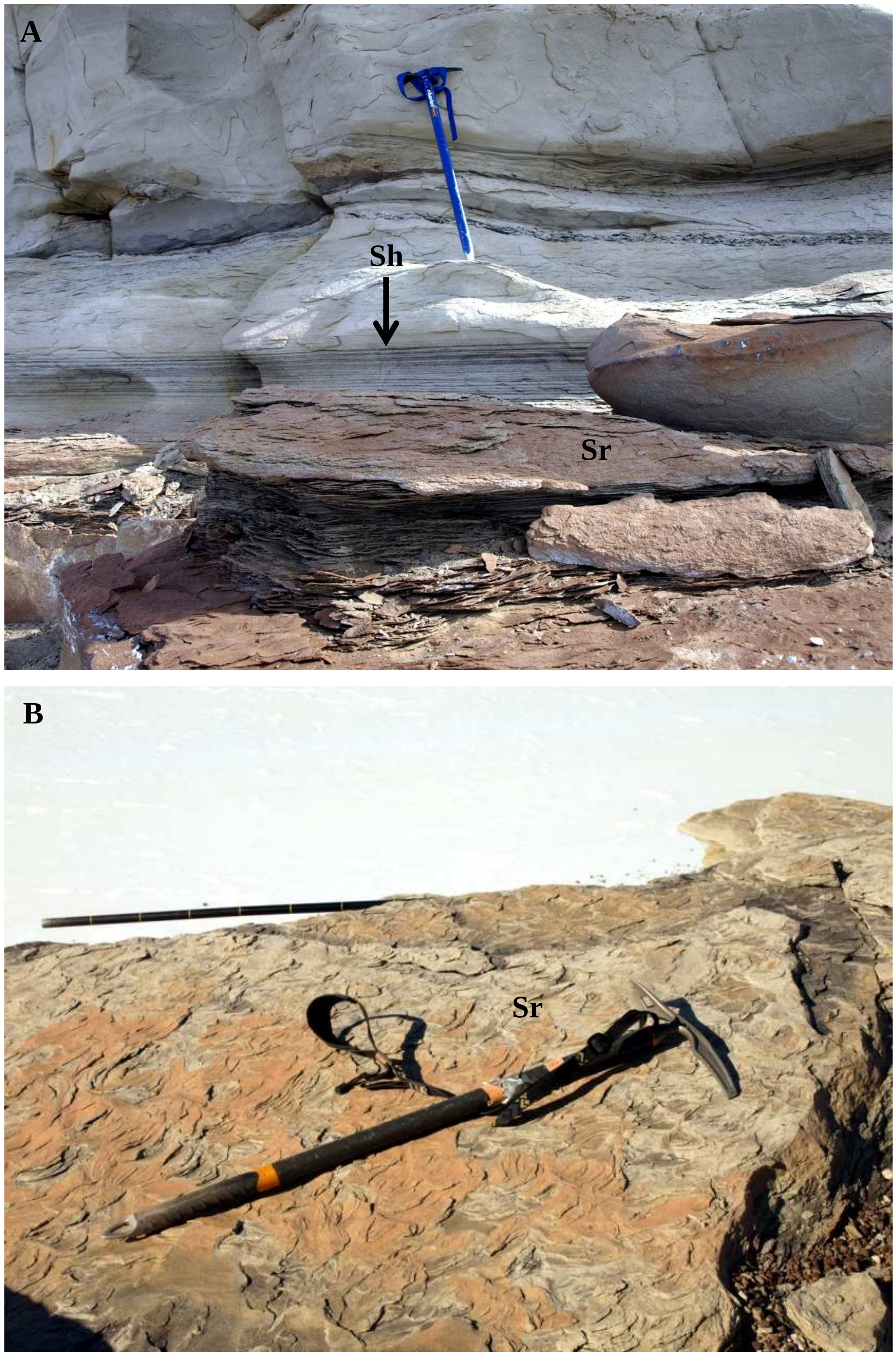



Figure 26. Sedimentary structures of the fine-grained sandstone facies in the Lower Buckley Formation at Mt. Bowers. (A) Photo showing ripples overlain by horizontal lamination in and (B) a rippled surface in plain view (rib and furrow) with ice axes for scales.

Figure 27. Climbing ripples (within black bracket) within the FGSF in the lower Buckley Formation at Mt. Bowers. Jacob staff for scale.


Figure 28. Fine-grained sandstone facies in the lower Buckley Formation at Mt. Bowers. Photo shows Wavy (W), lenticular (L), and horizontal laminations (Sh) with a Jacob staff for scale.

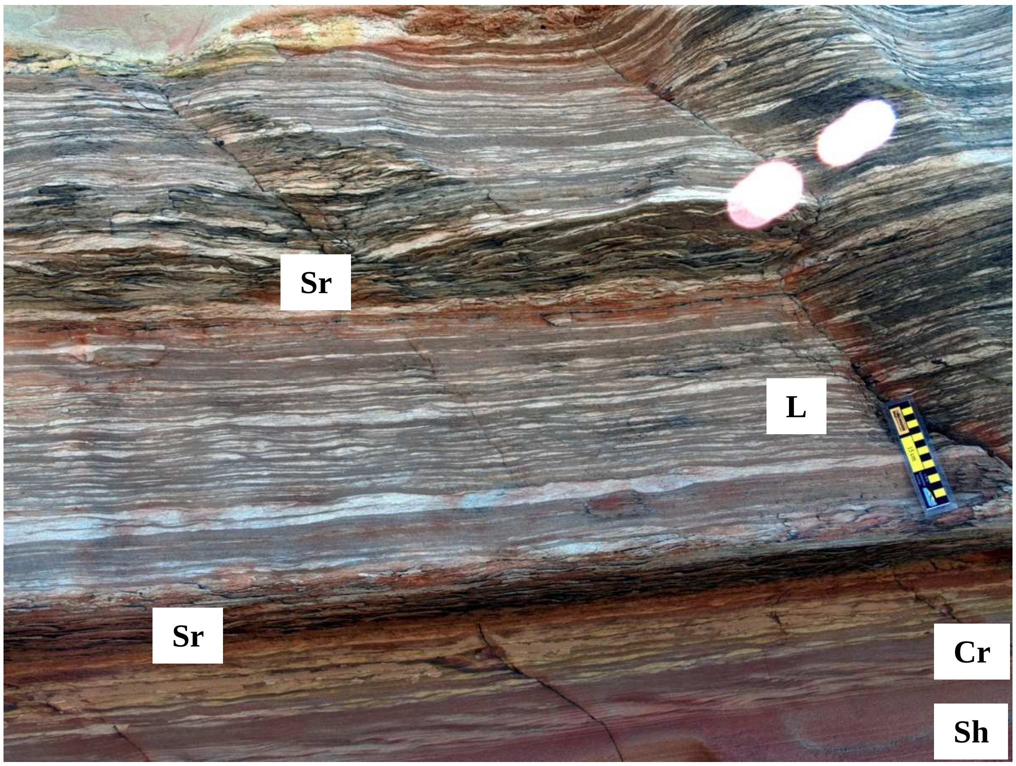



Figure 29. Close up photo of the fine-grained sandstone facies in the lower Buckley Formation at Mt. Bowers. Photo shows a transition from horizontal laminations (Sh), to climbing ripples (Cr), to ripples (Sr), to lenticular bedding (L), to ripples, and back to lenticular bedding. Scale is 15 cm long.


Figure 30. Fine-grained sandstone facies (FGSF) interbedded with mudrock and an erosional upper contact with the medium- to coarse-grained sandstone facies (CGSF) at Mt. Bowers. Sandstone is ~ 10 m thick.

are subparallel to perpendicular to the paleoslope obtained from channel sandstone facies (Figure 31). The packets contain medium-grained small-scale trough cross-beds, plane beds, and ripple laminations, which may fine upward from fine-to very fine-grained ripple cross laminations, climbing ripples, horizontal lamination, and wavy/lenticular bedding. Some sand sheets, within the siltstones, were observed to coarsen upward. Massive bedding also occurs in some medium-grained sandstone sheets. Furthermore, tetrapod burrows have been recorded at Wahl Glacier in the Fremouw Formation (Sidor et. al., 2008). Deformational structures such as flame structures and convolute bedding are also common (Figure 32). Where present, fining upward (FUS) sandstone packets are up to 4 m thick and can be traced laterally 10's to 100's of m across outcrops. Individual packets are either truncated by the basal erosional surface of the next overlying FUS package or show a sharp planar contact when overlain by floodplain deposits. The upper contact is sharp when overlain by fine-grained facies associations, but can be erosional when overlain by coarse-grained facies associations.

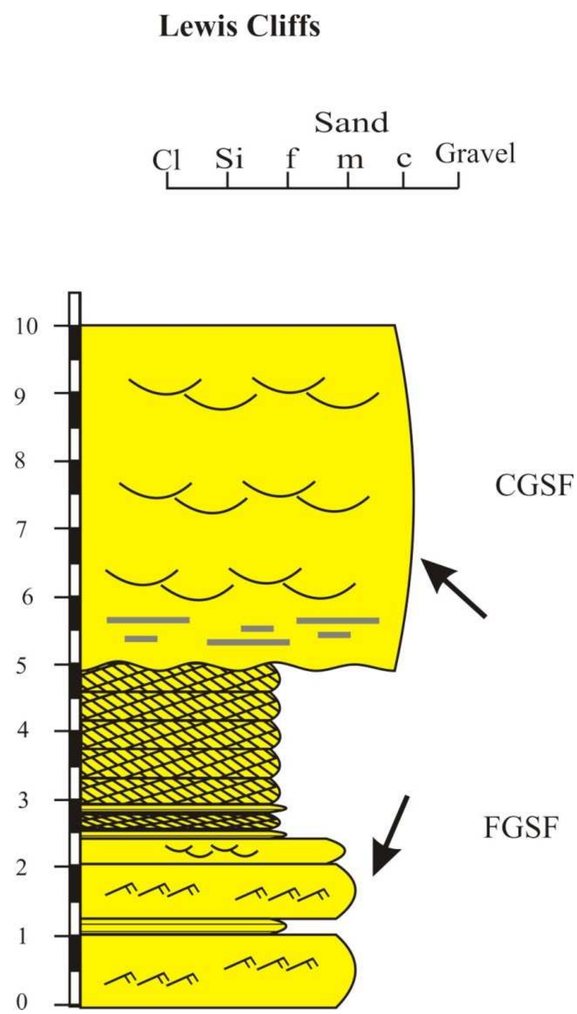



Figure 31. Stratigraphic column from Lewis Cliffs showing the coarse-grained sandstone facies erosionally truncating the fine-grained sandstone facies. Photo shows variation in paleocurrent directions between the two facies.

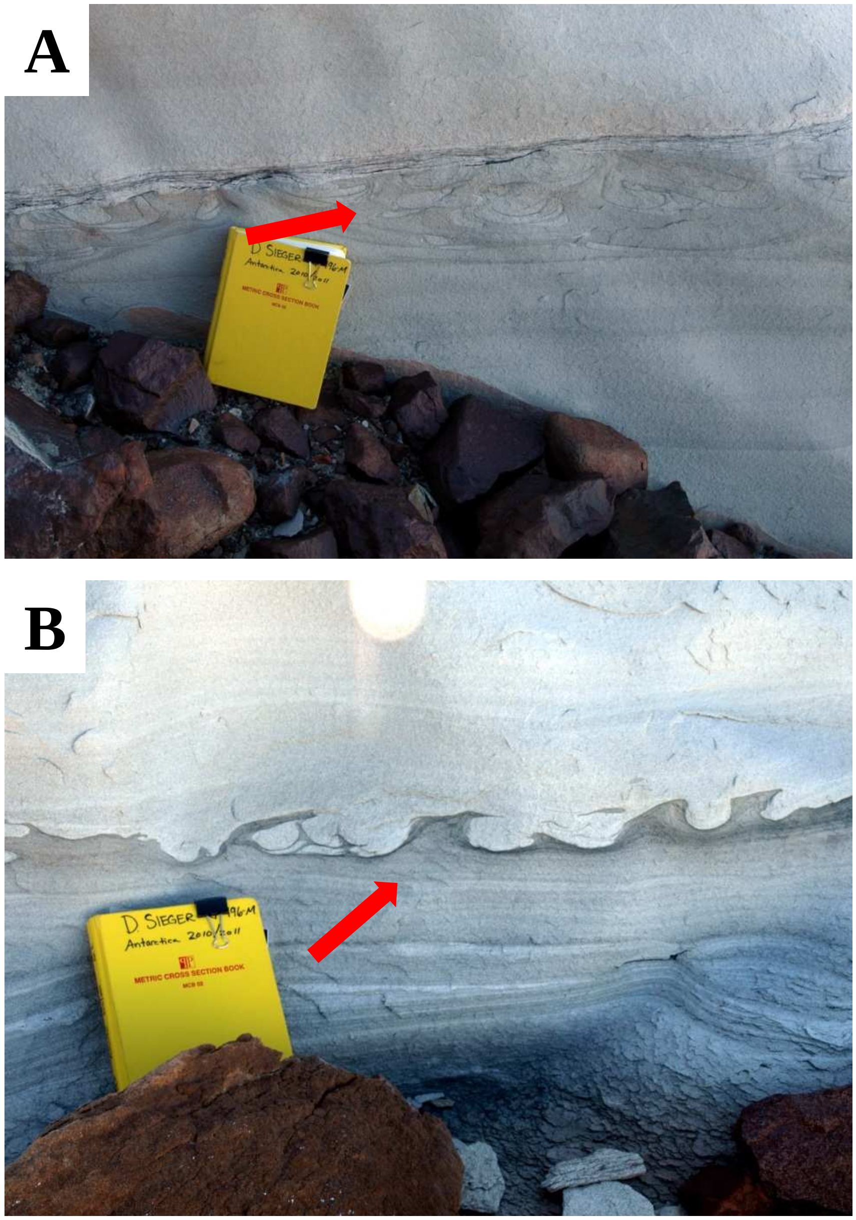



Figure 32. Soft sediment deformation structures in the lower Buckley Formation, Mt. Bowers. Photo shows: Convolute bedding (A) and flame structures (B) with a notebook ~17cm wide for scale.

*Interpretation:* Fine- to medium-grained sandstone sheet facies associations contain characteristics that suggest deposition from crevasse channels, crevasse-splays, splay complexes, and sheet floods. The sand sheets thin and fine in the direction of paleoflow and pinch out into floodplain deposits, suggesting the sandstone sheets are unconfined flows distal to the main channel. The FUS likely represents gradual abandonment of individual crevasse channels and splay complexes possibly due to avulsion of the main channel body (Bridge, 1984) or due to waning flood conditions. CUS packages suggest progradation of the crevasse-splay complexes. The paleocurrent orientation within the sandstones are variable and often occur at high angles to the channels and are subparallel to perpendicular to the paleoslope, suggesting the sandstone sheets were deposited in crevasse channels and as splays. The changes in sedimentary structures and grain size within the sheets in the direction of flow—from medium-grained, small-scale, trough cross-beds deposited in close proximity to the main channel and fine-grained cross laminations and climbing ripple stratification occurring distal from the main channel—indicate decreases in flow velocity and suggests flow expansion and a decrease in flow power as the flood waters exited the channel and expanded as a sheet flood onto the adjacent floodplain (cf., Allen, 1964,1965; Bridge & Diemer, 1983; Miall, 1996).

**2.3 Fine-grained facies associations**

*Siltstone:* The siltstone deposits of the fine-grained facies associations are up to 12 m thick in some locations (e.g., Mt. Achenar and Clarkson Peak) and extend 100’s of

m across the outcrop.  Siltstones occur in both the Buckley and Fremouw formations, but are more prevalent in the Buckley Formation and in the middle member of the Fremouw Formation.  The siltstone deposits in the Fremouw Formation are greenish in color, whereas they are grey to white in the Buckley Formation.

In the Buckley Formation siltstone deposits have a sharp basal contact and occur thin  deposits that interfinger with or pinch out into mudrock deposits, or occur as thick (12 m) laterally continuous deposits interbedded with fine-grained sandstone lenses and sheets (Figures 33 and 34), carbonaceous mudrock, and coal.  Siltstones are typically found beneath and cut into by fine- to coarse-grained sandstone sheets facies associations (see section 2.1 and 2.2).  Siltstones are also interstratified with coal seams and may grade vertically into coal or be underlain gradationally, or occur in sharp contact with coal deposits (Figure 35).  Sedimentary structures within the siltstone include horizontal laminations (silt and clay), small ripple cross-laminations (interbedded sandy intervals), and deformation structures, which include deformed sandy foreset beds, flame structures, and convolute bedding.  The deformed foreset beds occur in the fine-grained sandstone interbeds and extend laterally for 10’s of m (e.g., Mt. Achenar; and Clarkson Peak).  The horizontal laminations extend laterally for 10’s to 100’s of m across the outcrop.  Plant material, *Glossopteris* leaves, wood impressions, paleosols, and fossil forests are commonly found in, or associated with, the siltstone deposits (Figure 36).  The upper contact can be sharp or gradational.


Figure 33. Siltstone interbedded and overlain by coal in the upper Buckley Formation at Mt. Achenar, which is in turn, overlain by a medium- to coarse grained sheet sandstone. Person for scale.

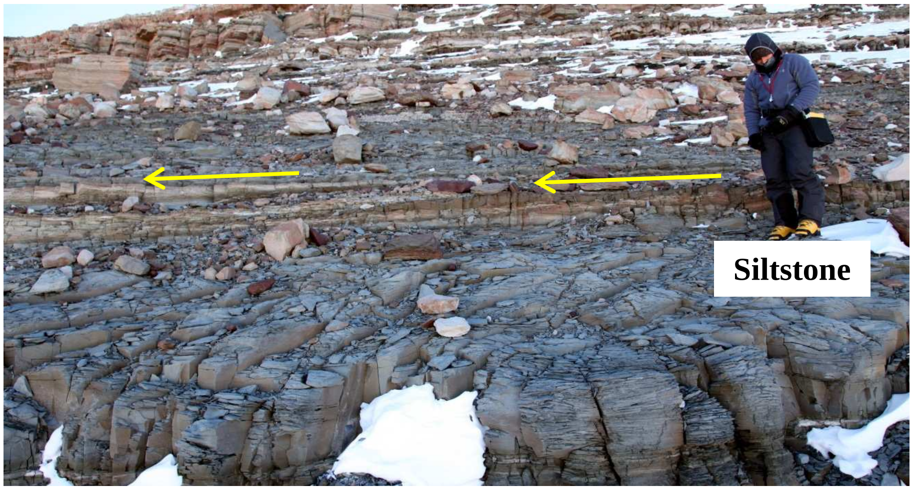



Figure 34. Siltstone with thin downlapping sandstone (yellow arrows) from the lower Buckley Formation, Clarkson Peak (Photograph courtesy of John Isbell). Person for scale.

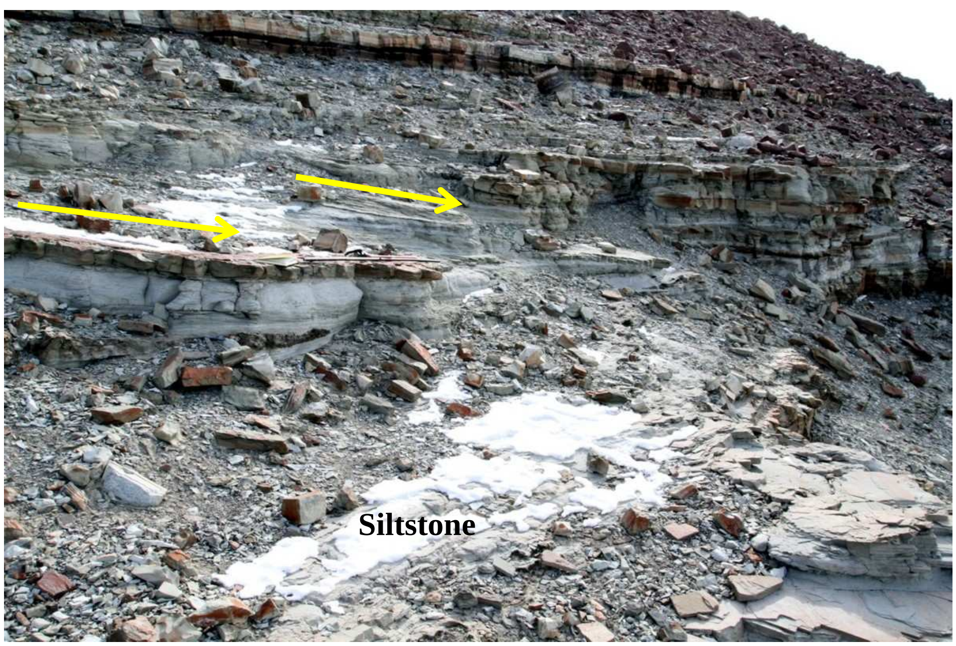



Figure 35. Siltstone overlain by thin downlapping fine-grained sandstone (yellow arrows; ~0.5 m thick) from the upper Buckley Formation, Mt. Achenar.

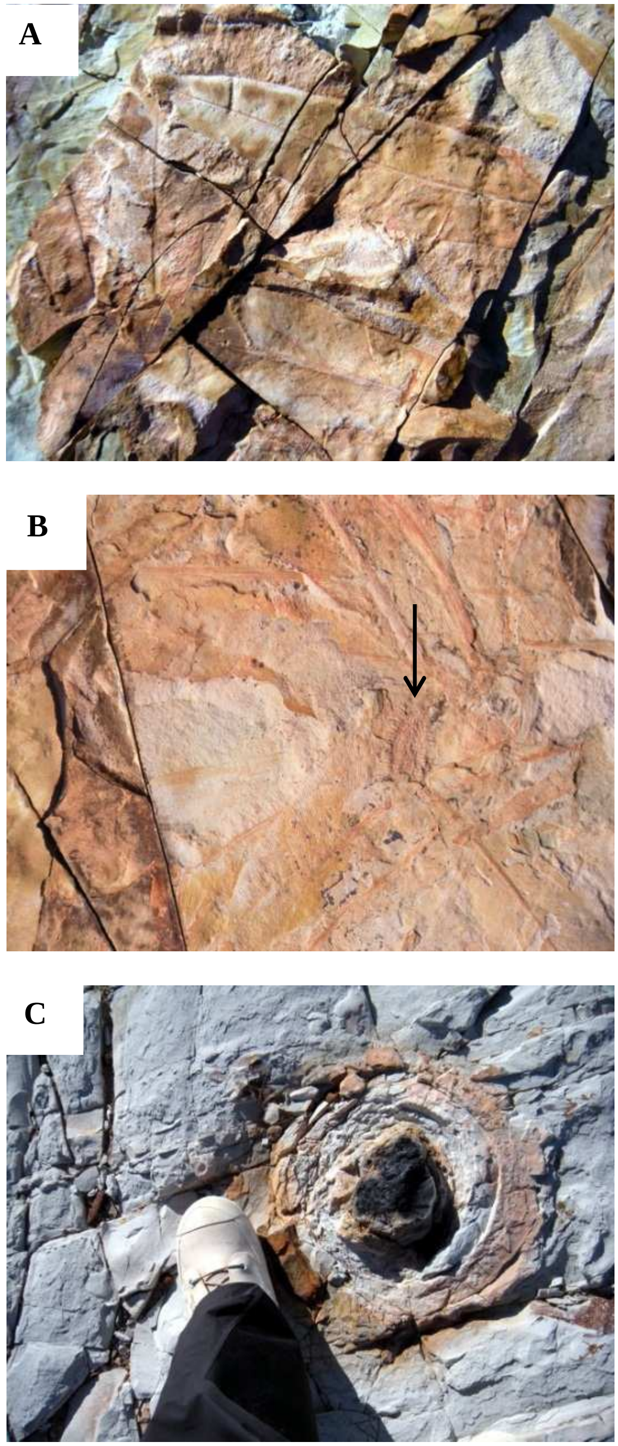



Figure 36. Siltstone facies showing excellent preservation of (A) *Glossopteris* leaves (~15 cm), (B) *Plumsteadia* (below arrow; ~1 cm) and *Glossopteris*, and (C) *in situ* tree stump (~40 cm in diameter). Photograph is from the upper Buckley Formation, Mt. Achenar.

*Interpretation:* The extensive horizontally laminated siltstone represents lacustrine intervals with sandstone interbeds representing crevasse-splay complexes (Barrett et al., 1986; Isbell, 1990) or small delta deposits (cf. Reineck & Singh, 1980; Reading, 1996; Nichols & Fisher, 2006). The horizontal laminations are of alternating clay and silt settling out from suspension (cf. Franco, 1995). Fossil leaves often occur as leaf mats on bedding planes. These have been interpreted as annual leaf falls from deciduous vegetation (Taylor et al., 2000; Gulbranson et al., 2012). The deformation structures within the fine-grained sandstone interbeds (i.e., flames and folds) are likely dewatering structures caused by soft sediment mixing and rapid sedimentation into the lake or due to small scale slumping of sediment (Reineck & Singh, 1980; Franco, 1995). The presence of coal and deformed fine-grained sandstone foreset beds suggest the occurrence of mires and deltas along the margins of the lake, and that water levels within the lake fluctuated due to sedimentation and/or subsidence. At Mt. Achenar, plant material, including fossil *in situ* tree stumps and *Glossopteris* leaves, is found in some horizons that also suggest close proximity of these sites to lake margins.

*Mudrock:* Mudrock deposits are thin bedded (up to 1.5 m thick) and interbedded with siltstone, coal, and very fine-grained sandstone sheets. The mudrocks can be traced laterally across the outcrop for 100's of m. Mudrocks are often overlain by siltstones and fine-grained sandstone sheets (Figure 37). These units may contain gradational or sharp basal boundaries and erosional, sharp, gradational, or bioturbated (rooted surfaces) upper contacts. Mudrock in the Buckley Formation are carbonaceous (black and grey). The

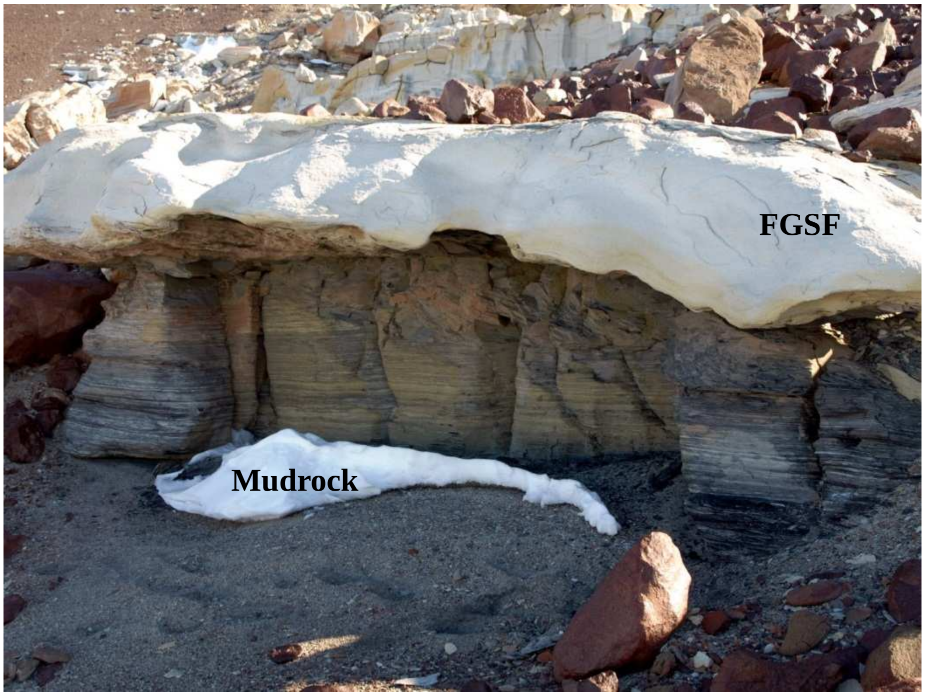



Figure 37. Mudrock (~1 m thick) overlain by the fine- to medium-grained sand sheet facies association (FGSF) from the lower Buckley Formation, Mt. Bowers

carbonaceous-rich mudrock also contains rooted horizons and fossil forests in the upper Buckley Formation.

Mudrocks in the lower member (Figure 38) of the Fremouw Formation are non-carbonaceous (green, brown, and occasionally red). Darker mudrocks in the middle and upper Fremouw members are the result of an increase in carbonaceous material (Barrett et al., 1986). The mudrock is a few meters thick and are often interbedded with thin sand sheets and sand lenses (Barrett et al., 1986). Mudrocks typically have a sharp basal contact and tend to grade upward from siltstone facies (Flaig, 2005). Upper contacts are typically erosional (Flaig, 2005). The middle and upper members contain occasional rootlets (Barrett et al., 1986; Flaig, 2005).

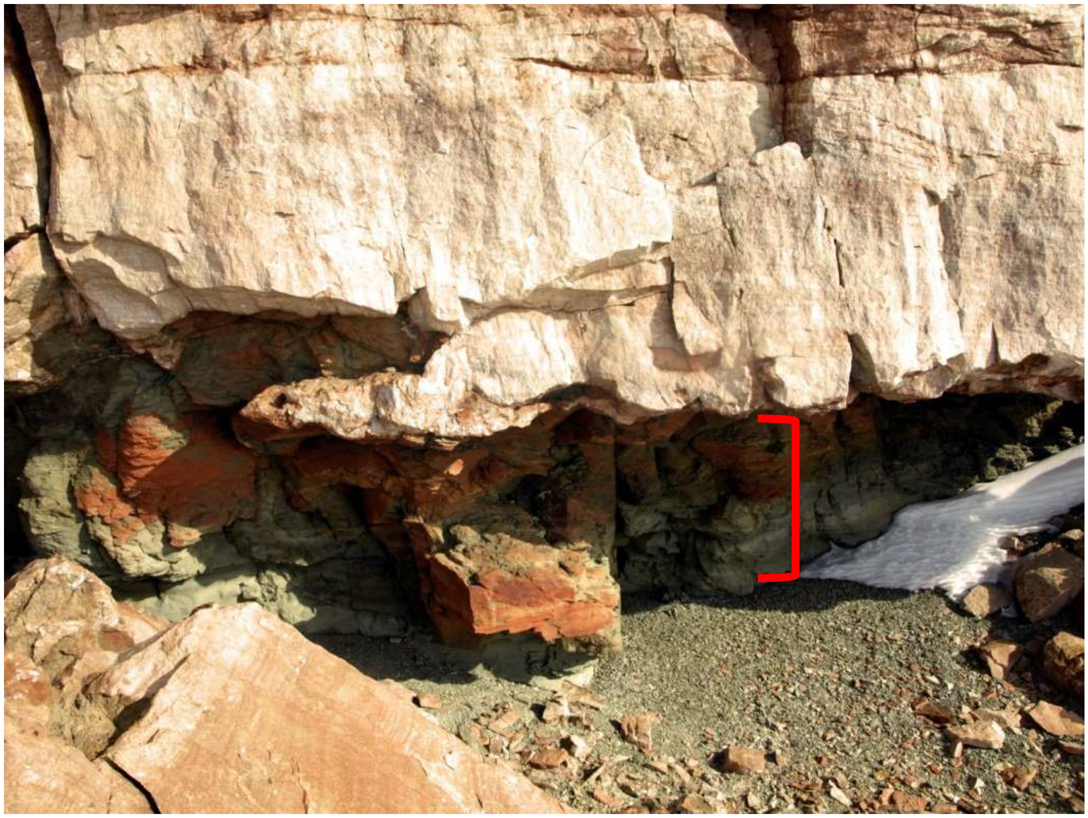


Figure 38. Greenish brown mudrock facies association (red bracket; ~0.5m thick) truncated by the coarse-grained sand sheet facies of the lower Fremouw Formation, Wahl Glacier.

*Interpretation:* The mudrock deposits of the fine-grained facies associations represent low-energy depositional environments associated with fluvial floodplain and lacustrine deposits (Barrett et al., 1986; Isbell, 1990; Collinson et al., 1994; Flaig, 2005). The carbonaceous mudrocks of the Buckley Formation are typically black and grey in color. The black and dark grey color is likely due to organic carbon and the lighter colors are likely due to a more silty composition (cf., Stow, 2005). The color of mudrock is also

indicative of the drainage. For example, the Buckley Formation is grey and black consisting of preserved organic material; grey mudrocks contain less carbon compared to black mudrocks, and may represent better drained areas on the floodplain. This indicates the floodplains of the Buckley Formation were poorly drained and formed in water-logged conditions where the decay of organic matter was inhibited (cf., Reading, 1996). This is also supported by the occurrence of coal, which indicates the presence of mires (i.e., poorly drained floodplains). The presence of rooted horizons and fossil forests in the Buckley Formation intercalated with siltstones and coals suggest deposition in close proximity to lacustrine, marsh, and mire environments.

The Fremouw Formation is green, red, and brown in color, which suggests that the floodplains were well drained and the sediment was oxidized (cf., Reading, 1996; Collinson, 1997). Mudrocks in the Fremouw Formation have been interpreted as over-bank deposits within a swamp setting with small amounts of standing water compared to the extensive lake deposits in the Buckley Formation (Barrett et al., 1986; Flaig, 2005). Retallack & Krull (1999) have interpreted the mudrock with rooted horizons as paleosols.

*Coal:* Coal varies in from <0.5 m to several meters thick and can be traced 100's of m to km across the outcrop. Coal only occurs in the upper Buckley Formation and in the upper member of the Fremouw Formation. The coal may contain silicified peat and is often interbedded with siltstone and mudrock (Figure 39). Barrett et al. (1986) reported coal from the Buckley Formation to have low ash and sulfur content and ranked it as mostly low-volatile bituminous coal. A study by Coates et al. (1990) reports Permian coals in the CTM with ash contents up to 50% with a mean of 15.3%. Coates et al.

(1990) also ranked the Permian coals ranging from high-volatile bituminous coal to anthracite, with most of the coal samples having apparent ranks of low-volatile bituminous coal and semianthracite. High grades of coal occur in close proximity to dikes and sills of Jurassic age, which occur throughout the Transantarctic Mountains (Barrett et al., 1986). Coal typically overlies siltstone and is overlain either by siltstone or fine- to medium-grained sandstone sheets (Figure 33). Contacts can be sharp or gradational with carbonaceous mudrocks and siltstones, and may be overlain erosionally or abruptly by the fine- to medium-grained facies associations.

*Interpretation:* The coal deposits represent mire and marsh environments. Thin coal deposits interbedded with siltstone and mudrock suggest deposition in close proximity to lakes and crevasse-splay channels.

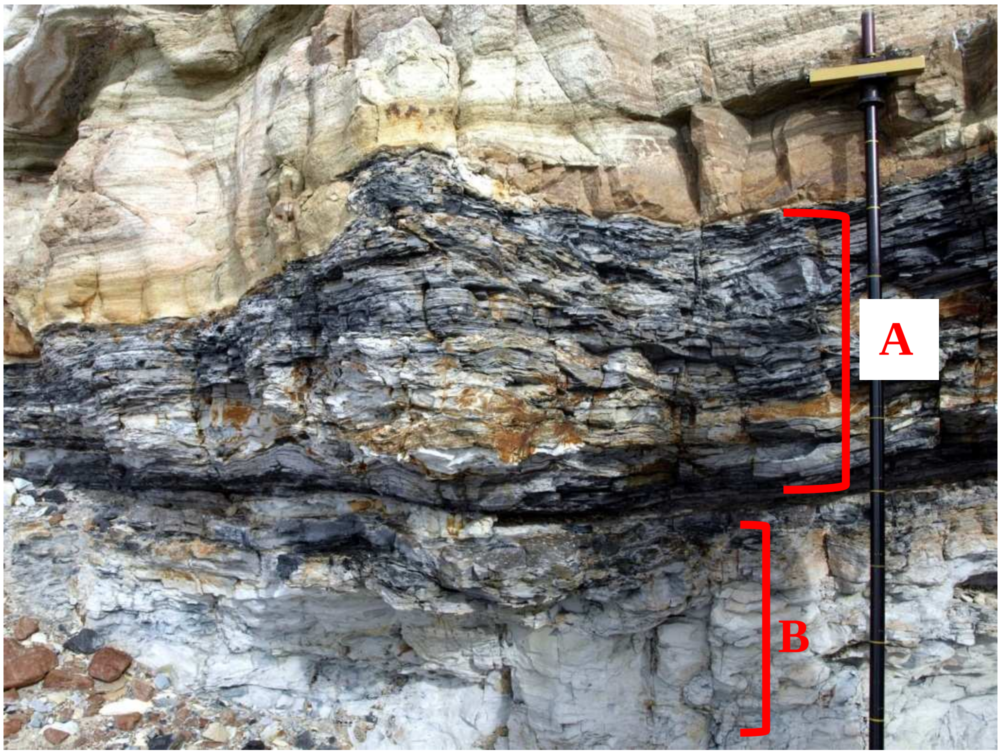



Figure 39. Photograph showing siltstone (B) overlain by a poorly developed coal (A) interbedded with siltstone and overlain by the FGSF in the upper Buckley Formation at Mt. Achenar. Jacob staff for scale.

## 2.3 Discussion of Depositional Settings

The facies associations of the Buckley Formation consist of coarse-grained sandstone facies associations, fine- to medium-grained sandstone sheets, and fine-grained facies associations (siltstone, coal, and mudrocks). The deposition of the coarse-grained facies associations occurred in braided stream settings based on the occurrence of: 1) laterally extensive sand sheets with erosive basal contacts suggesting deposition in a mobile channel belt; 2) low paleocurrent dispersion suggesting deposition by bars oriented parallel to the trend of channels within a relatively straight channel system; 3) fining-upward macroforms with abundant downstream accretion suggesting deposition as longitudinal bars; and 4) sandstone-filled abandoned channels suggesting gradual channel abandonment like those typical of channel abandonment in braided streams.

The fine- to medium-grained sandstone sheet associations were deposited as crevasse channels based on the occurrence of: 1) sandstone sheets that thin and fine in the direction of paleoflow and pinch out into floodplain deposits suggesting that the sand sheets were deposited by unconfined flows distal to the main channel; 2) paleocurrent orientations derived from primary sedimentary structures that are variable and at high angles to the channels suggesting that waters flowed out of the channels and spread onto the adjacent alluvial surface as unconfined flows adjacent to the fluvial channel belt; and 3) the sedimentary structures in the FUS and CUS packets that decrease in the direction of paleoflow suggesting decreases in flow velocity and flow expansion and a decrease in flow power as the flood waters exited the channel and expanded as a sheet flood onto the adjacent floodplain. The fine- to medium-grained sandstone sheet associations also

represent crevasse splay complexes based on the vertical stacking of multiple crevasse splays.

The fine-grained facies associations consist of siltstone, mudrock, and coal. The siltstones were deposited in lakes based on the occurrence of horizontal laminations of alternating clay and silt suggesting deposition in standing water with the particles settling from suspension. The carbonaceous mudrocks were deposited in low lying flood basins on poorly drained floodplains. Grey mudrocks suggest deposition on better drained portions of the floodplain compared to that of the organic-rich mudrocks. The relatively thick coals were deposited in mires and peat swamps (cf., McCabe, 1984) in distal locations on the flood plains away from channel belts where there was a the lack of coarse-grained sediment influx and where there was a high abundance of organic material. Interbedded fine-grained sandstone sheets suggest that crevasse splay complexes prograded into these low lying areas.

The lower Buckley Formation consists of abundant medium- to coarse-grained sandstone sheet facies association, abundant fine- to medium-grained sandstone sheet facies association, and lacks an abundance of the fine-grained facies associations. At Lewis Cliffs the only mudrock (~4 m thick) was located above and interbedded with the fine- to medium-grained sandstone sheet facies associations. Only 0.5 m thick coal was found above the mudrock and was overlain sharply by the fine- to medium-grained sandstone facies association. At Mt. Bowers the first appearance of coal and mudrock occurred at 127 m above the base of the Fairchild□Buckley contact. The coal was bound by sharp upper and lower contacts with the fine- to medium-grained sandstone sheet facies association. Isbell (1990) also found a high abundance (~95%) of channel-form

sheet sandstones and only small amounts of coal (0-3%) and mudstone (2-15%) in the lower Buckley Formation.

In contrast, the upper Buckley Formation consists of medium- to coarse-grained sandstone sheet facies association that decreases up section, common fine- to medium-grained sandstone sheet facies association, and an abundance of the fine-grained facies association compared to that of the lower Buckley Formation. At Mt. Achenar two sections were measured in the upper Buckley Formation and are separated by a Jurassic dolerite sill. Below the dolerite sill towards the base of the upper Buckley Formation the medium- to coarse-grained sandstone sheet facies association is multistoried and erosionally overlies siltstone and interbedded coal. Above the dolerite sill the medium- to coarse-grained sandstone sheet facies is absent and the section consists primarily of laminated siltstones, carbonaceous mudrock, and thin sandstone sheets interbedded with siltstone.

Overall, the Buckley Formation was formed by a complex of braided river systems containing channel sandstones consisting of the medium- to coarse-grained sandstone sheet facies, crevasse channels and crevasse splay complexes consisting of the fine- to medium-grained sandstone sheet facies, and floodplain deposits consisting of siltstone, mudrock and coal facies. The channel form sandstone sheets are abundant and commonly multistoried and multilateral suggesting they were laterally mobile on their alluvial surface (cf., Friend, 1983). The crevasse splays are abundant and often organized into stacked deposits similar to splay complexes. There is also a change in grain size and paleocurrent direction between splays (Isbell, 1990). Crevasse channels that fine upward suggest gradual abandonment and crevasse -splays that coarsen upward suggest

progradation of splays into the floodplain and development of crevasses splay complexes (cf., Bridge, 1984). These channel and crevasse splay deposits alternate vertically through the section with floodplain deposits. For example, the channel deposits often occur overlying coal deposits. The coals are formed in the lowest part of the floodplain, so this suggests that channels were diverted into the lowest parts of the floodplain. This alternating of overbank floodplain deposits and channel deposits within the vertical section is similar to characteristics that Bridge (1984) identifies as avulsing systems.

Avulsion is defined by Bridge & Leeder (1979) as a river's relatively sudden abandonment of a channel belt in favor of a new course. According to Bridge (1984) evidence of an avulsing system includes: 1) evidence of major overbank floods (i.e. laterally extensive units with erosive bases, coarser and thicker than underlying beds; 2) abrupt changes in thickness and grain size between splay packages (complexes) that represent whether a channel has avulsed closer or farther away from a given area on the floodplain; 3) changes in paleocurrent directions between individual splay complexes suggesting major reorganization of the floodplain before and after the avulsion event and the direction to the new channel systems; and 4) abrupt changes in lithofacies (i.e. channel sandstones overlying coal).

Evidence for avulsion in the Buckley Formation includes (Figure 10 and 40): 1) evidence of major overbank floods and abrupt changes in lithofacies where fine- to medium-grained sandstones erosionally overlie siltstone, mudrock, and coal suggesting that the channel has moved closer and is being diverted into low lying areas of the floodplain; 2) abrupt changes in grain size with siltstone abruptly overlying medium-grained sand bodies suggesting that the channel has moved further away from the flood

basin leaving the area isolated from clastic influx; 3) variations in paleocurrent directions between individual splay complexes suggesting reorganization before and after the avulsion; and 4) sandstone sheets interfingering with overbank deposits suggesting the channels were in close proximity to the floodplain. A modern analogue comparable to the Buckley Formation is the South Saskatchewan River and the Cumberland Marshes.

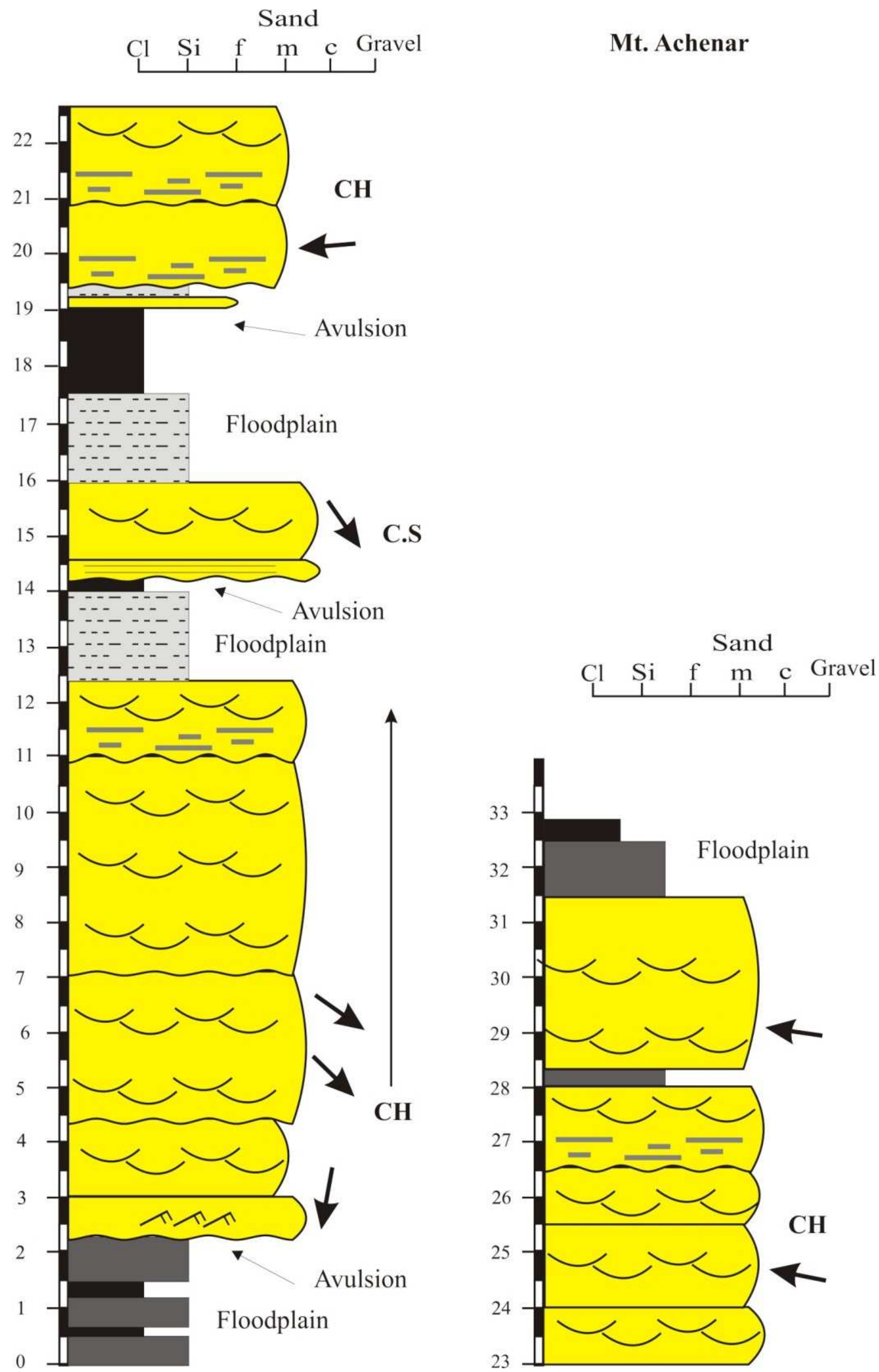



Figure 40. Stratigraphic column from the upper Buckley Formation at Mt. Achenar showing floodplain deposits overlain by and overlying a multistoried channel deposit (CH) with avulsion events marked. Above the CH deposit is a crevasse splay (C.S) encased within floodplain deposits. The thick black arrows represent paleocurrent orientations.

The South Saskatchewan River and Cumberland Marshes in Saskatchewan, Canada covers an 8000 $km^2$ area consisting of wetlands intersected by small tributaries, active, and abandoned channels of the Saskatchewan River (Figure 41; Smith et al., 1989, 1998; Pereze-Arlucea & Smith, 1999; Morozova & Smith, 2003; Smith & Perez-Arlucea, 2004). Regional aggradation has resulted in periodic shifting of major river channels by avulsion (Cazanacli & Smith, 1998; Asselen et al., 2010). At least nine avulsions have occurred in this region in the past 5400 years (Morozova & Smith, 1999, 2000) with the most recent avulsion occurring in the 1870s due to an ice jam (Smith et al., 1989, 1998; Morozova & Smith, 1999). During the 1870 avulsion, the Old Channel diverted northward resulting in the Saskatchewan River diverting most of its flow into parts of the Cumberland Marshes. This diversion has altered over 500 $km^2$ of wetlands in favor of a new course (present day New Channel) or developing course for the avulsed Saskatchewan River (Smith et al., 1989, 1998; Morozova & Smith, 1999). The “breakout area” is the main area of avulsive flooding and consists of active and abandoned anastomosed channels (Perez-Arlucea & Smith, 1999; Smith & Perez-Arlucea, 2008). The avulsion belt contains straight to sinuous isolated active and abandoned channels, anastomosed channel systems, levees, crevasse splays, marshes, lakes, and bogs (Smith et al., 1989; Smith & Perez-Arlucea, 2004). Presently, the Old Channel is slowly being abandoned and only carries 5-10% of the annual flow with the majority of the discharge being carried by a single-thread, relatively straight channel called the New Channel in the upper reach and the Centre Angling Channel in the medial reach (Smith & Perez-Arlucea, 2008). The new channel is 200 to 350 m wide and ~5 to 7.5 m deep (mean bankfull depth) (Perez-Arlucea & Smith, 1999).

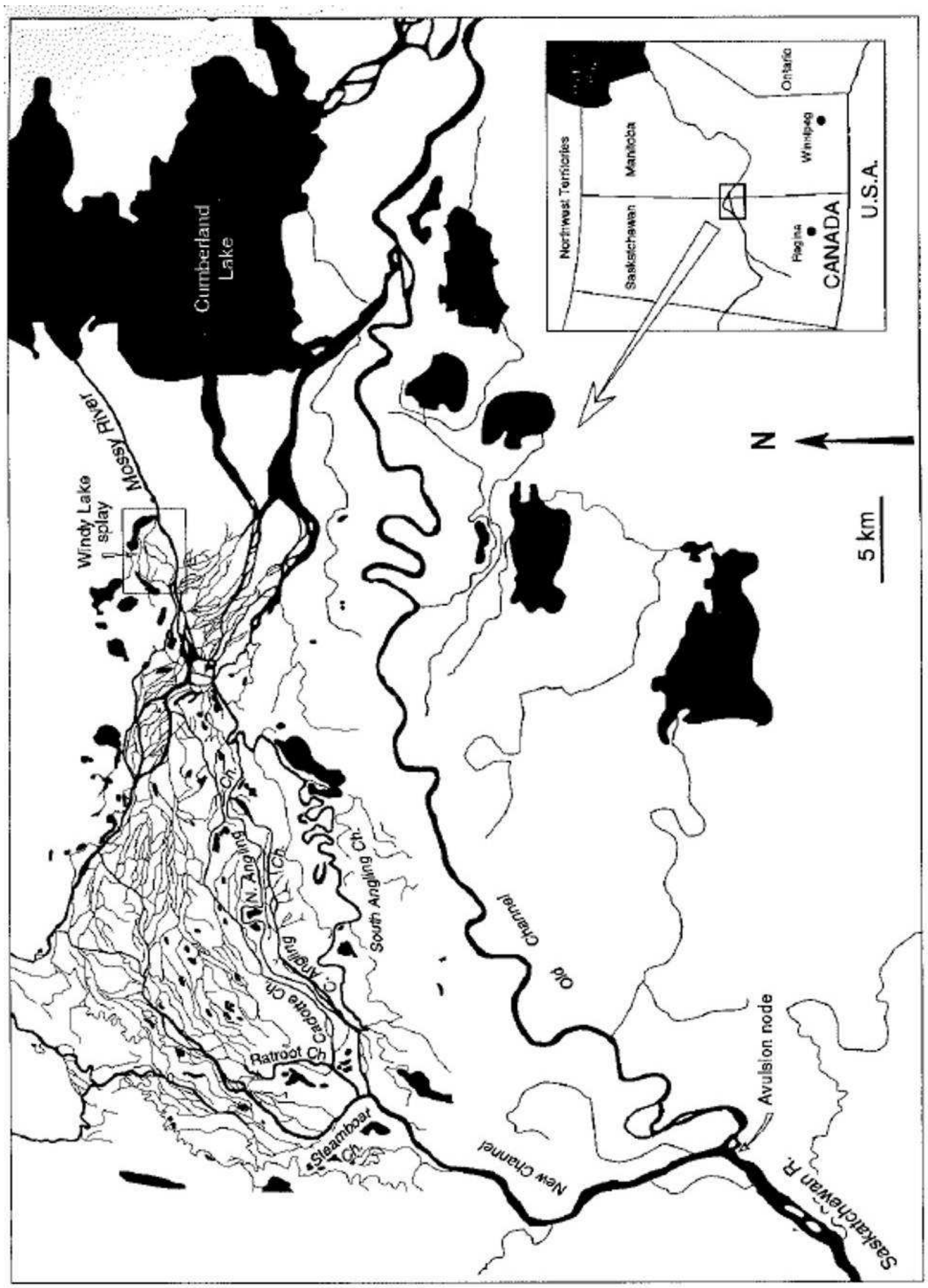



Figure 41. Map showing the location of the Saskatchewan River avulsion belt with main channels labeled and lakes represented by black areas (Smith & Perez-Arlucea, 1994).

Facies of this region have been described by Smith (1982, 1989), Smith & Perez-Arlucea, 1994, Morozova & Smith (1999), and Perez-Arlucea & Smith (1999). The main facies consist of channels, crevasse splays, lakes, marshes, wetlands, and peat. The channel facies consist of well to moderately sorted fine to medium grained sand with rippled and planar tabular cross-bed sets (Smith, 1982; Perez-Arlucea & Smith, 1999). Channels are typically underlain by wetland environments represented by very fine silts, clays, and organic rich sediments (Morozova & Smith, 1999). Channel fills form fining upward sequences due to gradual channel abandonment (Perez-Arlucea & Smith, 1999). The crevasse splays consist of moderately to very poorly sorted medium- to fine-grained sand to sandy silt with sedimentary structures such as, ripples, flaser and lenticular bedding, and horizontal laminations (Smith, 1982; Morozova & Smith, 1999; Perez-Arlucea & Smith, 1999). The crevasse splays generally coarsen upward then fine upward and thin distally (Smith, 1982; Perez-Arlucea & Smith, 1999). Smith et al. (1989) characterized the crevasse splay deposits into three stages (Figure 42). The stage one splay are the smallest and form simple lobate bodies deposited by sheet and channelized flow (Smith et al., 1989). Stage two splays are larger and more complex than stage one splays. The stage two splays have a high channel density and are elliptical to elongate bodies composed of complex networks of active and abandoned channels fed by multiple crevasse channels, which have been termed crevasse splay complexes (Smith et al., 1989). The stage three splays are also elliptical to slightly elongate bodies with lower channel densities than stage two splays (Smith et al., 1989). Stage three splays are the most stable of the different splay types and typically interfinger with silt, clay, and organic sediment of the wetland environments or form where splays prograded into

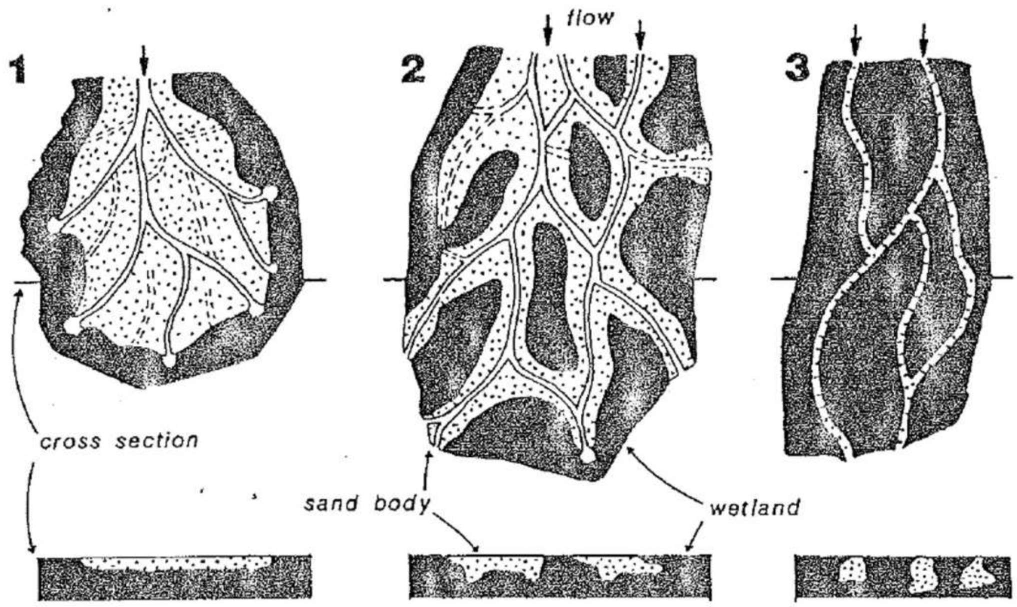



Figure 42. Generalized relationships between splay types and sand-body geometry with wetland sediments shown in black (Smith et al., 1989).

shallow lakes (Smith et al., 1989). Lakes are extensive and shallow with round to irregular shapes (Smith, 1982). The lake deposits are either massive due to bioturbation or laminated with alternating clay and silty clay (Smith, 1982; Perez-Arlucea & Smith, 1999). Marshes (referred to as backswamps by Smith & Smith, 1980) contain bioturbated and thin laminated organic and clastic mud deposits (Smith, 1982). The marshes form in highly vegetated, low lying areas typically near lakes and contain water year round (Smith, 1982; Perez-Arlucea & Smith, 1999). Wetlands described by Smith & Perez-Arlucea (1994) are characterized as low, open, densely vegetated areas below or slightly above mean water elevation including interchannel ponds, marshy pond borders, and higher fen meadows. The sediments are organic-rich clay to medium silt (Smith &

Perez-Arlucea, 1994). Peat either consists of fine-grained, dark brown to dark gray fine silt and clay or is reddish brown and coarser with many roots and fragments of marsh, bog, and fen vegetation and forms in areas isolated from fluvial deposition (Smith et al., 1989; Smith & Perez-Arlucea, 1994; Perez-Arlucea & Smith, 1999).

Avulsions of the Saskatchewan River are mainly nodal avulsions (Leeder, 1978; Mackey & Bridge, 1995) and consist of two basic avulsion styles (Morozova & Smith, 1999). The first type of avulsion style is by channel reoccupation, which has been noted widely by Smith et al. (1998), Aslan & Blum (1999), Morozova & Smith (1999, 2000), Mohrig et al. (2000), Stouthamer (2001), Makaske et al. (2002), Aslan et al. (2003, 2005), and Jerolmack & Paola (2007) and is represented by multistoried stacking pattern in the stratigraphic record but doesn't result in significant floodplain aggradation. The New Channel developed by reoccupation of a former Saskatchewan River course and by interception and enlargement of the Torch River (Smith et al., 1998; Perez-Arlucea & Smith, 1999). The second type of avulsion style, and dominant type, is progradational with extensive deposition of crevasse splay complexes into the adjacent floodbasin resulting in significant floodplain aggradation (Morozova & Smith, 1999). If an intercepted channel is too small to accommodate the entire flow the channel will either enlarge, overflow, or form crevasses by breaching its levee (Perez-Arlucea & Smith, 1999). Smith (1989) summarized the sequence of avulsive events and channel evolution into four stages. The first stage is the initial avulsion resulting in deposition of shallow, wide unstable distributary channels of stage 1 splays. Stage two is referred to as the anastomosed stage, which results in the enlargement of stage 1 splays into stage 2 or stage 3 splays with relatively stable anastomosed channels (Smith, 1989). During stage 2

channels will become abandoned as new channels form and this stage will continue as long as the avulsive flow continues (Smith, 1989). Stage 3 is the reversion stage where the rate of channel abandonment exceeds the rate of channel formation resulting in fewer but larger channels (Smith, 1989). The final stage is when splays no longer form and a single meander channel belt forms (Smith, 1989).

Vertical sections from borings show a general coarsening upward succession with peat occurring beneath 80% of avulsion belt deposits overlain by a thin lacustrine deposit representing the initial flooding due to the avulsion, which is overlain by fine- to medium-grained sand sheet deposits interpreted as crevasse splays (Smith et al., 1989; Smith & Perez-Arlucea, 1994, 2004; Perez-Arlucea & Smith, 1999). The medium-grained crevasse channels are typical of stage 1 splays and occur close to the trunk channel. Finer grained crevasse channels of stage 2 and 3 splays occur further from the trunk channel and often prograded into lacustrine facies (Perez-Arlucea & Smith, 1999). The crevasse splays either coarsen upward suggesting progradation into the floodbasin or fine upward suggesting channel abandonment with peat at the top as the abandoned channels become vegetated and isolated from the trunk channel (Perez-Arlucea & Smith, 1999). It is suggested by Davies-Vollum & Smith (2008) if the fine-grained deposits of the Saskatchewan River avulsion belt are preserved in the geologic record they would be seen as mudrock between thin beds of coal or carbonaceous shale.

The Buckley Formation is similar to the Saskatchewan River and Cumberland Marshes regions in many ways. Both areas contain avulsive deposits likely resulting from both channel reoccupation and progradation due to splay complexes. Both areas consist of similar facies such as, channel fills, crevasse splays, lacustrine, and marsh

facies. The channel fills for both consist of sand and generally form fining upward successions with planar and trough cross-stratification suggesting channel abandonment. The channel fill, in the case of the Buckley Formation, is a little coarser (mainly medium-grained sand) than the Saskatchewan channel fills (mainly very-fine to fine-grained sand). The crevasse splay deposits in both cases occur as sand sheets, which show sedimentary structures such as, ripple cross-lamination, lenticular, flaser and wavy bedding, and horizontal laminations. The crevasse splays show both fining upward and coarsening upward successions suggesting abandonment of crevasse channels and progradation of crevasse splays into the flood basin respectively. The breakout region of the Saskatchewan River consists primarily of stage 2 and 3 splays (Smith et al., 1989; Perez-Arlucea & Smith, 1999), which are finer grained than stage 1 splays and are commonly narrower and more stable than stage 1 splays. It is likely that the crevasse splay deposits of the Buckley Formation represent stage 1 and stage 2 splays as the crevasse splays are coarser grained and are typically similar in grain size to the channel fill facies. Lacustrine deposits occur in both areas as laminated clay and silt deposits settling from suspension as well as massive deposits likely due to heavy bioturbation. The Buckley Formation contains abundant laterally extensive carbonaceous mudrock and coal deposits interpreted as floodplain and marsh deposits. The avulsion belt of the Saskatchewan River contains abundant marshes and peat, which if they were to be preserved in the stratigraphic record would be interbedded mudrocks and coals (Davies-Vollum & Smith, 2008). Facies distributions are also similar in the Buckley Formation and the Saskatchewan avulsion belt. In the Buckley Formation channels are both multistoried, likely due to channel reoccupation, or overlie coal and carbonaceous

mudrock deposits suggesting channels have been diverted into the low lying floodplain. Likewise, the majority of channels in the break out area (abandoned and active) overlie peat or thin lacustrine deposits also suggesting the flow was diverted into the low lying floodplains. The crevasse splays of both the Buckley Formation and Saskatchewan River (active and abandoned) thin away from the trunck channel and either interfinger with floodplain deposits or form lacustrine deltas as they enter floodbasin lakes (i.e. Mt. Achenar). In contrast, the channel deposits of the Buckley Formation show characteristics for being braided stream deposits, whereas channel deposits within the breakout area north of the Saskatchewan River are commonly anastomosed channels. Overall, both the Buckley Formation and the Saskatchewan River region represent floodplain aggradation due to channel avulsion.

The Fremouw Formation consists primarily of the medium- to coarse-grained sandstone sheet facies associations and mudrocks of the fine-grained facies associations. The medium-to coarse-grained sandstone sheet facies associations has been interpreted as braided stream deposits based on: 1) The presence of multi-storied sheet sandstones with planar erosional bases suggests that these streams may have migrated laterally on their alluvial surface within the channel belt; 2) Low paleocurrent dispersion suggesting deposition by bars oriented parallel to the trend of channels within a relatively straight channel system; 3) The presence of vertical accretion, downstream accretion suggesting deposition along the front of the migrating macroforms or bars, and upstream accretion suggesting the bars are overtaking bars located downstream, and the bars are climbing up the back of the downstream bars within the multi-storied packages; 4) Reactivation surfaces are common representing a halt in sedimentation following individual floods

especially where fine-grained facies associations drape the accretionary surfaces of these macroforms and; 5) Sand-filled abandoned channels suggesting gradual channel abandonment that exhibits lateral accretion indicating the channels were filled by bank attached bars, vertical accretion suggesting gradual abandonment of channels , and downstream accretion representing the downstream migration of bars

The mudrock, of the fine-grained facies association, is most abundant in the middle Fremouw and increases in carbonaceous material into the upper Fremouw. The mudrock has been interpreted as well-drained floodplain deposits in low lying areas based on the color (brownish, green, red) and grain size. Interbedded fine-grained sandstone sheets suggest that crevasse-splay complexes prograded into these low lying areas.

The lower and upper Fremouw Formation (for this study) consists primarily of the coarse-grained sandstone sheet facies association that has been interpreted as channels deposited by a braided stream. The channels have erosive to planar basal contacts and are multistoried containing abundant large and small-scale cross-stratification and reactivation surfaces containing fine-grained drapes suggesting a fluctuating discharge regime. The multistoried channels either show little variation in grain size or fine upwards within the last FUS into fine-grained sand, siltstone, or mudrock. The channel sandstones are typically found truncating underlying floodplain deposits of the fine-grained facies association and laterally pinch out into floodplain deposits. At other locations in CTM the lower Fremouw Formation is composed of fine-grained floodplain deposits representing levee and crevasse splay deposits making up about half of the lower Fremouw Formation (Collinson et al., 1994). The siltstone and mudrock of the fine-

grained facies association is non-carbonaceous in the lower member but increases in carbonaceous material in the upper member (Barrett et al., 1986).

In contrast, the middle Fremouw Formation is primarily composed of fine-grained floodplain rather than channel deposits (Barrett et al., 1986; Collinson et al., 1994). The channel sand bodies are much smaller and less sheet like than the lower and upper Fremouw Formation (Collinson et al., 1994).

The upper Fremouw Formation at Gordon Valley consists primarily of the coarse-grained sandstone sheet facies association that has been interpreted as channels deposited by a braided stream. The channels are similar to the lower Buckley Formation at Wahl Glacier, Lewis Cliffs, and Mt. Bowers and have erosive to planar basal contacts and are multistoried containing abundant large and small-scale cross-stratification. Unlike Wahl Glacier, the channel sandstone deposits at Gordon Valley lack reactivation surfaces draped with silt or mudrock. The multistoried channels either show little variation in grain size or fine upwards within the last FUS into fine-grained sand, siltstone, or mudrock. The channel sandstones are typically found truncating underlying floodplain deposits of the fine-grained facies association. In contrast to the lower Fremouw Formation, the mudrock and siltstones are brown in color compared to green mudrock and siltstone at Wahl Glacier.

These facies associations of the Buckley and Fremouw Formations are part of avulsive fluvial systems that were deposited in the Transantarctic basin during the Permian and Triassic. Variations of these facies and how they are stacked or arranged together regionally across the basin into basin-scale (large-scale) facies motifs, are here

referred to as alluvial channel stacking patterns. These stacking patterns are the subject of the next chapter. Because fluvial stacking patterns are the result of the relationship between the rate of the creation of accommodation space and avulsion rates (Allen, 1978; Bridge & Leeder, 1979), which are a proxy for sedimentation rates/sediment influx rates, these avulsive systems (fluvial strata in the Buckley and Fremouw formations) reveal important information on driving mechanisms in the development of fluvial foreland basin stratigraphy.

### 3. Fluvial Stacking Patterns

Fluvial stacking patterns or fluvial architecture is defined as the density and interconnectedness of channel sandstone bodies relative to floodplain deposits in time and space (Leeder, 1978; Bridge & Leeder, 1979; Miall, 1983; Einsele, 1992; Bridge & Mackey, 1993). Controls on fluvial stacking patterns are linked to changes in the rate of accommodation (e.g., subsidence, eustatic changes in sea level) (Flemings & Jordan, 1990; Frostick & Steel, 1993; Shanley & McCabe, 1994; DeCelles & Giles, 1996; Miall, 1996, 1997; Blum & Törnqvist, 2000; Clevis et al., 2004) relative to changes in the volume of channel-bound sediment deposited through time, which is also referred to as sediment influx rates (e.g., tectonism, climate changes, loss of vegetation; Schumm, 1977; Bridge & Leeder, 1979; McLoughlin et al., 1997; Cecil, 2003). In reality, it is a comparison between the creation of accommodation space relative to channel avulsion rates where avulsion rates are a function of the rate of sediment influx (Allen, 1978; Bridge & Leeder, 1979). In terrestrial foreland basins, like the Middle Permian to Triassic Transantarctic basin, changing accommodation rates are the result of changing subsidence rates, which vary across the basin due to different degrees of flexural loading (differential subsidence; cf. Flemings & Jordan, 1990; Jordan, 1995). Fluvial stacking patterns are described as being low, moderate, or high depending on the density of channel bodies in a particular area within a specific stratigraphic interval. Low stacking patterns consist of abundant floodplain deposits containing single ribbon or sheet sandstone bodies, which occur as channel bodies isolated within thicker floodplain successions. This pattern is the result of accommodation being greater than sediment influx. Moderate stacking patterns consist of equal volumes of channel sandstone bodies

and floodplain deposits, which result from the creation of accommodation space being equal to sediment influx for a given interval of time. High stacking patterns consist of thick multi-lateral multistoried channel sandstone bodies with little to no preserved floodplain deposits due to sediment flux being greater than the rate of the creation in accommodation (sediment flux > creation of accommodation space). In this scenario, fine-grained sediments are eroded and flushed through the system. Sediment starvation or a lack of channel sandstones (no stacking pattern) results from very high rates of the creation of accommodation under scenarios of extremely low clastic sediment influx. In such cases either lakes (extremely low influx) or mires (no clastic sediment influx) dominate the landscape and are represented in the rock record by predominantly lacustrine and/or coal facies. This chapter focuses on and discusses changes in fluvial stacking patterns through time and space across the Transantarctic foreland basin in the Beardmore Glacier region (BGR) during deposition of the Permian Buckley and Triassic Fremouw formations.

The Transantarctic basin evolved into a foreland basin during the mid-Permian suggested by paleocurrent changes showing a change from a uniformly dipping paleoslope (dipping towards the Weddell Sea) during the deposition of the Fairchild Formation to an asymmetric trough shaped basin orientated oblique to the present day trend of the CTM (Isbell, 1990; Collinson et al., 1994; Isbell et al., 1997). The basin margins of the foreland basin are located on the East Antarctica craton (present day polar plateau) and in the direction of the paleo-Pacific margin of Antarctica (present day Ross Sea), which was in the direction of the orogenic belt (Isbell, 1990). The lower Buckley Formation is separated from the upper Buckley Formation by a 140˚ reversal in

paleoslope with regional paleoflow to the ESE (toward the present Weddell Sea) during the lower Buckley, which changed to NW (toward Victoria Land) regional paleoflow during the upper Buckley Formation (Figure 43; Isbell, 1990, 1991; Collinson et al., 1994; Isbell et al., 1997). The change in regional paleoflow is accompanied by a change in sandstone composition from quartzo-feldspathic sandstone in the lower Buckley Formation to volcaniclastic sandstone in the upper Buckley Formation (Figure 44; Barrett et al., 1986; Isbell, 1990, 1991; Collinson et al., 1994). The paleoflow for the Fremouw Formation is transverse to oblique across the present trend of the Transantarctic Mountains with flow oriented toward the craton (NW in the BGR). Sandstone composition in the Fremouw Formation is quartz rich in the BGR, but contains volcaniclastic grains in the Shackleton Glacier region (Figure 44; Barrett et al., 1986; Isbell 1990; Collinson et al., 1987, 1994; Vavra 1984). Paleocurrent orientations show a change from both transverse (flowing perpendicular across the basin margins) and longitudinal flow down the basin axis during the Middle and Late Permian to transverse flow during the Early Triassic suggesting a change from an under-filled to an over-filled foreland basin (cf. Vavra, 1984; Isbell, 1990, 1991; Collinson et al., 1994). Flemings & Jordan (1990) and Jordan (1995) define an under-filled basin as a topographic trough with both transverse and longitudinal drainage patterns and an over-filled basin as a basin that displays no topographic depression within the basin. For an over-filled foreland basin, drainage is transverse away from an orogenic belt (Figure 45).

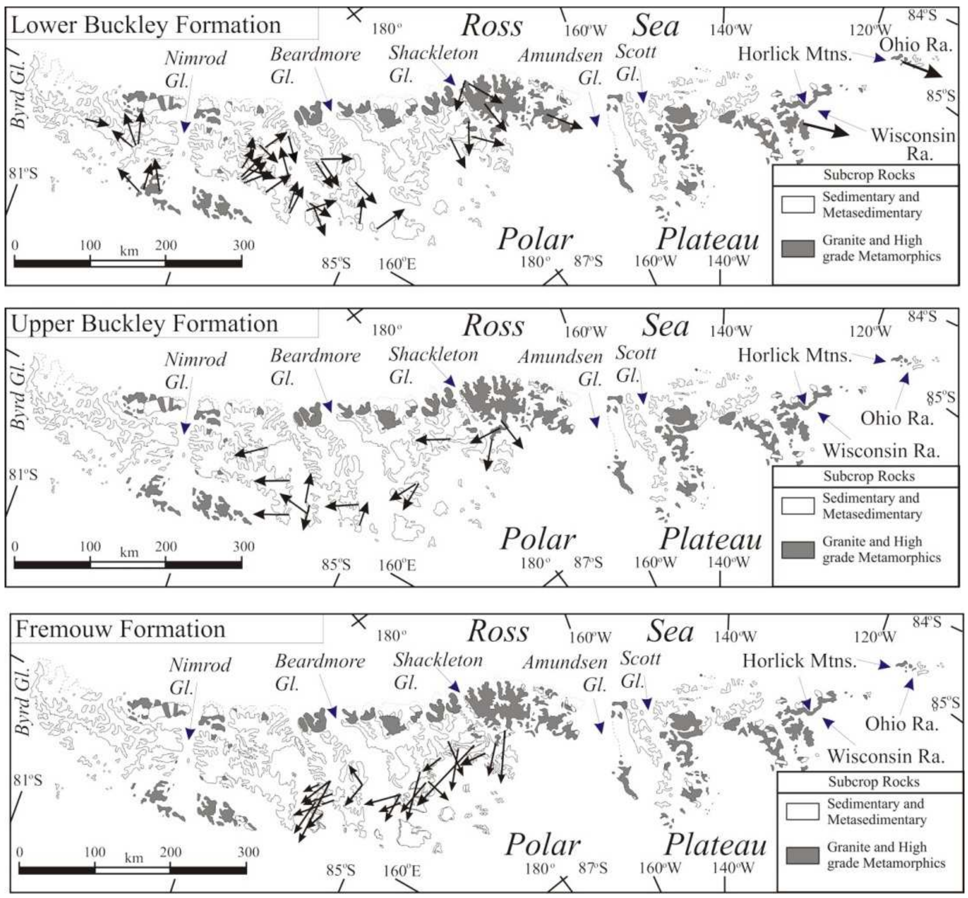



Figure 43. Diagrams showing changes from both longitudinal and transverse paleoflow in the lower and upper members of the Buckley Formation and a dominant longitudinal paleoflow in the Fremouw Formation in the CTM, Antarctica. Note that north changes as you move across the map (Modified from Isbell, 2005; data from Barrett 1968; Vavra, 1981; Barrett et al., 1986; Isbell 1990, 1991, 2005; Isbell et al., 1997).

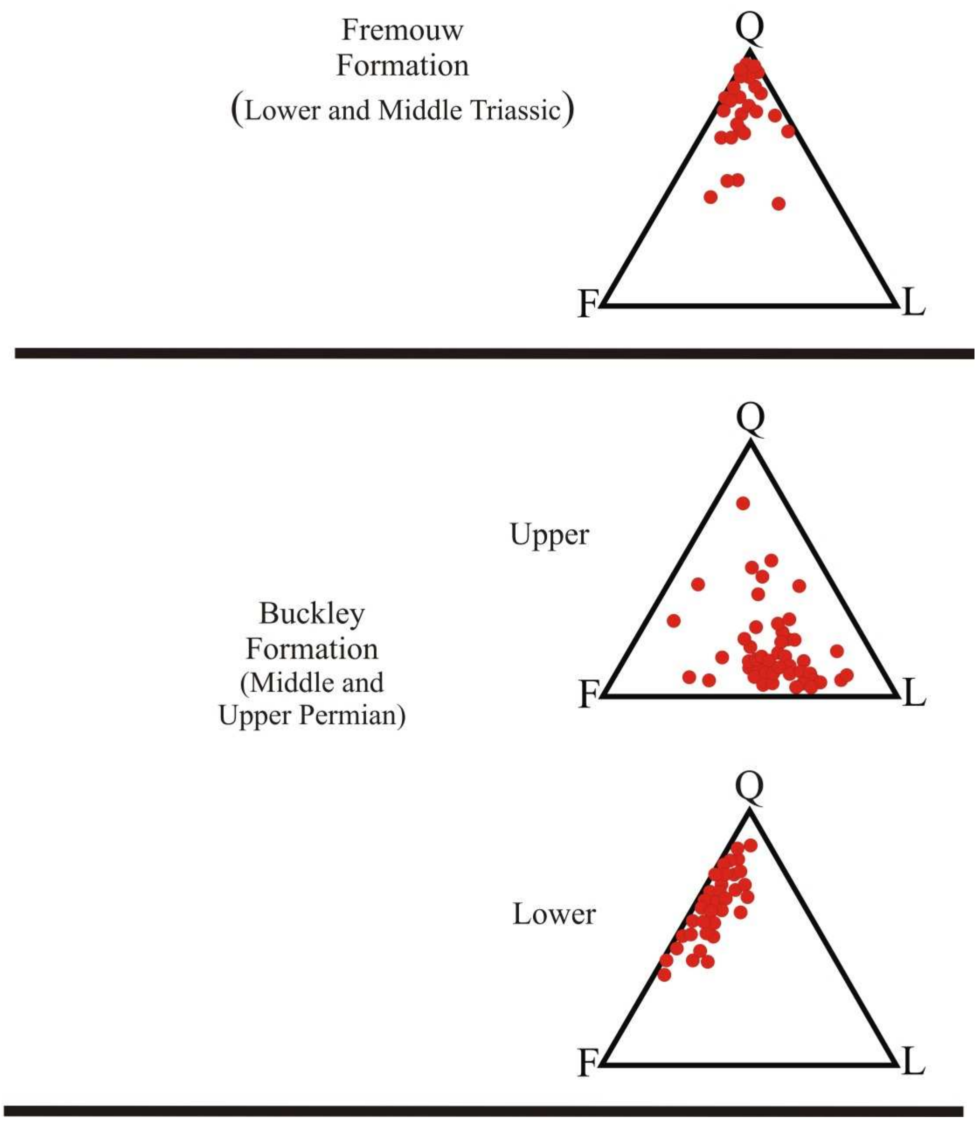



Figure 44. Ternary diagrams showing changes in sandstone composition in the BGR from quartzo-feldspathic in the lower Buckley, to volcaniclastic sandstones in the upper Buckley, to quartz rich sandstones in the Fremouw Formation (Modified from Isbell, 2005; Data from Barrett, 1968; Vavra, 1981; Barrett et al., 1986; Isbell, 1990).

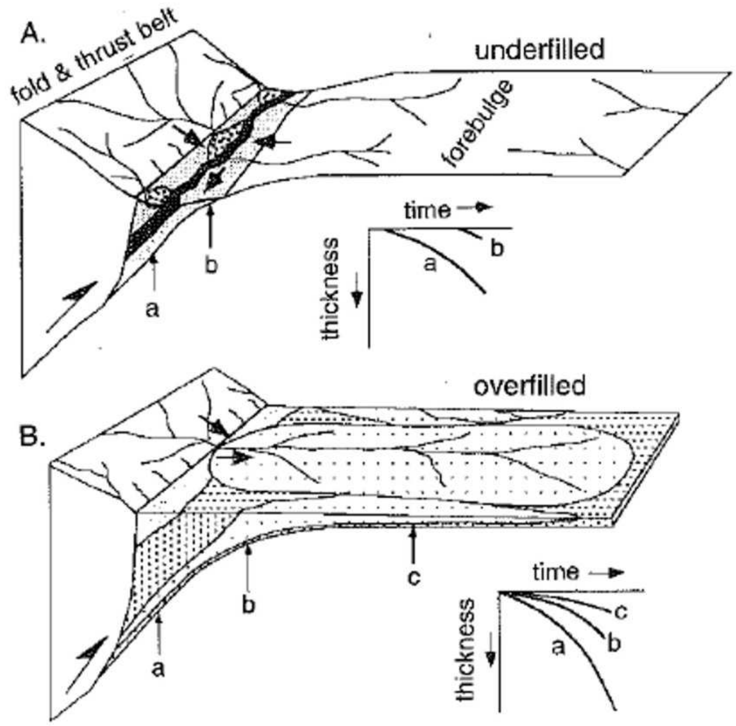



Figure 45. Diagram comparing geometry and lithofacies of under-filled and over-filled foreland basins. A) Under-filled basins form a trough and receive sediment from the fold & thrust belt and the forebulge. B) Over-filled basins show no topographic expression of a depression and receive sediment solely from the fold & thrust belt (Jordan, 1995).

Previous studies state that strata in the Middle to Upper Permian Buckley Formation are characterized by thick floodplain deposits (low stacking pattern); in contrast to the Triassic Fremouw Formation, which contains stacked interconnected sandstone bodies with rare floodplain deposits (high stacking pattern; cf. Barrett et al., 1986). The low stacking pattern however, is not consistent throughout the Buckley Formation either in a vertical succession or laterally across the depositional basin (Isbell & Macdonald, 1991a, 1991b; Collinson et al., 1994; Isbell et al., 1997; 2005). Similarly, the high stacking pattern is not consistent throughout the entire Fremouw Formation

either vertically or regionally (Isbell et al., 2005). This chapter will identify and discuss the various stacking pattern hypotheses for the Transantarctic basin by determining the stacking patterns at various sites within the basin, discuss how the stacking patterns change through time across the Transantarctic basin (Figure 46 a), and identify what controlled the change in fluvial stacking patterns. Middle to Late Permian strata of the Buckley Formation are described from Mt. Bowers, Lewis Cliffs, Mt. Achenar, and Clarkson Peak, and Early Triassic strata of the Fremouw Formation are described from Wahl Glacier and Gordon Valley (Figure 46 b). Other sites in the Beardmore Glacier Region, such as Wyckoff Glacier and Graphite Peak, will also be described where data are available from the literature to discern fluvial stacking patterns.

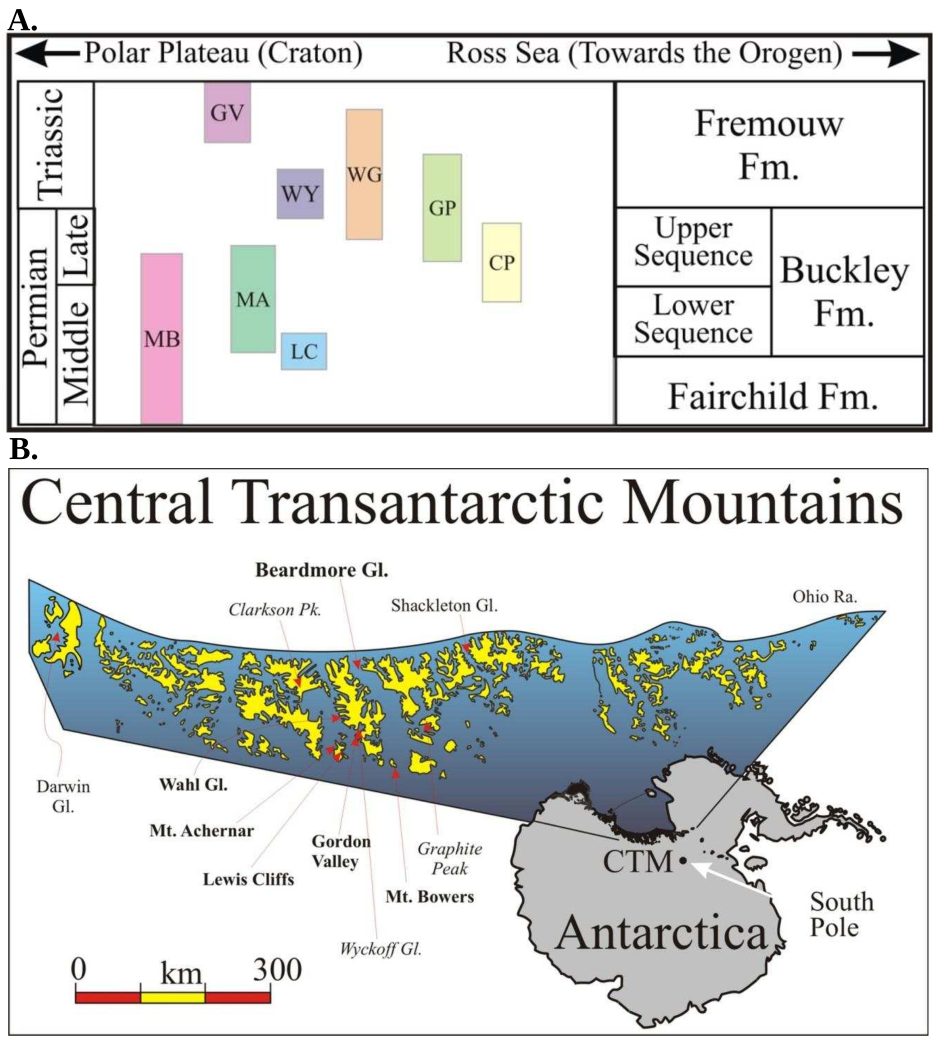



Figure 46. A) Cross sectional diagram displaying the relative position of study locations through time and space perpendicular to the trend of the Transanatarctic basins. Sites include: Mt. Bowers (MB), Lewis Cliffs (LC), Mt. Achenar (MA), Clarkson Peak (CP), Wahl Glacier (WG), Gordon Valley (GV), Wyckoff Glacier (WY), and Graphite Peak (GP) (Modified from Isbell, 2005; Based on data from Barrett, 1968; Vavra, 1981; Barrett et al., 1986; Isbell 1990, 1991, 2005; Isbell et al., 1997; Flaig, 2005). B) Location map of study sites showing the paleogeography of the basin (Modified from Isbell, 2005).

### 3.1 Permian Buckley Formation Fluvial Stacking Patterns

#### Mt. Bowers

*Description:* Strata exposed at Mt. Bowers were deposited on the cratonic side of the foreland basin adjacent to the Polar Plateau (Figure 43 and 46). The stratigraphic succession consists of thick stacked channel deposits of the Lower to Middle Permian Fairchild Formation overlain by a large-scale fining upward succession of the Middle Permian lower Buckley Formation (392 m thick), and 74 m of the Upper Permian upper Buckley Formation preserved near the top of Mt. Bowers (Figure 47).

The 250 m thick Fairchild Formation is dominated by a high fluvial stacking pattern represented by overlapping multistoried, interconnected channel sandstone bodies (Isbell, 1990). These sandstones were interpreted as sandy braided stream deposits that lack fine-grained floodplain deposits (Isbell, 1990). Paleocurrent orientations are towards 160˚ (towards the present day Weddell Sea; Isbell, 1990). The sandstone composition of the Fairchild Formation is quartzo-feldspathic with an average composition of $Q_{58}$ $F_{39}$ $R_3$ (Isbell, 1990).

The contact between strata of the Fairchild and the lower Buckley formations is sharp, possibly erosional with an abrupt increase in floodplain deposits above the contact. However, the basal ~150 m of section in the lower Buckley member consists of abundant, relatively thick, medium- to coarse-grained sandstone sheet facies association, fine- to medium-grained sandstone sheet facies associations, and moderate fine-grained facies association. The medium- to coarse-grained sandstone sheet facies association is interpreted as channel sandstone bodies deposited by braided rivers. The fine- to medium-grained sandstone sheet facies association is interpreted as crevasse-splay

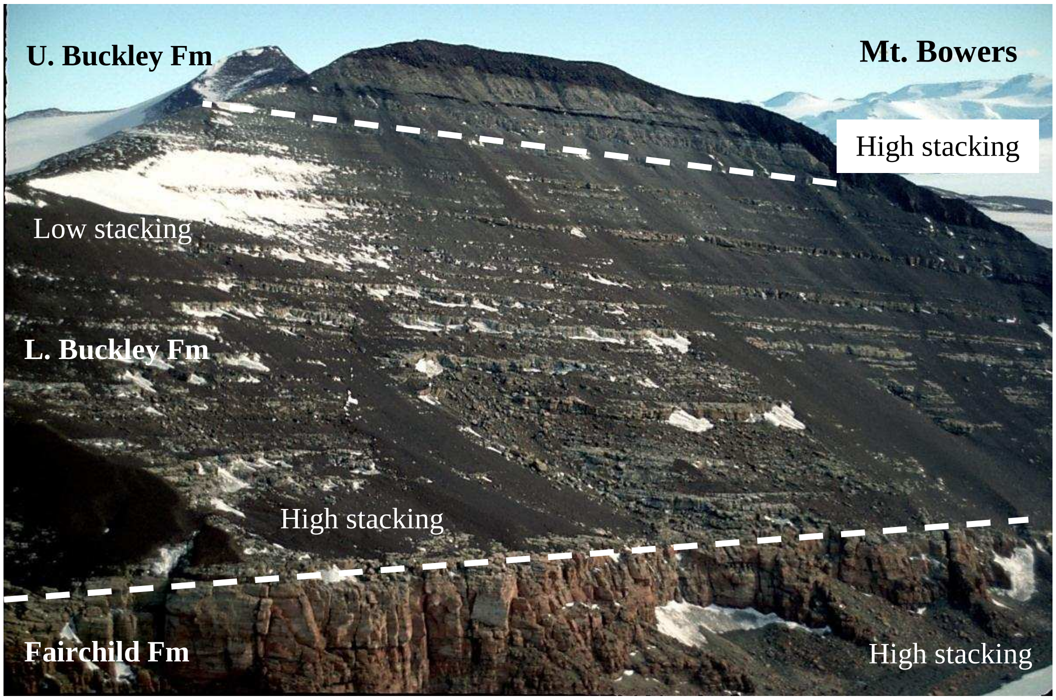



Figure 47. Photograph of the outcrop at Mt. Bowers showing a sharp contact between the Fairchild and Buckley formations. An abrupt increase in floodplain deposits occur at the base of the lower Buckley Formation with a gradual decrease in channel sandstone bodies upward showing lower fluvial stacking patterns. A sharp contact occurs between the lower and upper Buckley formations with an abrupt change back to a high fluvial stacking pattern. Basal sandstone in lower Buckley Formation is ~ 15 m thick.

deposits and the moderate amounts of thin beds of the fine-grained facies association is interpreted as floodplain deposits occurring at the top of fining upward successions of the channel deposits and between crevasse-splay deposits. There is a gradual change up section from a high fluvial stacking pattern to a low fluvial stacking pattern in the lower Buckley member. This change occurs at ~170 m and low stacking patterns occur throughout the rest of the lower member (Figure 48). The low fluvial stacking pattern consists of abundant floodplain deposits with an increase in coal deposits and crevasse-

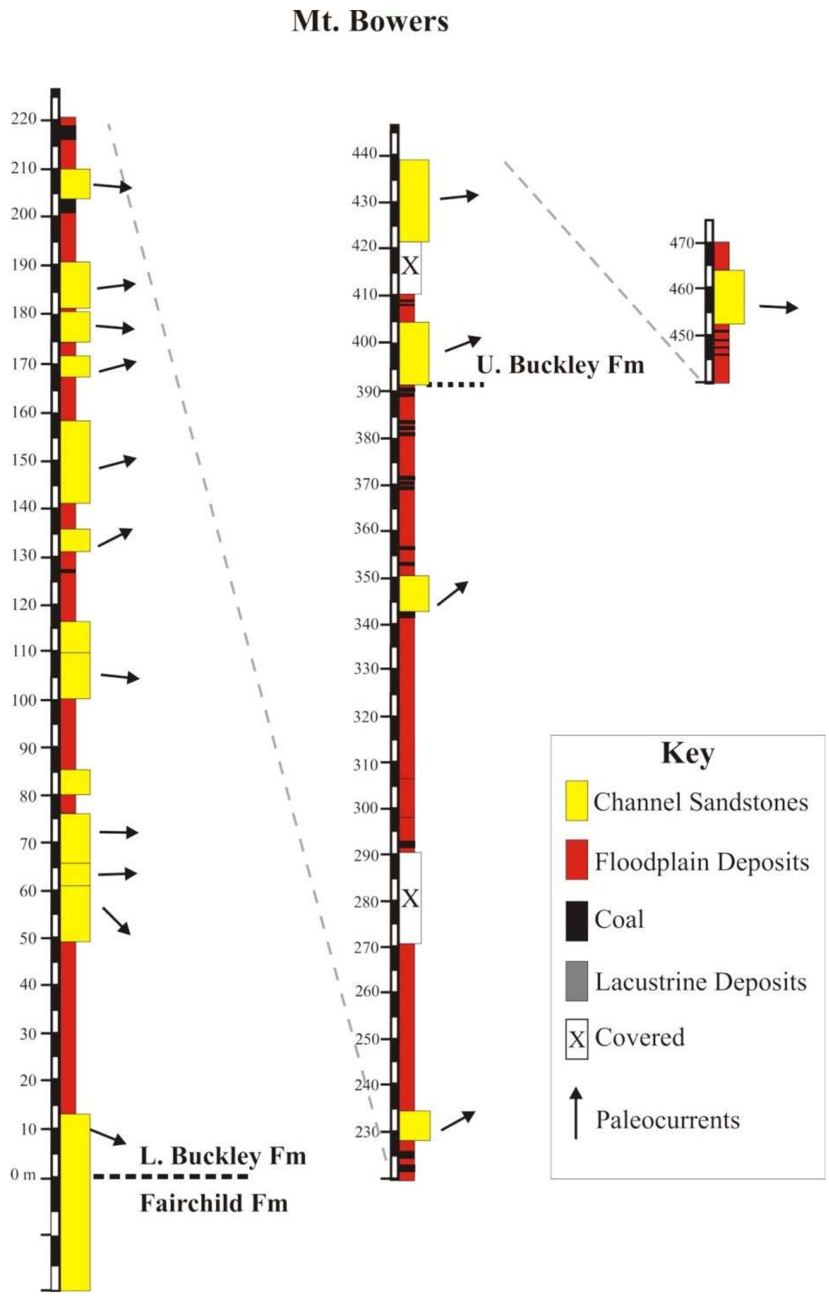



Figure 48. General stratigraphic column of strata deposited at Mt. Bowers showing changes in facies upward from the Fairchild Formation through the upper Buckley Formation. The direction of the paleocurrent arrows are an azimuth orientation with north towards the top of the page.

splay deposits relative to channel sandstone bodies in the upper portion of the lower member. Paleocurrent orientations in the lower Buckley member at Mt. Bowers are to the E at 86˚. Isbell (1990) reported paleocurrent orientations towards 109˚. The sandstone composition of the lower Buckley Formation is more feldspathic than that in the underlying Fairchild Formation with an average composition of $Q_{49}$ $F_{49}$ $R_2$ (Isbell, 1990).

An erosional contact occurs between the lower and upper members of the Buckley Formation (392 m above the base of the Formation). An abrupt change in facies occurs across this contact with interbedded siltstones and mudrocks deposited on floodplains below (lower Buckley) the contact and abundant, medium-grained, volcaniclastic channel sandstone deposits with rounded siltstone rip up clasts above (upper Buckley member). Thin coal deposits and floodplain deposits also occur within the upper member, but are less common when compared to that found in the lower Buckley Formation at this site. The upper member is characterized by a high fluvial stacking pattern represented by abundant channel sandstone bodies relative to floodplain and coal deposits (Figure 47). The upper Buckley Formation sandstones are volcaniclastic (Elliot, personal communication) and the paleocurrent data record an average flow direction of 82˚.

*Interpretation:* The high density stacking pattern displayed throughout the 250 m thick Fairchild Formation, which consist of overlapping and interconnected sandstone bodies without floodplain deposits, is characteristic of basinal settings where long-term sediment influx is greater than the creation of accommodation space. This suggests deposition in a slowly subsiding, cratonic basin with a constant sediment flux (cf. Isbell, 1990). The abrupt increase in fine-grained material above the Fairchild-Buckley contact

suggests that an increase in the rate of the creation of accommodation space relative to sedimentation rates occurred. The gradual transition from a high stacking pattern at the base to a low stacking pattern upward in the lower Buckley Formation indicates a transition from low accommodation to higher accommodation at the top of the lower member, which was likely the result of increased subsidence rates in this portion of the basin through time. Near the top of the lower member, floodplain and coal deposits become more abundant than the channel sandstone deposits. Such a change is consistent with an increase in subsidence upward relative to sediment influx. The presence of coal deposits at the top of the member suggests that basin subsidence rates were high enough at times to result in sediment starvation.

The change in fluvial stacking patterns and changing accommodation rates in the lower member of the Buckley Formation were accompanied by a major change in paleocurrent orientation within channel sandstones and by a change in sandstone composition. Paleoslope changes occur due to changing subsidence patterns within basins (Isbell, 1990). The change from a uniform dipping paleoslope during the Fairchild Formation to transverse flow at the base of the lower Buckley Formation marks the transition from a broad cratonic basin to a narrow trough-shaped foreland basin (Isbell, 1990; Isbell et al., 1997). The paleoslope indicators at Mt. Bowers (cross-beds from channel sandstone bodies) indicate transverse flow off of the cratonic basin margin suggesting deposition in the foreland basin close to a forebulge. Relatively low subsidence rates typically occur in this position within foreland basins. However, subsidence rates increase from the cratonic margin inward towards the basin axis (Flemings & Jordan, 1990; Jordan, 1995). Regional paleocurrent patterns identified by

Isbell (1990, 1991) and Isbell et al. (1997, 2005) indicate that drainage during deposition of the lower Buckley Formation at Mt. Bowers fed trunk streams farther to the east that flowed longitudinally down the axis of the basin toward the present location of the Weddell Sea. The increase in feldspar content, particularly K-feldspars, across the Fairchild-lower Buckley contact suggests that uplift occurred on the craton. Such uplift was likely the result of development of a forebulge associated with the development of the foreland basin (cf. Isbell, 1990).

The abrupt change in facies from interbedded siltstone and coal deposits in the lower Buckley Formation to coarse-grained channel sandstone bodies with rounded siltstone rip-up clasts in the upper Buckley Formation suggests possible massive erosion of the pre-existing floodplain deposits and an abrupt change in the rate of the creation of accommodation space within the basin. The change in facies across the lower-upper Buckley formational contact signals a return to a high fluvial channel sandstone stacking pattern, which indicates a decrease in accommodation space. The high stacking pattern suggests that following development of the unconformity, low accommodation rates relative to sediment influx rates prevailed. This change is accompanied by a slight change in paleoslope orientation from an easterly dipping slope to a north easterly dipping slope. These orientations indicate that this site was located along the cratonic basin margin at this time.

**Lewis Cliffs**

*Description:* The lower Buckley Formation at Lewis Cliffs is located 83 km from Mt. Bowers and ~ 12 km south of Mt. Achenar (Figure 49). It is located near the polar plateau and, at the time of deposition, was located on the cratonic side of the Transanatarctic Basin. However, it was likely located closer to the basin axis than Mt. Bowers as suggested by regional paleocurrent orientations in the lower Buckley Formation (Figure 43 and 46). Three stratigraphic sections were measured at Lewis Cliffs totaling ~ 32 m (Figure 50). Isbell (1990) measured ~ 90 m of section at this site (Figure 51). The measured sections occur within the high fluvial stacking pattern (Figure 49) and contain abundant medium- to coarse-grained channel sandstone bodies relative to floodplain deposits. In section 1, only 5.5 m of channel sandstone bodies were measured due to the occurrence of steep scree covered slopes underlain by ice. Section 2 was measured 23 m to the north of section 1. In section 2 ~10 m of channel sandstone bodies are overlain by floodplain deposits, which is capped by a coal seam. The coal is erosionally overlain by a crevasse-splay channel, which, for this study, is considered a floodplain deposit. Section 3 consists of floodplain deposits overlain by a crevasse-splay channel, which is also considered as a floodplain deposit. The stratigraphic section measured by Isbell (1990) shows abundant channel sandstone bodies relative to floodplain deposits and coal at the base of the succession overlain by thick crevasse-splay bearing floodplain deposits to the top of the section. The channel sandstone deposits at the base of the section are separated by floodplain deposits, which thin upward until ~ 68 m above the base of the section where thick (~42 m thick) floodplain deposits cap the stratigraphic section, which ends at a Jurassic age dolerite sill. Thin coals are encased

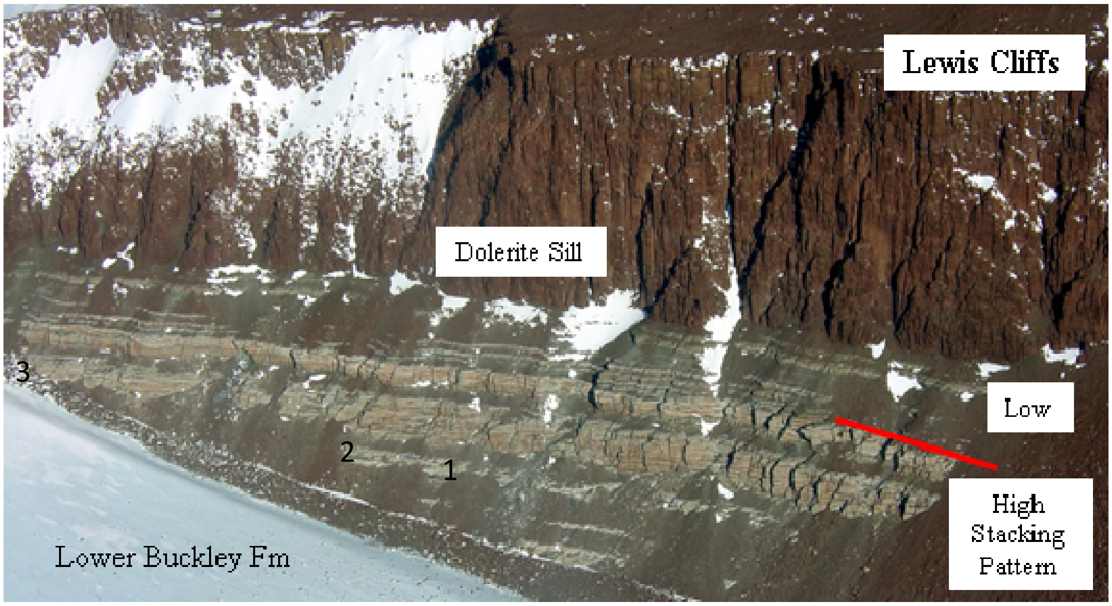



Figure 49. Photograph of the lower Buckley Formation outcrop at Lewis Cliffs. Photograph shows an abrupt transition (red line) from a high stacking pattern to a low stacking pattern. The thick sandstone bodies in the lower half of the photo are channel deposits while the thin sheet sandstones in the upper half of the sedimentary succession are crevasse-splay deposits. The thick red rock at the top of the photo is a Jurassic Dolerite sill. See figure 25 for additional information on this section. Approximate locations of the three measured stratigraphic sections are noted as 1, 2, and 3, respectively. Stratigraphic section is ~125 m thick.

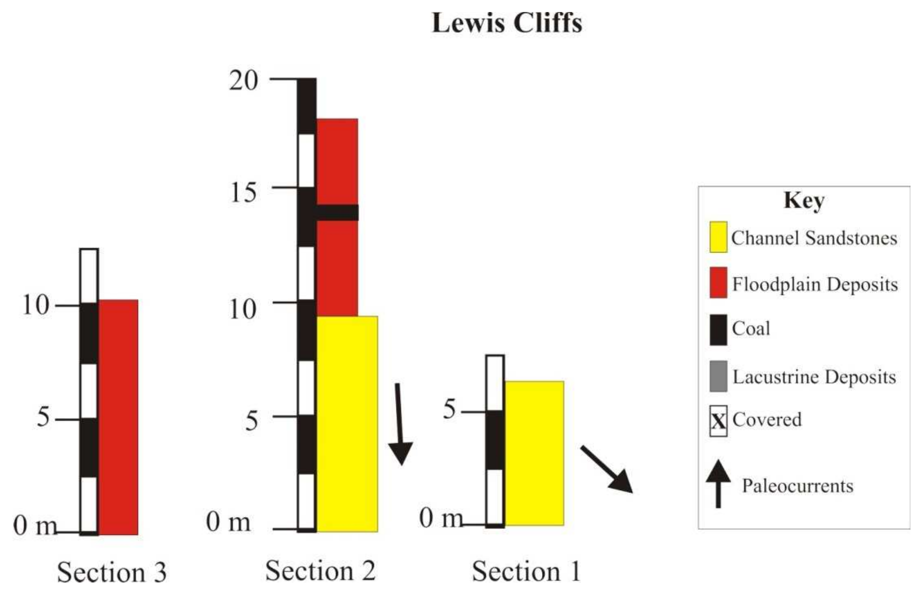



Figure 50. Generalize stratigraphic column of strata deposited a Lewis Cliffs (this study) showing vertical changes in facies and paleocurrents. The direction of the paleocurrent arrows are an azimuth orientation with north towards the top of the page. The floodplain deposit in section 3 consists almost exclusively of sandstone deposited as crevasse-splays and crevasse channels.

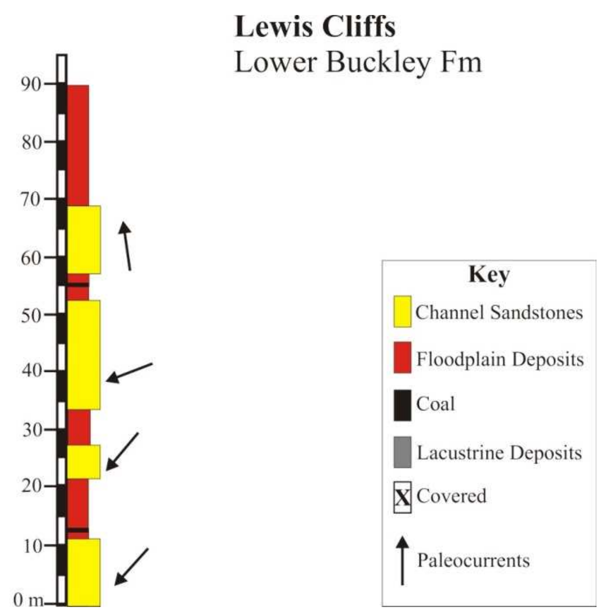



Figure 51. Generalized stratigraphic column of strata deposited in the lower Buckley Formation at Lewis Cliffs (Paleocurrent data from Isbell, 1990). The stratigraphic column shows vertical changes in lithofacies and paleocurrents.

within the floodplain deposits. Overall, the fluvial stacking pattern abruptly changes from a high to low pattern (Figure 49). The abrupt change from a high to a low fluvial stacking pattern is represented by the change from thick amalgamated channel sandstone bodies at the base of the section to thin crevasse-splay deposits encased within thick floodplain deposits upward.

The average sandstone composition for the lower Buckley Formation at this site is $Q_{69}$ $F_{18}$ $R_{13}$ (Isbell, 1990). Isbell (1990) reported paleocurrent orientations to the SW at 222˚. For this study the average paleocurrent orientation is to the SE at 152˚, however variations in paleocurrent orientations occur in the measured sections. In the first measured section paleocurrents are to the SE at 131˚. In section two, paleocurrents are to the SSE at 174˚.

*Interpretation:* The high fluvial stacking pattern at the base of the succession at Lewis Cliffs is dominated by multistoried, fining-upward, channel sandstone bodies and crevasse-splay deposits. These sandstones are interpreted to have been deposited in sandy braided streams with splay deposits extending outward from the channel bodies as thin crevasse splay wings with encased splay channels. This high stacking pattern suggests that the creation of accommodation space was less than the rate of sediment influx. However, an abrupt transition at ~ 68 m above the base of the section records a sharp change to floodplain dominated deposits and development of an extremely low stacking pattern. These floodplain deposits are composed of thin crevasse-splay deposits encased in floodplain deposits that are interbedded with coal seams. The presence of coal

suggests that this location in the basin was, at times, sediment starved. Therefore, the top of the section at Lewis Cliffs represents a change to high accommodation rates relative to sediment influx rates.

The paleocurrent orientations taken from large-scale cross-stratification suggest that the lower Buckley Formation at Lewis Cliffs was flowing almost perpendicular to the lower Buckley paleoflow at Mt. Bowers, which was flowing transversely across the basin margins off of the East Antarctic Craton. At Lewis Cliffs, flow is towards the SE and SW. Such flow was parallel to sub-parallel to regional longitudinal flow within the basin. In under-filled foreland basins, major trunk streams typically flow down the axis of the basin in the topographically lowest point within these trough-shaped depressions. The lower part of the section at Lewis Cliffs may represent deposition from such trunk streams. The abrupt change to low stacking may represent avulsion of this system out of the area, or a shift in the position of the topographic low. Thus, in the absence of trunk streams, higher subsidence rates near the basin axis would produce thick floodplain deposits and a low stacking pattern.

**Mt. Achenar**

*Description:* Mt. Achenar is located 102 km north of Mt. Bowers at the northeastern end of the MacAlpine Hills near where the Law Glacier joins the Polar Plateau. Strata on the outcrop consist of the lower and upper members of the Buckley Formation (Figure 52). A ~100 m thick dolerite sill separates the upper member of the Buckley Formation into two successions.

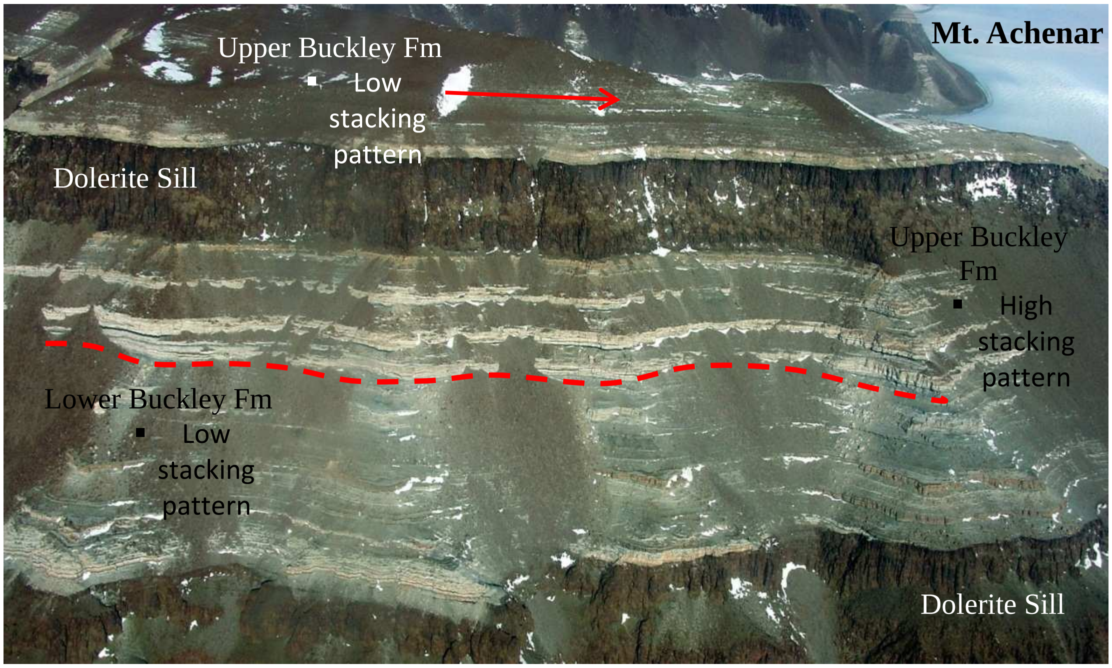



Figure 52. A photograph of the outcrop at Mt. Achenar showing a change from a low fluvial stacking pattern in the lower Buckley Formation to a high fluvial stacking pattern in the upper Buckley Formation beneath the dolerite sill back to a low fluvial stacking pattern above the dolerite sill. The stratigraphic section between the two sills is 175 m thick.

The lower member of the Buckley Formation at this site consists of abundant floodplain deposits and a single channel sandstone deposit occurring 20 m above the base of the section. The first appearance of coal deposits, at this location, occurs as thin beds at ~ 73 m above the base within the floodplain deposits. The lower Buckley Formation at this site is characterized by a low fluvial sandstone stacking pattern. The sandstone composition is quartzo-feldspathic with an average composition of $Q_{68}$ $F_{29}$ $R_3$ and paleocurrent directions are to the S at 180˚ (Isbell, 1990).

The contact between the lower and upper Buckley formations is erosional and occurs at the base of a channel sandstone. This sandstone has an erosional contact with an underlying coal seam. The lower portion of the upper Buckley Formation consists of three channel-form sandstones interbedded with mudrock and coal deposits (Figure 53). The sandstones were deposited in braided stream systems, while mudrock and coal represent floodplain and mire deposits respectively. Thick laminated siltstone deposits interstratified with thin carbonaceous mudrock, coal, and thin fine-grained sandstone deposits continue at the top of the section (Figure 53). The siltstones are interpreted as having been deposited in lakes while the sandstones represent thin splays or deltas prograding into the lake. Interbedding of the siltstones with carbonaceous mudrocks and coals suggest that mires formed along the edge of the lake and suggest expansion and contraction of the lake margins. The abundance of channel sandstone bodies at the base of the upper Buckley formation indicate a moderately high fluvial stacking pattern, whereas upward, the occurrence of lakes and mires indicate a dramatic decrease in fluvial stacking patterns through time. The sandstone composition of the upper Buckley Formation is volcaniclastic with an average sandstone composition of $Q_{8.5}$ $F_{37.5}$ $R_{54}$ (Isbell, 1990). Paleocurrent orientations derived from cross-bedding within the channel-form sandstones vary upward within the upper Buckley Formation. The lowest of the three channels displays a mean paleoflow direction to the SW at 212˚, while the second and third channels return average paleoflow directions of 256˚ and 261˚ respectively (Isbell, 1990).

*Interpretation:* The sandstone composition of the lower Buckley Formation is quartzo-feldspathic and paleocurrent orientations taken from large-scale trough cross-

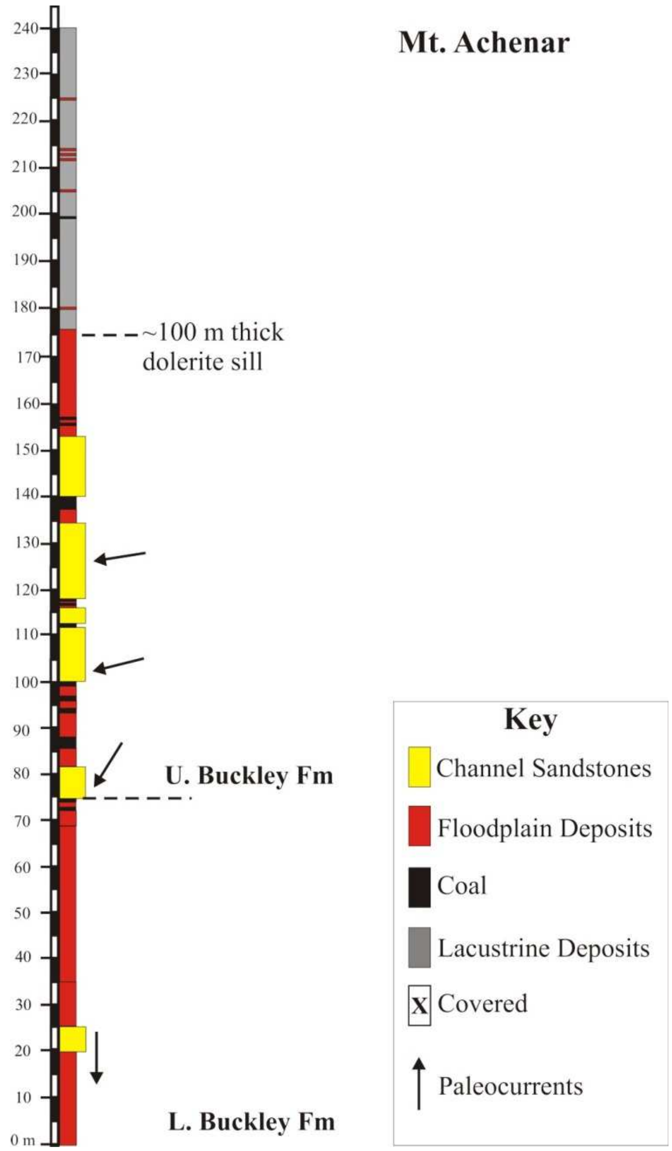



Figure 53. Generalized stratigraphic column of strata deposited at Mt. Achenar showing vertical changes in lithofacies, paleocurrents, and fluvial stacking patterns. The direction of the paleocurrent arrows are an azimuth orientation with north towards the top of the page (Paleocurrent data from Isbell, 1990 and this study).

stratification are orientated towards the S suggesting deposition at or near the southward oriented basin axis. The low stacking pattern is consistent with deposition under high rates of the creation of accommodation space, which is to be expected near the basin axis,

where, outside of the area of major trunk stream, subsidence is greater than the rate of clastic influx.

An abrupt change in sandstone composition from quartzo-feldspathic in the lower Buckley Formation to volcaniclastic sandstones during deposition of the upper Buckley Formation indicates that new sediment sources were supplying clastics to the basin. This change, along with a change in paleocurrent orientation from S in the lower Buckley Formation to SW and then to W in the upper Buckley Formation suggests major reorientation of the basin during deposition of the upper member from that which occurred during the lower member. The presence of westward flow is transverse (perpendicular) to regional flow. This along with the change in sandstone composition could be explained if this was part of a clastic wedge migrating into the basin axis from the orogenic basin margin. However, paleocurrent orientations at this site are anomalous when compared to regional flow directions from nearby sites. The vertical increase in channel sandstone deposits in the upper Buckley Formation suggests an increase in sediment influx relative to accommodation. In contrast, the abundance of interbedded coal seams in the same interval suggests alternation between periods of high clastic influx and periods of sediment starved conditions. Upward, thick lacustrine deposits suggest high accommodation rates due to an increase in subsidence relative to clastic influx. The gradual change from a high fluvial stacking pattern at the base of the upper member to a low stacking pattern near the top of the succession suggests a change from low accommodation to high accommodation.

**Clarkson Peak**

*Description:* Clarkson Peak is located on the eastern margin of the basin, which would have been closer to the orogenic margin than other sites in the basin. Stratigraphic sections were measured during the 1985-86 and the 2010-11 field seasons by John Isbell. A photograph of the outcrop shows a initial high fluvial stacking pattern at the base of the upper Buckley Formation overlain by a thick interval characterized by a low stacking pattern with an increase in channel sandstone abundance near the top of the measured section and a further increase in stacking pattern above the dolerite sill (Figures 54 and 55; Isbell, 1990). Isbell (1990) reported an average volcaniclastic sandstone composition of $Q_{15}$ $F_{31}$ $R_{54}$ with an increase in volcanic rock fragments upward. Isbell (1990) found paleocurrent orientation for the formation changed upward within the section from north to northwestward in the lower part of the succession to westward near the top of the measured section (Figure 56). The upper Buckley Formation consists of channel sandstones, floodplain, coal, and lacustrine deposits. A high fluvial channel sandstone stacking pattern occurs at the base of the outcrop decreasing in density upward. At ~ 180 m above the base of the section a low density stacking pattern occurs throughout the middle of the exposure where floodplain and lacustrine deposits dominate. The remainder of the section shows a gradual, but dramatic increase in the abundance of channel sandstone deposits upward. These channel deposits are separated by interbedded floodplain, coal, and lacustrine deposits. Above the uppermost dolerite sill, thick channel sandstone deposits with rare floodplain deposits dominate (Figure 55).

*Interpretation:* The change in paleocurrent orientations upward from N to W suggests a change from longitudinal flow down the basin axis for sandstones near the

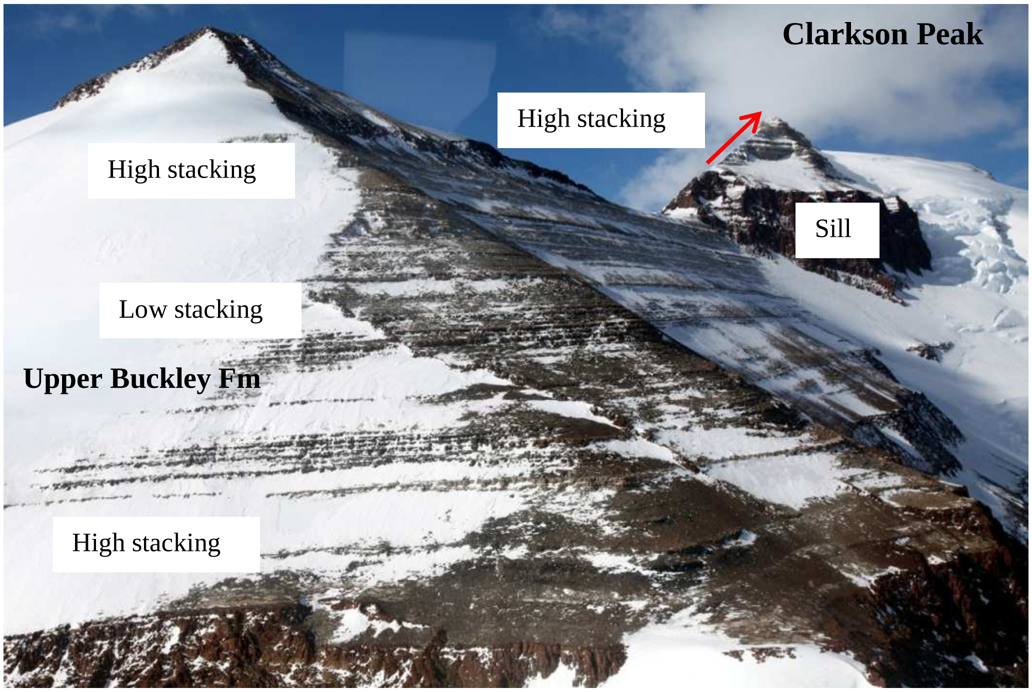



Figure 54. Outcrop photograph of the upper Buckley Formation at Clarkson Peak. The photograph shows a change from a high stacking pattern at the base of the section to a low stacking pattern near the middle of the measured section back to a high stacking pattern near the top of the measured section and above the dolerite sill (Photograph courtesy of John Isbell). The stratigraphic section near the end of the ridge (closest to the camera) is ~ 320 m thick.

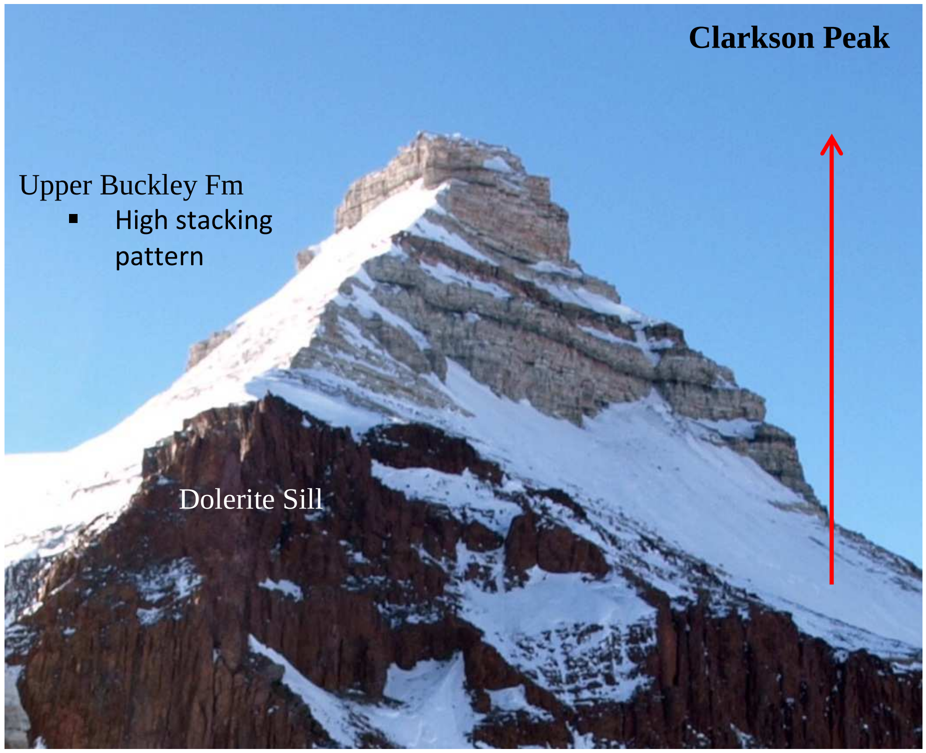



Figure 55. Close up of the high fluvial stacking pattern (indicated by the red arrow) of the upper Buckley Formation above the dolerite sill at Clarkson Peak. The sandstone body is ~100 m thick.

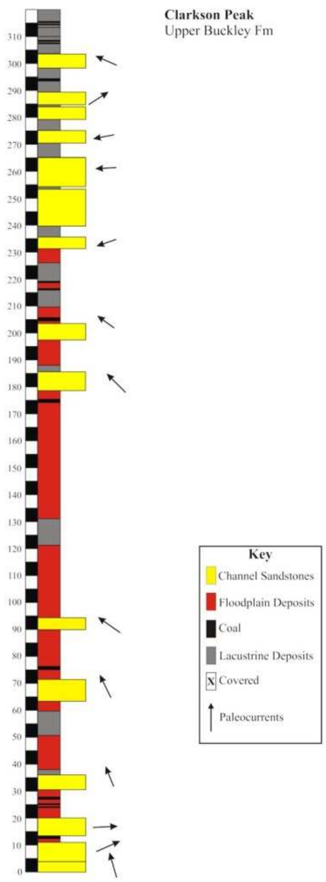



Figure 56. Generalized stratigraphic column of strata deposited at Clarkson Peak. Column shows vertical changes in lithofacies and paleocurrents. The direction of the paleocurrent arrows are an azimuth orientation with north towards the top of the page (Stratigraphic column and paleocurrent data from Isbell personal communication, 2010).

base of the exposure to transverse flow across basin margins for sandstones near the top of the outcrop. The high stacking pattern at the base of the exposure may represent axial trunk streams flowing down a topographic axis within the basin. The low stacking pattern near the middle of the measured section suggests that high accommodation rates relative to clastic influx occurred during deposition of these thick mudrocks and siltstones; lacustrine deposits suggest very low sediment rates. This low stacking pattern is interpreted to be due to high subsidence rates relative to sediment influx.

An increase in channel sandstone bodies occurs towards the top of the Upper Buckley Formation at this site. This suggests that clastic influx increased through time to conditions where clastic influx was greater than the rate of the creation of accommodation space. However, these channel deposits are interbedded with lacustrine siltstones and coal deposits. This can be explained by the progradation of clastic wedges into the basin from the orogenic basin margin (Isbell, 1991) where high subsidence rates along the distal portion of the clastic wedges allowed for interfingering of lake and channel deposits. The gradual increase in channel sandstone bodies upward suggests no hiatus is present and continuous deposition occurred. Thus, the transition to a high stacking pattern with abundant channel sandstone deposits relative to floodplain deposits above the dolerite sill is interpreted as an increase in sedimentation rate relative to subsidence. This would likely have occurred due to continued progradation of the clastic wedges. The gradual increase in volcaniclastic material up section and the gradual change to paleocurrent orientations coming from the direction of the orogenic belt is consistent with the interpretation of the upper sandstones as prograding clastic wedges.

## 3.2 Triassic Fremouw Formation Fluvial Stacking Patterns

Originally the Fremouw Formation was described as having a high fluvial channel stacking pattern with amalgamated channel sand bodies and very little fine-grained deposits (Barrett et al., 1986; Collinson et al., 1994, 2006; Isbell et al., 1997). The abrupt change from a low fluvial stacking pattern represented by the deposition of laminated siltstones, carbonaceous shales, coals, and individual thin sandstone bodies isolated in thick floodplain deposits previously reported in the literature for the Buckley Formation to a high fluvial stacking pattern in the Fremouw Formation characterized by stacked interconnected sandstone bodies with rare floodplain deposits was attributed to catastrophic loss of land plants during the end-Permian Mass Extinction resulting in a dramatic increase in erosion rates (cf. Retallack, 1995; Retallack et al., 1996, 2005; Ward et al., 2000; Michaelsen, 2002). However, the type of stacking pattern exhibited by channel sandstone bodies in the Fremouw Formation is dependent on stratigraphic position and location within the basin as will be shown in this section. The field localities discussed in this section includes Wahl Glacier and Gordon Valley.

### Wahl Glacier

*Description:* At Wahl Glacier the outcrop consists of ~ 330 m of the upper Buckley Formation, ~ 100 m of the lower Fremouw Formation, and ~ 210 m of the middle Fremouw Formation (Figure 57; Barrett et al., 1986). The average sandstone composition of the upper Buckley Formation is $Q_{10}$ $F_{25}$ $R_{65}$ (Flaig, 2005). A vector mean of all paleocurrents in the Buckley Formation is towards 267 ˚ (Flaig, 2005). At the base of the stratigraphic section the vector mean is to the SW at 203˚, with a change in the

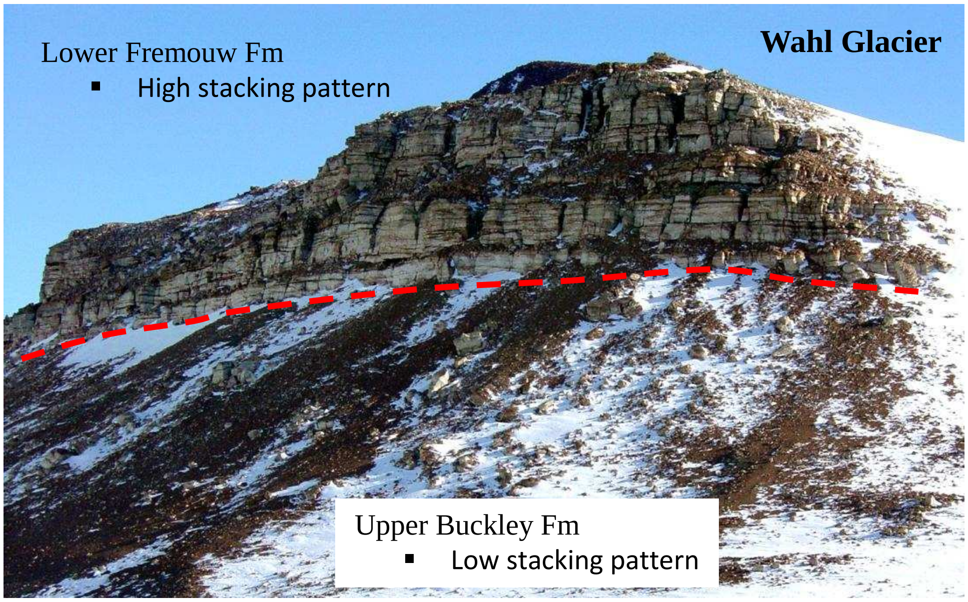



Figure 57. Photograph showing a change from a low stacking pattern in the upper Buckley Formation (mudrock covered scree slope) to a high stacking pattern in the lower Fremouw Formation at Wahl Glacier. Stratigraphic section is ~ 640 m thick.

vector mean to the NW at 289 ˚ at the top of the upper Buckley Formation (Flaig, 2005). Based on a generalized stratigraphic column using data from Barrett (1968) and Flaig, (2005) the upper Buckley Formation consists of channel sandstone deposits and floodplain deposits interbedded with coals (Figure 58). Channel sandstone deposits are rare at the base of the section with only two significant channel sandstone deposits in the first 100 m above the base of the section (Figure 58). At 110 m above the base there is a slight increase in channel sandstone deposits represented by three channel sandstone deposits separated by thin floodplain deposits. At 155 m above the base channel sandstone deposits become rare for the remainder of the upper Buckley Formation and

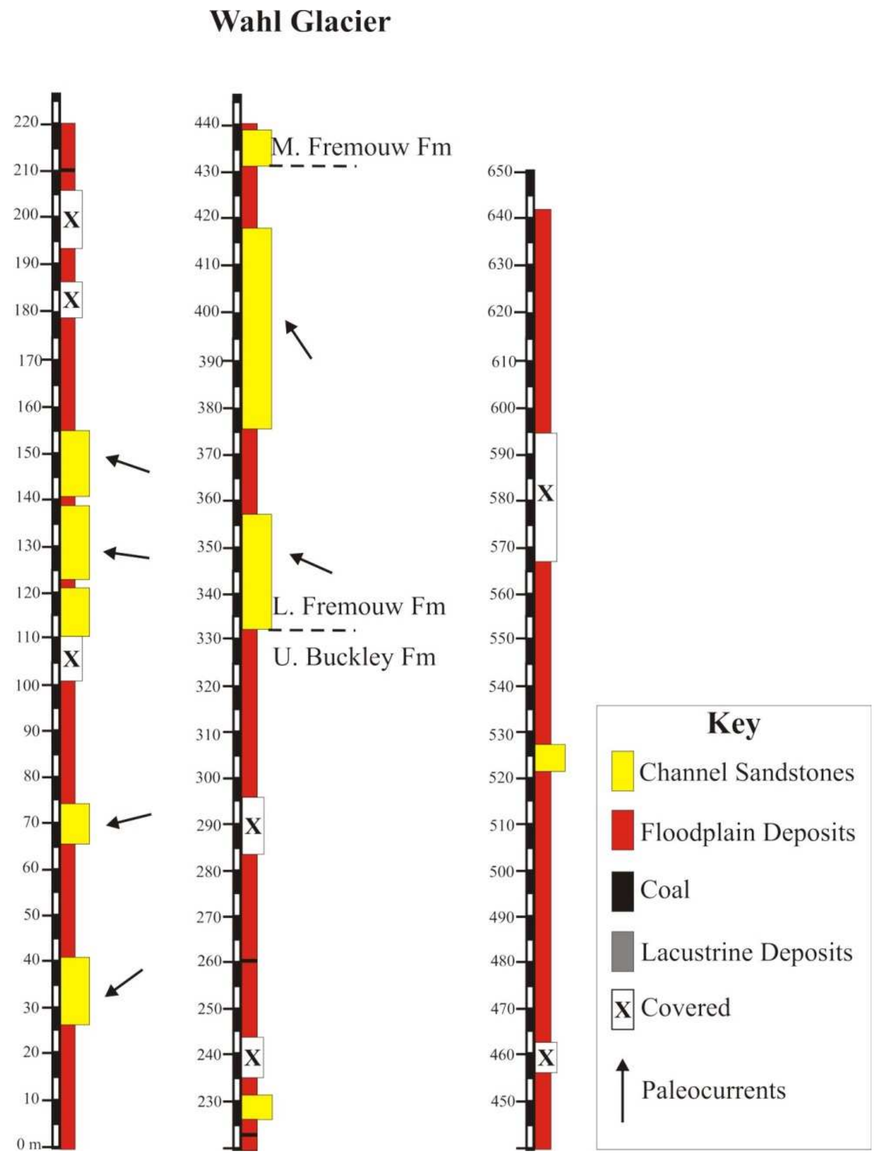



Figure 58. Generalized stratigraphic column of strata deposited at Wahl Glacier. Column shows vertical changes in lithofacies and paleocurrent orientations. The direction of the paleocurrent arrows are an azimuth orientation with north towards the top of the page (Facies based off of Barrett, 1968 and paleocurrent data from Flaig, 2005).

floodplain deposits interbedded with thin rare coal deposits occur. The contact between the upper Buckley Formation and lower Fremouw Formation is erosional. The average sandstone composition ($Q_{81}$ $F_{14}$ $R_5$) changes drastically by a decrease in lithics (less than

12%) and an increase in quartz (~ 80%) at the base of the lower Fremouw Formation (Flaig, 2005). Paleocurrents vary slightly from 294 ˚ to 326 ˚ over an interval of 27 m but remain to the NW for the remainder of the Fremouw Formation (Flaig, 2005). The lower Fremouw Formation predominately consists of thick multistoried channel sandstone deposits and is separated by non-carbonaceous floodplain deposits (Figure 57). The contact between the lower and middle Fremouw is gradational (Barrett, 1968). Sandstone composition returns to volcaniclastic in the middle Fremouw Formation and paleocurrent orientations remain to the NW (Barrett et al., 1986). The middle Fremouw Formation contains abundant thick floodplain deposits and rare relatively thin channel sandstone deposits. An 8 m thick channel sandstone occurs at the base of the middle Fremouw Formation and another ~ 7 m thick channel sandstone deposit occurs at about ~520 m above the base of the section. The remainder of the middle Fremouw Formation consists of thick floodplain deposits including thin crevasse-splay deposits. Overall, the fluvial stacking pattern abruptly changes from a low stacking pattern in the upper Buckley Formation, to a high stacking pattern in the lower Fremouw Formation back to a low stacking pattern in the middle Fremouw Formation.

*Interpretation:* At Wahl Glacier the abrupt change in sandstone composition from volcaniclastic sandstone in the upper Buckley Formation to quartz rich sandstone in the lower Fremouw Formation along with an abrupt change in lithofacies suggest that a disconformity occurs between the formations. The change from a low to high fluvial stacking pattern is recognized by the abrupt increase in density and interconnectedness of channel sandstone bodies in the lower Fremouw Formation. This change is due to high rates in the creation of accommodation space relative to clastic influx rates during

deposition of the upper Buckley Formation changing to high rates of clastic influx relative to accommodation rates in the lower Fremouw Formation. The low stacking pattern in the middle Fremouw represents high accommodation rates relative to clastic influx rates.

**Gordon Valley**

*Description:* The outcrop at Gordon Valley consists of the middle Fremouw Formation (90 m thick) overlain by the upper Fremouw Formation (80 m thick; Figure 59; Isbell & Macdonald, 1991). The sandstone composition in the middle Fremouw Formation is volcaniclastic (Barrett et al., 1986; Collinson et al., 1994; Isbell & Macdonald, 1991) and paleocurrent directions have not been noted likely due to the lack of channel sandstone deposits. The middle Fremouw Formation consists of greenish-gray siltstone, very fine sandstone, silty mudstone, and rare black shale and has been interpreted to be fluvial overbank deposits, crevasse-splay and crevasse-channel sandstones (Barrett et al., 1986; Isbell & Macdonald, 1991). The contact between the middle and upper Fremouw Formations is erosional with an abrupt increase in thick abundant channel sandstone deposits. The sandstone composition of the upper Fremouw Formation is a very coarse- to pebbly granular-grained, quartzose sandstone (Isbell & Macdonald, 1991). Paleocurrents for this study trend to the NW at 332˚. Similarly, Isbell & Macdonald (1991) reported paleocurrent orientations toward the north. A 27 m stratigraphic section was measured in the upper Fremouw Formation. The measured section consists of abundant thick channel sandstone deposits and subordinate floodplain deposits (Figure 60). The floodplain deposits occur above the measured channel sandstone and are, in turn, overlain by channel sandstone deposits. Such interbedding of

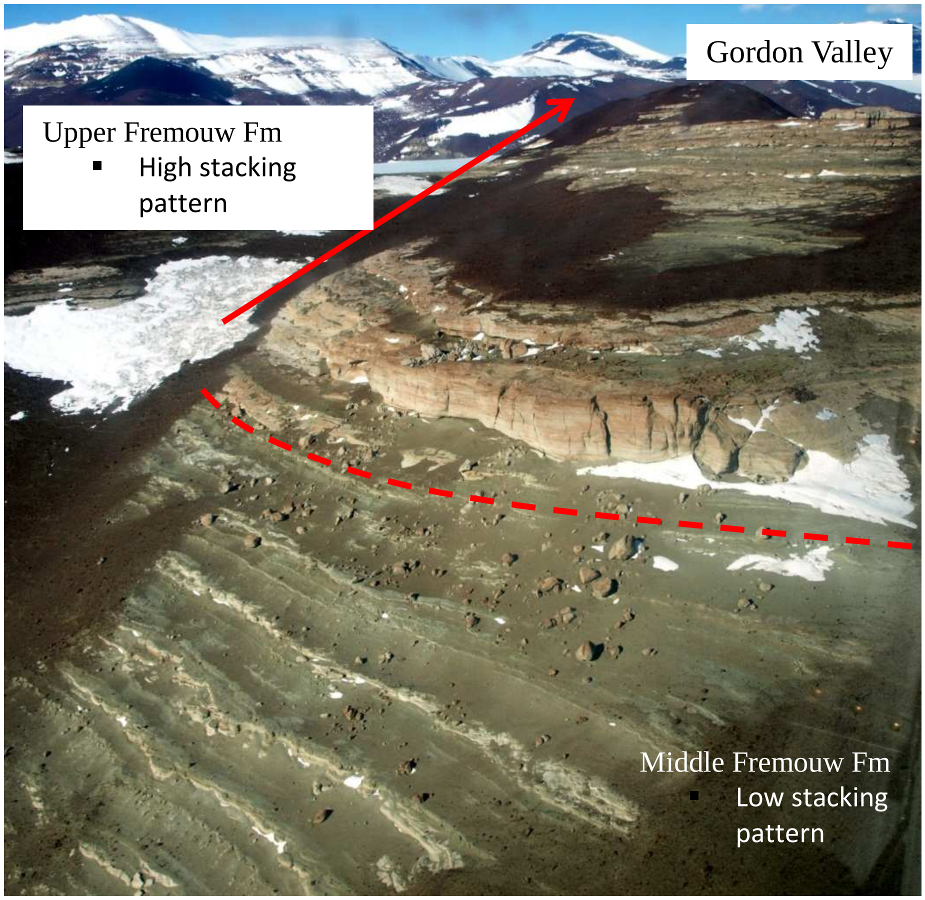



Figure 59. Photograph showing the abrupt transition from a low fluvial stacking pattern in the middle Fremouw Formation to a high fluvial stacking pattern in the upper Fremouw Formation at Gordon Valley. The sandstone at the base of the upper Fremouw Formation is ~20 m thick.

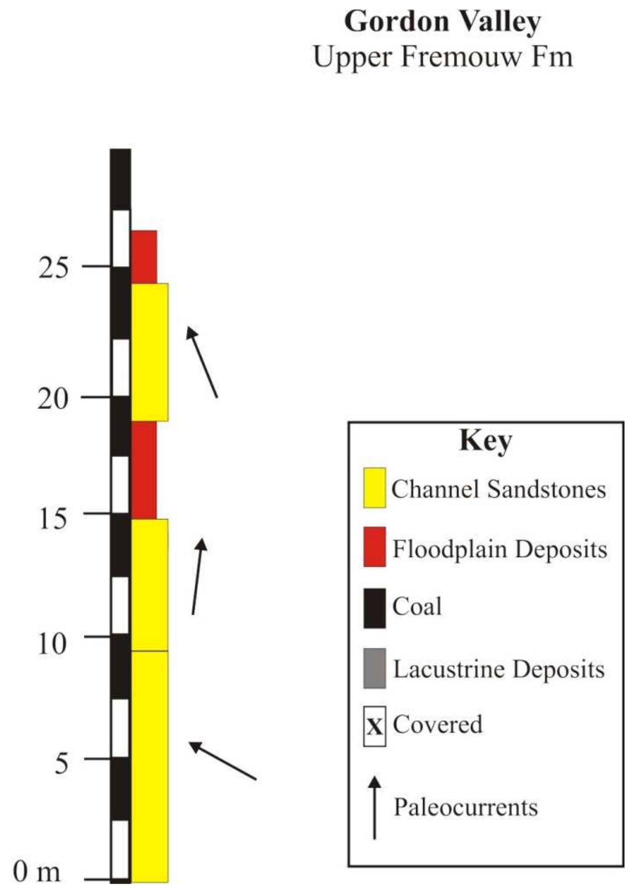



Figure 60. Generalized stratigraphic column showing vertical changes in facies and paleocurrent directions from the upper Fremouw Formation at Gordon Valley. The direction of the paleocurrent arrows are an azimuth orientation with north towards the top of the page.

thick channel sandstone deposits and floodplain deposits continue for several hundred meters into the overlying upper Fremouw Formation. The measured channel sandstone deposits are interpreted to be deposited by a low sinuosity braided river. Overall, there is an abrupt change from a low fluvial stacking pattern in the middle Fremouw Formation to a high fluvial stacking pattern in the upper Fremouw Formation.

*Interpretation:* Paleocurrent orientations to the NW suggest Gordon Valley is located proximal to the craton with paleoslope orientation in the Fremouw Formation, and paleocurrent orientations at this site oriented transverse to the direction of the Transantarctic basin. In the middle Fremouw Formation the low stacking pattern is represented by abundant floodplain and crevasse-splay deposits relative to channel sandstone bodies suggesting high accommodation due to subsidence greater than sediment supply. An abrupt change from abundant floodplain deposits relative to channel sandstone bodies to abundant channel sandstone bodies relative to floodplain deposits occurs in the upper Fremouw Formation. The change from a low to high fluvial stacking pattern with amalgamated bar forms and channel sand bodies with little floodplain deposits suggests a change from high accommodation to low accommodation relative to clastic influx/avulsion rates. This could be due to a decrease in subsidence or due to an increase in sedimentation rates.

**3.3 Discussion**

Sedimentary basins are characterized by either under-filled or over-filled conditions (Jordan, 1995). In under-filled basins, a topographic depression occurs in areas where the creation of accommodation space is greater than sedimentation rates. These basins are characterized by trough shaped depressions where dispersal systems flowing across basin margins are transverse to longitudinal drainage systems flowing down the basin axis. In over-filled basins, topography does not reflect subsidence in the basin and drainage flows across an alluvial ramp that dips transversely away from upland areas (cf., Jordan, 1995). Cratonic basins are typically characterized by slow subsidence

rates; whereas, foreland basins display much higher subsidence rates. Flexural subsidence in foreland basins is greatest directly beneath tectonic loads and decreases in magnitude towards the craton where flexure results in uplift of a forebulge (Jordan, 1995). Sediment loading within the basin and continued migration of the tectonic load will cause expansion of the basin and displacement of the zone of maximum subsidence and the zone of subsidence cratonward through time.

In under-filled foreland basins, the relationship between accommodation and sedimentation define three different basinal environments and associated large-scale basinal facies or motifs. These settings and facies are: the orogenic basin margin, the basin axis, and the cratonic basin margin (Figure 61). The orogenic basin margin is characterized by high subsidence rates, and although the creation of accommodation space in the basin is high, right at the basin margin, the rate of the influx of clastics from the orogenic belt exceeds the creation of accommodation space. These conditions favor erosion of the orogen and progradation of clastic wedges into the depositional basin. Sedimentation rates decrease outward towards the basin axis. Here high subsidence rates allow accommodation to exceed sedimentation rates. However, the topographic axis will be the locus for trunk streams flowing longitudinally down the basin axis (Flemings & Jordan, 1990; Jordan, 1995). Along the cratonic basin margin, subsidence rates are low as are sedimentation rates. However, material eroded from the forebulge allows sedimentation rates to exceed accommodation rates (cf. Flemings & Jordan, 1990). Clastic wedges will also prograde basinward away from the cratonic basin margin. Sediment loading and continued migration of the tectonic load result in cratonward

displacement of these environments and facies through time (Flemings & Jordan, 1990; Jordan, 1995).

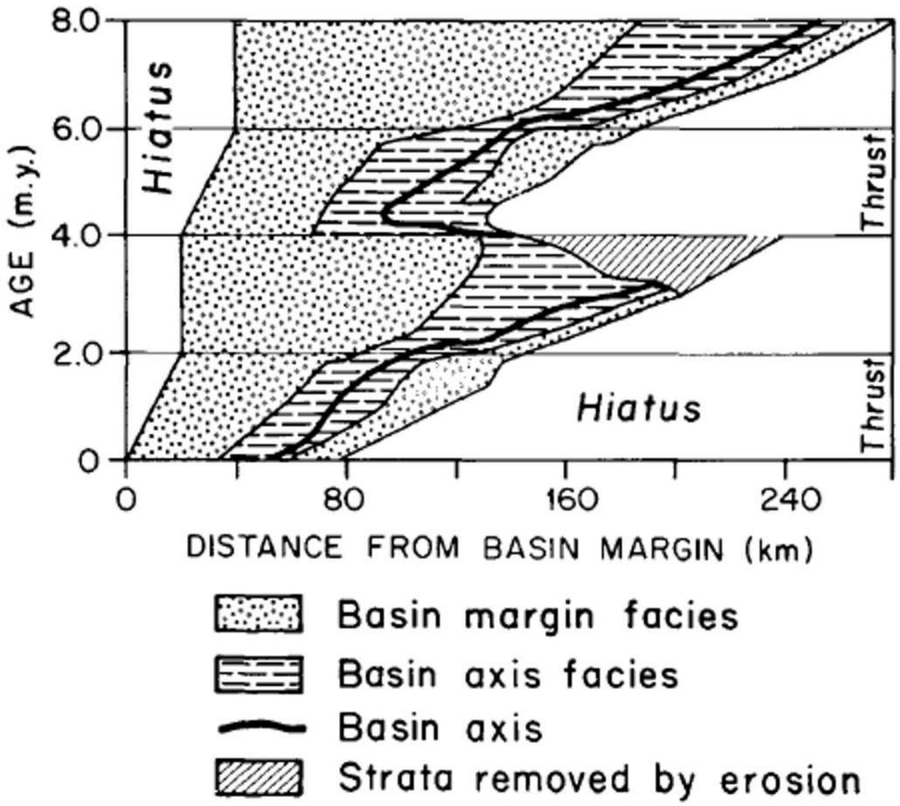



Figure 61. Chronostratigraphic plot showing migration of an underfilled foreland basin and associated facies through time due to tectonic activity and loading of the basin by sediment (Flemings & Jordan, 1990).

In the BGR of the CTM there is a regional change in sandstone composition, paleocurrent orientations, and stacking patterns from strata of the Fairchild Formation, to that of the Buckley Formation. A change also occurs between the lower and upper members of the Buckley Formation. Another change occurs between strata of the upper Buckley Formation and the Fremouw Formation.

Sandstones in the Fairchild Formation are quartzo-feldspathic. An increase in feldspar, primarily K-feldspar occurs upward into the lower Buckley Formation along the cratonic side of the basin, while upper Buckley Formation sandstones are volcaniclastic. In the BGR, lower and upper Fremouw sandstones are sub-arkosic quartz arenites. However, volcaniclastic sandstones occur in the lower, middle, and upper Fremouw Formation sandstones outside of the study region (up the paleoslope) in the Shackleton Glacier region, and throughout CTM, including the BGR during deposition of the middle and upper members. These changes signal the occurrence of changing tectonic conditions and changing source terrains.

The abrupt change from a uniformly dipping paleoslope showing SSE flow in the Fairchild Formation to transverse and longitudinal flow in the Buckley Formation suggests that, at that time, the basin evolved from a uniformly southward dipping cratonic basin to a trough-shaped, under-filled foreland basin (Isbell, 1990, 1991; Isbell et al., 1997). This change coincides with the increase in feldspar content of sandstones derived from the craton indicating uplift along the cratonic basin margin, possibly due to development of a forebulge. Sandstone composition and transverse paleocurrent orientations at Mt. Bowers support this interpretation. Also, abrupt changes in sandstone composition and paleocurrent orientations like those that occur at Mt. Bowers typically signal the occurrence of disconformities between sedimentary packages (Miall, 1984). At Lewis Cliffs, SW flow perpendicular to that at Mt. Bowers suggests regional flow down the basin axis. A reversal in regional paleocurrent orientations, from SE to NW, occurs between paleoflow indicators in the lower and upper Buckley Formation. This change occurs at the same time as the change to volcaniclastic sandstone, which indicates

major rearrangement of the basin due to changing subsidence regimes or due to massive influx of volcaniclastic detritus (Isbell 1990; Collinson et al., 1994; Isbell et al. 1997). Paleocurrent orientations in the upper Buckley Formation indicate the continued occurrence of a trough shaped, under-filled, foreland basin. At Mt. Bowers flow was transverse (E) across basin margins off of the craton; whereas, at Clarkson Peak, on the opposite (orogenic) side of the basin, drainage was transverse (W) across the orogenic basin margin. Regional flow during deposition of the upper Buckley member was toward the northwest down the axis of the basin (Isbell, 1990; Isbell et al., 1997, 2005). Predominately transverse flow in the Fremouw Formation with paleocurrent orientations to the NW indicates that the basin became an over-filled foreland basin in the Triassic with dispersal of clastics flowing transversely away from the orogenic belt and across the structural trend of the Transantarctic basin (cf. Barrett et al., 1986; Vavra 1984; Isbell, 1990, 1997, 2005; Collinson et al., 1994). Paleoslope changes occur due to changing subsidence patterns within basins (Isbell, 1990). Therefore, paleocurrent orientations indicate that the Fairchild Formation was deposited in an over-filled cratonic basin (no trough-shaped depression) while lower and upper members of the Buckley Formation were deposited in an under-filled, trough-shaped, foreland basin (Isbell, 1990; Isbell et al., 1997; 2005). Paleocurrent orientations for the Fremouw Formation indicate that it was deposited on a paleoslope that dipped uniformly away from an orogenic belt. Such conditions indicate that the Transantarctic basin was an over-filled foreland basin at that time (Isbell et al., 2005).

The high fluvial stacking pattern in the Fairchild Formation supports the interpretation of deposition in a slowly subsiding, over-filled basin. An abrupt change

from a high fluvial stacking pattern lacking fine-grained floodplain deposits in the Fairchild Formation to a high fluvial stacking pattern with an increase in the abundance of floodplain deposits characterizes the base of the lower Buckley Formation at Mt. Bowers. Although not reported here, but observed at Tillite Glacier, a low stacking pattern at the base of the lower Buckley Formation overlies the high stacking pattern in the Fairchild Formation (Isbell, 1990, and this thesis). These changes indicate that a slight increase in the rate of the creation of accommodation space verses sedimentation rates occurred at Mt. Bowers, whereas, a much higher increase in accommodation space to sedimentation occurred at Tillite Glacier. These changing stacking patterns in time and space suggest development of differential subsidence in the basin with Mt. Bowers characterized by cratonic basin margin subsidence and sedimentation patterns while Tillite Glacier may have been characterized by basin axis conditions. The S and SW paleocurrent orientations at the base of the lower Buckley Formation at Lewis Cliffs suggest deposition along the basin axis. However, the high stacking pattern is similar to that at Mt. Bowers. The stacking pattern at Lewis Cliffs may represent deposition from trunk streams located near the topographic axis of the basin. At Lewis Cliffs and at Mt. Bowers there is a dramatic decrease in the density of fluvial stacking patterns upward in the lower Buckley member. The top of the lower Buckley Formation at Mt. Achenar is also characterized by an extremely low density fluvial stacking pattern. The southerly paleoflow at Mt. Achenar suggest that the basin axis was located in that position during deposition of the upper portion of the lower Buckley member. As foreland basins fill, sediment loading in the basin results in a cratonward expansion of the basin and concomitant cratonward shift in the location of the basin axis and its associated facies.

The stacking patterns at these three sites (Mt. Bowers, Lewis Cliffs, and Mt. Achenar), can be explained by a cratonward shift in the position of the basin axis through time (Isbell et al., 2005). Higher subsidence rates and lower sedimentation rates (outside of the area of the location of the trunk stream) would produce a low density fluvial stacking pattern (Figure 62). Differential subsidence would have characterized the basin during Buckley deposition with high subsidence rates occurring near the orogenic basin margin decreasing outward toward the craton and onto the forebulge. Fluvial stacking patterns, paleocurrent orientations, and sandstone composition all suggest tectonism and basin subsidence were driving sedimentation patterns in the lower Buckley Formation.

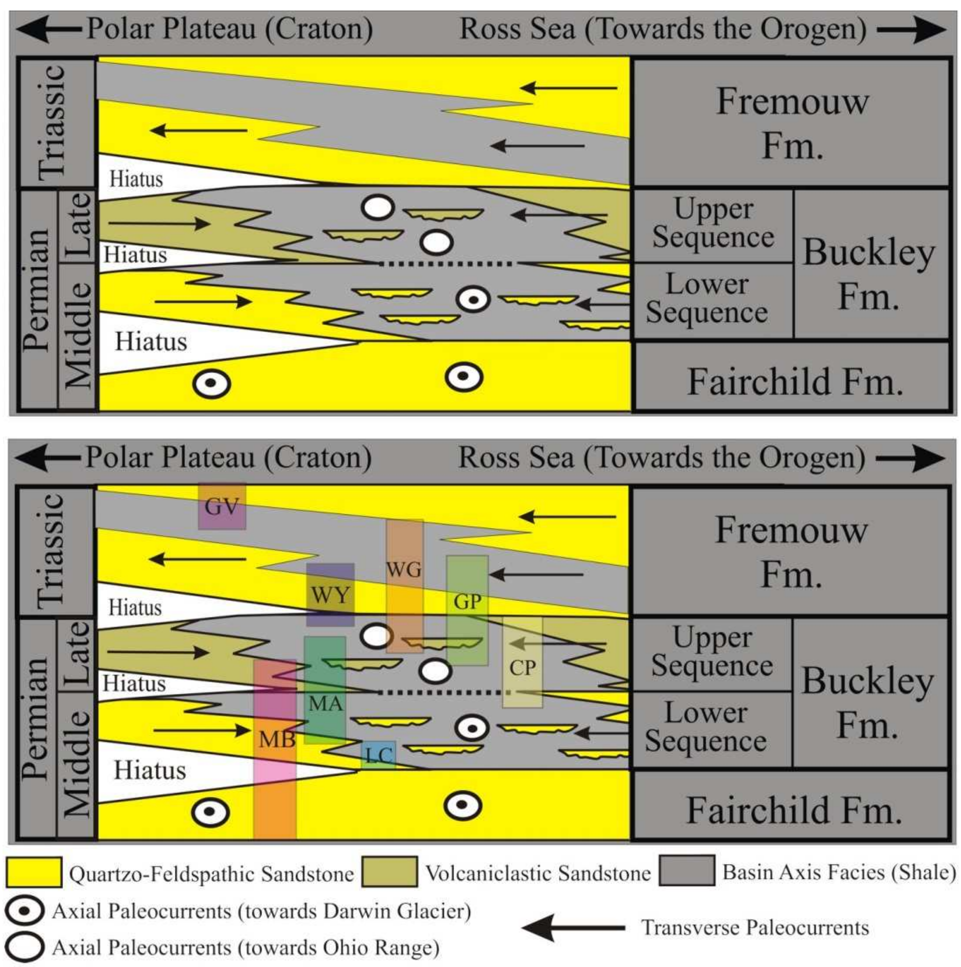


Figure 62. Time space diagram displaying large-scale lithofacies patterns and paleocurrent orientations, which define an under-filled basin during the deposition of the Late Permian Buckley strata and a transition to an over-filled basin during the deposition of the Triassic Fremouw strata. Study locations include Mt. Bowers (MB), Lewis Cliffs (LC), Mt. Achenar (MA), Clarkson Peak (CP), Wahl Glacier (WG), Gordon Valley (GV), Wyckoff Glacier (WY), and Graphite Peak (GP) (Modified from Isbell, 2005; based on data from Barrett, 1968; Vavra, 1981; Barrett et al., 1986; Isbell, 1990, 1991, 2005; Isbell et al., 1997, and this thesis).

The abrupt change in sandstone composition from quartzo-feldspathic to volcaniclastic, and a regional reversal in paleocurrent orientations indicate rearrangement of subsidence and depositional patterns in the basin in the BGR. Along the cratonic side of the basin, where paleoslope orientations towards the E indicate flow off of the craton transversely into the basin, an abrupt change in sandstone composition from quartzo-feldspathic to volcaniclastic sandstones, the abrupt change in fluvial channel stacking patterns and the occurrence of a siltstone intraformational conglomerate at the lower-upper Buckley contact suggest that the two units are separated by a disconformity. At this site, the high density fluvial stacking pattern in the upper Buckley member suggest that avulsion rates, and therefore sedimentation rates were greater than the creation of accommodation space. Such conditions are typical of low subsidence rates and the progradation of a clastic wedge into the basin from the cratonic basin margin (Flemings & Jordan, 1990). The fluvial stacking pattern also shows a similar abrupt change at Mt. Achenar. However, paleocurrent orientations suggest flow towards the craton, which is opposite from that at the base of the upper Buckley member at Mt. Bowers. Such flow is difficult to explain, unless, the basal portion of the upper Buckley Formation at Mt. Achenar represents the distal portion of a clastic wedge prograding off of the orogenic margin, or that it represents partitioning within a sub-basin within the larger foreland

basin. Because Mt. Achenar is located relatively close to the cratonic margin, it likely represents local flow within a sub-basin. Upward at Mt. Achenar, a change to a low density fluvial stacking pattern and the abundance of lacustrine facies signals an increase in the rate of the creation of accommodation space, which is likely the result of increasing subsidence rates through time, or a decrease in sedimentation rates/avulsion rates. In contrast, at Clarkson Peak, which is located closer to the orogenic margin, the fluvial stacking pattern changes gradually from a high stacking pattern at the base of the upper Buckley with flow towards the north down the axis of the basin to a low stacking pattern with rotation of the paleoflow direction towards the NW in the middle of the upper member to an extremely high density stacking pattern with paleoflow directions towards the west at the top of the member. The gradual change in stacking pattern suggests the strata are conformable, as does a gradual increase in the abundance of volcaniclastic framework grains within sandstones upward in the section. The high stacking pattern with northward flow at the base of the upper member suggests deposition in trunk streams along the basin axis. The gradual westward rotation in paleocurrent orientations and the change to a low density stacking pattern also suggest deposition in the high subsidence sediment starved zone just off of the drainage axis and in front of a prograding clastic wedge. Continued rotation of the paleoflow indicators and an increase in the density of the fluvial stacking pattern suggest continued progradation of a clastic wedge into the basin from the orogenic basin margin. This orogenic side of the basin is the site of the highest subsidence rates and also the site of the highest sedimentation rates. Such conditions help to explain the interstratification of lacustrine facies, representing sediment starved conditions interspersed with channel sandstone deposits, which

represent the distal portion of the clastic wedge. Amalgamated channel sandstone bodies above the thick dolerite sill at the top of the exposure at Clarkson Peak represent a more proximal portion of the prograding wedge. Elsewhere in the basin, at Wahl Glacier, a low stacking pattern suggests deposition near the basin axis.

A regional change in sandstone composition and a change in fluvial stacking pattern from low to high density occur between the Buckley and Fremouw Formations in the BGR. A change from transverse and longitudinal flow in the Buckley Formation to transverse flow away from the orogenic belt characterizes paleoslope orientation in the Fremouw Formation across the CTM. This indicates that the Transanatarctic basin was over-filled during deposition of the Fremouw Formation (Figure 64?). Thus, throughout Fremouw time, sedimentation rates outpaced subsidence rates. At Wahl Glacier flow directions change from SW in the upper Buckley to NW in the Fremouw Formation. Also, there is an abrupt change in sandstone composition from volcaniclastic in the upper Buckley Formation to quartz rich sandstone at the base of the Fremouw Formation. An abrupt change from a low fluvial stacking pattern with predominately floodplain deposits in the Buckley Formation to a high fluvial stacking pattern with thick amalgamated channel sand bodies with minor floodplain deposits that occur at the top of fining upward channel bodies suggests the presence of a disconformity at Wahl Glacier (Barrett et al., 1986; Flaig, 2005). The stacking pattern at Wahl Glacier suggests that sedimentation/avulsion rates were much higher than the creation of accommodation space/subsidence at this site. This stacking pattern changes back to a low density stacking pattern in the middle member of the Fremouw Formation at Wahl Glacier (cf. Barrett, 1986), suggesting, slightly higher accommodation/subsidence rates relative to

sedimentation at that time. Flaig (2005) measured an 87 m thick stratigraphic section in the lower member of the Fremouw Formation at Wyckoff Glacier, which is located ~ 10 km south of Wahl Glacier. Sandstone composition was not completed at Wyckoff glacier, but Flaig (2005) reported paleocurrent orientations towards 296 ˚ (NW). A high stacking pattern occurs at Wyckoff Glacier, similar to at Wahl Glacier consisting of three fining-upward sequences of predominately multistoried, trough cross-bedded channel-form sandstones, which grade upward into fine-grained facies (Figure 63; Flaig, 2005). This suggests a similar scenario at Wyckoff Glacier where sedimentation/avulsion rates were much higher than accommodation rates. In contrast, at Graphite Peak a single 6 m thick fining-upward channel sandstone deposit occurs at the base of the lower Fremouw Formation (Collinson et al., 2006), which is overlain by a thick succession of floodplain deposits until the base of the middle Fremouw Formation where a second thick channel sandstone deposit occurs (Figure 64). The contrasting fluvial stacking patterns between Wahl Glacier, Wyckoff Glacier, and Graphite Peak is likely due to Graphite Peak being located in a more proximal position to the orogenic basin margin than the other two sites. In this position, even though the basin was over-filled, subsidence rates would have been higher resulting in greater preservation of floodplain facies. Whereas, Wahl Glacier and Wyckoff Glacier were located in a distal position relative to drainage on the cratonic side of the lower Fremouw basin. There, sediment influx would have been higher than subsidence rates. Throughout the middle Fremouw, low-density stacking patterns occur across the basin. Although this is likely a time transgressive facies (Isbell, 2005), sedimentation rates appear to have decreased relative to accommodation rates. Such lower rates may have been controlled by tectonic activity or due to climate. An abrupt

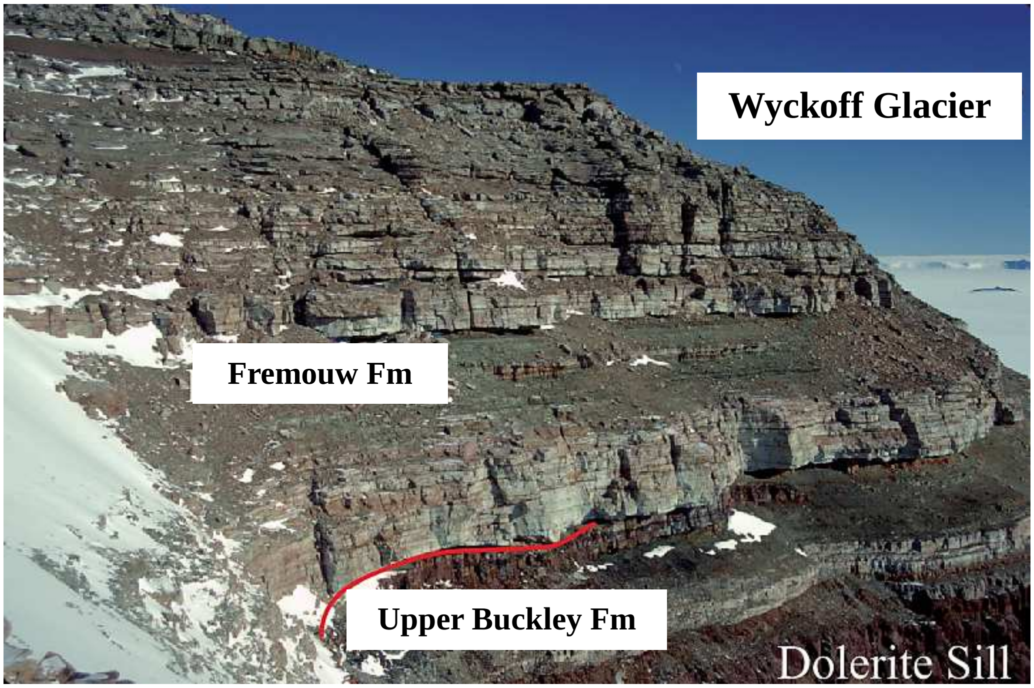



Figure 63. Photograph of the outcrop at Wyckoff Glacier showing the erosional contact between the upper Buckley Formation and the lower Fremouw Formation (Photograph courtesy of John Isbell).

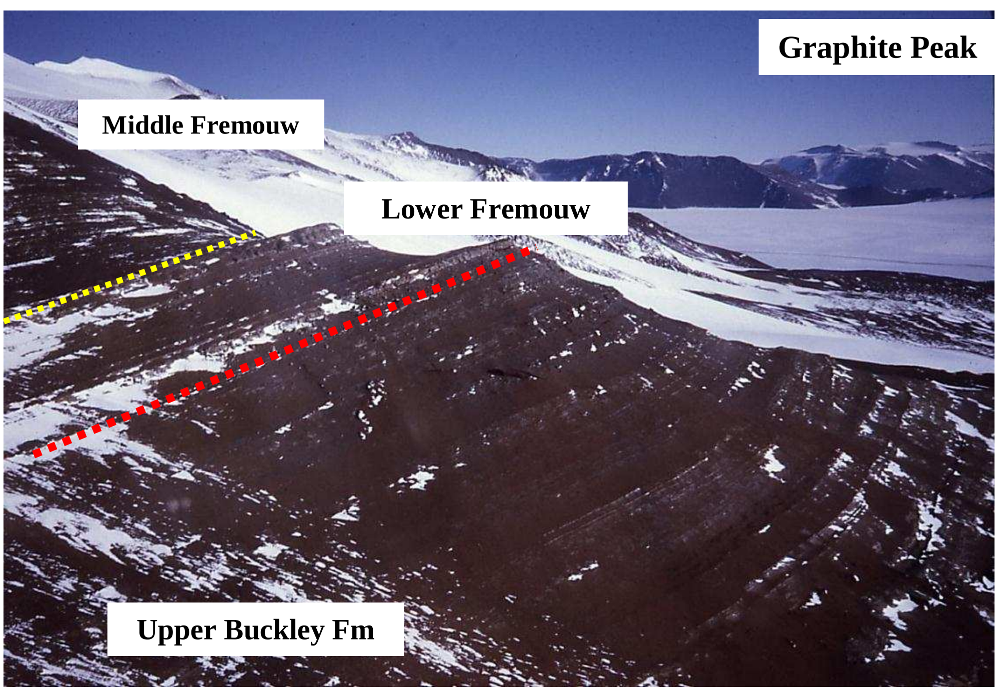



Figure 64. Photograph of the outcrop at Graphite Peak showing contrasting fluvial stacking patterns in the lower Fremouw Formation compared to Wyckoff Glacier

(Photograph from GSA data repository # 2006080 from Collinson et al., 2006). The sandstone bodie above the dotted red line is ~ 6 m thick.

change from a low to high fluvial stacking pattern occurs between the middle Fremouw Formation and upper Fremouw Formation. The middle Fremouw Formation consists of thin crevasse-splay deposits encased in thick floodplain deposits, which abruptly changes in the upper Fremouw Formation to thick fining upward channel sandstone bodies with thin floodplain deposits between channel sandstones.

## 3.4 CONTROLS ON FLUVIAL STACKING PATTERNS IN THE TRANSANTARCTIC BASIN

Changes in fluvial stacking patterns are controlled by changes in the rate of the creation of accommodation (e.g., subsidence, eustatic sea level) and changes in clastic influx rates (e.g., climate change, loss of vegetation). The previous hypothesis for the control of fluvial stacking pattern across the PTB has been accredited to the extinction of peat forming plants (cf. Retallack, 1995; Retallack et al., 1996, 2005; Ward et al., 2000; Michaelsen, 2002)). This section aims to discuss all possible controls on the stacking pattern (e.g., sea level, climate, plant extinction, and tectonism) and identify what is controlling the change in fluvial stacking patterns in the Transantarctic basin.

A change in sea level can cause changes in fluvial stacking patterns, but is not attributed to a change in fluvial stacking patterns in the Transantarctic basin. Changing sea level can be ruled out as a control of the fluvial stacking patterns in the BGR, as the nearest possible coastal plain to marine deposits are suggested to be in the Ohio Range (Bradshaw et al., 1984). In the BGR the latest possible marine influence is recorded by

the Mackellar Formation, which was deposited in the Early Permian. It has been debated whether the Mackellar Formation truly has a marine influence, however, recent studies by Hasiotis et al. (2012), Jackson et al. (2012), and Wildermuth et al. (2012) suggest that the Mackellar Formation in the BGR and Shackleton Glacier Region is a post-glacial marine environment with freshwater influence based on trace fossils. However, the Mackellar Formation is of Early Permian age (Miller and Isbell, 2010); whereas, the Buckley Formation is of middle to Late Permian and the Fremouw Formation is of Triassic age (Collinson et al., 1994; 2006).

A rapid change in climate can produce changes in fluvial stacking patterns by causing changes in precipitation and sediment yield (Langbein & Schumm, 1958; Schumm, 1977). A change in climate across the Permian-Triassic boundary in Antarctica has been noted by Collinson (1997), McLoughlin et al. (1997) and Retallack & Krull (1999). The climate is suggested to have changed from a cold glacial Early Permian to a cool humid climate in the Middle to Late Permian to a warmer more arid climate with seasonal moisture in the Triassic (Collinson et al., 1994; Collinson, 1997; Isbell et al., 2012). McLoughlin et al. (1997) noted the change from glossopterid-dominated vegetation in the Late Permian to peltasperm-, lycophyte-, and corystosperm-dominated flora was synchronous across Gondwana, whereas the demise of coal was diachronus across Gondwana from mid-Permian to mid-Triassic times suggesting climate change was gradual and was not a major factor contributing to changes of fluvial stacking patterns in the CTM. Characteristics suggesting a climate change from cool humid wet climate to a more arid climate would include: 1) an increase in sedimentation rates from the Late Permian to Early Triassic represented by an increase in channel sandstone

deposits showing a high fluvial stacking pattern; 2) sandstone compositions with an increase in percentage of feldspar and decrease in percentage of quartz; 3) a change from coal deposits to redbeds; and 4) consistent paleocurrent orientations. In the BGR only one of the four former characteristics is seen, which is an increase in channel sandstone deposits showing a change from a low to high fluvial stacking pattern from the Buckley to Fremouw formations. Although coal deposits are absent from the lower Fremouw Formation no redbeds were observed at these study locations. Furthermore, changes in sandstone composition from volcaniclastic in the upper Buckley Formation to quartz rich sandstones in the lower Fremouw Formation support evidence for changes in provenance, but don't show an increase in feldspar that you would expect to see in a change from a cool humid climate to a more arid climate. Lastly, a regional change in paleocurrent orientation from ESE in the lower Buckley Formation to N to NW in the upper Buckley and to WNW in the lower Fremouw Formations occurs, which is in contrast with what would be expected from climatic fluctuations. Overall, the changes in sandstone composition and paleocurrent orientations do not support climate change as being a major control on the fluvial stacking patterns.

An abrupt die-off of terrestrial vegetation causing a dramatic increase in erosion and therefore, sedimentation rates has been proposed by Retallack (1995) and Retallack et al. (1996) for the change in fluvial stacking patterns across the Permian–Triassic boundary in Antarctica, as well as in the Karoo Basin in southern Africa by Ward et al. (2000) and the Bowen Basin in Australia by Michaelsen (2002). The Permian–Triassic boundary is marked by a change from the Late Permian glossopterid-dominated floras, which were capable of living in a wide range of fluvial environments, including swamps

and channel margins to a Dicroidium-dominated fossil flora in the Triassic (Barrett et al., 1986; Veevers et al., 1994; Collinson & Hammer, 1996; Collinson et al., 2006). The glossopterid-dominated flora show a gradual taxonomic turnover during the Permian with notable increases in diversity of glossopterids, other gymnosperms, and herbaceous sphenophytes towards the end of the period (Retallack, 1980; McLoughlin, 1992). In lower paleolatitudes floristic changes appear to initiate earlier than in higher paleolatitudes (McLoughlin, 1997). A catastrophic loss of terrestrial vegetation would cause a synchronous change in fluvial stacking pattern from a low stacking pattern in the upper Buckley Formation to a high stacking pattern in the lower Fremouw Formation across the Transantarctic Basin and should have no effect on sandstone composition or paleocurrent orientations. Glossopteris flora, along with roots, leaves, and wood have been found in the Shackleton Glacier region in the Fremouw Formation, as well as Lower Triassic formations in Tasmania, South Africa, and India (McManus et al., 2002; Collinson et al., 2006). These findings suggest that the lower part of the Fremouw Formation in the Shackleton Glacier region maybe uppermost Permian (Collinson et al., 2006). This also implies that the Buckley-Fremouw formational contact is diachronus across the Transantarctic basin and that the change in fluvial stacking patterns is also time transgressive across the basin. The occurrence of the PTB with the formational boundary in the BGR is likely the result of basin fill pattern rather than plant extinction. Perhaps the change in landscape drainage and groundwater profile, due to a change in climate, affected the distribution and depth of rooting that may have had an effect on preservation bias and change in vegetation.

Evidence for active tectonism in the Transantarctic foreland basin includes variations in sandstone composition, paleocurrent orientation, and changes in fluvial architecture through space and time (Figure 62). Two different sandstone compositions occur in the Transantarctic basin suggesting two difference sources. The quartzo-feldspathic sandstones suggest a cratonic source, likely due to erosion off the forebuldge and the volcaniclastic sandstones are suggestive of an orogenic source. The occurrence of transverse and longitudinal paleocurrent orientations during the Middle to Late Permian suggests the foreland basin was under-filled. The presence of floodplain deposits including crevasse-splay, lacustrine, and coal deposits during the Buckley Formation also suggests an under-filled basin. The transition to uniform transverse flow off the orogenic belt towards the craton and lack of lacustrine and coal deposits during the Triassic Fremouw Formation suggests that the foreland basin became over-filled. The presence of disconformities at locations located on the cratonic margin between the lower and upper Buckley Formations and the upper Buckley and lower Fremouw Formations suggested by abrupt changes in sandstone composition, changes in paleocurrent orientations, and changes in lithofacies suggests the occurrence of two separate tectonic events.

Overall, data from this study corroborates the hypothesis from Isbell (1990, 2005), Isbell et al. (1997), and Flaig (2005) that the change in stacking patterns across the boundary is the result of tectonism (Table 4). No evidence was found to support the previous hypothesis for plant extinction as a control on the changes in fluvial stacking patterns in the BGR of the CTM.

Table 4. Summary of evidence for controls on fluvial stacking patterns

| **CTAM** | | | | |
|---|---|---|---|---|
| | Sea Level | Climate Change | Plant Extinction | Tectonism |
| Sandstone Composition | — | — | — | + |
| Paleocurrent Orientation | — | — | — | + |
| Fluvial Stacking Patterns | + | + | + | + |

## 4. Conclusions

The Transantarctic basin evolved during a time of great change where the climate was transitioning from the late Paleozoic ice age to the Late Permian-Mesozoic Hothouse, the EPME devastated marine and terrestrial life, and changing tectonic conditions were occurring along the Panthalassan margin of Antarctica and throughout Gondwana. Changes in fluvial stacking patterns from the Late Permian through the Triassic have been observed in the BGR of the CTM and throughout Gondwana. The previous hypothesis for the shift in fluvial style has been accredited to the extinction of peat forming plants. However, other controls of fluvial stacking patterns must be considered, such as, changes in sea level, climate change, and tectonism. This study reevaluated previous data and new data from well exposed sections of Permian and Triassic strata in order to determine what controlled the change in fluvial stacking patterns and how the fluvial stacking patterns changed across the basin and through time. This study found the following:

- Late Permian strata consist of three main facies associations including: medium- to coarse-grained sandstone sheet facies associations, fine- to medium-grained sandstone sheet facies associations, and fine-grained facies associations consisting of carbonaceous siltstone, mudrock, and coal deposits. The facies associations have been interpreted as channel sandstone bodies deposited by braided streams, crevasse-splays, and floodplain deposits consisting of lacustrine, and mire deposits respectively. The Late Permian strata are characterized as avulsion deposits based on the abrupt changes in grain size (coal and carbonaceous siltstone overlain by medium-grained sandstone), abrupt change in lithofacies

(coal overlain by channel deposits), and variations in paleocurrent directions between individual splay complexes.

- The Early Triassic strata consist of abundant medium- to very coarse-grained sandstone sheet facies association and moderate fine- to medium-grained facies and fine-grained facies consisting of predominately non-carbonaceous siltstone. The facies associations have been interpreted as channel sandstone bodies deposited by braided rivers, crevasse-splay deposits, and floodplain deposits respectively. The Fremouw strata also likely represent avulsion deposits based on abrupt changes in grain sizes (non-carbonaceous siltstone to coarse-grained sandstone), abrupt changes in lithofacies (floodplain deposits overlain by channel deposits), and variations in paleocurrent directions between splay complexes.
- The abrupt change from a uniform dipping paleoslope showing SSE flow in the Fairchild Formation to transverse and longitudinal flow in the Buckley Formation suggests that, at that time, the basin evolved from a uniformly southward dipping cratonic basin to a trough-shaped, under-filled foreland basin (Isbell, 1990, 1991; Isbell et al., 1997).
- Sandstone composition changes from quartzo-feldspathic sandstone in the Fairchild and lower member of the Buckley Formation to volcaniclastic sandstone in the upper Buckley Formation to quartz rich sandstone in the lower and upper member of the Fremouw Formation and volcaniclastic sandstone in the middle Fremouw Formation. These changes signal the occurrence of changing tectonic conditions and changing source terrains.

- Abrupt changes in paleocurrents, sandstone composition, and facies suggest major disconformities occur at locations proximal to the cratonic margin, such as Mt. Bowers and Mt. Achenar.
- Paleocurrent orientations for the Fremouw Formation indicate that it was deposited on a paleoslope that dipped uniformly away from an orogenic belt suggesting that the foreland basin transitioned from an under-filled to over-filled basin during the Early Triassic.
- A change in fluvial stacking pattern occurred from a low stacking pattern in the upper Buckley Formation to a high stacking pattern in the lower Fremouw Formation. Changing stacking patterns in time and space suggest development of differential subsidence in the Transantarctic basin. Locations proximal to the orogenic basin margin (i.e. Clarkson Peak), where the rate of sediment influx exceeds the rate of the creation of accommodation exhibit high stacking patterns. Similarly, locations proximal to the cratonic basin margin (i.e. Mt. Bowers), also exhibit high stacking patterns due to the rate of sediment influx exceeding the rate of the creation of accommodation. In contrast, the basin axis facies exhibit low stacking patterns due to the rate of creation of accommodation space exceeding the rate of sediment influx.
- Overall, re-examination of previous data and new data provided by this study corroborates with the hypothesis from Isbell (1990, 2005), Isbell et al. (1997), and Flaig (2005) for the main control on the fluvial stacking pattern to be tectonism. No evidence was found to support the previous hypothesis made by Retallack

(1995) and Retallack et al. (1996, 2005), which proposed plant extinction as the main cause for the change in fluvial stacking patterns across the PTB.

# References


Allen, J.R.L., 1964, Studies in fluviatile sedimentation: Cyclothems from the Lower Old Red Sandstone, Anglo-Welsh Basin: Sedimentology, v.3, p. 163-198.

Allen, J.R.L., 1965, A review of the origin and characteristics of recent alluvial sediments: Sedimentology, v. 5, p. 89-191.

Allen, J. R. L., 1978, Studies in fluviatile sedimentation: an exploratory quantitative model for the architecture of avulsion-controlled alluvial suites: Sedimentary Geology, v. 21, p. 129-147.

Allen, P.A., and Allen, J.R., 2005, Basin analysis: principles and applications. Malden, MA: Blackwell Publishing.

Aslan, A., Blum, M.D., 1999, Contrasting styles of Holocene avulsion, Texas Gulf Coastal Plain, USA: Special Publication of the International Association of Sedimentologists, v. 28, p. 195-209.

Aslan, A., White, W.A., Warne, A.G., and Guevara, E.H., 2003, Holocene evolution of the western Orinoco Delta, Venezuela: Geological Society of America, Bulletin, v. 115, p. 479-498.

Aslan, A., Autin, W.J., and Blum, M.D., 2005, Causes of river avulsion: insights from the late Holocene avulsion history of the Mississippi River, U.S.A.: Journal of Sedimentary Research, v. 75, p. 650-664.

Asselen, S., Stouthamer, E., and Smith, N.D., 2010, Factors controlling peat compaction in alluvial floodplains: A case study in the cold-temperate Cumberland Marshes, Canada: Journal of Sedimentary Research, v. 80, p. 155-166.

Barrett, P.J., 1968, The post glacial Permian and Triassic Beacon rocks in the Beardmore Glacier area, Central Transantarctic Mountains, Antarctica, [Ph.D. thesis]: Department of Geological Sciences, The Ohio State University, Columbus, OH, 510 p.

Barrett, P.J., 1972, Stratigraphy and petrology of the mainly fluviatile Permian and Triassic part of the Beacon Supergroup, Beardmore Glacier Area: International Union of Geological Sciences, series B1, p. 365-372.

Barrett, P.J., Elliot, D.H., and Lindsay, J.F., 1986, The Beacon Supergroup (Devonian-Triassic) and Ferrar Group (Jurassic) in the Beardmore Glacier Area, Antarctica: Geology of the Central Transantarctic Mountains Antarctic Research Series, v. 36, p. 339-428.

Barrett, P.J., 1991, The Devonian to Jurassic Beacon Supergroup of the Transantarctic Mountains and correlatives in other parts of Antarctica; in Tingey, R.J., ed., The Geology of Antarctica: Oxford University Press, Oxford, p. 120-152.

Berner, R.A., 2002, Examination of hypotheses for the Permo-Triassic boundary extinction by carbon cycle modeling: Proceedings of the National Academy of Sciences of the United States of America, v. 99, p. 4172-4177.

Blum, M.D., and Törnqvist, T.E., 2000, Fluvial responses to climate and sea-level change: a review and look forward: Sedimentology, v. 47, p. 2-48.

Blum, M.D., and Aslan, A., 2006, Signatures of climate vs. sea-level change within incised valley-fill successions; Quaternary examples for the Texas Gulf Coast: Sedimentary Geology, v.190, p. 177-211.

Blum, M.D., Martin, J., Milliken, K., and Garvin, M, 2013, Paleovalley systems: Insights from Quaternary analogs and experiments: Earth-Science Reviews, v. 116, p. 128-169.

Bowring, S.A., and Erwin, D.H., 1998, U/Pb zircon geochronology and tempo of the end-Permian mass extinction: Science, v. 280, p. 1039–1045.

Bradshaw, M.A., Newman, J., and Aitchison, J.C., 1984, Preliminary geological results of the 1983-84 Ohio Range Expedition: New Zealand Antarctic Record, v. 5, p. 1-17.

Bridge, J. S., and Leeder, M. R., 1979, A simulation model of alluvial stratigraphy: Sedimentology, v. 26, p. 617-644.

Bridge, J.S., and Diemer, J.A., 1983, Quantitative interpretation of an evolving ancient river system: Sedimentology, v. 30, p. 599-623.

Bridge, J.S., 1984, Large scale facies sequences in alluvial overbank environments: Journal of Sedimentary Petrology, v. 54, p. 583-588.

Bridge, J.S., 1985, Paleochannel patterns inferred from alluvial deposits; a critical evaluation: Journal of Sedimentary Petrology, v. 55, p. 579-589.

Bridge, J., S., and Mackey, S., D., 1993, A theoretical study of fluvial sandstone body dimensions: International Association of Sedimentologists Special Publications 15, p. 213-236

Bridge, J.S., and Demicco, R.V., 2008, Earth surface processes, landforms and sediment deposits: Cambridge University Press, Cambridge, United Kingdom, 815 p.

Bristow, C.S., 1987, Brahmaputra River: channel migration and deposition, in Ethridge, F.G., Flores, R.M., and Harvey, M.D., eds., Recent Developments in Fluvial Sedimentology: Society of Economic Paleontologists and Mineralogists, Special Publication 39, p. 63-74.

Bristow, C.S., Skelly, R.L., and Ethridge, F.G., 1999, Crevasse splays from the rapidly aggrading, sand-bed, braided Niobrara River, Nebraska: effect of base-level rise: Sedimentology, v. 46, p. 1029-1047.

Bryant, M., Falk, P., Paola, C., 1995, Experimental study of avulsion frequency and rate of deposition: Geology, v. 23, p. 365-368.

Catuneanu, O., and Elango, H.N, 2001, Tectonic control of fluvial styles; the Balfour Formation of the Karoo Basin, South Africa: Sedimentary Geology, v. 140, p. 291-313.

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

Catuneanu, O., 2006, Sequence stratigraphy of clastic systems: concepts, merits, and pitfalls: Journal of African Earth Sciences, v. 35, p. 1-43.

Cazanacli, D., and Smith, N.D., 1998, A study of morphology and texture of natural levees—Cumberland Marshes, Saskatchewan, Canada: Geomorphology, v. 25, p. 43-55.

Cecil, B.C., and Dulong, F.T., 2003, Precipitation models for sediment supply in warm climates: Special Publication Society for Sedimentary Geology, v. 77, p. 5-8.

Cecil, B.C., 2003, The concept of autocyclic and allocyclic controls on sedimentation and stratigraphy, emphasizing the climatic variable: Special Publication Society for Sedimentary Geology, v. 77, p. 13-20.

Church, K.D., and Coe, A.L., 2003, Processes controlling relative sea-level change and sediment supply. The Sedimentary Record of Sea-Level Change: Cambridge University Press, Cambridge, p. 99-117.

Clevis, Q., De Boer, P.L., and Nijman, W., 2004, Differentiating the effect of episodic tectonism and eustatic sea-level fluctuations in foreland basins filled by alluvial fans and axial deltaic systems: insights from a three-dimensional stratigraphic forward model: Sedimentology, v. 52, p. 809-835.

Coates, D.A., Stricker, G.D., and Landis, E.R., 1990, Coal geology, coal quality, and coal resources in Permian Rocks of the Beacon Supergroup, Transantarctic Mountains, Antarctica: Mineral Resources Potential of Antarctica Research Series, v. 51, p. 133-162.

Collinson, J.W., 1970, Bedforms of the Tana River, Norway: Geografiska Annaler, v. 52A, p. 31-56.

Collinson, J.W., and Isbell, J.L., 1986, Permian-Triassic sedimentology of the Beardmore Glacier region: Antarctic Journal of the United States, v. 21, p. 29-30.

Collinson, J.W., 1990, Depositional setting of Late Carboniferous to Triassic *biota* in the Transantarctic Basin, in Taylor, T.N., and Taylor, E.L., eds., Antarctic Paleobiology; its role in the reconstruction of Gondwana, New York, NY, Springer-Verlag, p. 13-15.

Collinson, J.W., Isbell, J.L., Elliott, D.H., Miller, M.F., and Miller, J.M.G, 1994, Permian-Triassic Transantarctic Basin, in Veevers, J.J., and Powell, C.M., eds., Permian-Triassic Pangean basins and fold belts along the Panthalassan Margin of Gondwanaland: Boulder, CO, Geological Society of America Memoir, v. 184, p. 173-222.

Collinson, J.W., and Hammer, W.R., 1996, New observations on the Triassic stratigraphy of the Shackleton Glacier region: Antarctic Journal of the United States, v. 31, p. 9-11.

Collinson, J.W., 1997, Paleoclimate of Permo-Triassic Antarctica: The Antarctic Region: Geological Evolution and Processes, p. 1029-1034.

Collinson, J.W., Hammer, W.R., Askin, R.A., and Elliot, D.H., 2006, Permian-Triassic boundary in the Central Transantarctic Mountains, Antarctica: GSA Bulletin, v. 118, p. 747-763.

Davies-Vollum, K.S., and Smith, N.D., 2008, Factors affecting the accumulation of organic-rich deposits in a modern avulsive floodplain: examples from the Cumberland Marshes, Saskatchewan, Canada: Journal of Sedimentary Research, v. 78, p. 683-692.

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

Domeier, M., Van der Voo, R., Tohver, E., Tomezzoli, R. N., Vizan, H., Torsvik, T. H., and Kirshner, J., 2011, New Late Permian paleomagnetic data from Argentina: Refinement of the apparent polar wander path of Gondwana: Geochemistry, Geophysics, Geosystems, v. 12, p. 1-21.

Einsele, G., 1992, Sedimentary basins; evolution, facies, and sediment budget. Berlin, United Kingdom: Springer-Verlag.

Erwin, D.H., 1994, The Permo-Triassic extinctions: Nature, v. 367, p. 231-236.

Fielding, C.R., Sliwa, R., Holcombe, R.J., and Jones, A.T., 2001, A new palaeogeographic synthesis for the Bowen, Gunnedah and Sydney Basins of Eastern Australia: Petroleum Exploration Society of Australia Special Publication, Eastern Australasian Basins Symposium, Melbourne, Vic, November 25-28, p. 269-278.

Flaig, P.P., 2005. Changing fluvial style across the Permian-Triassic boundary: Beardmore Glacier region, Central Transantarctic Mountains, Antarctica [M.S. thesis]: Department of Geological Sciences, The University of Wisconsin-Milwaukee, Milwaukee, WI, 288p.

Flemings, P.B., and Jordan, T.E., 1990, Stratigraphic modeling of foreland basins; interpreting thrust deformation and lithosphere rheology: Geology, v. 18, p. 430-434.

Friend, P.F., Slater, M.J., and Williams, R.C., 1979, Vertical and lateral building of river sandstone bodies, Ebro Basin, Spain: Journal of the Geological Society of London, v. 136, p. 39-46.

Friend, P.F., 1983, Towards the field classification of alluvial architecture or sequence, in Collinson, J.D., and Lewin, J., Modern and ancient fluvial systems: International Association of Sedimentologists Special Publication 6, p. 345-354.

Frostick, L.E., and Steel, R.J., 1993, Tectonic signatures in sedimentary basin fills: an overview: International Association of Sedimentologists Special Publication 20, p. 1-9.

Gastaldo, R.A., Adendorff, R., Bamford, M., Labandeira, C.C., Neveling, J., and Sims, H., 2005, Taphonomic trends of macroforal assemblages across the Permian-Triassic boundary, Karoo Basin, South Africa: Palaios, v. 20, p. 479-497.

Gulbranson, E.L., Isbell, J.L., Taylor, E.L., Ryberg, P.E., Taylor, T.N., and Flaig, P.P., 2012, Permian polar forests: deciduousness and environmental variation: Geobiology, v. 10, p. 479-495.

Hasiotis, S.T., Flaig, P.P., Jackson, A., and Isbell, J.L., 2012, First report of marine trace fossils from fluvio-deltaic-shallow marine deposits in the Lower Permian Mackellar Formation, Beardmore Glacier area, central Transantarctic Mountains, Antarctica; salinity stressed aquatic conditions: International Geological Congress, Abstracts, p. 4018.

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

Holmes, W.B.K., 1992, *Glossopteris*-like leaves from the Triassic of eastern Australia, in Venkatachala, B.S., Jain, K.P., Awasthi, N., eds., Geophytology, v. 22, p. 119-125.

Isbell, J.L., 1990. Fluvial sedimentology and basin analyses of the Permian Fairchild and Buckley formations, Beardmore Glacier region, and the Weller Coal Measures, Southern Victoria Land, Antarctica [Ph.D. thesis]: Department of Geological Sciences, The Ohio State University, Columbus, OH, 347 p.

Isbell, J.L., 1991, Evidence for a low-gradient alluvial fan from the palaeo-Pacific margin in the Upper Permian Buckley Formation, Beardmore Glacier region, Antarctica, in Thomson, M.R.A., Crame, J.A., and Thomson, J.W., eds., Geological evolution of Antarctica; proceedings of the Fifth international symposium on Antarctic earth sciences, Cambridge, Cambridge University Press, p. 215-217.

Isbell, J.L., and Collinson, J.W., 1991, Sedimentological significance of fluvial cycles in the Permian of the Central Transantarctic Mountains, Antarctica: Seventh International Gondwana Symposium, Sao Paulo, Brazil, p. 189-199.

Isbell, J.L., and MacDonald, D.I.M., 1991, Alluvial stratigraphic sequences within the Permian Transantarctic foreland basin, Beardmore Glacier region, Antarctica: Antarctic Journal of the United States, v. 26, p. 15-16.

Isbell, J.L., and MacDonald, D.I.M., 1991, Lithofacies analysis of the Triassic Fremouw Formation at the Gordon Valley vertebrate site, Antarctica: Antarctic Journal of the Unite States, v. 26, p. 15-16.

Isbell, J.L., and Askin, R.A., 1999, Search for evidence of impact at the Permian-Triassic boundary in Antarctica and Australia; discussion and reply: Geology, v. 27, p. 859-860.

Isbell, J.L., Seegers, G.M., and Gelhar, G.A., 1997, Upper Paleozoic glacial and postglacial deposits, Central Transantarctic Mountains, Antarctica, in Martini, I.P., ed., Late Glacial and Postglacial Environmental Changes: Quaternary, Carboniferous-Permian, and Proterozoic: Oxford, Oxford University Press, p. 230-242.

Isbell, J.L., 1999, The Kukri Erosion Surface; a reassessment of its relationship to rocks of the Beacon Supergroup in the central Transantarctic Mountains, Antarctica: Antarctic Science, v. 11, p. 228-238.

Isbell, J.L., Flaig, P.P., Miller, M..F., Sidor, C.A., 2005, Fluvial architecture of Permian and Triassic strata in the Beardmore Glacier region, Antarctica: was plant extinction or tectonics the cause of fluvial changes across the P-T boundary? In: Pankhurst, R.J. and Veiga, G.C. (eds) Gondwana 12: Geological and Biological

Heritage of Gondwana, Abstracts, Academia Nacional de Ciencias, Cordoba, Argentina, p. 209.

Isbell, J.L., Koch, Z.J., Szablewski, G.M., and Lenaker, P.A., 2008, Permian glacigenic deposits in the Transantarctic Mountains, Antarctica, in Fielding, C.R., Frank, T.D., and Isbell, J.L., eds., Resolving the Late Paleozoic Ice Age in Time and Space: The Geological Society of America Special Paper 441, p. 59-70.

Isbell, J.L., 2010, Environmental and paleogeographic implications of glaciotectonic deformation of glaciomarine deposits within Permian strata of the Metschel Tillite, southern Victoria Land, Antarctica: Geological Society of America Special Paper 468, p. 81-100.

Isbell, J.L., Henry, L.C., Gulbranson, E.L., Limarino, C.O., Fraiser, M.L., Koch, Z.J., Ciccioli, P.L., and Dineen, A.A., 2012, Glacial paradoxes during the late Paleozoic ice age: Evaluating the equilibrium line altitude as a control on glaciation: Gondwana Research, v. 22, p. 1-19.

Jacob, K., 1952, A brief summary of the stratigraphy and palaeontology of the Gondwana system: 19th International Geology Congress: Symposium sur les Series de Gondwana, Algiers, p. 153-174.

Jackson, A.M., Hasiotis, S.T., Flaig, P.P., and Isbell, J.L., 2012, Preliminary report of trace fossils and sedimentology indicating a shallow marine deltaic environment for the Lower Permian Mackellar Formation at Turnabout Ridge and Buckley Island, Beardmore Glacier, Central Transantarctic Mountains, Antarctica: Geological Society of America Abstract, v. 44, p. 289.

Jerolmack, D.J., and Paola, C., 2007, Complexity in a cellular model of river avulsion: Geomorphology, v. 91, p. 259-270.

Jervey, M.T., 1988, Quantitative geological modeling of siliciclastic rock sequences and their seismic expression. Sea-Level Changes- An Integrated Approach: SEPM Special Publication 42, p. 74-79.

Jordan, T.E., 1995, Retroarc foreland and related basins, in Tectonics of sedimentary basins, eds. C. J. Busby and R. V. Ingersoll Cambridge, Massachusetts: Blackwell Science, p. 331-362.

Langbein, W.B., and Schumm, S.A., 1958, Yield of sediment in relation to mean annual precipitation: American Geophysical Union, v. 39, p. 1076-1084.

Leeder, M.R., 1978, A quantitative stratigraphic model for alluvium, with special reference to channel deposit density and interconnectedness, in Miall, A.D., ed., Fluvial Sedimentology: Canadian Society of Petroleum Geologists, Memoir 5, p. 587-596.

Leeder, M.R., 1999, Tectonic subsidence and deposition. Sedimentology and Sedimentary Basins from Turbulence to Tectonics: Wiley-Blackwell Publications, p. 497-506.

Lopez-Gamundi, O.R., 1997, Glacial-postglacial transition in the late Paleozoic basins of southern South America, in Martini, I.P., ed., Late Glacial and postglacial environmental changes: Quaternary, Carboniferous-Permian, and Proterozoic. Oxford University Press, Oxford, UK, p. 147-168.

Mackey, S.D., and Bridge, J.S., 1995, Three-dimensional model of alluvial stratigraphy: theory and application: Journal of Sedimentary Research, v. B65, p. 7-31.

Makaske, B., Smith, D.G., and Berendsen, H.J.A., 2002, Avulsions, channel evolution and floodplain sedimentation rates of the anastomosing upper Columbia River, British Columbia, Canada: Sedimentology, v. 49, p. 1049-1071.

McCabe, P.J., 1984, Depositional environments of coal and coal-bearing strata, in Rahmani, R.A., and Flores, R.M., eds., Sedimentology of Coal and Coal-Bearing sequences: Oxford, Blackwell Scientific Publications, International Association of Sedimentologist Special Publication 7, p. 13-42.

McLoughlin, S., 1992, Glossopterid megafossils in Permian Gondwanic non-marine biostratigraphy, in Findlay, R.H., Unrug, R., Banks, M.R., and Veevers, J.J., eds., Gondwana Eight: assembly, evolution and dispersal. Rotterdam: A.A. Balkema, p. 253-264.

McLoughlin, S., Lindstrom, S., and Drinan, A.N., 1997, Gondwanan floristic and sedimentological trends during the Permian-Triassic transition: new evidence from the Amery Group, northern Prince Charles Mountains, East Antarctica: Antarctic Science, v. 9, p. 281-298.

McManus, H.A., Taylor, E.L., Taylor, T.N., and Collinson, J. W., 2002, A petrified *Glossopteris* flora from Collinson Ridge, central Transantarctic Mountains: Late Permian or Early Triassic?: Review of Palaeobotany and Palynology, v. 120, p. 233-246.

Miall, A.D., 1983, Basin analysis of fluvial systems, in Collinson, J.D., and Lewin, J., Modern and ancient fluvial systems: International Association of Sedimentologists Special Publication 6, p. 345-354.

Miall, A.D., 1984, Principles of Sedimentary Basin Analysis: New York, NY, Springer-Verlag, 490 p.

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

Miall, A.D., 1996, The Geology of Fluvial Deposits: Springer-Verlag Berlin Heidelberg, 582 p.

Miall, A.D., 1997, The Geology of Stratigraphic Sequences: Springer-Verlag Berlin Heidelberg, 433 p.

Miall, A.D., 2003, Fluvial architecture of the Hawkesbury sandstone (Triassic), near Sydney, Australia: Journal of Sedimentary Research, v. 73, p. 531-545.

Michaelsen, P., 2002, Mass extinction of peat-forming plants and the effect on fluvial styles across the Permian-Triassic boundary, northern Bowen Basin, Australia: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 179, p. 173-188.

Miller, M.F., and Collinson, J.W., 1994, Trace fossils from Permian and Triassic sandy braided stream deposits, central Transantarctic Mountains: Palaios, v. 9, p. 605-610.

Miller, M.F., and Isbell, J.L., 2010, Reconstruction of a high-latitude, postglacial lake: Mackellar Formation (Permian), Transantarctic Mountains: Geological Society of America Special Paper 468, p. 193-207.

Mohrig, D., Heller, P.L., Paola, C., and Lyons, W.J., 2000, Interpreting avulsion process from ancient alluvial sequences: Guadalope-Matarranya system (northern Spain) and Wasatch Formation (western Colorado): Geological Society of America, Bulletin, v. 112, p. 1787-1803.

Morozova, G.S., and Smith, N.D., 1999, Holocene avulsion history of the lower Saskatchewan fluvial system, Cumberland Marshes, Saskatchewan-Manitoba, Canada, in Smith, N.D., and Rogers, J., eds., Fluvial Sedimentology VI: International Association of Sedimentologists, Special Publication 28, p. 231-249.

Morozova, G.S., and Smith, N.D., 2000, Holocene avulsion styles and sedimentation patterns of the Saskatchewan River, Cumberland Marshes, Canada: Sedimentology Geology, v. 130, p. 81-105.

Morozova, G.S., and Smith, N.D., 2003, Organic matter deposition in the Saskatchewan River floodplain (Cumberland Marshes, Canada): effect of progradational avulsions: Sedimentary Geology, v. 157, p. 15-29.

Mundil, R., Ludwig, K.R., Metcalfe, I., and Renne, P.R., 2004, Age and timing of the Permian mass extinctions: U/Pb dating of closed-system zircons: Science, v. 305, p. 1760-1763.

Newell, N.D., 1973, The very last moment of the Paleozoic Era, in Logan, A., and Hills, L.V., eds., The Permian and Triassic systems and their mutual boundary: Calgary, Canadian Society of Petroleum Geologists Memoir 2, p. 1-10.

Newell, A.J., Tverdokhlebov, V.P., and Benton, M.J., 1999, Interplay of tectonics and climate on a transverse fluvial system, Upper Permian, southern Uralian foreland basin, Russia: Sedimentary Geology, v. 127, p. 11-29.

Nicohls, G.J., and Fischer, J.A., 2007, Processes, facies and architecture of fluvial distributary system deposits: Sedimentary Geology, v. 195, p. 75-90.

Pant, D.D., and Pant, R., 1987, Some *Glossopteris* leaves from Indian Triassic beds: Palaeontographica, v. 205B, p. 165-178.

Perez-Arlucea, M., and Smith, N.D., 1999, Depositional patterns following the 1870s avulsion of the Saskatchewan River (Cumberland Marshes, Saskatchewan, Canada): Journal of Sedimentary Research, v. 69, p. 62-73.

Puigdefabregas, C., 1973, Miocene point-bar deposits in the Ebro Basin, northern Spain: Sedimentology, v. 20, p. 133-144.

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

Reichow, M.K., Saunders, A.D., White, R.V., Pringle, M.S., and Al'Mukhamedov, A.I., Medvedev, A.I., and Kirda, N.P., 2002, 40Ar/39Ar dates from the West Siberian Basin: Siberian Flood Basalt Province doubled: Science, v. 296, p. 1846-1849.

Reineck, H.E., and Singh, I.B., 1980. Depositional Sedimentary Environments with Reference to Terrigenous Clastics: Springer-Verlag Berlin Heidelberg, 549 p.

Retallack, G.J., 1980, Late Carboniferous to Middle Triassic megafossil floras from the Sydney Basin, in Herbert, C., and Helby, R., eds., A guide to the Sydney Basin: Geological Survey of New South Wales Bulletin, v. 26, p. 384-430.

Retallack, G.J., 1995, Permian-Triassic life crisis on land: Science, v. 267, p. 77-80.

Retallack G.J., Veevers, J.J., and Morante, R., 1996, Global coal gap between Permian-Triassic extinction and Middle Triassic recovery of peat-forming plants: Geological Society of America Bulletin, v. 108, p. 195-207.

Retallack G.J., Seyedolali, A., Krull, E.S., Holser, W.T., Ambers, C.P., and Kyte, F.T., 1998, Search for evidence of impact at the Permian-Triassic boundary in Antarctica and Australia: Geology, v. 26, p. 979-982.

Retallack, G.J., 1999, Search for evidence of impact at the Permian-Triassic boundary in Antarctica and Australia; discussion and reply: Geology, v. 27, p. 859-860.

Retallack, G.J., and Krull, E.S., 1999, Landscape ecological shift at the Permian-Triassic boundary in Antarctica: Australian Journal of Earth Sciences, v. 46, p. 786-812.

Retallack, G.J., Jahren, A.H., Sheldon, N.D., Chakrabarti, R., Metzger, C.A., and Smith, R.M.H., 2005, The Permian-Triassic boundary in Antarctica: Antarctic Science, v. 17, p. 241-258.

Retallack, G.J., and Krull, E.S., 2006, Carbon isotopic evidence for terminal-Permian methane outbursts and their role in extinctions of animals, plants, coral reefs, and peat swamps: Geological Society of America Special Paper 399, p. 249-268.

Schumm, S.A., 1977. The Fluvial System: New York, NY, John Wiley & Sons, 338 p.

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, U.S.A., in Flint, S., and Bryant, I., eds., Quantitative modeling of clastic hydrocarbon reservoirs and outcrop analogues: International Association of Sedimentologists Special Publication 15, p. 21-55.

Shanley, K.W., and McCabe, P.J., 1994, Perspectives on the sequence stratigraphy of continental strata: The American Association of Petroleum Geologists Bulletin, v. 78, p. 544-568.

Smith, D.G., and Smith, N.D., 1980, Sedimentation in anastomosed river systems: Examples from alluvial valleys near Banff, Alberta: Journal of Sedimentary Petrology, v. 50, p. 157-164.

Smith, D.G., 1982, Anastomosed fluvial deposits: modern examples from Western Canada: International Association of Sedimentologists Special Publication 6, p. 155-168.

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

Smith, N.D., and Perez-Arlucea, M., 1994, Fine-grained splay deposition in the avulsion belt of the lower Saskatchewan River, Canada: Journal of Sedimentary Research, v. B64, p. 159-168.

Smith, N.D., Slingerland, R.L., Perez-Arlucea, M., and Morozova, G.S., 1998, The 1870s avulsion of the Saskatchewan River: Canadian Journal of Earth Science, v. 35, p. 453-466.

Smith, N.D., and Perez-Arlucea, M., 2004, Effects of peat on the shapes of alluvial channels: Examples from the Cumberland Marshes, Saskatchewan, Canada: Geomorphology, v. 61, p. 323-335.

Smith, N.D., and Perez-Arlucea, M., 2008, Natural levee deposition during the 2005 flood of the Saskatchewan River: Geomorphology, v. 101, p. 583-594.

Stouthamer, E., 2001, Sedimentary products of avulsions in the Holocene Rhine-Meuse Delta, The Netherlands: Sedimentary Geology, v. 145, p. 73-92.

Stouthamer, E., and Berendsen, H.J.A., 2007, Avulsion: The relative roles of autogenic and allogenic processes: Sedimentary Geology, v. 198, p. 309-325.

Stow, D.A.V., 2005, Sedimentary Rocks in the Field a Color Guide: Manson Publishing Ltd., 320 p.

Taylor, E.L., Taylor, T.N., and Cuneo, R.N., 2000, Permian and Triassic high latitude paleoclimates: evidence from fossil biotas, in Huber, B.T., MacLeod, K.G., Wing, S.L., eds., Warm climates in earth history, Cambridge University Press, Cambridge, p. 321-350.

Thomas, H.H., 1952, A *Glossopteris* with whorled leaves: Birbal Sahni Memorial Volume, Palaeobotanist, v. 1, p. 435-438.

Thomas, H.H., 1958, *Lidgettonia,* a new type of fertile *Glossopteris*, Bull Br. Mus, (Nat. Hist.) Geology, v. 3, p. 179-189.

Törnqvist, T.E., 1994, Middle and late Holocene avulsion history of the River Rhine: Geology, v. 22, p. 711-714.

Törnqvist, T.E., and Bridge, J.S., 2002, Spatial variation of overbank aggradation rate and its influence on avulsion frequency: Sedimentology, v. 49, p. 891-905.

Vavra, C.L., Stanley, K.O., and Collinson, J.W., 1981, Provenance and alteration of Triassic Fremouw Formation, Central Transantarctic Mountains, in Creswell, M.M., and Vella, P., eds., Gondwana Five: Selected papers and abstracts of papers presented at the Fifth International Gondwana Symposium, Wellington, 1980: Rotterdam, A.A. Balkema, p. 149-153

Vavra, C.L., 1984, Provenance and alteration of the Triassic Fremouw and Falla Formations, Central Transantarctic Mountains, Antarctica: Report- Institute of Polar Studies, v. 87, 98 p.

Veevers, J.J., Powell, C.M., Collinson, J.W., and Lopez-Gamundi, O.R., 1994, Synthesis, in Veevers, J.J., and Powell, C.M., eds., Permian Triassic Pangean basins and

foldbelts along the Panthalassan Margin of Gondwanaland: Boulder, CO, Geological Society of America Memoir, v. 184, p. 331-353.

Visscher, H., Looy, C.V., Collinson, M.E., Brinkuis, H., Van Konijnenburg-van Cittert, J.H.A., Kurschner, W.M., and Sephton, M.A., 2004, Environmental mutagenesis during the end-Permian ecological crisis: PNAS, v. 101, p. 12952-12956.

Ward, P.D., Montgomery, D.R., and Smith, R., 2000, Altered river morphology in South Africa related to the Permian-Triassic extinction: Science, v. 289, p. 1740-1743.

Waterhouse, J.B., and Shi, G.R., 2010, Evolution in a cold climate: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 298, p. 17-30.

Wignall, P.B., and Hallam, A., 1992, Anoxia as a cause of the Permian/Triassic mass extinction: facies evidence from northern Italy and the western United States: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 93, p. 21-46.

Wignall, P.B., and Twitchett, R.J., 2002, Extent, duration, and nature of the Permian-Triassic super anoxic event: Special Papers-Geological Society of America, p. 395-414.

Wildermuth, S.C., Hasiotis, S.T., Isbell, J.L., and Miller, M.F., 2012, A diverse high-latitude ichnofauna from the Permian Pagoda and Mackellar Formations (Beacon Supergroup), Shackleton Glacier area, Central Transantarctic Mountains, Antarctica: Geological Society of America Abstracts, v.44, p. 90.

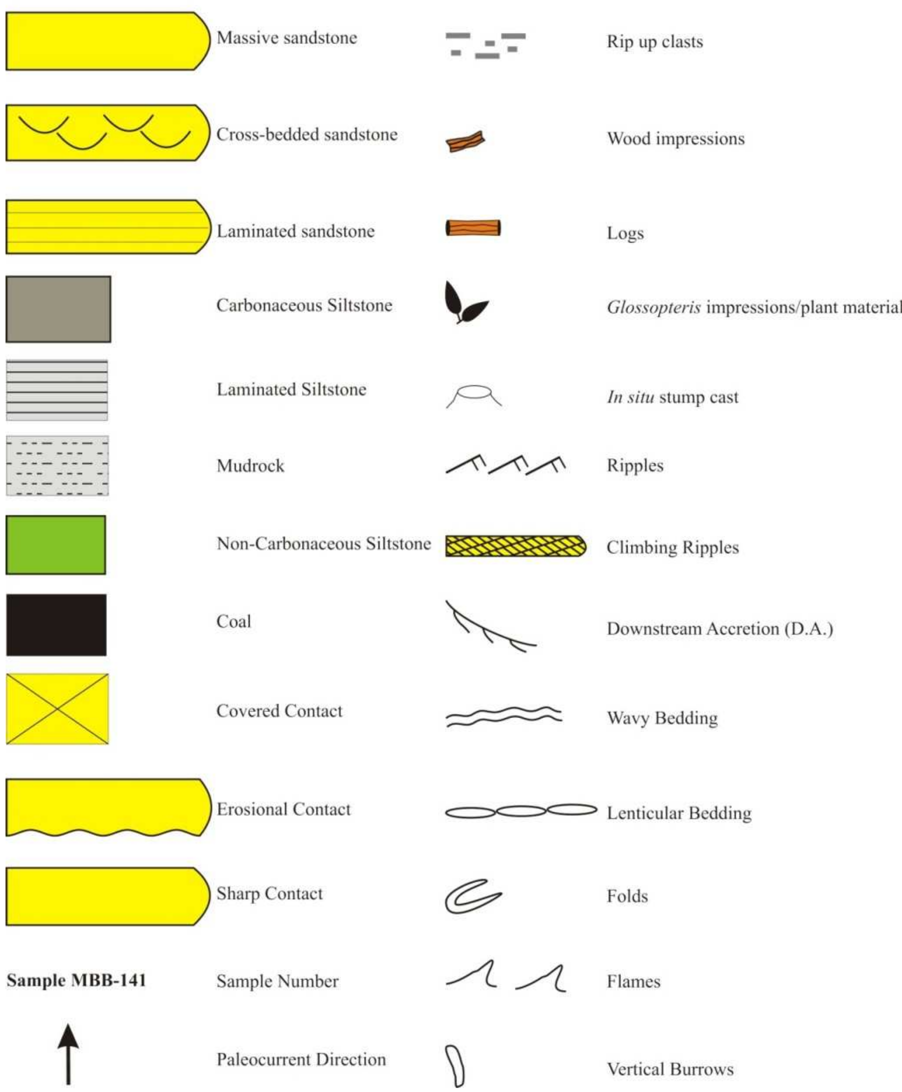
Williams, P.F., and Rust, B.R., 1969, The sedimentology of a braided river: Journal of Sedimentary Petrology, v. 39, p. 649-679.

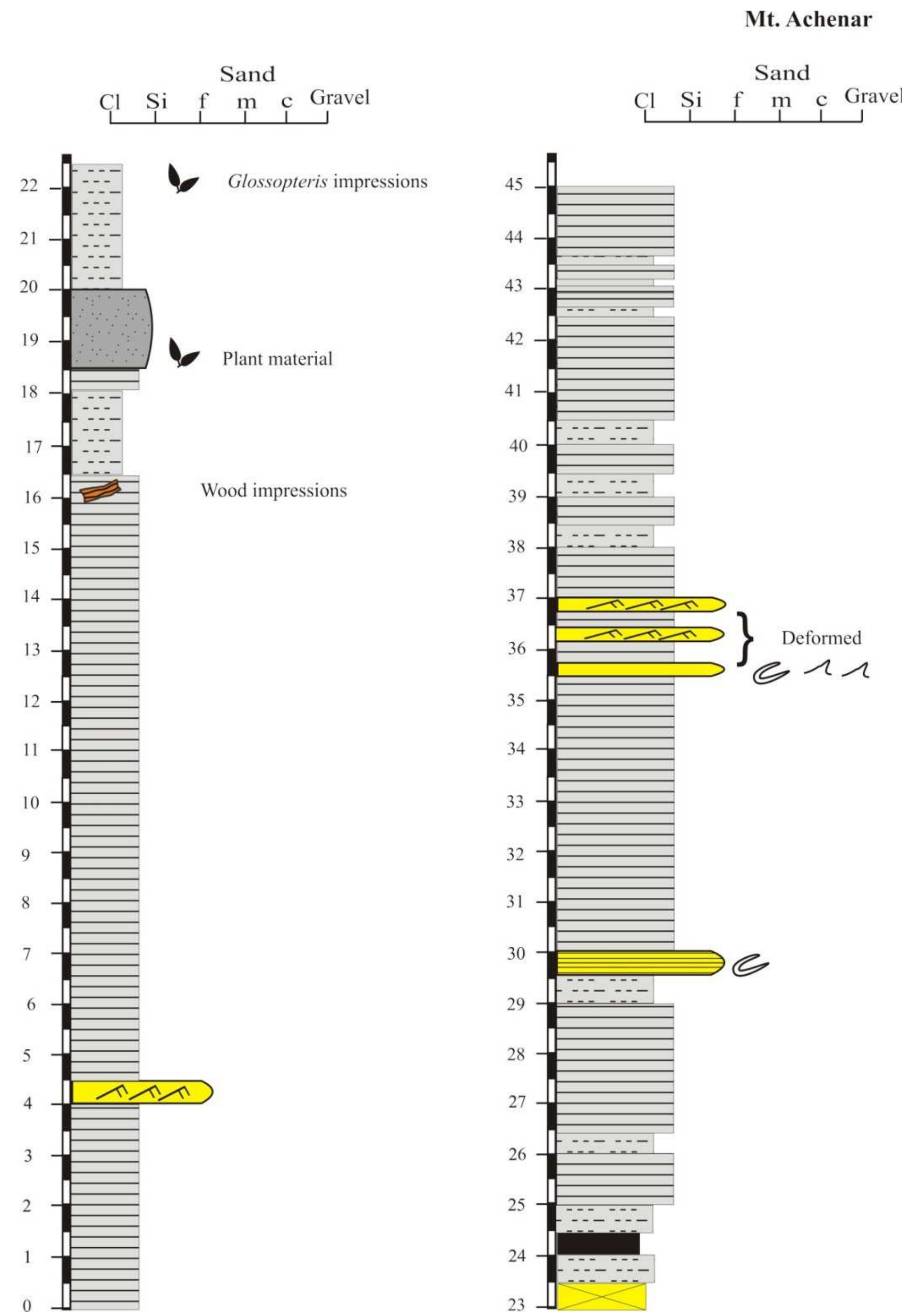
Appendix A:  Stratigraphic Sections:  Mt. Achenar, Mt. Bowers, Lewis Cliffs, Wahl Glacier, and Gordon Valley

Table 5.  GPS Coordinates of field locations in the CTM

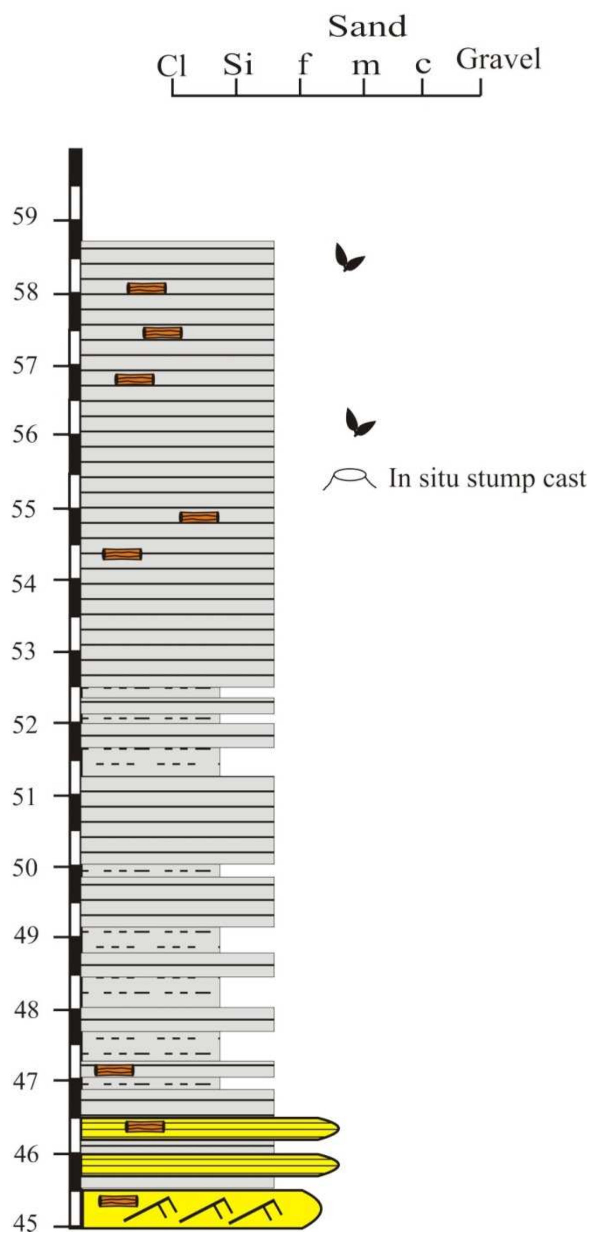
| GPS Coordinates | | |
|---|---|---|
| Location | S | E |
| Mt. Achenar (above the sill) | 84□ 11.353’ | 160□ 57.849’ |
| Mt. Achenar (below the sill) | 84□ 11.789’ | 160□ 58.535’ |
| Mt. Bowers | 85°00.122’ | 164°09.606’ |
| Lewis Cliffs | 84°15.620’ | 161°11.615’ |
| Wahl Glacier | 84□ 5.028’ | 165□ 58.535’ |
| Gordon Valley | 84□ 21.196’ | 163□50.929’ |

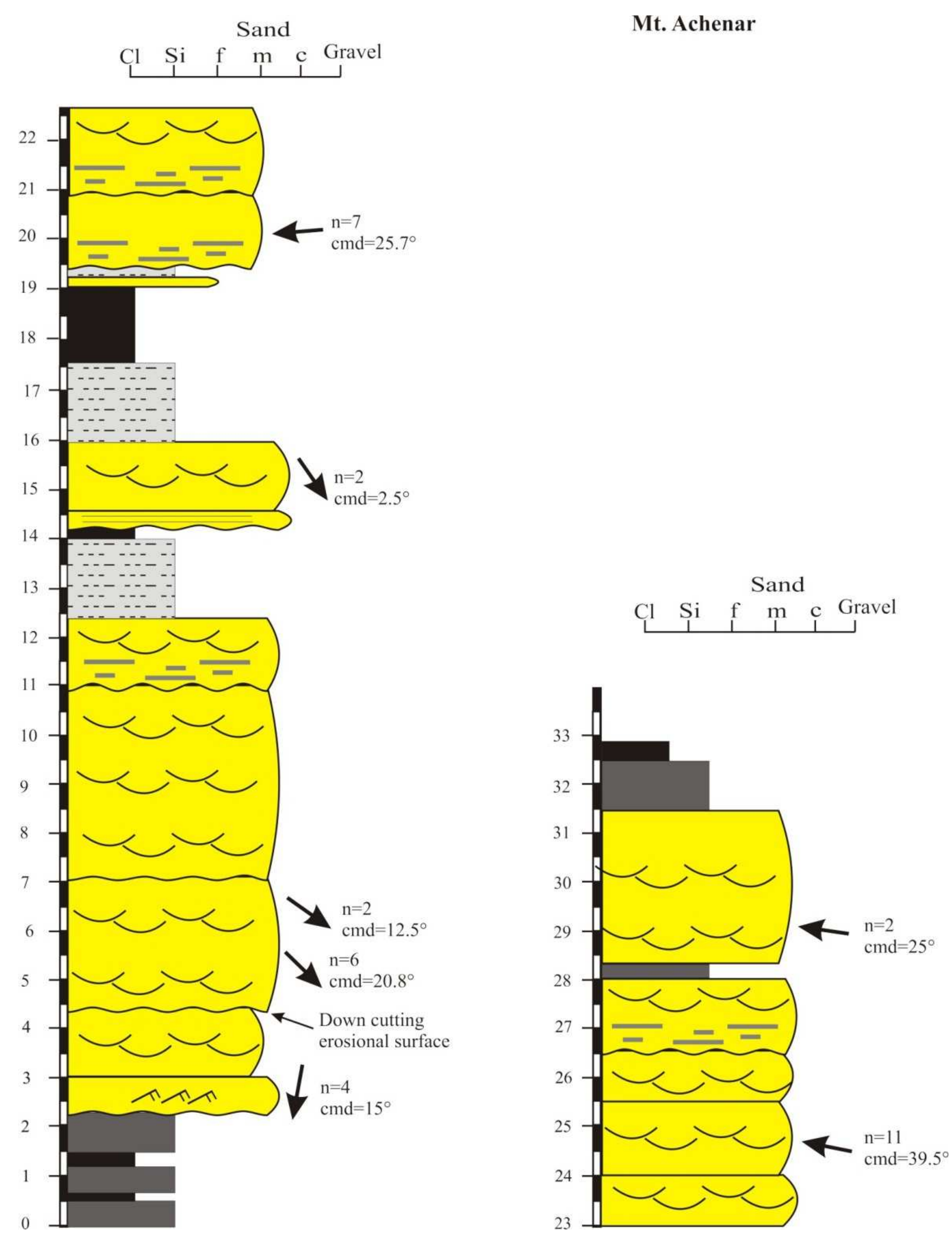
**Key**

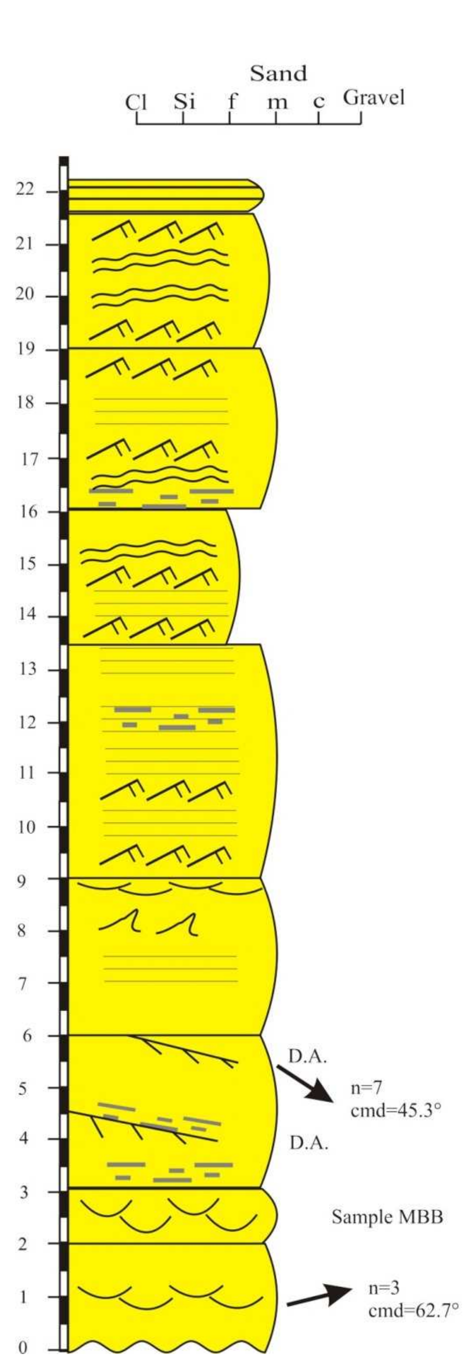

Massive sandstone
Cross-bedded sandstone
Laminated sandstone
Carbonaceous Siltstone
Laminated Siltstone
Mudrock
Non-Carbonaceous Siltstone
Coal
Covered Contact
Erosional Contact
Sharp Contact
Sample MBB-141
Sample Number
Paleocurrent Direction
Rip up clasts
Wood impressions
Logs
*Glossopteris* impressions/plant material
*In situ* stump cast
Ripples
Climbing Ripples
Downstream Accretion (D.A.)
Wavy Bedding
Lenticular Bedding
Folds
Flames
Vertical Burrows


Mt. Achenar
Sand
Cl Si f m c Gravel
Glossopteris impressions
Plant material
Wood impressions
Deformed
0
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
27
28
29
30
31
32
33
34
35
36
37
38
39
40
41
42
43
44
45

**Mt. Achenar**

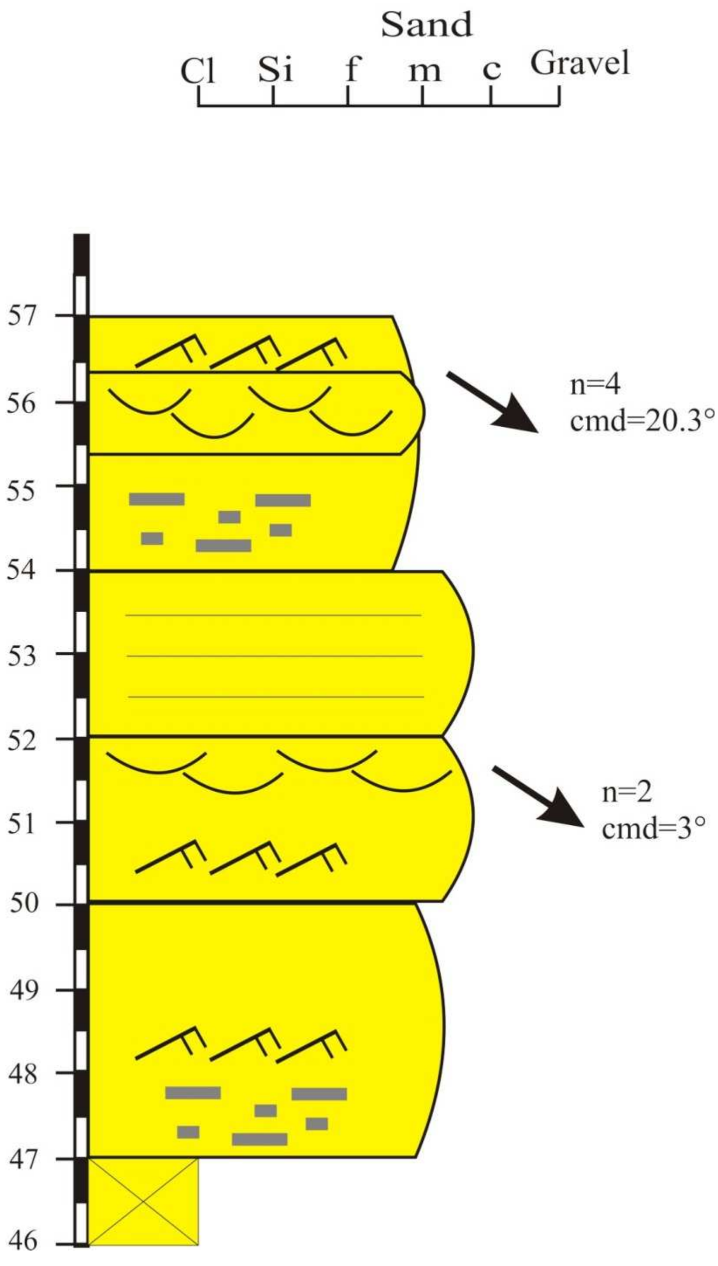

Sand
Cl
Si
f
m
c
Gravel
59
58
57
56
55
54
53
52
51
50
49
48
47
46
45
In situ stump cast


Mt. Achenar
Sand
Cl Si f m c Gravel
22
21
20
19
18
17
16
15
14
13
12
11
10
9
8
7
6
5
4
3
2
1
0
n=7
cmd=25.7°
n=2
cmd=2.5°
n=2
cmd=12.5°
n=6
cmd=20.8°
Down cutting
erosional surface
n=4
cmd=15°
Sand
Cl Si f m c Gravel
33
32
31
30
29
28
27
26
25
24
23
n=2
cmd=25°
n=11
cmd=39.5°

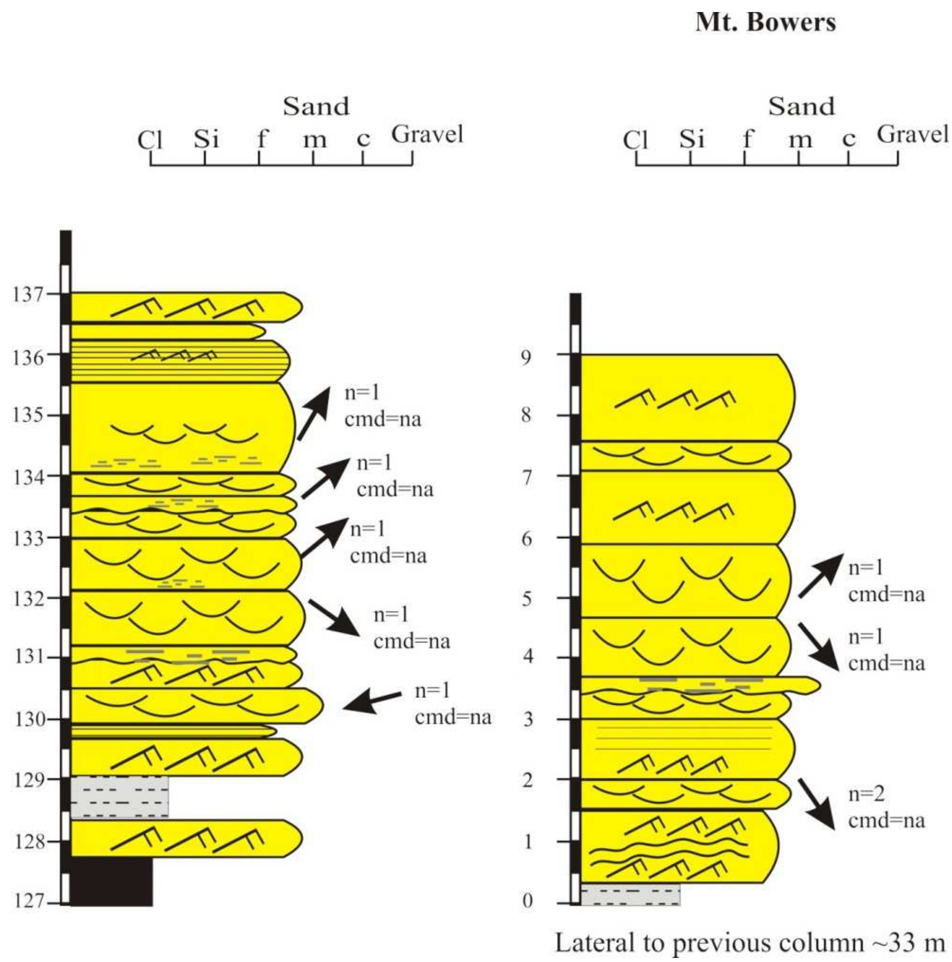


Mt. Bowers
Sand
Cl Si f m c Gravel
Sand
Cl Si f m c Gravel
22
21
20
19
18
17
16
15
14
13
12
11
10
9
8
7
6
5
4
3
2
1
0
D.A.
n=7
cmd=45.3°
D.A.
Sample MBB
n=3
cmd=62.7°
45
44
43
42
41
40
39
38
37
36
35
34
33
32
31
30
29
28
27
26
25
24
23

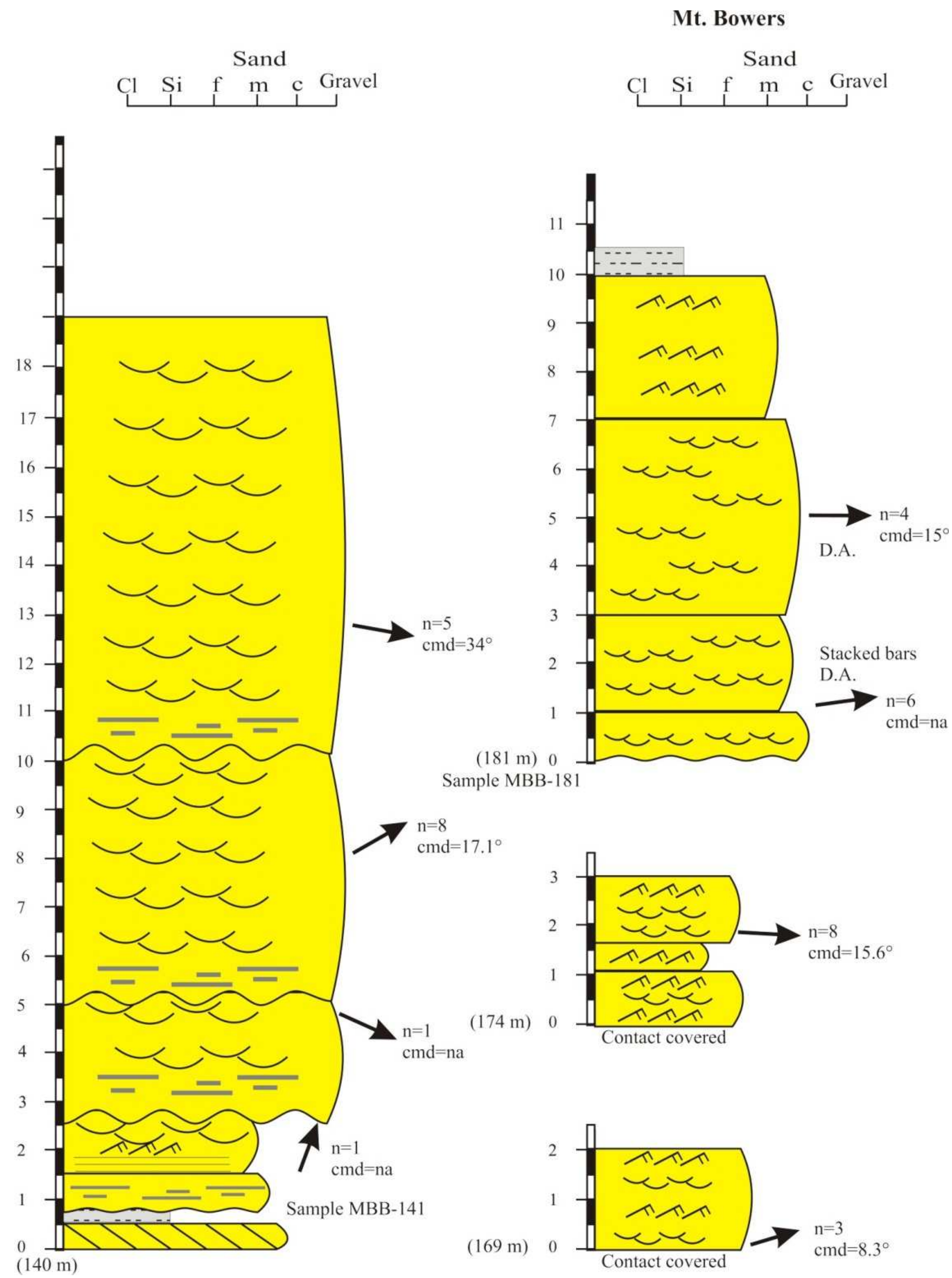
**Mt. Bowers**



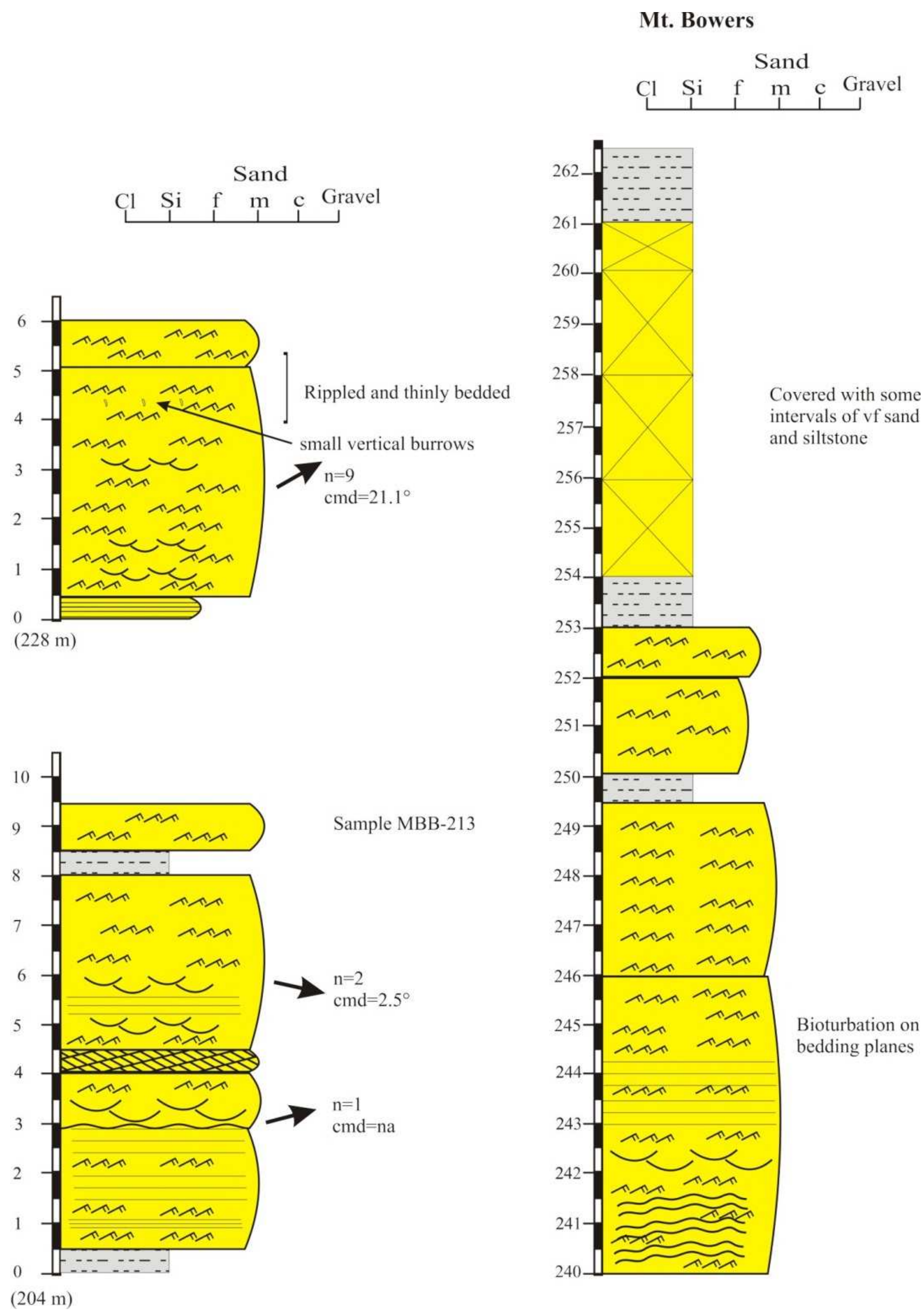
Sand
Cl
Si
f
m
c
Gravel
57
56
55
54
53
52
51
50
49
48
47
46
n=4
cmd=20.3°
n=2
cmd=3°



Mt. Bowers
Sand
Cl
Si
f
m
c
Gravel
137
136
135
134
133
132
131
130
129
128
127
n=1
cmd=na
9
8
7
6
5
4
3
2
1
0
n=2
cmd=na
Lateral to previous column ~33 m

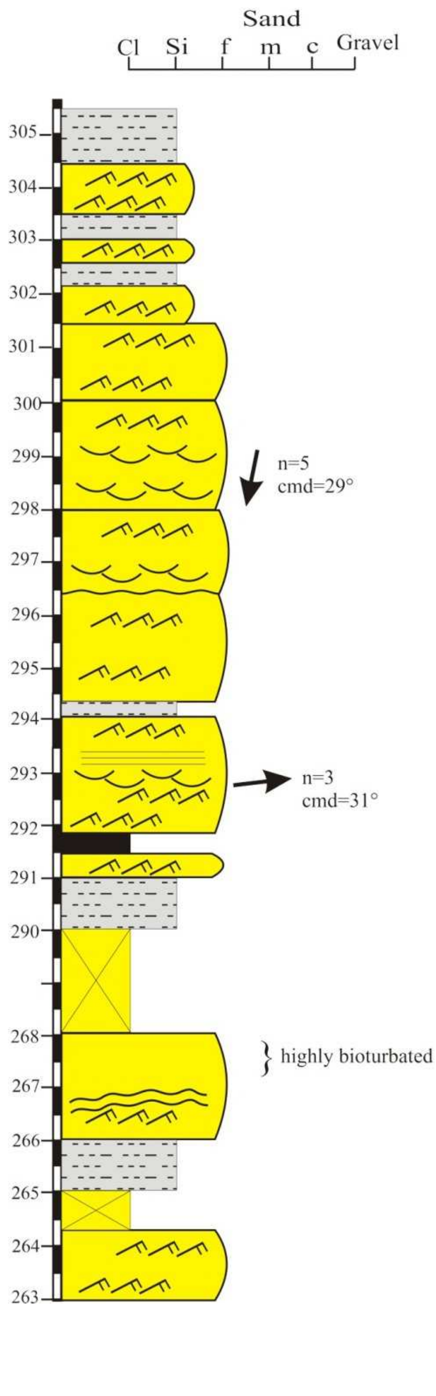


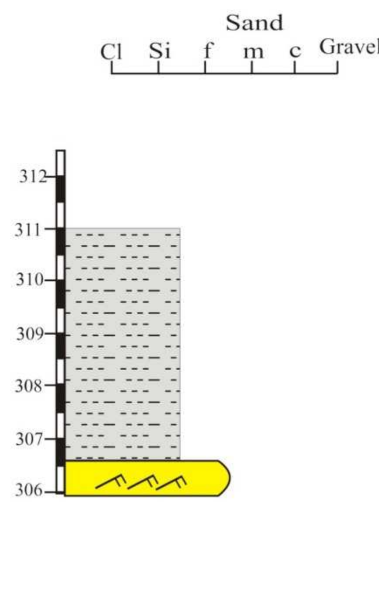
Mt. Bowers
Sand
Cl Si f m c Gravel
Sand
Cl Si f m c Gravel
18
17
16
15
14
13
12
11
10
9
8
7
6
5
4
3
2
1
0
(140 m)
n=5
cmd=34°
n=8
cmd=17.1°
n=1
cmd=na
n=1
cmd=na
Sample MBB-141
11
10
9
8
7
6
5
4
3
2
1
(181 m) 0
Sample MBB-181
n=4
cmd=15°
D.A.
Stacked bars
D.A.
n=6
cmd=na
3
2
1
(174 m) 0
Contact covered
n=8
cmd=15.6°
2
1
(169 m) 0
Contact covered
n=3
cmd=8.3°

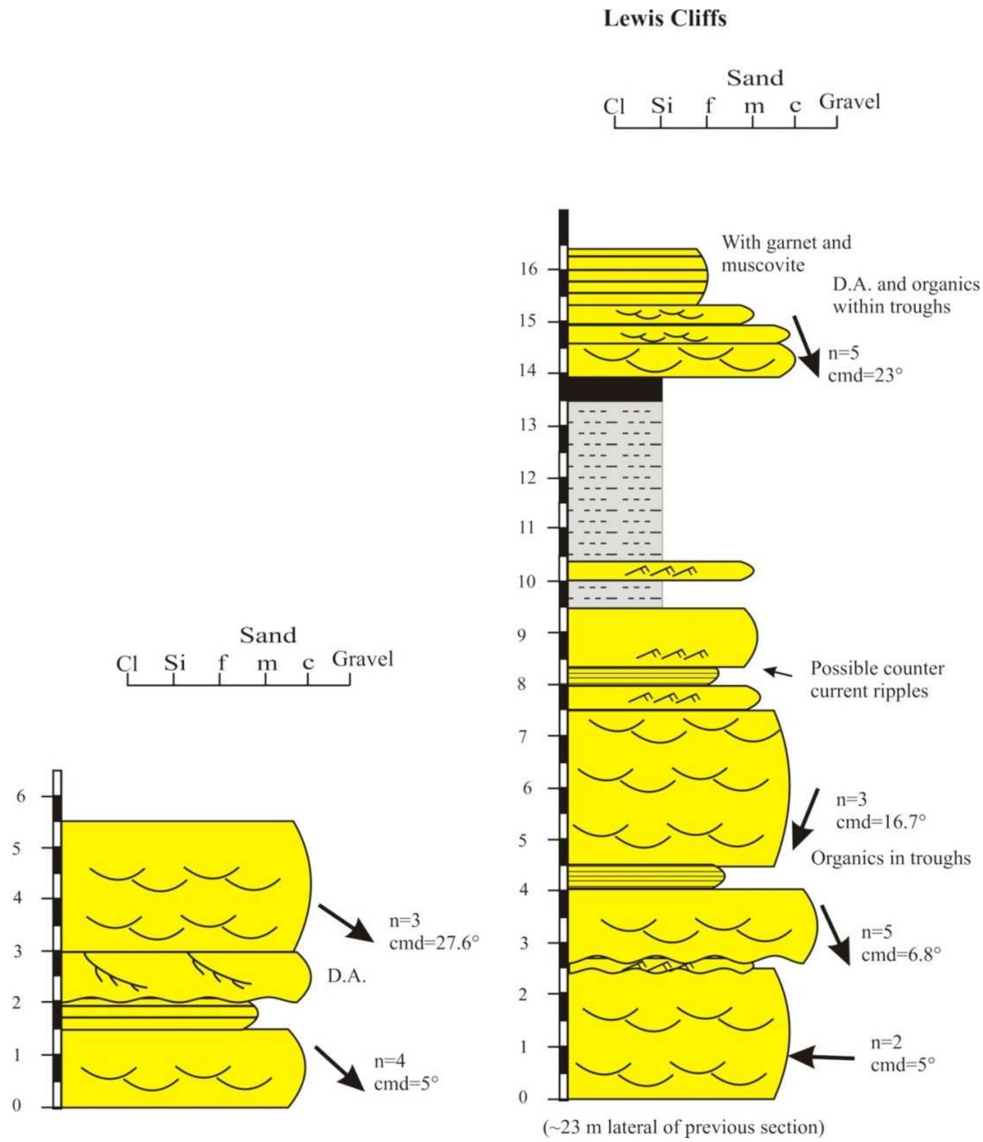


Mt. Bowers
Sand
Cl Si f m c Gravel
Rippled and thinly bedded
small vertical burrows
n=9
cmd=21.1°
(228 m)
Sample MBB-213
n=2
cmd=2.5°
n=1
cmd=na
(204 m)
Covered with some intervals of vf sand and siltstone
Bioturbation on bedding planes

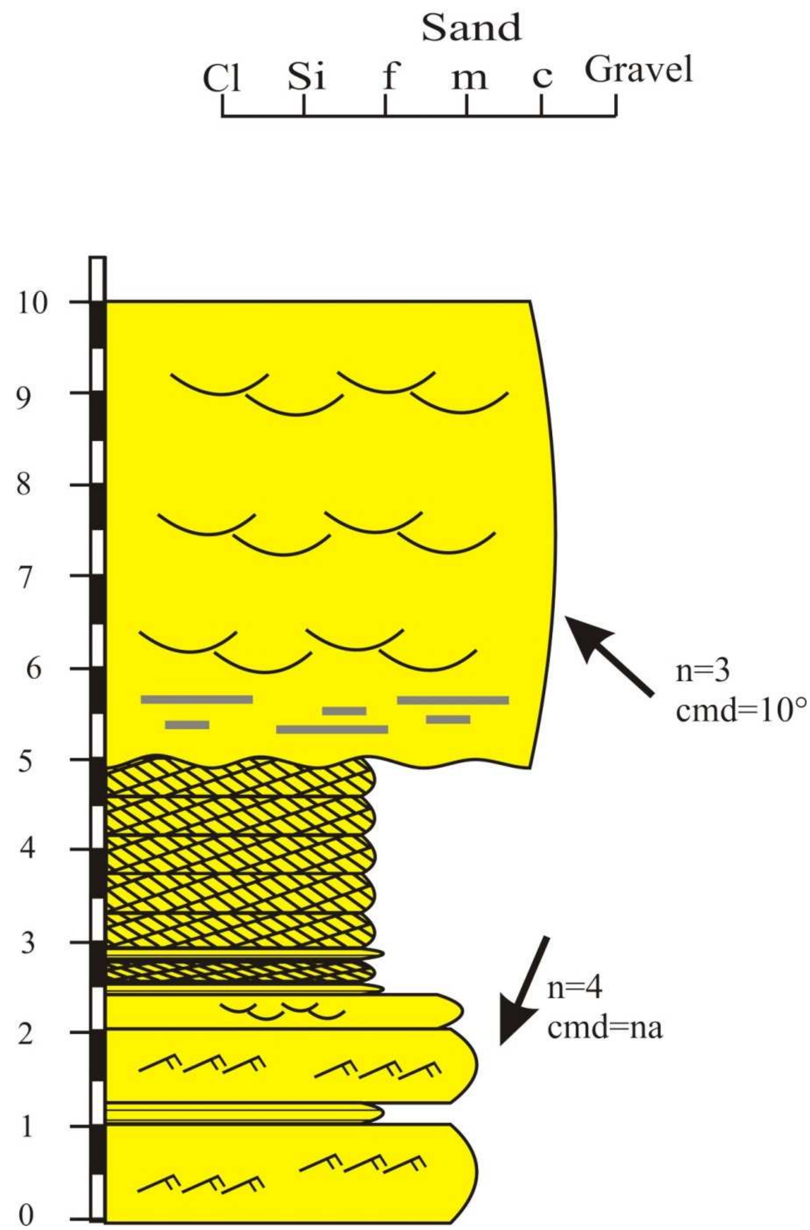
**Mt. Bowers**


Sand
Cl Si f m c Gravel
305
304
303
302
301
300
299
298
297
296
295
294
293
292
291
290
268
267
266
265
264
263
n=5
cmd=29°
n=3
cmd=31°
highly bioturbated

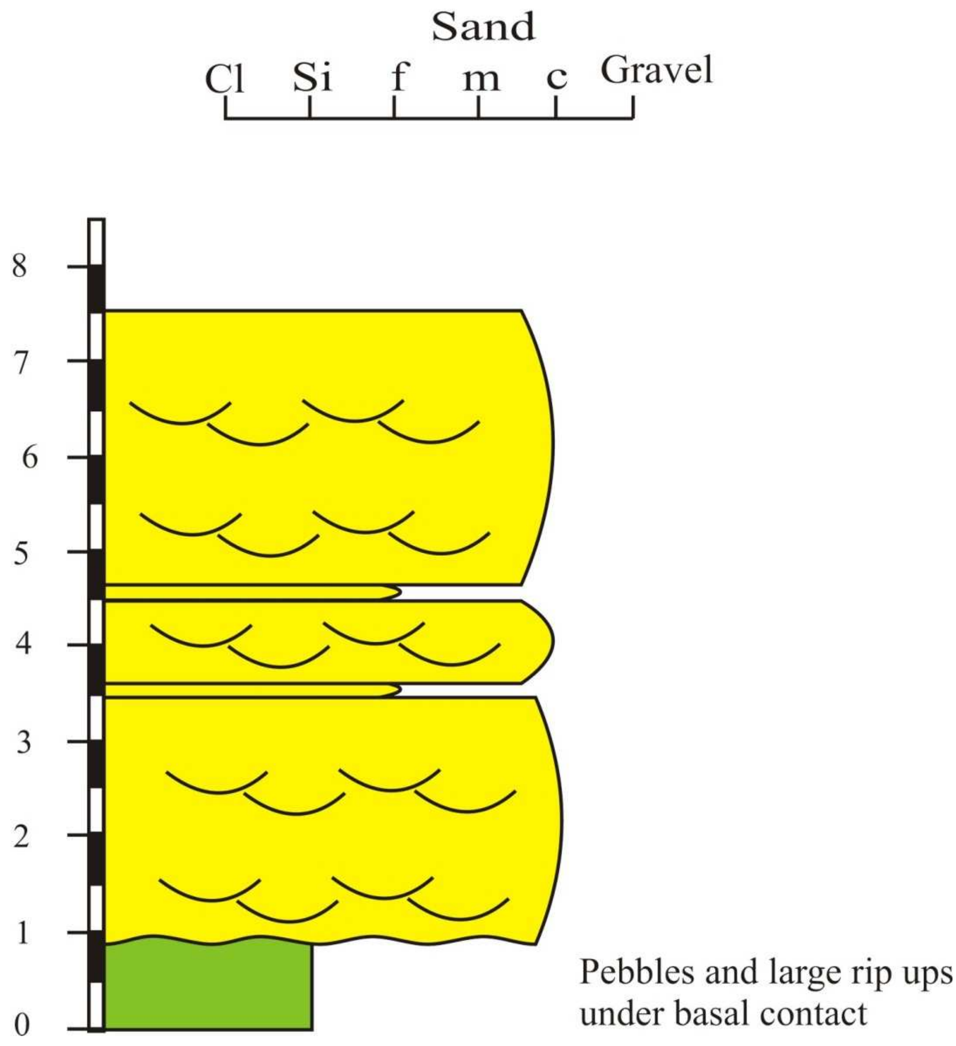

Sand
Cl Si f m c Gravel
312
311
310
309
308
307
306


Lewis Cliffs
Sand
Cl Si f m c Gravel
With garnet and muscovite
D.A. and organics within troughs
n=5
cmd=23°
Possible counter current ripples
n=3
cmd=16.7°
Organics in troughs
n=5
cmd=6.8°
n=2
cmd=5°
(~23 m lateral of previous section)
Sand
Cl Si f m c Gravel
n=3
cmd=27.6°
D.A.
n=4
cmd=5°
0
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16

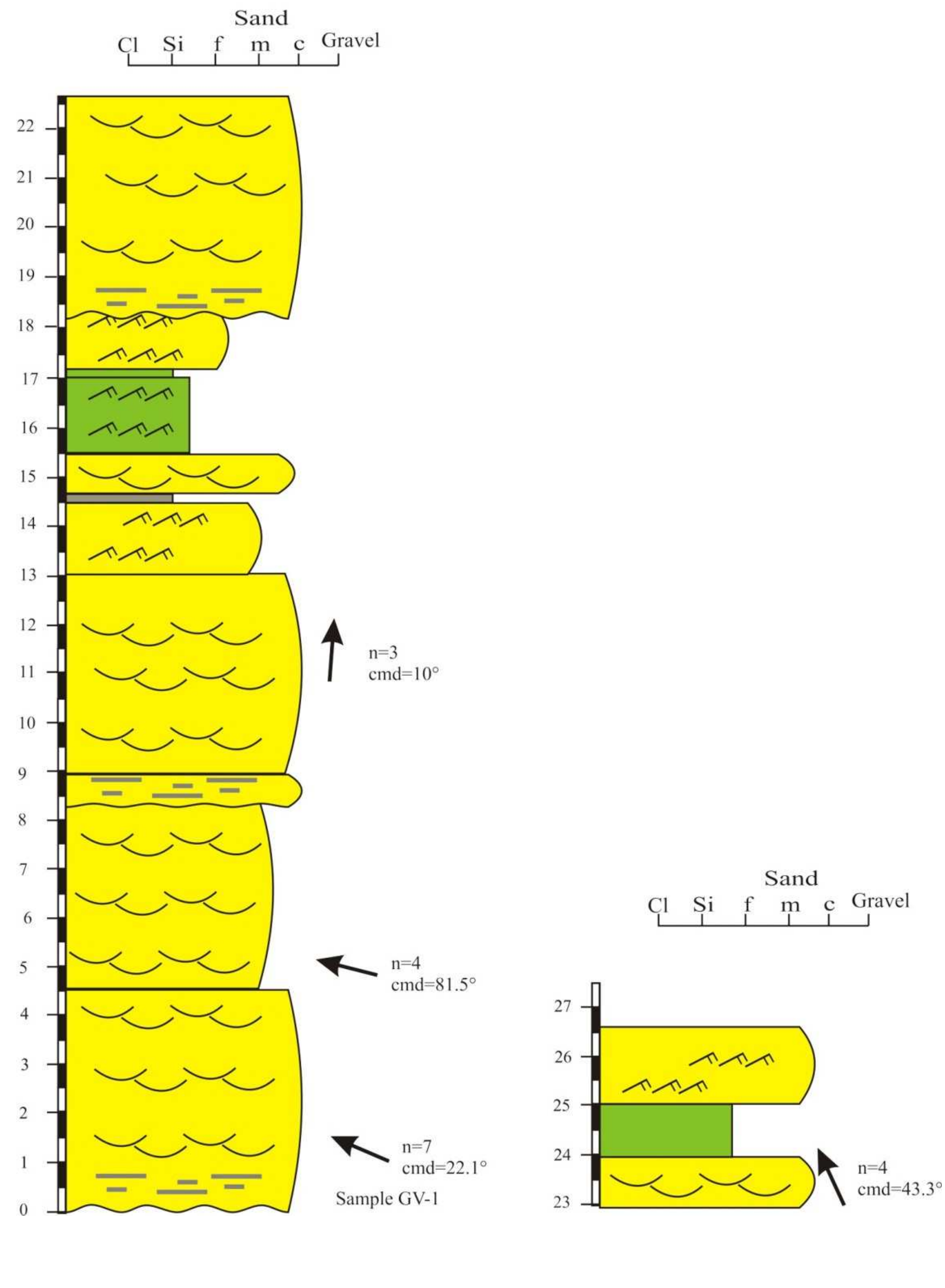
**Lewis Cliffs**



**Wahl Glacier**



**Gordon Valley**



Appendix B: Paleocurrent data: Mt. Bowers,

Lewis Cliffs, Mt. Achenar, and Gordon Valley

Table 6. Lower Buckley Formation, paleocurrent data

| Lower Buckley Formation | | | | | |
|---|---|---|---|---|---|
| Location | n | Azimuths | Declination | Vector Mean | Circle Variance |
| Mt. Bowers | | | 162 | | |
| Channel 1<br>0-6 m | 10 | 144, 76, 16, 92, 88, 78, 126, 122, 222, 197 | | 112 | 0.41 |
| Channel 2<br>50-57 m | 4 | 128, 122, 154, 136 | | 135 | 0.02 |
| Channel 3<br>67 m | 3 | 113, 68, 83 | | 88 | 0.05 |
| Channel 4<br>78 m | 2 | 78, 104 | | 91 | 0.03 |
| Channel 5<br>100-105 m | 9 | 134, 77, 122, 80, 74, 107, 97, 91, 86 | | 96 | 0.06 |
| Channel 6<br>131-135 m | 4 | 127, 52, 52, 32 | | 63 | 0.19 |
| Channel 7<br>142-159 m | 15 | 22, 116, 107, 84, 72, 57, 44, 57, 35, 47, 72, 142, 162, 67, 77 | | 75 | 0.2 |
| Channel 8<br>169-173 | 3 | 67, 72, 87 | | 75 | 0.01 |

| | | | | | |
|---|---|---|---|---|---|
| Channel 9<br>175-180 m | 8 | 77, 77, 92, 72, 102, 112, 107, 122 | | 95 | 0.05 |
| Channel 10<br>181-190 | 19 | 92, 87, 87, 87,97, 87, 62, 77, 72, 72, 62, 42, 77, 87, 102, 87, 117, 92, 77, 102 | | 83 | na |
| Channel 11<br>204-210 m | 3 | 77, 107, 102 | | 95 | 0.03 |
| Channel 12<br>228-234 | 7 | 67, 32, 47, 87, 102, 37, 57, | | 61 | 0.09 |
| Channel 13<br>343-350 m | 8 | 67, 122, 102, 142, 32, 7, 297, 322 | | 52 | 0.6 |
| Total Data:<br>Average Paleoflow | 13 | 112, 135, 88, 91, 96, 63, 75, 75, 95, 83, 95, 61, 52 | | 86 | 0.07 |
| Lewis Cliffs | | | 164 | | |
| Section 1<br>0-5.5 m | 7 | 129, 139, 129, 139, 134, 154, 89 | | 131 | 0.05 |
| Section 2<br>3-9 m | 13 | 159, 164, 149, 149, 164, 224, 214, 174, 189, 184, 134, 169, 124 | | 174 | 0.11 |
| Total Data: Average Paleoflow | 2 | 131, 174 | | 152 | 0.07 |

Table 7. Upper Buckley Formation, paleocurrent data

| Upper Buckley Formation | | | | | |
|---|---|---|---|---|---|
| Location | n | Azimuths | Declination | Vector Mean | Circle Variance |
| Mt. Achenar | | | 165 | | |
| 3-7 m | 12 | 210, 180, 205, 175, 170, 160, 140, 130, 110, 105, 115, 140 | | 153 | 0.17 |
| 20 m | 7 | 265, 355, 255, 210, 265, 260, 280 | | 267 | 0.21 |
| 24-29 m | 13 | 230, 355, 335, 295, 325, 250, 235, 325, 275, 250, 255, 255, 305 | | 283 | 0.23 |
| Total Data: Average Paleoflow | 3 | 153, 267, 283 | | 245 | 0.44 |

Table 8. Upper Fremouw Formation, paleocurrent data

| Upper Fremouw Formation | | | | | |
|---|---|---|---|---|---|
| Location | n | Azimuths | Declination | Vector mean | Circle variance |
| Gordon Valley | | | 161 | | |
| Channel 1<br>0-9 m | 10 | 316, 291, 271, 271, 281, 301, 296, 316, 11, 281 | | 298 | 0.11 |
| Channel 2<br>10-15 m | 3 | 356, 1, 21 | | 6 | 0.02 |
| Channel 3<br>19-24 m | 4 | 21, 11, 281, 311 | | 337 | 0.25 |
| Total Data: Average Paleoflow | 3 | 298, 6, 337 | | 334 | 0.11 |