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POSTGLACIAL GEOMORPHIC RESPONSE AND ENVIRONMENTAL CHANGE IN SOUTHEASTERN ALBERTA, CANADA

BY

CELINA CAMPBELL

A THESIS
SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH
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FOR THE DEGREE OF

DOCTOR OF PHILOSOPHY

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The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled **POSTGLACIAL GEOMORPHIC RESPONSE AND ENVIRONMENTAL CHANGE IN SOUTHEASTERN ALBERTA, CANADA** submitted by **CELINA CAMPBELL** in partial fulfilment of the requirements for the degree **DOCTOR OF PHILOSOPHY.**

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ABSTRACT

This thesis tests the hypothesis that the tempo and pattern of geomorphic changes in southeastern Alberta over the postglacial period reflects the nature and intensity of climatic forcing as expressed principally through hydrological response. In order to test this hypothesis, five separate studies were conducted in different geomorphic settings to investigate a wide range of dominant geomorphic processes which operated through the varying climates of the postglacial period.

This first Chapter introduces the study and study area, and develops an *a priori* empirical model through a synthesis of published and unpublished calibrated radiocarbon dates of postglacial geomorphic events from southeastern Alberta and surrounding plains. Chapter 2 examines the variability of lake area - a simple indicator of hydrology - across a climatic transect extending through the study area and further north. Chapter 3 is a study of the sediments retrieved from the Tide Lake basin, a playa in the south of the study area. Chapter 4 examines the Klassen Site, an alluvial fan developed on an abandoned meander of the Red Deer river. Chapter 5 develops a 4000+ year paleohydrology from the sediments of Pine Lake, in the northernmost part of the study area, and uses it for an examination of the pattern of terrace formation in streams tributary to the Red Deer River. Chapter 6 studies the aeolian deposits of the area. Finally, in Chapter 7, the regional chronology developed in Chapter 1 is revised and refined in light of the knowledge and understanding gained in this study. The thesis culminates with the proposal of a synthetic paleohydrology curve for the past 20,000 years developed from all the available information from the study area and surrounding

regions.

In addition to testing the thesis hypothesis, this research yields two principle by-products: the first, is a refined and more accurate interpretation of the existing spatial and temporal models of the pattern of geomorphic response to climatic forcing during the postglacial in the plains of southeastern Alberta; the second, is an improved basis for identifying the nature of these geomorphic responses in the context of data from a variety of previously unreported, and selected sites in the region.

Perhaps the most interesting results of this research are: (1) the close association of geomorphic patterns when dated in cal yr with insolation; (2) the brevity and timing of ice-cover in this region; (3) the importance of the Younger Dryas aged climatic reversal; (4) the emergence of a coherent pattern of geomorphic activity from a synthesis of seemingly disparate studies, implying a regional forcing mechanism, most likely variations in effective moisture, and; (5) the identification of where geomorphic activity seems to reflect climate, where it doesn't and why.

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To the memory of Kent Holden.

Ian, my partner in life, has contributed to this work through endless hours of child care, drafting several of the figures and putting up with me (despite my best efforts).Our wonderful children Jacob and Avery have been real troupers. I owe them all big-time, and I dedicate this thesis to them with love.

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CHAPTER 1

1.1 INTRODUCTION AND STATEMENT OF OBJECTIVES

Landscapes are a palimpsest of the "superimposed effects of climatic and tectonic vicissitudes" (Chorley *et al.* 1984:3). Climatic changes can affect the magnitude and rates of weathering, erosion, and material transportation and deposition. These processes leave an imprint on the landscape. While some landforms clearly show the effects of climate change on geomorphic processes, correlation and causation between climate and geomorphic response are problematic. In some landforms, extrapolation of the geomorphic response to climatic change remains elusive because climate is only one of many variables, both independent (external variables) and dependent (internal variables), which can affect a geomorphic system (Schumm 1981; Bull 1990).

Landform development is further complicated by factors such as complex responses, geomorphic thresholds, allometry, magnitude and frequency relationships, catastrophic versus gradual change, and patterns of equifinality and scale- related issues (for example Wolman and Miller 1960; Schumm and Lichty 1965; Schumm 1973; Schumm and Parker 1973; Bull 1975, 1991; Wolman and Gerson 1977; Brunsden and Thornes 1979; Schumm 1979; I.A. Campbell and Honsaker 1982; Schumm *et al.* 1987; de Boer and Campbell 1989; de Boer 1990, 1992a, 1992b). Thus, although similar sites in similar settings tend to show similar responses to a particular stimulus, each site constitutes an individual geomorphic system with individual thresholds of response; accordingly, correlations often remain elusive.

Superimposed on this assortment of controls are a host of diverse theoretical approaches which impose different structures and emphasis on geomorphic methodology, description and interpretation (for example Brunsden 1990; Baker and Twidale 1991; Rhoads and Thorn 1993). The work of Buttimer (for example 1982, 1986) eloquently shows that no research is without bias, and that research is the direct outcome of the meaning, milieu and metaphor of an individual researcher's personal and working world.

The principle goal of geomorphological research is to understand the processes which mould the earth's landforms, and to provide a theoretical basis for explaining how such processes operate in the context of both spatial and temporal variations in the forces which control them. In so doing, geomorphologists may devise models in order to define the linkages and attributes of the process-landscape system they are investigating (Thorn 1988). Any model is an abstraction of reality - a simplification of a set of complex interrelationships (Harvey 1969) - which forms the basis for the creation of new theory or the restructuring of existing theory. As Graf explains however, a model may be either "...in the form of an hypothesis, law or theory, depending on its degree of confirmation" (1988:26). It is in Graf's sense that the term model is used in this thesis; i.e., the model to be developed forms part of a more complex hypothetical structure which seeks to define the relationships between the interactive variables of climate and geomorphic response. The specific hypothesis which will be examined throughout this thesis proposes that the tempo and pattern of geomorphic changes in southeastern Alberta over the postglacial period reflects the nature and intensity of climatic forcing as expressed principally through hydrological response. This is reflected in a synthetic regional chronology of geomorphic activity in the region, developed in Chapter 1.

In order to test this hypothesis, five separate studies were conducted in different geomorphic settings to investigate a wide range of dominant geomorphic processes which operated through the varying climates of the postglacial period. The different geomorphic settings were selected for study because each covers different periods of postglacial time in varying degrees of detail (Chapters 2-6). Furthermore, each site represents a different type of geomorphic system which may show differing types of response and sensitivities to climatic change. Thus each site can potentially serve as a test-case and be used to refine different aspects of the general model presented in Chapter 1.

A holistic research approach employing a variety of methods in sedimentary core analysis was used, as advocated by the International Geological Correlation Programme (IGCP) subproject 158B, "Paleohydrology of the Temperate Zone During the Last 15,000 years - Lake and Mire Environments" (Berglund 1986). Sedimentary descriptions (structure, wet and dry Munsell colours), grain size, grain roundness, loss on ignition, bulk geochemistry, elemental charcoal quantification, charcoal identification, X-ray diffraction, and tephra analysis of ^{14}C, tephra and ^{210}Pb-dated postglacial terrestrial and limnic sediments from different types of landforms are described and interpreted, although not all analyses were performed on each site.

Bull (1991), Clayton (1983), and Sauchyn (1993), amongst others, discuss the difficulties inherent in assigning stratigraphic units to particular climatic episodes, or conversely, assigning paleoclimates to stratigraphic units. There are four major facets to this problem. Firstly, as noted above, a landscape is a palimpsest, and any climatic episode can only overprint the signatures of prior climatic episodes and in turn be overprinted by subsequent episodes. Thus it is common for a given landscape to include features derived from, or modified by, the variety of climates which have occurred in the area through time.

The second facet is that most terrestrial sites are non-depositional, or even erosional, at any given moment. As a result only unusual sites are likely to retain a complete, or even nearly complete, record of change. The third facet is that the more extreme and rare the event, the more likely it is that it will leave a record; conversely, non-extreme events will leave weaker records.

Finally, the fourth facet, often least recognized (although briefly explored in Sauchyn 1993), is that geomorphic events are the result of instabilities. This implies that long periods of unvarying climate will be represented by diminishing geomorphic activity through time as the system approaches equilibrium with the climatic conditions. Similarly, the greatest geomorphic activity may occur during or immediately following the greatest or most rapid climatic change, i.e., at times of greatest disequilibrium between the geomorphic system and climate.

Despite these complicating factors, records of past climatic events have been obtained from studies of erosional and depositional landforms. Although certain processes appear to operate within relatively narrow bands of climatic domains, others remain effective over a wide climatic spectrum. In such cases regional geomorphic trends can reflect a similar response to external variables. For example: (1) dust storm activity is highest in areas with 100-200 mm annual rainfall (Goudie 1983); (2) barchan dunes and yardangs are formed in areas with minimal to no vegetation (Thomas 1989); and (3) continuous permafrost develops when the mean annual temperature is below -6 - -8 ^{0}C (Sugden 1982). In comparison, (4) coastal dunes can occur in all climates. The key is to seek geomorphic process-climate response patterns in order to ascertain geomorphic 'indicators' of climate.

In humid environments, where precipitation exceeds potential evapotranspiration in most years, hydrologically-driven geomorphic processes will show relatively constant behaviour through time. Similarly, in hyperarid environments, where precipitation is a rare event, there will be very little hydrologically-driven geomorphic activity except during those rare events. In the intermediate, semiarid zone, where precipitation is common but not normally sufficient to balance potential evapotranspiration, hydrologically-driven geomorphic processes will be highly variable, and may therefore potentially be good indicators of climate change (for example C. Campbell *et al.* 1994 - Chapter 3). Semiarid terrains are thus particularly sensitive to minor changes in weather / climate (seasonal, annual, decadal or long-term) (Graf 1988; Vance and Wolfe 1996). A historical example of this occurred in the semiarid northern Great Plains in the 1920s and 1930s when perturbations in precipitation resulted in dust storms, which were further exacerbated by poor land management techniques (for example Gray 1978). In this way, semiarid environments form not only vegetation ecotones, but the geomorphic equivalents of ecotones - boundaries between regions of differing geomorphic regimes.

"If geomorphic processes can be characterized as the interplay between force and resistance..., then the place and time at which the two are equal is a threshold" (Graf 1988:50). Landscapes operate within a complex set of thresholds rather than a single threshold. Each landscape unit will respond to a given climate or climatic event in a different way, depending on the nature and magnitude of its disequilibrium. Thus the same type of geomorphic event (e.g., permafrost initiation) may occur in different areas at different times in response to a monotonic regional climate change, i.e., may be time-transgressive. This means that the use of geomorphic events to reconstruct paleoclimate, requires not only an understanding of the geomorphic system but also a reliable dating of each event considered.

We must therefore view the geomorphic record as a function of discrete threshold crossings, such that each geomorphic event is viewed as a discrete datum, possibly unrelated to any other geomorphic events, no matter how similar they may be. Only by determining temporal and spatial patterns of events can we hope to unravel the relationship between paleoclimate and landscape evolution. While unique geomorphic events carry significant information, because of the possible effects of complex responses it is next to impossible to ascertain the relative roles of the various external and internal variables which may have led to the threshold crossing represented by the unique event. Throughout this thesis, the emphasis will accordingly be placed on the identification and analysis of spatial and temporal patterns, rather than on unique geomorphic events.

In addition to testing the thesis hypothesis, this research yields two principle by-products: the first, is a refined and more accurate interpretation of the existing spatial and temporal models of the pattern of geomorphic response to climatic forcing during the postglacial in the plains of southeastern Alberta; the second, is an improved basis for identifying the nature of these geomorphic responses in the context of data from a variety of previously unreported, and selected sites in the region.

This first Chapter[1] introduces the study and study area, and develops an *a priori* empirical model through a synthesis of published and unpublished calibrated radiocarbon dates of postglacial geomorphic events from southeastern Alberta and surrounding plains. Chapter 2 examines the variability of lake area - a simple indicator of hydrology - across a climatic transect extending through the study area and further north. Chapter 3 is a study of the sediments retrieved from the Tide Lake basin, a playa in the south of the study area. Chapter 4 examines the Klassen Site, an alluvial fan developed on an abandoned meander of the Red Deer river. Chapter 5 develops a 4000+ year paleohydrology from the sediments of Pine Lake, in the northernmost part of the study area, and uses it for an examination of the pattern of terrace formation in streams tributary to the Red Deer River. Chapter 6 studies the aeolian deposits of the area. Finally, in Chapter 7, the regional chronology developed in Chapter 1 is revised and refined in light of the knowledge and understanding gained in Chapters 2 through 6. The thesis culminates with the proposal of a synthetic paleohydrology curve developed from all the available information from the study area and surrounding regions. From this evidence an assessment is made of the validity of the hypothetical model which forms the focus of this thesis.

1.2 GENERAL DESCRIPTION OF THE STUDY AREA

1.2.1 GEOLOGY AND PHYSIOGRAPHY

Intensive geomorphological research began in southeastern Alberta in the late 1960s, and since then substantial amounts of information have been accumulated on the nature and rates of processes of geomorphic processes and the patterned of landscape evolution (for example McPherson 1968; I.A. Campbell 1970a, 1970b, 1973, 1974, 1977a, 1977b, 1981, 1982, 1985, 1987a, 1987b, 1989, 1992; Bryan *et al.* 1978, 1984, 1987, 1988; Bryan and Campbell 1980, 1982, 1984, 1985, 1986; I.A. Campbell and Honsaker 1982; Hodges 1982; Hodges and Bryan 1982; Bryan and Hodges 1984; Harty 1984; Bryan and Harvey 1985; O'Hara 1986; I.A. Campbell and de Boer 1988; Sutherland and Bryan 1988; de Boer and Campbell 1989, 1990; de Boer 1990; I.A. Campbell and Evans 1990, 1993; de Lugt and Campbell 1992; Evans and Campbell 1992, 1995; Seemann 1993; I.A. Campbell *et al.* 1993; O'Hara and Campbell 1993; Rains *et al.* 1994; Barling 1995; Paterson 1996).

This part of Alberta offers four distinct advantages for geomorphological studies:

1

Parts of each chapter are written in individual paper format, inevitably producing some redundancies. Also, parts of this thesis have been written with co-authors; their contributions are noted in the text as appropriate.

(1) the absence of forest vegetation cover means that landform features can be readily discerned; (2) there are extensive areas of exposed bedrock and surficial deposits (for example, valleys, badlands, etc.) (3) there have been a number of regional environmental models proposed which, when combined, span the entire postglacial period, and can thus be synthesised and tested; and (4) according to Vance (1996) the study area is generally perceived as being a paleoenvironmental data gap, this present study intends to help close.

Figure 1.1 maps the locations of all the sites analysed in this thesis. The majority of sites studied are located just east of Dinosaur Provincial Park. Dinosaur Provincial Park is an extensive area of badland development associated with the incision of spillways during the latest Wisconsinan deglaciation which exposed strata within the highly erodible Upper Cretaceous Judith River Group (Bryan *et al.* 1987; Eberth *et al.* 1996). Pine Lake is located north of the rest of the study area and is discussed separately in Chapter 5.

The study area is located in the Western Canada Sedimentary Basin, a thick wedge of Phanerozoic strata overlying Precambrian basement rock (Lemmen, in press). The geological evolution of the region has been recently summarised by Leckie and Smith (1993), Dawson *et al.* (1994), Mossop and Shetsen (1994) and Lemmen (in press).

The modern surface is covered by prairie grassland chernozemic soils, and Holocene and Pleistocene deposits which are underlain by horizontally bedded, poorly consolidated, Upper Cretaceous Judith River Group and Bearpaw Formation sandstones, siltstones, and mudstones (Eberth *et al.* 1996; Lemmen, in press). These rocks were deposited during a series of transgressive / regressive cycles within the marine basin of the Western Seaway, associated with the uplift in the Rocky Mountains to the west (Laramide Orogeny) and post-tectonic adjustments during the Tertiary (Leckie and Smith 1993). The extremely gentle regional dip of the bedrock is to the north-northeast. The Quaternary evolution of the area is reviewed in Section 1.2.

The postglacial geomorphology of the region primarily reflects the interaction of fluvial and aeolian processes on a variety of unconsolidated surficial deposits of variable thicknesses derived principally from glacial, glaciofluvial, and glaciolacustrine sources (for examples see: Berg and McPherson 1972; Kjearsgaard *et al.* 1983; Shetsen 1987; Kjearsgaard 1988; Evans and Campbell 1992; Kwiatkowski and Marciak 1994; Lemmen, in press). Postglacial geomorphic processes, such as slopewash, piping, mass movements and aeolian activity continuously alters the landscape (for example O'Hara 1986; Bryan *et al.* 1987; I.A. Campbell and Evans 1990; de Lugt and Campbell 1992; Evans and Campbell 1992; O'Hara and Campbell 1993).

1.2.2 CLIMATE

The region has a midlatitude, semiarid, and continental climate with short warm summers and long cold winters (Environment Canada 1993) and is designated as BSk (cold winter steppe) in the Köppen Climatic Classification (I.A. Campbell 1974). The interaction between Arctic and Pacific air masses produces great variability in both weather and climate on an annual and interannual basis. Air mass interaction is controlled mainly by geographic setting and seasonal variation in the latitudinal position of the core of the jet stream. The Rocky Mountains to the west serve as a barrier to moist mild Pacific air (Gullet and Skinner 1992), although in the summer modified Pacific air predominates. In winter, the jet stream coupled with the relatively flat prairie surface, facilitates incursion of shallow, cold, dry Arctic air masses. Low amounts of winter precipitation is due to the dominance of dry Arctic air masses.

Mean monthly temperatures for thirty-year normals (1961-1990) for July and January range from *ca.* 19°C in the south to 16°C in the north in the study area, and *ca.* -12°C in the south and -14°C in the north, respectively (Environment Canada 1993; Figure 1.1; modifed from Vance *et al.* in press). Mean annual precipitation for 1961-1990 ranges from *ca.* 350 mm in the south to 400 mm in the north (Environment Canada 1993; Figure 1.1). Average potential evapotranspiration often exceeds precipitation by >300 mm yr^{-1} (Winter 1989). Much of the annual rainfall (*ca.* 70 %) falls during the summer as short, relatively intense convective storms, or as more prolonged low intensity storms (Bryan *et al.* 1988). Approximately 35 % of the annual precipitation falls as snow, but sublimation / chinook melts commonly deplete the snow-pack. Mean annual windspeed can exceed 20 km hr^{-1} (Walmsley and Morris 1993), with significant fetch across the treeless, flat terrain. Strong westerly winds prevail (Environment Canada 1993). In the winter and early spring chinook winds with speeds in excess of 90 km hr^{-1} have been recorded (Vance *et al.* in press).

During the summer, warm temperatures, strong winds and extended daylight cause potential evapotranspirational losses to far exceed mean annual precipitation; the result is a pronounced moisture deficit (Hogg 1994). Vance *et al.* (in press) note potential evapotranspiration of 520-580 mm (Figure 1.1). Widespread drought has occurred frequently throughout the historic record (Chakravarti 1976). Historic droughts have been linked to the development of stable high pressure cells displacing cyclonic tracks, moist air masses and fronts northward, resulting in clear skies, high temperatures, limited convective storm development (Dey 1982; Dey and Chakravarti 1976), reduction in spring runoff, and the desiccation of sloughs and small lakes (Wheaton and Chakravarti 1990). The severe 1988 drought is an example of this synoptic pattern (Vance *et al.* in press). Vance *et al.* (in press) note, however, that expressions of drought prior to 1950 (the date at which collection of upper air data started), especially during the early to mid-Holocene, may have been a function of different synoptic patterns.

A statistically significant regional warming of 0.9°C has occurred since the late 1800s (Gullet and Skinner 1992). Although historic droughts of the 1920s, 1930s and 1980s greatly affected both the landscape and its inhabitants, these droughts appear to represent normal patterns of variability in the longer term mean of regional climate rather than climate change. Increased precipitation appears to be correlated with below normal temperatures (Vance, in press).

1.2.3 VEGETATION

Excluding agricultural crops, which presently cover most of the prairie surface, the "northern mixed grass prairie" (Risser *et al.* 1981) or "mixed prairie" (Coupland 1950, 1961) is dominated by grasses, xerophytic flowering plants and shrubs. Seven major grass species dominate the region including blue grama (*Bouteloua gracilis*), needle and thread grass (*Stipa comata*), porcupine grass (*Stipa spartaea*), wheat grass (*Agropyron smithii*), June grass (*Koeleria cristata*) and Sandberg bluegrass (*Poa secunda*) (Coupland 1950, 1961). Xerophytic flowering plants such as prickly pear (*Opuntia polyacantha*) and cushion cacti (*Mamillaria vivipara*), and shrubs including sagebrush (*Artemisia cana*), and pasture sagewort (*Artemisia frigida*) are also found. Cottonwoods (*Populus acuminata, P. angustifolia*, and *P. deltoides*) grow along the edge of river valleys, coulees, and streams (Coxson and Looney 1986).

1.3 POSTGLACIAL GEOMORPHIC / ENVIRONMENTAL CHANGE MODELS IN SOUTHEASTERN ALBERTA

1.3.1 PREGLACIAL

Quaternary studies in southeastern Alberta during the first half of this century were dominated by surficial mapping and the establishment of Quaternary stratigraphy (for example Dowling 1917; Williams 1929; Williams and Dyer 1930; Johnson and Wickenden 1931; Bretz 1943; Stalker 1960, 1963, 1965, 1969, 1973). Until the application of oxygen isotope analysis to deep sea cores by Emiliani (1955) and the development and widespread application of absolute dating techniques, the Pleistocene of North America was subdivided into four major glacial phases: Nebraskan, Kansan, Illinoian and Wisconsinan (parallelling the fourfold European glacial sequence: Günz, Mindel, Riss and Würm). Regionally, the multiple glaciation model was then subjected to testing, modification (for example Horberg 1954; Stalker 1963, 1983), and eventual rejection. The two most recent provide evidence supporting: (1) a two-fold glaciation in southeastern Alberta (Proudfoot 1985; Evans and Campbell 1992, 1995); and, (2) a single late Wisconsin Laurentide glaciation in northwestern Alberta (Liverman *et al.* 1989; Young *et al.* 1994) which followed a very long depositional period of Empress Formation, formerly termed Saskatchewan Gravels and Sands (Rutherford 1937; Evans and Campbell 1995).

Based on 29 radiocarbon dates from bones found within the terminal Empress Gravel Formation in central Alberta, Young *et al.* (1994) suggest ice-free conditions in the Edmonton area prior to *ca.* 24,820 cal yr B.P. (21,330-42,910 ^{14}C yr B.P., AECV-1664C, AECV-1658C; linear extrapolation of calibration, after Bard *et al.* 1990). A date obtained on gastropod shells *ca.* 30,840 cal yr B.P. (27,420 $\pm$ 240 ^{14}C yr B.P., TO-1464, Evans and Campbell 1992; linear extrapolation of calibration, after Bard *et al.* 1990) from a preglacial fluvial deposit in Dinosaur Provincial Park, lends additional support for ice-free conditions during this time period.

The preglacial, Canadian Shield clast-free, Empress Formation was deposited in vast valley systems up to 215 m deep (Farvolden 1963; Carlson 1970; Tokarsky 1986; Evans and Campbell 1995; Pawlowicz and Fenton 1995). Several of these preglacial valleys cross the study area at various depths from the surface. The effect of these valleys on glacial and postglacial landscape evolution is potentially large (for example Evans and Campbell 1995). Late Wisconsinan Laurentide glaciation disrupted the preglacial drainage network, partly by infilling and partly by incising new channels via subglacial and proglacial glaciofluvial activity. In both cases, channels would have been forced / drawn away from the old drainage network wherever these actions combined with strong subglacial hydraulic flow or where the moving ice-front created a different optimal flow path. Thus apart from a few minor streams the modern drainage network is an amalgam of re-excavated preglacial, subglacial, and proglacial ice-marginal drainage.

1.3.2 GLACIAL

Based on 31 stratigraphic logs in Dinosaur Provincial Park, Evans and Campbell (1992) identified seven major and five minor lithofacies associations (LFA). LFA 1 (the oldest LFA), with a gastropod shell radiocarbon date of *ca.* 30,840 cal yr B.P. (27,420 $\pm$ 240 ^{14}C yr B.P., TO-1464; linear extrapolation of calibration, after Bard *et al.* 1990) which the authors suggest may be contaminated as there is only a 50% chance of shell dates >25,000 B.P. being uncontaminated, records preglacial deposition of sandy and then braided fluvial floodplains. LFA 1 grades into LFA 2, recording the formation of an ice-dammed lake as Laurentide ice advanced into the region. Ice rafted shield clasts were deposited at the top of LFA 2. The contact between LFA 2 and LFA 3 is erosional. LFA 3 is a grey till, recording ice movement from the northeast. LFA 3 grades into LFA 4.

LFA 4 is described as interbedded planar, trough cross-stratified gravels, ripple and planar cross-stratified sands, yellow / orange in colouring, disturbed by glaciotectonic folding, thrusting and minor normal faults. Evans and Campbell (1992) interpret LFA 4 sediments as supraglacially-derived diamictons and outwash. Its oxidized colour is interpreted as indicating either a considerable period of subaerial weathering or Fe-rich groundwater activity. The deposit is then tentatively assigned an interstadial or interglacial age. The contact between LFA 4 and LFA 5 is sharp.

LFA 5 is a heterogeneous diamicton interpreted as recording a period of subglacial moulding and then cavity filling by fluvial debris flow deposits. The contact between LFA 5 and LFA 6 is sharp. LFA 6 is a heterogeneous diamicton recording lodgement and melt-out of glacier ice. According to Evans and Campbell (1992:535) "this sequence of moulding - cavity fill - tectonism - melt-out represents large scale decoupling of the glacier from its bed and later surges, creating extensive deformation of the substrate." In places, melt-out was arrested by remobilisation due to surges which continued throughout ice retreat. LFA's 7a-7e chronicle postglacial glaciolacustrine, fluvial, pond and aeolian depositional environments.

Evans and Campbell (1992) reject their date from LFA 1 of *ca.* 30,840 cal yr B.P.(TO-1464), then fit their evolutionary sequence into Proudfoot's (1985) model and regional dates, spanning from *ca.* 610,000 yr B.P. through the Holocene. However, LFA 4 is also consistent with a model in which deposition of diamictons and outwash is by tunnel flow or slow melt out under ice which has decoupled from its bed. If this was the case, oxidation would be a function of postglacial Fe-rich groundwater, rather than subaerial oxidation. If the rejected date is valid, then the entire sequence records: (1) late mid-Wisconsinan, preglacial Empress Formation valley fill (LFA 1), (2) late Wisconsinan Laurentide glaciation (LFA 2-6; with perhaps a minor retreat or ice decoupling in LFA 4), followed by (3) terminal Pleistocene - Holocene postglacial landscape evolution (LFA 7a-7e). Thus, there would be evidence for only one Laurentide ice advance in this region as well as further north (Liverman *et al.* 1989; Young *et al.* 1994).

Based on landform evidence (Shaw and Kvill 1984; Rains *et al.* 1993; Shaw *et al.* 1996) it has been suggested that a short duration (16-162 days, Shaw 1989), large subglacial megaflood (the Livingstone Lake event) transmitted discharges from the Northwest Territories and northern Saskatchewan through Alberta, Montana, the Missouri-Mississippi drainage basins, Mississippi delta and the Gulf of Mexico (Rains *et al.* 1993; Shaw *et al.* 1996). This fluvial event is believed to have scoured a path directly through the study area, resulting in: giant flutings, some drumlins, tunnel channels, scoured bedrock tracts and the exposure of Cretaceous bedrock allowing the postglacial development of solonetzic soils. D. Sjogren and M. Munro (north and south of the study area respectively) are investigating the effect of subglacial sheet flood erosion on moraine fields (D. Sjogren and M. Munro, pers. comm. 1996). As the timing of Evans and Campbell's (1992) LFAs, and the timing of the Livingstone Lake event are uncertain, just where (or if) in the Evans and Campbell (1992) LFA sequence the Livingston Lake event occurred is unknown. Rains *(et al.* 1993) suggest that the Livingstone event occurred sometime *ca.* 21,650-17,830 cal yr B.P. (18,000-15,000 yr B.P.); and, J. Shaw (pers. comm. 1996) suggests that the event would have been contemporaneous with maximum glacial expansion and the first meltwater spike in the Gulf of Mexico, *ca.* 21,650-18,780 cal yr B.P. (18,000-16,000 yr B.P.) (Leventer *et al.* 1982; Kennett *et al.* 1985; (Dyke and Prest 1987).The above ^{14}C yr B.P. dates have been calibrated to cal yr B.P. (before A.D. 1955) using CALIB Rev 3.0.3 (Stuiver and Reimer 1993). CALIB Rev 3.0.3 assigns probabilities of the calibrated date falling into various intervals, and the date range reported here is that which yields a 100% probability when 1 standard deviation is used.

1.3.3 DEGLACIAL

As the ice sheet decayed in a north-eastward direction, lakes which were formerly partially ice-bordered decanted into lower level lakes further east (for example. St.-Onge 1972; Quigley 1980; Shetsen 1987; Teller 1987; Vreeken 1989; Paterson 1996) (Figure 1.2). Based on elevations, stratigraphy and sedimentology, Paterson (1996) reconstructed the evolution of Glacial Lake Bassano. Glacial Lake Bassano acted as the regional base level for all discharge from the western margin of the Laurentide ice sheet around Grande Prairie (a distance >1000 km) and the Cordilleran areas now drained by the North Saskatchewan, Red Deer and Bow rivers, from its establishment at *ca.* 945 m until ice retreat allowed it to drain at *ca.* 690 m. The height of the lake was regulated primarily by ice barriers and the regional topography. The initial lake margin was constrained by the Buffalo Lake moraine to the west, the Lethbridge moraine to the south, the Suffield moraine to the east, and the Laurentide ice sheet to the north-east. Ice retreat exposed progressively lower outlets further north; when the outlet dropped below 747 m, the area south of the Brooks divide (Glacial Lake Tilley) was abandoned, and the divide became the new southern lake margin. Glacial Lake Bassano eventually emptied with the deepening of the Red Deer valley at 690 m (Broad Valley Stage, Bryan *et al.* 1987). Sediments brought into the Bassano basin were of two distinct types: (1) silt sized sediments from Glacial Lake Drumheller deposited during the early Crawling Valley phase; and (2) coarser sand sized sediments following the rapid, high discharge drainage of Glacial Lake Drumheller, that were deposited during the later Red Deer Valley phase (Paterson 1996).

It is tempting to relate all the Glacial Lake Bassano complex to LFA 7a (postglacial proglacial lake) of Evans and Campbell (1992). However, Shetsen (1987) indicates a number of eskers angling into Crawling Valley, indicating that the valley was subglacial at some point in its history. Other valleys in the area also have incurving eskers, and many show uphill flow directions indicative of subglacial drainage (interpreted from Shetsen 1987). In any event, it appears likely that some degree of subglacial activity was responsible for modifying the topography of the area.

As the ice margin and its associated proglacial lakes receded, the Red Deer River valley was incised (or reincised where following the older preglacial Calgary valley) (Farvolden 1963; Carlson 1970; Evans and Campbell 1995; Pawlowicz and Fenton 1995). Bryan *et al.* (1987) suggest the initial postglacial incision proceeded in two steps: first, immediate postglacial meltwater released on a wide front combined with proglacial lake drainage removed surficial materials over an extensive area in and around Dinosaur Provincial Park; this flow formed the "broad" valley stage. Later, continued meltwater and meteoric water inputs caused incision of the narrower valley.

McPherson (1968) produced the first historical synthesis of the lower Red Deer River Valley, which continues to be tested and refined (for example O'Hara 1986; Bryan *et al.* 1987; Evans and Campbell 1992, 1995; O'Hara and Campbell 1993; Barling 1995). Based on surficial mapping (by airphoto and field verification), well logs and borehole data, McPherson (1968) noted five stages of postglacial development. During Stage 1, late Wisconsinan glacial meltwater cut through *ca.* < 16 m of glacial drift, incising a V-shaped valley into the Cretaceous bedrock some 90-120 m below the prairie surface. Incision was followed by aggradation (Stage 2) as meltwater deposited sand and gravel up to 46 m above the present Red Deer River channel. Alluvial fans began to form during this second stage. In Stage 3, trenching of the sand and gravel deposited in Stage 2 occurred to a depth of 30 m or more below the modern floodplain; this is believed to have occurred sometime in the early postglacial, possibly due to climatic change and / or isostatic readjustment. The Klassen site (see Chapter 4) is described by McPherson (1968) as a terrace remnant from this period. During Stage 4, the aggrading Red Deer River deposited up to 46 m of fine grained alluvium. Evidence of Stage 5 occurs east of the study area, with river downcutting dominating the last several thousand years.

There are three main difficulties with this interpretation: (1) there is little or no temporal control; (2) there were relatively few well / borehole logs available at the time; and, (3) the deepest sections on which the reconstruction is based are from areas where the preglacial Calgary valley cross-cuts the present Red Deer River valley - suggesting possible confusion of preglacial and postglacial gravels (see Evans and Campbell 1995).

Based on the sedimentology, stratigraphy and paleoecology of 29 stratigraphical logs from buried valleys of the lower Red Deer area, Evans and Campbell (1995) describe a five stage evolutionary model. Initially a pre-ice advance coulee with locally derived fill developed (Stage 1). Stage 2 records evidence of proglacial advance lacustrine fill followed by glacial till deposition. Subglacial / interglacial downcutting in

Stage 3 is characterized by channel cutting followed by deposition of gravels and sands. Stage 4 records postglacial lake deposition, followed by the formation of surface lag by spillway sheetfloods in Stage 5.

1.3.4 POSTGLACIAL

Paleoevegetation, primarily reconstructed from pollen, has traditionally formed the basis of most postglacial paleoenvironmental change models in North America. Unfortunately, there is a paucity of paleovegetation records from the study region. This problem is exacerbated by: (1) the ephemeral nature of most of the regional lakes resulting in truncated or restricted records; (2) high lake area variability (C. Campbell *et al.* 1994 - Chapter 3) which can redeposit old material; and, (3) poor dating control. Furthermore, grassland pollen spectra tend to be poor proxies for paleoclimate, except at a crude scale (Ritchie 1987; but see McAndrews 1966). This is shown in the pollen from the oldest dated record in the Canadian plains, the Horseman site in southern Saskatchewan (*ca.*17,330-10,380 cal yr B.P; 14,340 ± 100 - 9500 ± 80 ^{14}C yr B.P., TO-310, GSC-4098, Klassen 1994; calibrated after Stuiver and Reimer 1993). The pollen at this site has been interpreted by MacDonald as indicating open vegetation dominated by sage and grasses throughout the early postglacial (in Klassen 1994). However, ostracods from the same site were interpreted by Delorme as having been deposited in conditions similar to but cooler than present (in Klassen 1994). An analogous situation is exhibited at Chappice Lake (Vance 1991; Vance *et al.* 1992, 1993) where halophyte macrofossils proved to be useful in determining relative water levels and thus relative paleoclimate, while pollen showed little variation. Although a number of postglacial climatic changes appear to have occurred, prairie pollen spectra do not appear to have been altered in any significant way by these changes (Vance 1991).

Barnosky *et al.* (1987), Barnosky (1989), Schweger and Hickman (1989), Sauchyn (1990), Sauchyn and Sauchyn (1991), Beaudoin (1992, 1993, 1995, 1996), Beaudoin *et al.* (1996) and Last and Sauchyn (1993) note that two major climatic trends have dominated the postglacial history of northwestern North America. The first trend is decreasing humidity which occurred following deglaciation until the early to mid-Holocene. The second period is dominated by greater effective moisture *ca.* mid-Holocene - present.

Beaudoin *et al.* (1996) proposed the following model of postglacial vegetation development for a transect moving away from the Laurentide ice from southern Saskatchewan, through southern Alberta to northern Montana, *ca.* 12,970-9960 cal yr B.P. (11,000-9000 B.P.):

"(a) abundant residual ice and proglacial lakes; (b) a belt of open spruce forest in newly deglaciated terrain beyond the ice margin; (c) hummocky terrain with abundant kettles and melting ice with perennial wetlands, surrounded by aspen; (d) perennial wetlands, probably surrounded by open ? grasslands, perhaps with some aspen in particularly sheltered or moist locations; and, (e) open parkland or grassland terrain." This period was followed by a dry "Hypsithermal," characterised by drying and sediment redistribution, followed by increasing moisture and cooler temperatures in the late Holocene (Beaudoin *et al.* 1996).

A composite paleoclimatic record based on the lithostratigraphic, plant macrofossil and pollen records from Chappice Lake (lat 50°10'N, long 110°22'W; Vance 1991; Vance *et al.* 1992; Vance and Wolfe 1996) and lithostratigraphy and pollen from Harris Lake (in the Cypress Hills, Saskatchewan; Sauchyn 1990; Sauchyn and Sauchyn 1991; Last and Sauchyn 1993) has been developed by Vance and Wolfe (1996). Vance and Wolfe (1996) suggest that extremely arid conditions, prevalent through the mid-Holocene (*ca.* 8480-6800 cal yr B.P; 7700-6000 ^{14}C yr B.P.), were replaced by less severe conditions (but more arid than present) from *ca.* 6860-5050 cal yr B.P. (6000 to 4500 ^{14}C yr B.P.). Increasing effective moisture is indicated by *ca.* 5740-5730 cal yr B.P. (5000 B.P. ^{14}C yr B.P.) and is succeeded by the onset of cool and moist conditions characteristic of recent climate by *ca.* 3450-3380 cal yr B.P. (3200 ^{14}C yr B.P.) at Harris Lake. At Chappice Lake, decreasing mid-Holocene aridity (beginning *ca.* 5250-5050 cal yr B.P.; 4500 ^{14}C yr) was followed by a period of peak effective moisture *ca.* 2780-930 cal yr B.P. (2700-1000 ^{14}C yr B.P.). A brief reversion to more arid conditions, *ca.* 980-680 cal yr B.P. (1100-800 B.P.), was followed by a return to cool and moist conditions *ca.* 546-20 cal yr B.P. (550-150 ^{14}C yr). The historic period (beginning *ca.* 1880 A.D.) has been more arid, in general reflecting historically recorded drought events of the 1880s, 1920s, 1930s and 1980s (Vance and Wolfe 1996). Vance and Wolfe (1996) suggest that transfer functions developed on pollen records from Lofty Lake in central Alberta and Riding Mountain in southwestern Manitoba may be used to assess the magnitude of Holocene climatic changes. They propose that *ca.* 9990-3160 cal yr B.P. (9000-3000 B.P.) there was a *ca.* 1.5-3°C increase in growing season temperature, with an estimated growing season precipitation deficit of 50 mm. Zoltai and Vitt (1990) presented similar estimates based on changes in the distribution of peatlands in the western Canadian interior, and suggest that prior to 6860-6800 cal yr B.P. (6000 B.P.), mean July temperature was about 0.5°C warmer, and mean annual precipitation was 65 mm lower than present. The above scenario (in various earlier versions, for example Vance *et al.* 1992, 1993, 1995), has formed the working environmental model for a number of research projects such as O'Hara and Campbell (1993) and the Palliser Triangle Integrated Research and Monitoring Area (IRMA) project (Lemmen *et al.* 1993).

1.4 CALIBRATION, REVIEW AND GEOMORPHIC IMPLICATIONS OF POSTGLACIAL RADIOCARBON AGES IN SOUTHEASTERN ALBERTA, CANADA

A version of section 1.3 has been published.
Campbell, C., and Campbell I.A. (1997). Quaternary Research, 47:37-44.

1.4.1 INTRODUCTION

Widely accepted models of late Wisconsinan Laurentide deglaciation in central and southern Alberta (for example St.-Onge 1972; Christiansen 1979; Dyke and Prest 1987; Klassen 1989) generally posit ice-retreat phases from a glacial maximum in northern Montana (Fullerton and Colton 1986) which is assumed to date to *ca.* 18,000 ^{14}C yr B.P. (Dyke and Prest 1987). Ice-retreat phases are limited by a widely scattered group of deposits with radiocarbon dates, few of which are corrected for δ^{13}C fractionation or calibrated. These variations hinder accurate sequencing of dated events, and their comparison with sequences in other regions or with theoretical models.

Radiocarbon time deviates significantly from calendar time (Stuiver *et al.* 1986; Bard *et al.* 1990; Bard *et al.* 1993;). Variations in radiocarbon concentration in the atmosphere have led to 'time warps' (Bard *et al.* 1993) which preclude a simple linear conversion from radiocarbon age to calendar age (Figure 1.3b). Atmospheric radiocarbon concentration variations may be caused by secular variations in geomagnetic intensity governing the rate of ^{14}C production, and variations in the rate of deep-sea venting of old carbon (Olsson 1970; Stuiver *et al.* 1991; Stuiver and Braziunas 1993). Calibration of radiocarbon dates is necessary in order to: (1) determine true timing of events; (2) compare radiocarbon dates with dates obtained from other dating techniques such as TL, OSL, and dendrochronology; (3) establish meaningful regional chronologies; and (4) relate dated events to possible causal processes including insolation variations produced by variations of the Earth's orbital elements (Milankovitch cycles; Lowell and Teller 1994; Bartlein *et al.* 1995). Calibration procedures developed by Stuiver and Reimer (1993; program CALIB Rev 3.0.3) have made possible the calibration of radiocarbon ages over the past 18,400 ^{14}C yr B.P.

In this paper, calibrated radiocarbon and other dates of geomorphic events from Dinosaur Provincial Park, southeastern Alberta, and the surrounding plains (Figure 1.4), are compared in order to establish the calendric timing, duration and sequence of deglaciation and postglacial events. Such a reassessment is a necessary first step towards developing a more comprehensive understanding of the spatial and temporal sequences of postglacial landscape evolution in this region.

1.4.2 STUDY AREA

The study area is in the northern Great Plains, it is located in a chernozem-soil dominated mixed-grass prairie (Strong and Leggat 1992). Climatically the area is midlatitude, semiarid, and continental with long cold winters and short warm summers (Environment Canada 1993). The regional mean annual temperature is *ca.* 3.8°C, with typical mean values of -13°C for January and 19°C for July (Environment Canada 1993). Winds come dominantly from the west and southwest and winter temperatures are often ameliorated by chinook winds (Walmsley and Morris 1993). Regional mean annual precipitation ranges from about 300 to 400 mm, with the maximum in summer and the minimum in winter (Environment Canada 1993). Potential evapotranspiration often exceeds precipitation by >300 mm / yr (Winter 1989). Historically the area has been subjected to recurrent droughts associated with the development of stable high pressure ridges that displace cyclonic tracks, moist air masses, and fronts northward (Dey and Chakravarti 1976). Early and mid-Holocene arid phases may have resulted from different synoptic patterns (Vance 1984).

Regional drainage and topography are regulated by horizontally bedded, poorly consolidated late Cretaceous formations underlying the flat to undulating prairie surface. The postglacial geomorphology of the region primarily reflects the interaction of fluvial and aeolian processes on unconsolidated surficial deposits derived principally from glacial, glaciofluvial, and glaciolacustrine sources (for example, Evans and Campbell 1992). Since extensive European settlement in the late 1800s, there has been widescale modification of the prairie landscape by wheat growing, irrigation agriculture and ranching. As in other semiarid terrains this region is particularly sensitive to fluctuations in climate (particularly precipitation), but attempts to correlate climate change and geomorphic response are problematic because of the largely unknown effects of factors such as local baselevel changes and complex responses in drainage basin systems (Rains and Welch 1988; O'Hara and Campbell 1993).

Since deglaciation two major climatic phases have been identified in northwestern North America. The first was towards increasing aridity, peaking in the early Holocene, followed by one of increasing effective moisture (Barnosky *et al.* 1987). These major climatic trends appear to be closely linked to orbitally induced variations in insolation (Schweger and Hickman 1989; Thompson *et al.* 1993).

Figure 1.3a shows summer and winter insolation values for 50°N latitude for the last 25,000 years interpolated from Berger (1978). The difference between winter and summer insolation was at a minimum *ca.* 25,000 years ago., with deviations of solar radiation from their A.D. 1950 (0 yr B.P.) values of *ca.* 4.5 $W.m^2$ in winter and -5.8 $W.m^2$ in summer. At about 12,000 years ago, the seasonality of insolation peaked, with deviations of solar radiation of *ca.* -21.8 $W.m^2$ in winter and 17 $W.m^2$ in summer, followed by diminishing seasonality to the present. In the northern Great Plains, this sequence appears to have caused, relative to modern conditions, very cold dry winters and very hot dry summers by *ca.* 12,000-10,000 years ago. There was then a gradual change to the relatively cool moist winters and warm moist summers experienced in the region at present (Kutzbach and Guetter 1986; Schweger and Hickman 1989). Paleoecological investigations in central Alberta (Schweger and Hickman 1989) found a general lowering of lake levels *ca.* 12,000-6000 ^{14}C yr B.P., indicating increased moisture deficits. Following this early Holocene aridity, lakes began to fill.

1.4.3 POSTGLACIAL GEOMORPHIC CHRONOLOGY AND REGIONAL CORRELATIONS

Figure 1.3c shows geomorphic events by calibrated date. Selected dated events from sites adjacent to the study area, but still located in the northern Great Plains, are included to place the study area within its wider context. All ^{14}C yr B.P. dates used have been corrected for $\delta^{13}C$ fractionation (unless otherwise noted with an * in the text and a grey line in Figure 1.4) and have been calibrated to cal yr B.P. (before A.D. 1955) using CALIB Rev 3.0.3 (Stuiver and Reimer 1993). CALIB Rev 3.0.3 assigns probabilities of the calibrated date falling into various intervals, and the date range reported here is that which yields a 100% probability when 1standard deviation is used. In the text, sites / dates are referred to by the site number in Figure 1.3 in brackets. While evidence of more recent phenomena is more abundant, groupings of dominant geomorphic event-related deposits appear to form a distinct pattern over time.

Dyke and Prest (1987) suggest that deglaciation commenced in Montana at or close to the time of maximum ice advance *ca.* 21,650-21,310 cal yr B.P. (18,000 ^{14}C yr B.P.). If this is correct, then it occurred at a time when seasonal insolation patterns were almost identical to present (Berger 1978). The retreat of the late Wisconsinan Laurentide ice was accompanied by the formation of extensive, proglacial, ice-dammed lakes

(Stalker 1963, 1969, 1983; St.-Onge 1972; Dyke and Prest 1987; Kehew and Teller 1994). The oldest basal dates from proglacial lacustrine sediments calibrate to as early as 20,210-19,430 cal yr B.P. in Dinosaur Provincial Park (#1) and 17,080-15,450 cal yr B.P. in southern Saskatchewan (#2, #3). The validity of the date from Dinosaur Provincial Park has been questioned as it was believed to be too old to be deglacial (Evans and Campbell 1992).

Ice disintegration and recession, perhaps associated with ice dam failures, drained the proglacial lakes which decanted, forming numerous large meltwater channels (Kehew and Lord 1986; Bryan *et al.* 1987; Teller 1987). Glacial meltwater decanted down the Missouri River into the Mississippi River and finally into the Gulf of Mexico. Oxygen isotope analyses of planktonic foraminifera from the Gulf of Mexico record a negative isotopic anomaly *ca.* 16,500-16,000 ^{14}C yr B.P. (19,560-19,290 cal yr B.P. to 18,972-18,783 cal yr B.P.), attributed to Laurentide meltwater discharge (for example Leventer *et al.* 1982; Kennett *et al.* 1985).

Pond sediments over glacial lake deposits dated at 18,960-17,280* cal yr B.P. in southern Saskatchewan (#4, #5), show that water may have remained in localized depressions within glacial lake basins after proglacial lake drainage.

The oldest dates for river terraces in the major river valleys are: 18,260-12,910 cal yr B.P.* (#6, #8) on the South Saskatchewan River, and 13,740-11,880 cal yr B.P. (#7, #9) for the North Saskatchewan River. These dates are considered minimum ages, and therefore postdate, perhaps by some considerable time, the initial period of proglacial lake drainage.

In the Dinosaur Provincial Park region Bryan *et al.* (1987) suggest that as the ice margin disintegrated, meltwater supply to local drainage rapidly declined. However, continued inputs of meltwater from upstream (possibly accentuated by glacioisostatic rise) resulted in deep valley incision. McPherson (1968) identified gravel deposits associated with this phase of downcutting, which lie up to 40 m below the bed of the present Red Deer River. Waning discharges heralded valley aggradation and the formation of alluvial fans along the lower Red Deer Valley (McPherson 1968), and its tributaries in Dinosaur Provincial Park (O'Hara and Campbell 1993). Other deposits and events contemporaneous with ice disintegration, recession, and the development of the oldest dated river terraces include: (1) peat developed in a hummock from 17,330-17,050 to 12,220-11,230 cal yr B.P. in southern Saskatchewan (#12-#15); (2) valleys began to infill 13,660-13,120 cal yr B.P. in southern Saskatchewan (#26) (and continue to do so; Klassen 1993) and 12,220-11,230 cal yr B.P. in Alberta (#27); (3) sand dune activity started as soon as the proglacial sediments (outwash plains, deltas, lake basins, etc.) became exposed to the winds following deglaciation (David 1993); and (4) in central Alberta, at Lofty Lake *ca.* 11,400$\pm$190* ^{14}C yr B.P. (GSC-1049; Lichti-Federovich 1970; 13,400-13,000* cal yr B.P.) the development of lake sediments in an abandoned meltwater channel in central Alberta suggests that there was no glacial meltwater left to drain by this time.

From 12,000-10,000 years ago, a period of landscape stability was associated with maximum postglacial aridity in the Plains (Schweger and Hickman 1989). This phase was punctuated by short-term climatic perturbations. Geomorphic evidence of aridity includes: (1) the most strongly developed regional paleosols (#16-#19; Lowden *et al.* 1971; Turchenek *et al.* 1974; Valentine *et al.* 1987; David 1993 - but see Pennock and Vreeken 1986); (2) decreased fluvial erosional and depositional activity (Waters and Rutter 1984); (3) a period of non-deposition in formerly active floodplains in western Alberta (Waters and Rutter 1984); (4) alluvial fan accumulation in Dinosaur Provincial Park (O'Hara and Campbell 1993); and (5) drying of lake basins in central Alberta (Schweger and Hickman 1989). However this period of aridity included several reversals towards moister conditions. Vreeken (1989) notes that in the Lethbridge area six soil development-loess deposition cycles occurred between the deposition of Glacier Peak tephra, 13,460-12,980 cal yr B.P. (#98; Mehringer *et al.* 1977) and Mazama tephra, 7800-7480 cal yr B.P. (#99; Bacon 1983).

Around 10,000 years ago until just after the deposition of Mazama tephra (#99), there was a trend towards regionally moister climate, associated with decreasing extremes in winter and summer insolation (inferred from Berger 1978). Wetter conditions accelerated runoff resulting in: (1) terrace development in tributary river valleys (#33-#44); (2) infilling of local lakes (#83,#84, #86, #88); (3) dune stabilization by paleosol development (#23); (4) low terrace development in the North Saskatchewan River (#10); and (5) possible ponding events (#16, #20, #24, #86. #88; Evans and Campbell 1992; Campbell and Evans 1990). This humid episode was followed by a reversal to more arid conditions suggested by: (1) a reduction in the rate of tributary river valley incisions 8000 to 4000 cal yr B.P.; (2) loess deposition in Dinosaur Provincial Park *ca.* 5400±800 T.L. yr B.P. (#82; Bryan *et al.* 1987); and (3) an increase in salinity at Chappice Lake (Vance *et al.* 1993).

By *ca.* 4000 cal yr B.P., effective moisture reached levels similar to those experienced during the historic period. The moistening was not monotonic, but included several reversals towards drier conditions (Vance *et al.* 1992). Geomorphic processes during this period were characterised by increased variability dominated by: (1) repeated cycles of dune activation and stabilization (pedogenic intervals) (Wolfe *et al.* 1995); (2) renewed valley incision / terrace cutting and aggradation (#45-#79; Traynor and Campbell 1989; I.A. Campbell and Evans 1993; Rains *et al.* 1994; Barling 1995); (3) fluctuating lake levels (Vance *et al.* 1992); (4) multiple episodes of valley alluviation and incision (#80, #81; O'Hara 1986; O'Hara and Campbell 1993); (5) repeated cycles of loess deposition and paleosol development (Vreeken 1989; 1993); and (6) multiple episodes of landslides in the Cypress Hills (#89-#97, Goulden and Sauchyn 1986; Sauchyn and Lemmen 1996).

1.4.4 DISCUSSION

The geomorphology of the study area is heavily imprinted by regional changes in fluvial and aeolian activity which, in turn, appear to be linked to orbitally-forced climatic variations. The late glacial and postglacial period can be divided into four major episodes, each dominated by a particular set of geomorphic processes: glaciation, deglaciation, landscape stability and landscape instability.

The first period, glaciation (erosion and deposition by ice and subglacial meltwater), occurred prior to *ca.* 20,000 cal yr B.P. and does not appear to have lagged significantly behind the insolation seasonality minimum as the uncalibrated dates would imply. Young *et al.* (1994) report preglacial radiocarbon dates as young as 21,330±340 ^{14}C yr B.P. (26,290-25,630 cal yr B.P.; calibrated after Bard *et al.* 1990) indicating a relatively short duration of ice cover in southern Alberta.

The second period, deglaciation (erosion and deposition by water), occurred *ca.* 20,000-12,000 cal yr B.P. and was associated with increasing but not yet maximum seasonality of insolation. While in agreement with the timing of glacial maximum suggested by calibration of Dyke and Prest's (1987) date of *ca.* 21,650-21,310 cal yr B.P., calibration of the deglacial radiocarbon dates indicates no lag between glacial maximum and the onset of rapid deglaciation, requiring a reassessment of the timing, magnitude and sequence of late-glacial and postglacial events.

Calibration indicates that late Wisconsinan deglaciation commenced in northern Montana *ca.* 21,650-21,310 cal yr B.P. (Dyke and Prest 1987) and in the study area, perhaps as early as 20,210-19,430 cal yr B.P. (#1). This would allow, for a simple uniform model, about 1500-2000 years for the ice mass to thin and disappear over a latitudinal distance of *ca.* 350 km, retreating at an average rate of 1 km every 5 years.

If the calibrated dates are correct, much, if not most of the Laurentide ice sheet west of what later became the Glacial Lake Agassiz basin (Lowell and Teller 1994) disintegrated and disappeared from northern Montana, southern Alberta and southern Saskatchewan very rapidly, possibly in as little as *ca.* 1500-2000 years. Such a catastrophic collapse is consistent with the model of a very thin ice sheet in southern Alberta (Mathews 1974; Clayton *et al.* 1985; Dyke and Prest 1987; Rains *et al.* 1990). It is also consistent with the model of deglaciation in which large subglacial flood events (Shaw 1989; Rains *et al.* 1993; Shaw *et al.* 1996) lead to rapid melt, or vice versa.

The apparent long duration of the deglaciation period described here is an artifact of the inclusion of several major rivers in the study area, all of which have their headwaters in the Rocky Mountains or the foothills. Deglaciation of the study area itself was probably complete prior to 15,000 cal yr B.P. as indicated by dates on proglacial lakes; snow-fed and precipitation induced runoff from the headwater areas (Hickman and Schweger 1993) continued to influence the geomorphic development of the study area until *ca.* 13,000-12,000 cal yr B.P.

The third period, landscape stability (minimal deposition or erosion), occurred *ca.* 12,000-10,000 cal yr B.P. and was associated with maximum seasonality of insolation and thus peak postglacial aridity. This period was characterised by the most strongly developed regional paleosols, decreased fluvial erosional and depositional activity, a period of non-deposition in formerly active floodplains, alluvial fan accumulation and the drying of lake basins.

The final period was one of landscape instability, extending from *ca.* 10,000 cal yr B.P. to present. It was associated with decreasing seasonality of insolation. Geomorphic processes in semiarid regions like the study area operate within narrow sets of threshold conditions in which relatively small changes in environmental factors may induce disproportionately large responses (Schumm 1973; Bull 1991). The multiplicity of alternating phases of channel incision and aggradation and dune activation and stabilization, which are evident across the study area, provide some indication of these fluctuations.

Calibration of ^{14}C dates suggests that there are: (1) a closer temporal association between insolation and geomorphic trends than previously thought; and, (2) a need to re-examine the temporal scale and hence the rates of postglacial geomorphic processes. The postglacial pre-Holocene period in the northern Great Plains was roughly as long as the Holocene itself, *ca.* 10,000 years. This fact has profound implications; for example, there was a much larger temporal window than has been generally assumed for postglacial megafauna extinction, vegetation recolonization of deglaciated surfaces and various geomorphological events. Finally, atmospheric general circulation model simulations of paleoclimates, which use insolation variations based on calendric dates but other boundary conditions based on radiocarbon years, will need to be reassessed.

1.5 CHAPTER 1 - FIGURES

FIGURE 1.1

Locations of key sites analysed in this thesis (open square).

A. Mean January temperature (modified after Vance *et al.*, in press).

B. Mean July temperature (modified after Vance *et al.*, in press).

C. Mean annual precipitation (modified after Vance *et al.*, in press).

D. Mean annual potential evapotranspiration (modified after Vance *et al.*, in press).

FIGURE 1.2

Schematic diagram of the proglacial lake system in southern Alberta during the late Pleistocene showing the generalized lake areas superimposed over the South Saskatchewan drainage system (after Paterson 1996). Glacial Lake = G.L.

FIGURE 1.3

3A. Summer and winter insolation values for 50°N latitude for the last 25,000 years (Berger 1978).

3B. Relationship between cal yr B.P and ^{14}C yr B.P.(produced using file intcal93.14c, CALIB Rev 3.0.3 [Stuiver and Reimer 1993]).

3C. Dated postglacial geomorphic deposits in Dinosaur Provincial Park, southeastern Alberta and adjacent areas (all dates in calendar years). Black bars represent the 100% calibrated date range at 1σ. Dates uncorrected for $\delta^{13}C$ fractionation are indicated by a gray line. The numbers in brackets following the geomorphic event type (right hand column) refer to site numbers on Figure 1.4.

Proglacial Lake: 1. AECV-681C (Evans and Campbell 1992); 2. GSC-4675 (Vreeken 1989); 3. TO-694 (Vreeken 1989);

Collapsed Pond Deposits Over Glacial Lake Deposits: 4. S-300B (Christiansen 1979); 5. S-300A (Christiansen 1979);

Primary River (Meltwater Channel) Terraces: 6. GSC-1199 (Lowden and Blake (1975); 7. S-2385 (Rains and Welch 1988); 8.GSC-805 (Lowden and Blake 1968); 9. S-1923, 10. S-1706, 11. S-1799 (Rains and Welch 1988);

Robsart (Horseman) Site Peat: 12. TO-310, 13. GSC-4270, 14. GSC-4266, 15. GSC-4098 (Klassen 1993);

Paleosols: 16. GSC-1061 (Lowden *et al.* 1971) (Evans and Campbell [1992] suggest that this may be a pond rather than a paleosol deposit); 17. S-460 (Rutherford *et al.* 1975) (paleosol developed on loess); 18. S-442 (Turchenek *et al.* 1974) (paleosol developed on loess); 19. S-461 (Rutherford *et al.* 1975) (paleosol developed on loess); 20. GSC-1341 (Lowden *et al.* 1971) (Evans and Campbell [1992] suggest that this may be a pond rather than paleosol deposit); 21. GSC-1819 (In: Waters and Rutter 1984 - from Harrison 1973); 22. GSC-1944 (In: Waters and Rutter 1984 - from Reeves 1973); 23. S-443 (Turchenek *et al.* 1974) (paleosol developed on dune); 24. GSC-1302 (Lowden *et al.* 1971) (Evans and Campbell [1992] suggest that this may be a pond rather than a paleosol deposit); 25. TO-5478 (Barling 1995);

Valley Aggradation: 26. S-2932 (Klassen 1993); 27. AECV-1396C (Beaudoin 1992); 28. S-2911, 29. S-2820, 30. S-2821, 31. S-2819, 32. S-2930 (Klassen 1993);

Secondary River Terraces: 33. GSC-236 (In: Waters and Rutter 1984 - Dyke *et al.*

1965); 34. S-1798, 35. S-1926, 36. S-1787 (Rains and Welch 1988); 37. TO-1829 (Rains *et al.* 1994); 38. S-1789 (Rains and Welch 1988); 39. GSC-1735 (Lowden and Blake 1975); 40. S-1788, 41. S-2387, 42. S-1800, 43. S-2392, 44. S-1797, 45. S-1795, 46. S-2389, 47. S-1794, 48. S-1796, 49. S-2388 (Rains and Welch 1988); 50. AECV-1276C (I.A. Campbell and Evans 1993); 51. AECV-904C (I.A. Campbell and Evans 1993); 52. AECV-684C (Traynor and Campbell 1989); 53. AECV-911C (I.A. Campbell and Evans 1993); 54. AECV-9440 (Rains *et al.* 1994); 55. AECV-906C (I.A. Campbell and Evans 1993); 56. S-2390 (Rains and Welch 1988); 57. AECV-905C (I.A. Campbell and Evans 1993); 58. S-1783 (Rains and Welch 1988); 59. AECV-913C (I.A. Campbell and Evans 1993); 60. TO-4577 (Barling 1995); 61. S-1785 (Rains and Welch 1988); 62. AECV-682C (Traynor and Campbell 1989); 63. TO-4579 (Barling 1995); 64. AECV-604C (I.A. Campbell and Evans 1990); 65. TO-4576 (Barling 1995); 66. AECV-914C, 67. AECV-80 (I.A. Campbell and Evans 1993); 68. AECV-683C (Traynor and Campbell 1989); 69. AECV-912C (I.A. Campbell *et al.* 1993); 70. S-1793, 71. S-1786, 72. S-1790, 73. S-1784, 74. S-1791 (Rains and Welch 1988); 75. AECV-680C (I.A. Campbell, unpub.); 76. AECV-903C (I.A. Campbell and Evans 1993); 77. S-1792 (Rains and Welch 1988); 78. AECV-1278C, 79. AECV-1277C (I.A. Campbell and Evans 1993);
Channel Fill: 80. AECV-427C (O'Hara and Campbell 1993); 81. S-2546 (O'Hara and Campbell 1993);
Loess: 82. ALPHA-270, Bryan *et al.* 1987; TL date);
Basal Lake Dates / Lacustrine-Ponding Events: 83. S-2908 (Sauchyn 1990); 84. Clearwater Lake (Vance and Last 1994); 85. TO-959 (I.A. Campbell and Evans 1993) (in a series of lacustrine-ponding events); 86.Chappice Lake (Vance and Last 1994); 87. AECV-605.C (I.A. Campbell and Evans 1990) (in a series of lacustrine-ponding events); 88. Elkwater Lake (Vance and Last 1994);
Landslides: 89. Estimate from lake record; 90. TO-4479, 91. TO-4480, 92. S-2772, 93. S-2907, 94. S-2630, 95. S-2631, 96. Tree-ring date, 97. Historic observation date (Sauchyn and Lemmen 1996);
Regional tephra deposits: 98. WSU-1554, WSU-1548 (Mehringer *et al.* 1977) (Glacier Peak tephra bed); 99. W-4288, W-4290, W4255,W-4356,W-4295, USGS-870, (Bacon 1983) (Mazama tephra bed).

FIGURE 1.4

Study area showing locations of dated sites shown in Figure 1.3. The area of Dinosaur Provincial Park is indicated by a circle.

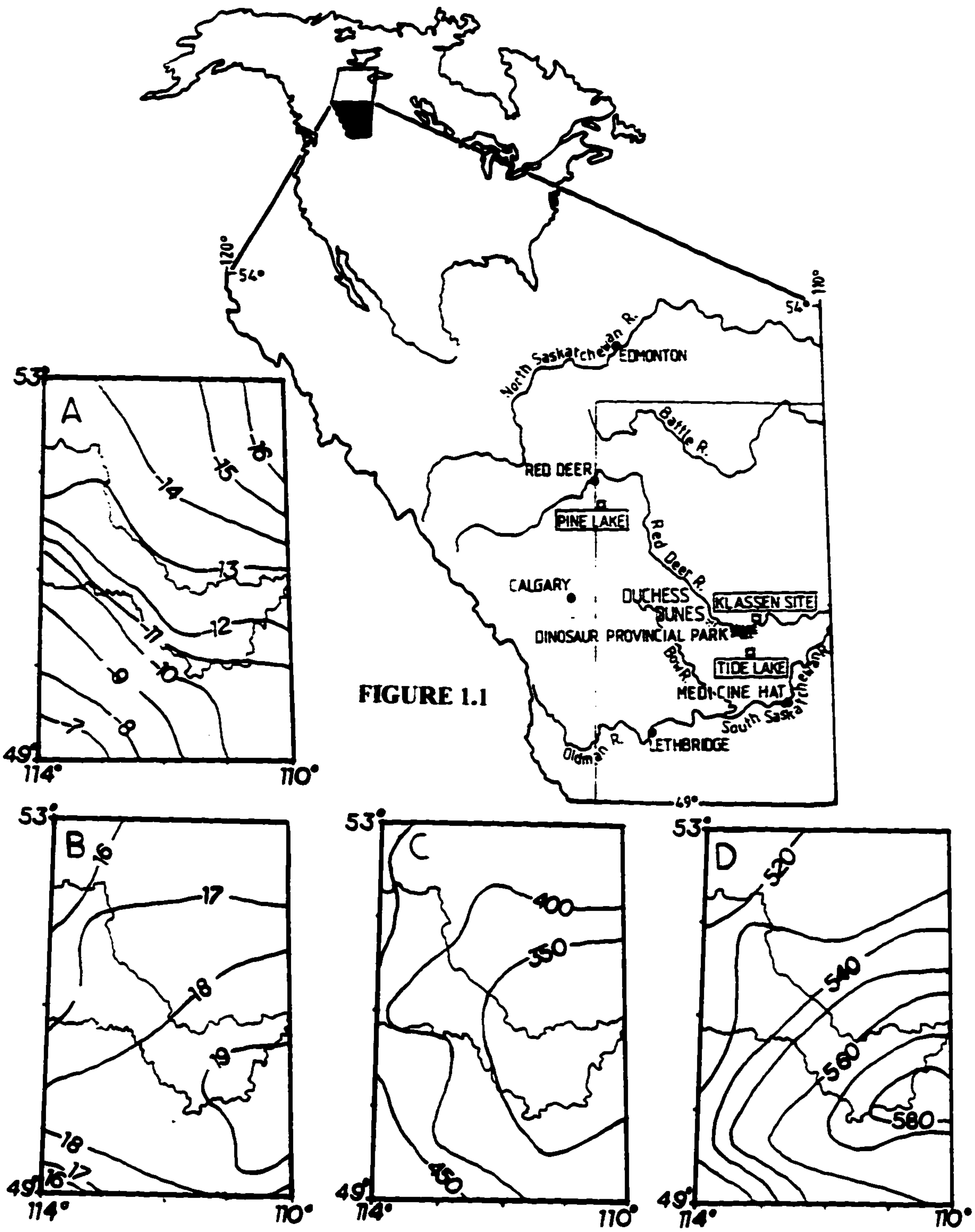
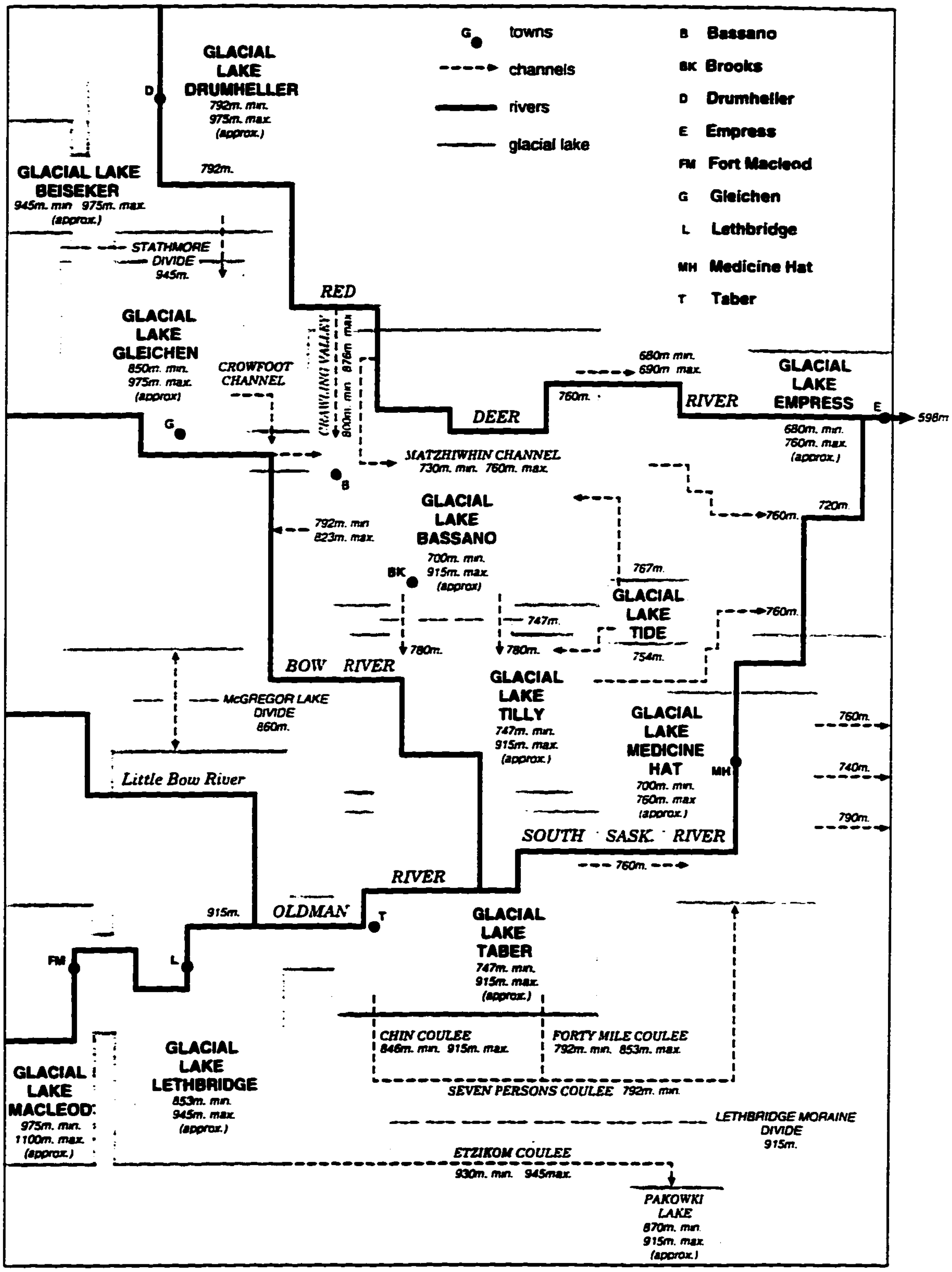

120°
54°
110°
54°
North Saskatchewan R.
EDMONTON
Battle R.
RED DEER
PINE LAKE
Red Deer R.
CALGARY
DUCHESS
DUNES
KLASSEN SITE
DINOSAUR PROVINCIAL PARK
Bow R.
TIDE LAKE
MEDICINE HAT
South Saskatchewan R.
Oldman R.
LETHBRIDGE
49°
A
B
C
D
53°
49°
114°
110°
16
15
14
13
12
11
10
9
8
7
17
18
19
400
350
450
520
540
560
580

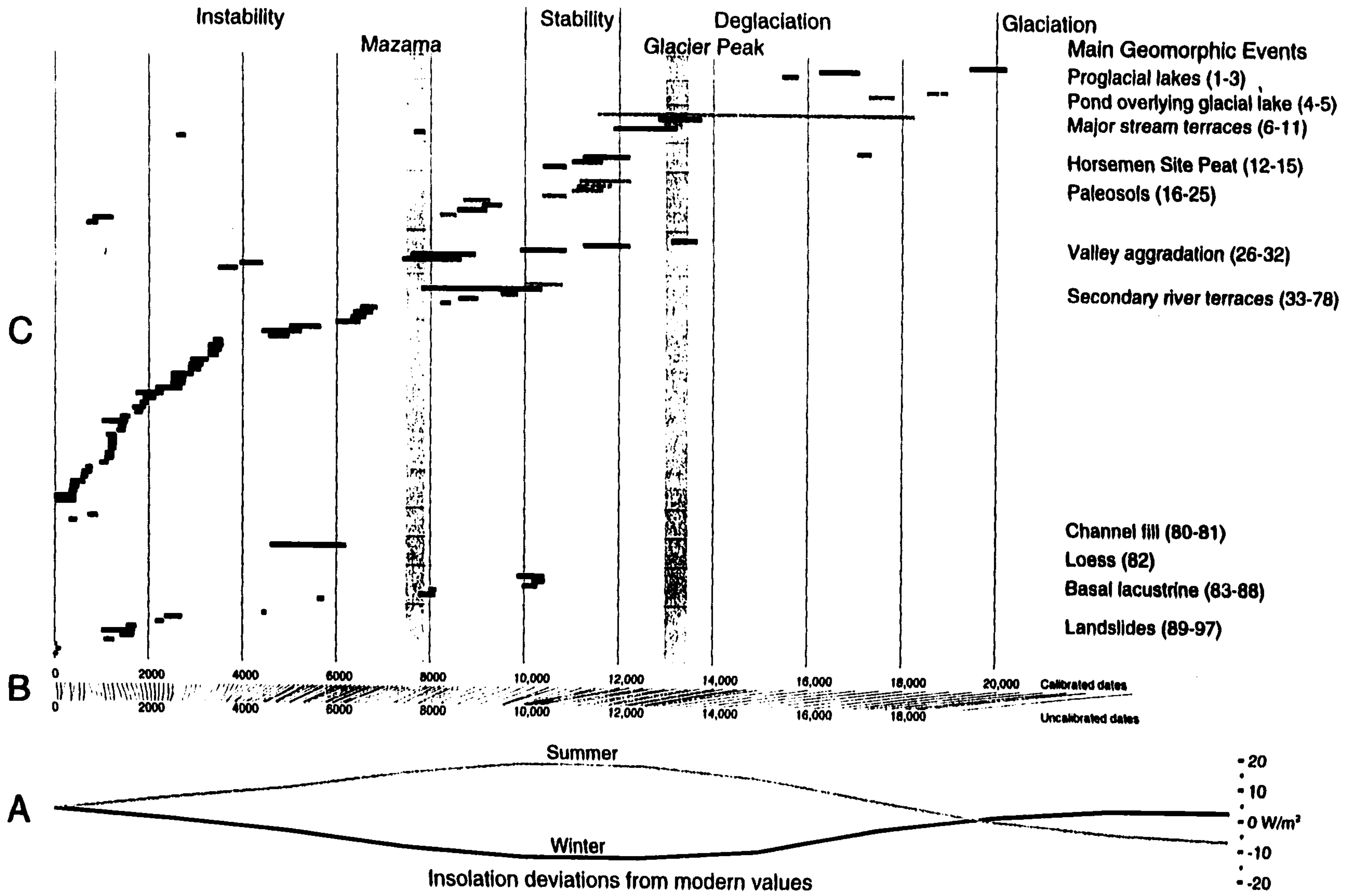
FIGURE 1.1

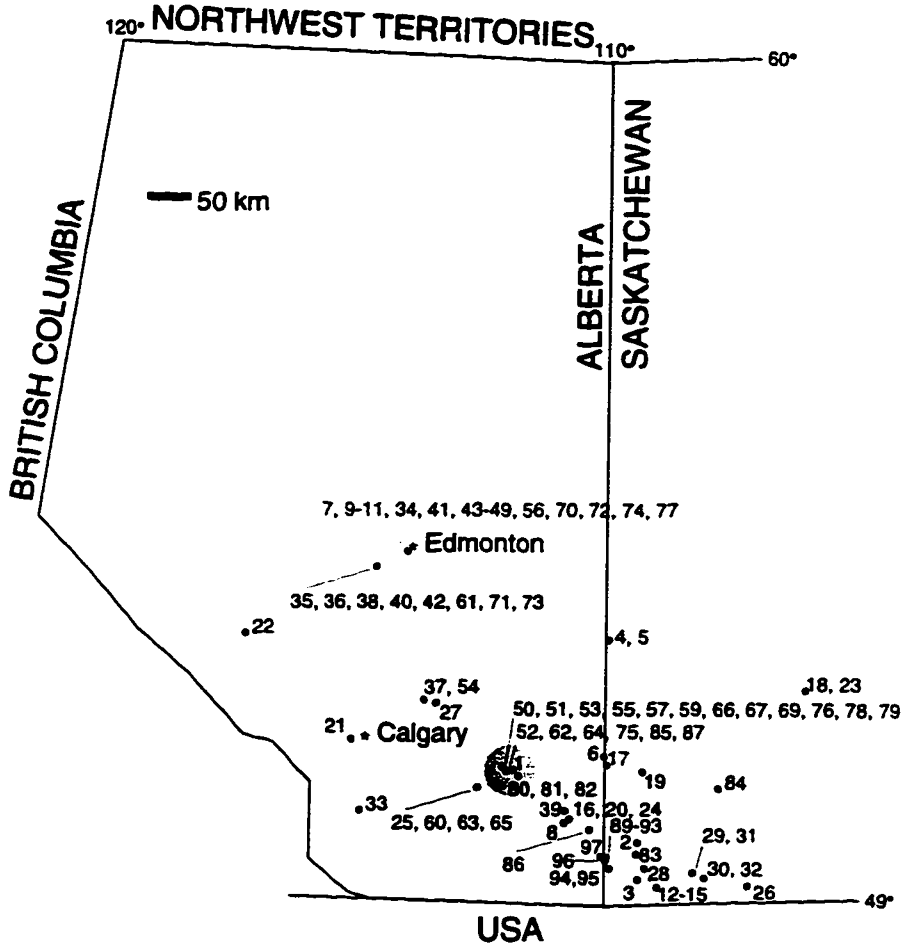
FIGURE 1.2



towns
channels
rivers
glacial lake
B Bassano
BK Brooks
D Drumheller
E Empress
FM Fort Macleod
G Gleichen
L Lethbridge
MH Medicine Hat
T Taber
GLACIAL LAKE DRUMHELLER
792m. min. 975m. max. (approx.)
792m.
GLACIAL LAKE BEISEKER
945m. min 975m. max. (approx.)
STATHMORE DIVIDE
945m.
RED
GLACIAL LAKE GLEICHEN
850m. min. 975m. max. (approx)
CROWFOOT CHANNEL
CRAWLING VALLEY
800m. min 876m. max
680m. min. 690m. max.
760m.
DEER
RIVER
GLACIAL LAKE EMPRESS
680m. min. 760m. max. (approx.)
598m
MATZHIWHIN CHANNEL
730m. min. 760m. max.
GLACIAL LAKE BASSANO
700m. min. 915m. max. (approx)
792m. min 823m. max.
760m.
720m.
767m.
GLACIAL LAKE TIDE
747m.
760m.
780m.
780m.
754m.
BOW RIVER
GLACIAL LAKE TILLY
747m. min. 915m. max. (approx.)
McGREGOR LAKE DIVIDE
860m.
Little Bow River
GLACIAL LAKE MEDICINE HAT
700m. min. 760m. max. (approx.)
760m.
740m.
790m.
SOUTH SASK. RIVER
760m.
RIVER
915m.
OLDMAN
GLACIAL LAKE TABER
747m. min. 915m. max. (approx.)
CHIN COULEE
846m. min. 915m. max.
FORTY MILE COULEE
792m. min. 853m. max.
SEVEN PERSONS COULEE 792m. min.
GLACIAL LAKE LETHBRIDGE
853m. min. 945m. max. (approx.)
GLACIAL LAKE MACLEOD
975m. min. 1100m. max. (approx.)
LETHBRIDGE MORAINE DIVIDE
915m.
ETZIKOM COULEE
930m. min. 945max.
PAKOWKI LAKE
870m. min. 915m. max. (approx.)

Calibrated Regional Chronology Figure 1.3





120° NORTHWEST TERRITORIES 110°
60°
BRITISH COLUMBIA
50 km
ALBERTA
SASKATCHEWAN
7, 9-11, 34, 41, 43-49, 56, 70, 72, 74, 77
Edmonton
35, 36, 38, 40, 42, 61, 71, 73
22
4, 5
37, 54
27
18, 23
50, 51, 53
52, 62, 64
55, 57, 59, 66, 67, 69, 76, 78, 79
75, 85, 87
21
Calgary
6
17
19
84
80, 81, 82
33
25, 60, 63, 65
39
16,
20, 24
89-93
8
97
2
83
29, 31
86
96
94,95
28
30, 32
3
12-15
26
49°
USA

FIGURE 1.4

1.6 CHAPTER 1 - REFERENCES

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APPENDIX 1.1 RADIOCARBON DATES USED IN FIGURE 1.3

Site Lab. #	^{14}C yr BP * Not corrected for $\delta^{13}C$ fractionation	Cal yr BP or other date * Not corrected for $\delta^{13}C$ fractionation	Location	Reference	Material and Comments
Proglacial Lake					
1. AECV-681C	16,790±270	20,214-19,431	Princess DPP, AB	Evans and Campell (1992)	Bone
2. GSC-4675	13,900±340	17,084-16,231	Maple Creek,Sask.	Vreeken (1989)	Shells
3. TO-694	13,120± 80	15,788-15,454	Maple Creek,Sask.	Vreeken (1989)	Shells
Collapsed Pond Deposits Over Glacial Lake Deposits					
4. S-300B	15,850±225*	18,9574-18,505*	Evesham, Sask.	Christiansen (1979)	Carbonaceous silt
5. S-300A	14,670±240*	17,831-17,282*	Evesham, Sask.	Christiansen (1979)	Carbonaceous silt
Primary River (Meltwater Channel) Terraces					
6. GSC-1199	14,200±1120*	18,260-11,532*	Empress, AB	Lowden and Blake (1975)	Bone
7. S-2385	11,345±420	13,739-12,828	N. Saskatchewan R. AB	Rains and Welch (1988)	Bone
8. GSC-805	11,200±200*	13,324-12,906*	Medicine Hat, AB	Lowden and Blake (1968)	Bone
9. S-1923	10,740±470	13,222-11,883	N. Saskatchewan R. AB	Rains and Welch (1988)	Bone
10. S-1706	6,955±80	7,878-7,655	N. Saskatchewan R. AB	Rains and Welch (1988)	Bone
11. S-1799	2,780±85	2,595-2,779	N. Saskatchewan R. AB	Rains and Welch (1988)	Bone
Robsart (Horseman) Site Peat					
12. TO-310	14,340±100	17,330-17,047	Robsart (Horseman),Sask.	Klassen (1993)	Peat
13. GSC-4270	10,000±130	11,635-11,005	Robsart (Horseman),Sask.	Klassen (1993)	Peat
14. GSC-4266	10,200±140	12,217-11,229	Robsart (Horseman),Sask.	Klassen (1993)	Peat
15. GSC-4098	9,500±80	10,863-10,377	Robsart (Horseman),Sask.	Klassen (1993)	Peat
Paleosol					
16. GSC-1061	10,200±240*	12,241-11,149*	Medicine Hat, AB	Lowden *et al.* (1971)	Gastropod shell. Note: Evans and Campbell (1992) suggest that this may be a pond rather than a paleosol deposit
17. S-460	10,060±160	11,835-11,080	S.Saskatchewan River Valley, Sask.	Rutherford *et al.* (1975)	Paleosol developed on loess
18. S-442	9,940±160	11,664-10,985	Beaver Creek, Sask.	Turchenek *et al.* (1973)	Paleosol developed on loess
19. S-461	9,500±150	10,881-10,365	N. of Great Sand Hills,	Rutherford *et al.* (1975)	Paleosol developed on loess
20. GSC-1341	8,120±170*	9,257-8,679*	Medicine Hat, AB	Lowden *et al.* (1971)	Gastropod shell. Note: Evans and Campbell (1992) suggest that this may be a pond rather than paleosol deposit
21. GSC-1819	8,400±150*	9,499-9,094*	Bowfort Road, AB	In: Waters and Rutter (1984) (from Harrison., 1973)	Wood

22. GSC-1944	8,030±200*	9,192-8,572*	Rough Cr., AB	In: Waters and Rutter (1984) (from Reeves, 1973)	Charcoal
23. S-443	7,640±150	8,549-8,191	Beaver Creek, Sask.	Turchenek *et al.* (1973)	Paleosol, developed on dune
24. GSC-1302	1,110±140	1,226-799	Medicine Hat, AB	Lowden *et al.* (1971)	Charcoal. Note: Evans and Campbell (1992) suggest that this may be a pond rather than a paleosol deposit
25. TO-5478	8,40±100	898-668	Matzhiwan Creek, AB	Barling (1995)	Bone
Valley Aggradation					
26. S-2932	11,460±250	13,660-13,115	Val Marie,Sask.	Klassen (1993)	Wood
27. AECV-1396C	10,200±130	12,216-11,230	Rowley, AB	Beaudoin (1992)	*Populus sp.* wood
28. S-2911	9,225±330	10,863-9,890	Cypress Lake,Sask.	Klassen (1993)	Wood
29. S-2820	7,395±590	8,947-7,589	Shaunavon ,Sask.	Klassen (1993)	Wood
30. S-2821	7,245±580	8,643-7,405	Climax,Sask.	Klassen (1993)	Wood
31. S-2819	3,800±165	4,410-3,934	Shaunavon,Sask.	Klassen (1993)	Wood
32. S-2930	3,440±165	3,876-3,475	Climax,Sask.	Klassen (1993)	Wood
Secondary River Terrace					
33. GSC-236	9,290±260*	10,799-9,984*	Willow Cr., AB	In: Waters and Rutter (1984) (Dyke *et al.*, 1965)	Gastropods
34. S-1798	8,195±1090	10,355-7,813	Whitemud Creek, AB	Rains and Welch (1988)	Bone
35. S-1926	8,660±125	9,838-9,492	Strawberry Creek,AB	Rains and Welch (1988)	Charcoal
36. S-1787	8,015±135	8,996-8,588	Strawberry Creek, AB	Rains and Welch (1988)	Mollusc Shell
37. TO-1829	7,610±70	8,427-8,217	Ghostpine Creek, AB	Rains *et al.* (1994)	Bone
38. S-1789	5,865±135	6,851-6,510	Strawberry Creek, AB	Rains and Welch (1988)	Bone
39. GSC-1735	5,760±170	6,752-6,351	Medicine Hat, AB	Lowden and Blake (1975)	Gastropod shell
40. S-1788	5,640±130	6,615-6,298	Strawberry Creek, AB	Rains and Welch (1988)	Bone
41. S-2387	5,490±230	6,496-5,988	Whitemud Creek, AB	Rains and Welch (1988)	Bone
42. S-1800	4,685±260	5,653-4,992	Strawberry Creek, AB	Rains and Welch (1988)	Bone
43. S-2392	4,220±250	5,242-4,411	Whitemud Creek, AB	Rains and Welch (1988)	Mollusc shell
44. S-1797	4,225±150	4,987-4,560	Whitemud Creek, AB	Rains and Welch (1988)	Bone
45. S-1795	3,255±90	3,569-3,374	Whitemud Creek, AB	Rains and Welch (1988)	Bone
46. S-2389	3,200±125	3,568-3,263	Whitemud Creek, AB	Rains and Welch (1988)	Wood
47. S-1794	3,200±80	3,545-3,276	Whitemud Creek, AB	Rains and Welch (1988)	Bone
48. S-1796	3,180±85	3,476-3,265	Whitemud Creek, AB	Rains and Welch (1988)	Wood
49. S-2388	2,940±125	3,254-2,894	Whitemud Creek, AB	Rains and Welch (1988)	Bone
50. AECV-1276C	2,880±110	3,152-2,862	Onetree Creek, AB	Campbell and Evans (1993)	Bone
51. AECV-904C	2,860±90	3,106-2,854	Onetree Creek, AB	Campbell and Evans (1993)	Bone
52. AECV-684C	2,650±140	2,935-2,484	Little Sandhill Cr., AB	Traynor and Campbell (1989)	Bone
53. AECV-911C	2,590±90	2,782-2,483	Onetree Creek, AB	Campbell and Evans (1993)	Bone
54. AECV-9440	2,580±90	2,774-2,481	Ghostpine Creek, AB	Rains *et al.* (1994)	Bone

55. AECV-906C	2,330±100	2,701-2,151	Onetree Creek, AB	Campbell and Evans (1993)	Bone
56. S-2390	2,025±205	2,305-1,724	Whitemud Creek, AB	Rains and Welch (1988)	Bone
57. AECV-905C	2,080±100	2,147-1,895	Onetree Creek, AB	Campbell and Evans (1993)	Bone
58. S-1783	1,965±75	1,989-1,822	Strawberry Creek, AB	Rains and Welch (1988)	Bone
59. AECV-913C	1,870±100	1,923-1,652	Onetree Creek, AB	Campbell and Evans (1993)	Bone
60. TO-4577	1,860±70	1,867-1,710	Matzhiwan Creek, AB	Barling (1995)	Bone
61. S-1785	1,625±80	1,603-1,406	Strawberry Creek, AB	Rains and Welch (1988)	Bone
62. AECV-682C	1,400±260	1,540-1000	Little Sandhill Cr., AB	Traynor and Campbell (1989)	Bone
63. TO-4579	1,550±60	1,506-1,358	Matzhiwan Creek, AB	Barling (1995)	Bone
64. AECV-604C	1,520±90	1,501-1,327	Little Sandhill Cr., AB	Campbell and Evans (1990)	Bone
65. TO-4576	1,320±90	1,306-1,099	Matzhiwan Creek, AB	Barling (1995)	Bone
66. AECV-914C	1,310±80	1,297-1,140	Onetree Creek, AB	Campbell and Evans (1993)	Bone
67. AECV-80	1,310±80	1,297-1,140	Onetree Creek, AB	Campbell and Evans (1993)	Bone
68. AECV-683C	1,310±80	1,1,297-1,140	Little Sandhill Cr., AB	Traynor and Campbell (1989)	Bone
69. AECV-912C	1,230±80	1,245-1,064	Onetree Creek, AB	Campbell *et al.* (1993)	Bone
70. S-1793	1,220±70	1,238-1,060	Whitemud Creek, AB	Rains and Welch (1988)	Bone
71. S-1786	1,135±80	1,132-951	Strawberry Creek, AB	Rains and Welch (1988)	Bone
72. S-1790	810±75	785-662	Whitemud Creek, AB	Rains and Welch (1988)	Bone
73. S-1784	760±85	770-573	Strawberry Creek, AB	Rains and Welch (1988)	Bone
74. S-1791	705±70	691-559	Whitemud Creek, AB	Rains and Welch (1988)	Bone
75. AECV-680C	490+80	632-336	Little Sandhill Creek, AB	Traynor and Campbell (unpub)	Bone
76. AECV-903C	370±90	495-316	Onetree Creek, AB	Campbell and Evans (1993)	Bone
77. S-1792	315±70	457-298	Whitemud Creek, AB	Rains and Welch (1988)	Wood
78. AECV-1278C	260±90	444-0	Onetree Creek, AB	Campbell and Evans (1993)	Bone
79. AECV-1277C	250±90	437-0	Onetree Creek, AB	Campbell and Evans (1993)	Bone
Channel Fill					
80. AECV-427C	870±70	901-703	DPP, AB	O'Hara and Campbell (1993)	Bone
81. S-2546	310±60	443-296	DPP, AB	O'Hara and Campbell (1993)	*Populus* log
Loess					
82. ALPHA-270		5,400±800	DPP, AB	Bryan *et al.* (1987)	Loess, TL date
Basal Lake Dates / Lacustrine-Ponding Events					
83. S-2908	9,120±250	10,426-9,853	Harris Lake, Sask.	Sauchyn (1990)	Lake sediment macrofossil
84.	9,340±70	10,419-10,163	Clearwater Lake,Sask.	Vance and Last (1994)	Macrofossil. Basal lake date
85. TO-959	9,080±70	10,271-9,971	Little Sandhill Cr., AB	Campbell and Evans (1992)	Bone in a series of lacustrine-ponding events
86.	7,325±70	8,137-7,995	Chappice Lake, AB	Vance and Last (1994)	Lake sediment macrofossil
87. AECV-605.C	7150±150	8,106-7,768	Little Sandhill Creek, AB	Campbell and Evans (1990)	Bone in a series of lacustrine-ponding events
88.	4,940±70	5,735-5,602	Elkwater Lake, AB	Vance and Last (1994)	Macrofossil, basal lake date

Landslides					
89. NA	ca. 4000	4,502-4,422	Harris Lake, Cypress Hills, Sask.	Sauchyn and Lemmen (1996)	Estimate from lake record
90. TO-4479	2,410±120	2,711-2,341	Cabin Cypress Hills, Sask.	Sauchyn and Lemmen (1996)	Gastropod
91. TO-4480	2,240±70	2,326-2,152	Roadside Cypress Hills, Sask.	Sauchyn and Lemmen (1996)	Gastropod
92. S-2772	1,745±85	1,731-1,537	Bendon Creek Cypress Hills, Sask.	Sauchyn and Lemmen (1996)	Bone
93. S-2907	1,445±320	1,693-996	Benson Creek Cypress Hills, Sask.	Sauchyn and Lemmen (1996)	Basal pond sediment
94. S-2630	1,635±105	1,686-1,403	Green Lake Cypress Hills, Sask.	Sauchyn and Lemmen (1996)	Organic soil
95. S-2631	1,234±100	1,264-1,062	Underdahl Creek Cypress Hills, Sask.	Sauchyn and Lemmen (1996)	Organic soil
96. NA	AD 1915±3	84-78	Nine Mile Cypress Hills, Sask.	Sauchyn and Lemmen (1996)	Treering date on landslide
97. NA	AD 1967	29	Police Point, Sask. Cypress Hills, Sask.	Sauchyn and Lemmen (1996)	Historic date on landslide
Regional tephra deposits					
98. WSU-1554	11,300±230	13,460-12,975	Loss Trail Pass Bog,	Mehringer *et al.* (1977)	Glacier Peak tephra bed
WSU-1548	11,200±100	13,224-12,998	Montana		
99. W-4288	6,780±100	7,665-7,478	Crater Lake National	Bacon (1983)	Mazama tephra bed
W-4290	6,830±110	7,722-7,533	Park, Washington		
W4255	6,880±70	7,721-7,584			
W-4356	7,000±60	7,889-7,716			
W-4295	6,840±100	8,825-7,543			
USGS-870	7,015±45	7,891-7,746			

Legend
Dinosaur Provincial Park = DPP

CHAPTER 2
LAKE AREA VARIABILITY ACROSS A CLIMATIC AND VEGETATIONAL TRANSECT IN SOUTHEASTERN ALBERTA, CANADA

A version of Chapter 2 has been published.
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2.1 INTRODUCTION

Lake level fluctuations have been extensively used as proxy indicators of paleoclimate (for example Butzer *et al.* 1972; Benson 1981; Harrison and Metcalfe 1985; Benson and Thompson 1987a, 1987b; Benson and Paillet 1989; Hickman *et al.* 1990; Vance 1991; Vance *et al.* 1992; Harrison *et al.* 1993). Benson and Paillet (1989) demonstrated that for paleoclimate studies in arid and semiarid regions, lake surface area is a better index of climate than lake level, because inflow and evapotranspiration occur mainly across the lake surface rather than through the lake volume. Of course, for a given lake morphometry, lake area, depth, and volume will be strongly correlated, although the relationships are not necessarily linear.

Although many studies in arid or semiarid regions have found lake levels to be highly sensitive to weather (for example Fritz and Krouse 1973; Laycock 1973; Morton 1978; Street-Perrott and Roberts 1983; Harrison and Metcalfe 1985; Street-Perrott and Harrison 1985; Digerfeldt 1986 1988; Harrison 1988; Stine 1990; Teller and Last 1990; Vance 1991; Vance *et al.* 1992), lakes in more humid areas are not generally as useful for paleoclimate studies (notable exceptions are Gaillard 1985; Digerfeldt 1988; Harrison 1989; Dooge 1992; Harrison *et al.* 1993). This is because humid region lakes, unlike arid region lakes, are controlled more by the level of the outlet sill than by weather.Although arid and semiarid region lakes may rarely fill sufficiently to overflow their sills, humid region lakes are maintained during dry periods by locally recharged groundwater and are rarely faced with a water deficit of sufficient duration to reduce them below their outlet sill levels. Only changes of the basin morphometry (including changes to the sill level such as beaver dams or outlet downcutting) or a change in climate sufficient to eliminate the normal excess of precipitation over evaporation will result in a significant change in water level.

There have been several correlative studies of lake level fluctuations crossing climatic gradients (for example Butzer *et al.* 1972; Street-Perrott and Roberts 1983; Gaillard 1985; Harrison 1989; Harrison and Metcalfe 1985; Schweger and Hickman 1989; Vance 1991). Benson (1981) developed a detailed mathematical model of the relationship between Lahontan Basin lake levels and climate, emphasizing the importance of cloud cover as an influence on both precipitation and evaporation. This model was used to evaluate possible paleoclimates corresponding to known paleolake levels, but did not address the issue of lake sensitivity to climate change. Almendinger (1990) did examine the sensitivity of lakes to climate change, but only as a function of physiography and soil permeability. Other models have successfully reproduced fluctuations of lakes in response to climate change (for example Hostetler and Bartlein 1990; Hostetler and Benson 1990), and Bowler (1986) studied the geomorphology and sedimentology of a transect of lakes from humid to arid regions in Australia. Despite the large number of studies of lake area and level changes, there are no published studies of geographic trends in lake sensitivity to less than century-scale climate changes.

Here we examine the variability of modern lakes in response to recent climate along a transect of sites extending from a semiarid region into a sub-humid region in Alberta, to assess the relationship between climate and lake level sensitivity to climatic variability. We further establish a correlation between lake level sensitivity and upland vegetation.

2.2 THEORY

Lake volume variability is the probability that the lake contains a significantly different amount of water from one observation to the next, and is a function of lake volume responsiveness to weather and the frequency of climatic events capable of producing a response. A large lake with a gently sloping bottom will show greater variability in surface area than will a small steep-sided lake under the same external forcing. Thus the sensitivity of lake surface area to small changes in climate is a function of both lake morphometry and lake volume responsiveness to weather.

In considering lake hydrology, it is important to separate the local and distal ground water inputs, because local ground water inputs may vary at short time-scales (seasonally to decadal) according to the local climate. Distal ground water, on the other hand, reflects the climate of sometime in the past at some distant location and is presumably more uniform at seasonal to decadal time-scales, but may nevertheless exert a strong influence on long term lake levels (Almendinger 1990; Digerfeldt *et al.* 1992). Almendinger (1990) demonstrates through an analytic ground water model that this local/distal distinction is important to understanding lake sensitivity to climate fluctuations, and notes that local ground water recharge has the effect of smoothing seasonal precipitation inputs, allowing a humid region lake to remain full through short periods of drought. Thus humid region lakes are relatively insensitive to short-term climate changes due to local ground water inputs, and are controlled by the level of the sill. That some humid region lakes do vary (for example Côté *et al.* 1990) or have varied in the past (for example Digerfeldt 1986, 1988; Harrison 1989; Harrison *et al.* 1993) may be a result of special conditions affecting these lakes such as exceptionally small drainage basins (Harrison 1989), influence of a nearby river (Côté *et al.* 1990), or paleoclimatic fluctuations such that at times in the past the region was semiarid rather than humid (Harrison *et al.* 1993). In addition, although the presence of an outlet dampens a lake's response to climate, the outflow volume may vary strongly, both seasonally and over longer timescales.

Where precipitation is a rare event, a lake will almost always be either dry or at a distally-recharged-ground-water-controlled-level, and thus show very little actual variation through time. It is in the intermediate, semiarid range that lakes should show the greatest sensitivity to short-term climate fluctuations (Figure 2.1). However, lakes which receive substantial distal ground water input will not be as sensitive as those which receive only local inputs; furthermore, lakes with steep sides will vary in volume but relatively little in area. A lake which receives little or no distal ground water, has a gently sloping bottom, and is in a semiarid region will therefore display the greatest sensitivity of surface area to small climate fluctuations.

2.3 METHODS

Thirty-four lakes forming a north-south transect through eastern Alberta were selected from a 1:1,000,000 scale map of Alberta (Figure 2.2). All lakes are in till on Mesozoic sedimentary bedrock; although morphometric and hydrologic information is not available for many of the lakes, most are less than ten metres deep and receive most of their water from precipitation within the watershed. The lakes selected were large enough to show on this scale map, but small enough to be covered by a single 1:31,680 scale aerial photograph (a range of about 100 - 1,000 ha). Lake areas were obtained from aerial photographs available at the Maps Alberta Reference Library. A total of 326 photographs or composite photographs were used; only lakes for which at least six usable, ice-free-season photos taken at different times spanning at least twenty years between 1949 and 1992 were available were included. The lake perimeters and tie points (road intersections or other distinctive fixed features) were traced onto acetate film. The scale and date of each photograph were noted. The lake area in each photograph was obtained using an Ushikata X-Plan 360d planimeter, and within each set of photographs (i.e., all the photographs for one lake) adjusted to a common scale using the tie-points.

For each lake, the standard deviation of the surface area, expressed as a percentage of its maximum recorded surface area, was calculated as an index of its variability. This index has high values when the lake is highly sensitive, and has low values when the lake sensitivity is low. However, since most months in most years were not covered by a photograph for a given site, there is no guarantee that the full dynamic range of lake area was found for a given lake. For this reason, no meaningful correlation can be calculated between lake sensitivity (as reflected by the standard deviation of lake area) and regional climate, because the regression would be leveraged by the large number of values representing less than the full dynamic range of lake areas; only minimum values are obtainable. Thus the results were interpreted using a bounding curve defined by the sites showing the greatest variability in each region rather than a regression line.

The degree of measurement error is very difficult to quantify because repeated measurements may include the same repeated biases such as smoothing of small embayments. However, it was found after the initial data were collected that in several instances photographs were taken of the same lake at different scales during the same month, and sometimes even the same week or day (they were excluded from the overall analysis). The variation in measured lake area values for these instances is less than 2%, suggesting that measurement error contributes a very small portion of the observed variation in lake area.

Climatic moisture index values for Alberta weather stations were calculated (Figure 2.2). For this climatic moisture index, climate normals from year-round climate stations in Alberta were accessed from a database provided by Environment Canada containing the 1951-1980 climate normals for temperature and precipitation (Environment Canada 1982a, 1982b). Maps showing monthly global solar radiation for the period 1967-1976 (Environment Canada 1985) were digitized on a 2° latitude x 4° longitude grid, and the global solar radiation for each station was then estimated by interpolation from these grid nodes.

The climatic moisture index was calculated for each station by subtracting the annual potential evapotranspiration (PET, in cm) from mean annual precipitation (P, in cm). PET was first calculated on a monthly basis (PET_m) using the Jensen and Haise method (Jensen 1973; Bonan 1989).

(1) $$PET_m = C_T (T - T_x) (R_s / \lambda)$$

where T is mean monthly air temperature (°C); R_s is mean monthly global solar radiation (MJ m^{-2} $month^{-1}$); and λ is latent heat of vaporization (24.54 MJ m^{-2} cm^{-1} at 20°C). The parameters C_T and T_X are calculated as:

(2) $$C_T = 1/(38-A/152.5+38/(e_2-e_1))$$

(3) $$T_X = -2.5-1.4(e_2-e_1)-A/550$$

where A is the station's altitude above sea level (m); e_1 and e_2 are the saturation vapour pressures (kPa) at the mean minimum and mean maximum July temperatures, respectively. Annual PET was then obtained by summing PET_m over all twelve months.

This index, hereafter called the P-PET moisture index, correlates strongly with the southern limit of the distribution of conifers throughout western Canada (Hogg 1994). It takes a value of 0 at the approximate location of the limit of conifers; negative values (indicating a normal excess of potential evapotranspiration over precipitation) occur in the aspen parkland and grassland, while positive values (indicating a normal excess of precipitation over potential evapotranspiration) occur in the boreal forest (Figure 2.2). For this paper, values of the P-PET moisture index were interpolated for each lake.

2.4 RESULTS AND DISCUSSION

The cloud of points formed by plotting the standard deviation of per cent lake area versus the P-PET climatic moisture index for 1951-1980 shows a constrained distribution (Figure 2.3). The points falling far below the bounding curve are most likely the result of imperfect aerial photography coverage (the lack of the full dynamic range of the lake being represented); other factors may include basin morphometry (although some of the lakes have steeper sides and so are less sensitive in area than others, there is no apparent tendency for steep-sided lakes to occur preferentially at the northern or southern end of the transect), and distal ground water inputs (although all the lakes are in till, bedrock aquifers carrying distally recharged ground water may outcrop in the bottoms of some lakes). Thus the bounding curve can be taken as the function relating lake variability to climatic moisture.

This bounding curve shows an asymptote at slightly positive values of standard deviation of per cent lake area in the subhumid end of the transect, and much higher values in the semiarid end. This apparent sigmoidal curve is in theory one end of a normal distribution, in which semiarid environments show the greatest lake level variability and absolute desert and humid environments show the least variability. This is because a lake in a humid climate will be controlled primarily by the sill, since evaporation rarely exceeds precipitation causing a nearly permanent water surplus which maintains the lake at its highest levels. In absolute desert, on the other hand, precipitation is an extremely rare event, so that incidences of rising lake level will be extremely rare, and lakes will nearly always be at their lowest levels, maintained exclusively (if at all) by distally-recharged (and therefore nearly unvarying in the short term) ground water. It is only in the semiarid regions, where precipitation events are not rare but evaporation nevertheless exceeds precipitation that neither the level of the sill nor the level of distally-recharged ground water will exert a controlling influence on lake level.

This agrees well with previous studies in Alberta, which taken together form a transect from the low boreal forest to the shortgrass prairie, in which the moister northern sites show less variation in lake level during the last few thousand years than the more southern sites (Schweger and Hickman 1989; Vance 1991; Hickman *et al.* 1990).

That semiarid regions have the most sensitive lakes is not a new conclusion (for example Street-Perrot and Harrison 1985; Harrison 1989; Almendinger 1990; Harrison *et al.* 1991; Harrison *et al.* 1993). However, our results indicate a very sharp change in the sensitivity-climate function near the boundary between the boreal forest and aspen parkland, near the zero isoline of the P-PET climatic moisture index (Figures 2.2 and 2.3). This implies that the lake sensitivity may be effectively binary: lakes are either sensitive (moisture index < 0) or they are not (moisture index >0). It is noteworthy that the mean annual water runoff (Fisheries and Environment Canada 1978) also shows a sharp regional gradient near the steepest portion of the lake sensitivity curve. Throughout Alberta and Saskatchewan, the quantity of mean annual runoff increases rapidly from less than 2.5 cm yr^{-1} over most of the grassland and aspen parkland regions to greater than 10 cm yr^{-1} over most of the boreal forest. Thus the reason for the steepness of the sensitivity gradient is clear: lakes in the boreal forest, where precipitation generally exceeds potential evapotranspiration and where runoff is substantial, will overflow their sills most years, and thus be controlled more by the sill than by climate. In contrast, lakes in the aspen parkland and grassland, where runoff occurs only in exceptionally wet years, will exhibit a high sensitivity to weather. However, such a high sensitivity is probably not applicable to hyperarid regions, for the reasons previously mentioned.

One caveat raised by the steepness of the sensitivity curve in the region of the parkland (transitional grassland) is that an abrupt change in the sensitivity of a lake in the past may not indicate an abrupt change in climate, but rather a crossing of this boundary.

2.5 APPLICATION

The lake sensitivity curve can be used to interpret paleovegetation and probable paleoclimatic conditions in southeastern Alberta.

Pollen zones 3 and 4 in the pollen diagram of Lofty Lake in central Alberta (Figures 2.2 and 2.4) were initially interpreted as birch forest and mixed wood forest respectively (Lichti-Federovich 1970). MacDonald and Ritchie (1986) reinterpreted this pollen diagram using a modern analogues-based approach, noting that while the analogues for these zones are poor, they most strongly resemble parkland pollen spectra. A core from the lake contains sedimentological evidence for a pronounced low water stand from 8700-6290 BP (Schweger and Hickman 1989). Such a low stand likely occurred during a period when the P-PET moisture index was negative. This would have made the region too arid to support a forest, and so the no-analogue pollen spectra most likely represented parkland rather than forest, in agreement with MacDonald and Ritchie (1986).

The correlation of lake stability with past climate conditions and vegetational zones may prove to be a useful tool in the study of past climate fluctuations in this and other regions.

2.6 CHAPTER 2 - FIGURES

FIGURE 2.1

Relative importance of different factors in controlling lake levels in different climate zones. Variability in response to climate will be greatest in the semiarid region where climate and locally recharged ground water exert the greatest control over lake levels.

FIGURE 2.2

Locations of 34 lakes studied (diamonds), climate moisture index values (numbers), vegetation regions (Ecoregions Workgroup 1989), and Lofty Lake, Alberta (star). Ecoregions: Ga = arid grassland; Gt = transitional grassland (aspen parkland); LBs = low boreal subhumid; MBs = mid-boreal subhumid; Gs = subhumid grassland; SC = southern cordilleran.

FIGURE 2.3

Standard deviation as a percentage of maximum lake area versus climatic moisture. The bounding curve is approximate. G = arid grassland; T = transitional grassland (aspen parkland); L = low boreal subhumid; M = mid-boreal subhumid.

FIGURE 2.4

Lofty Lake pollen diagram (modified from MacDonald and Ritchie 1986:256).

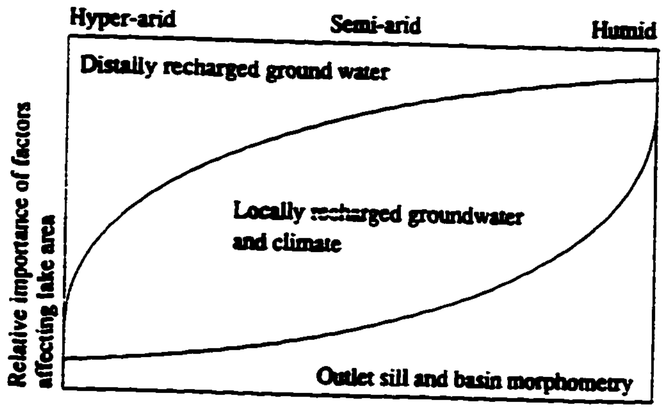

Hyper-arid
Semi-arid
Humid
Distally recharged ground water
Locally recharged groundwater and climate
Outlet sill and basin morphometry
Relative importance of factors affecting lake area

FIGURE 2.1

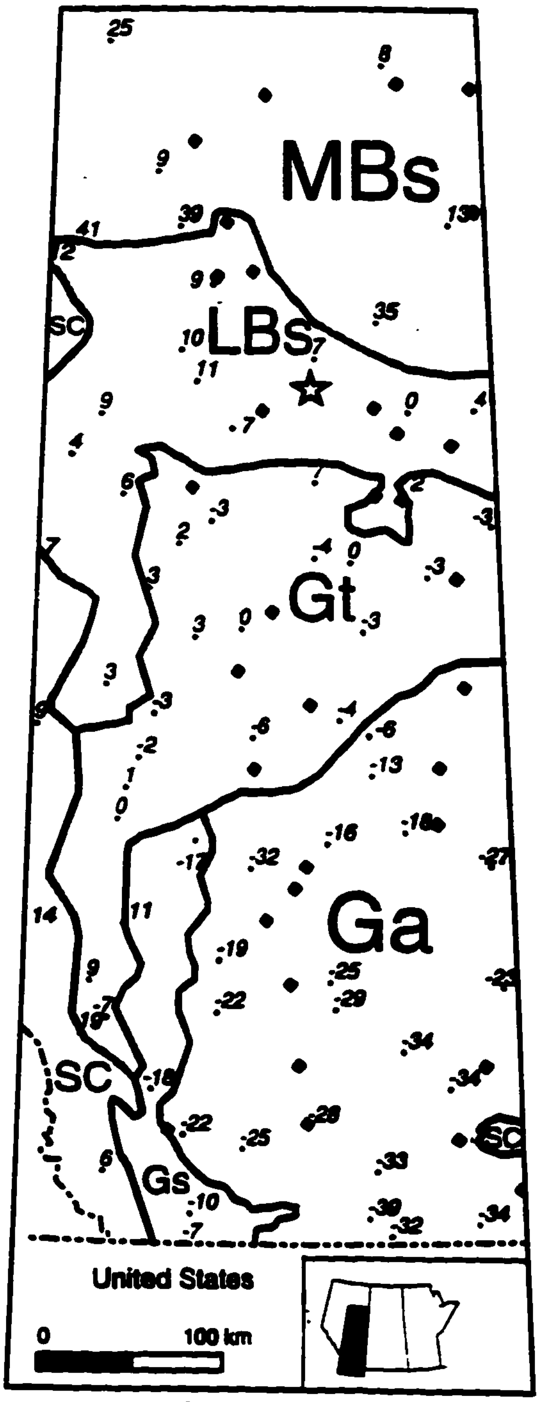
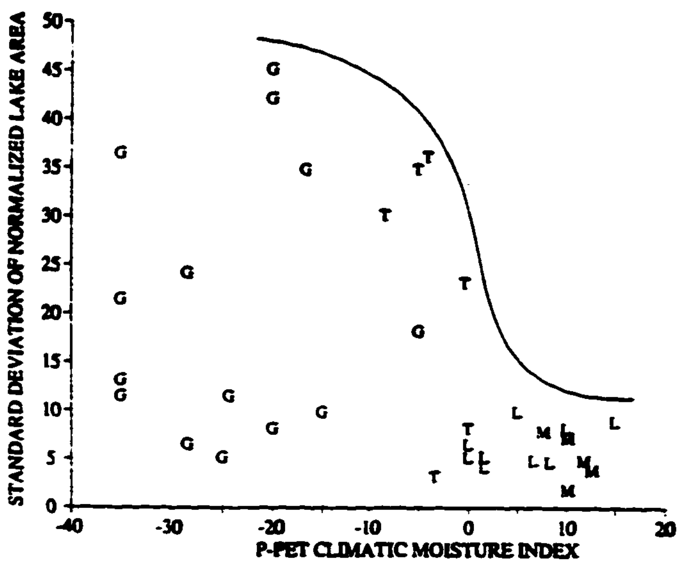


FIGURE 2.2



STANDARD DEVIATION OF NORMALIZED LAKE AREA
P-PET CLIMATIC MOISTURE INDEX

FIGURE 2.3

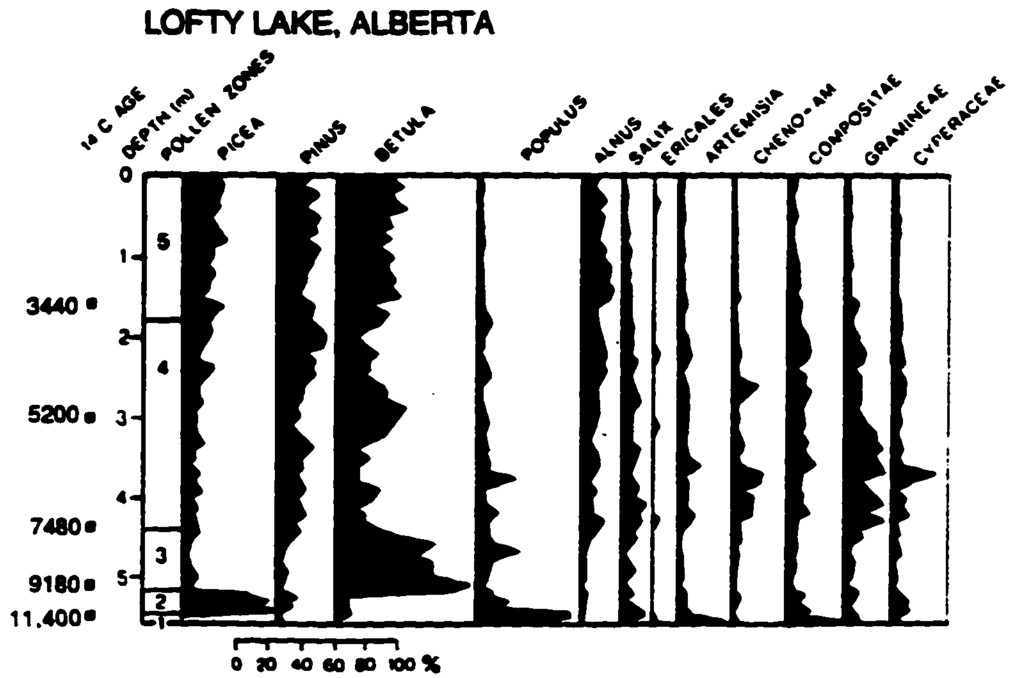

LOFTY LAKE, ALBERTA
14 C AGE
DEPTH (m)
POLLEN ZONES
PICEA
PINUS
BETULA
POPULUS
ALNUS
SALIX
ERICALES
ARTEMISIA
CHENO-AM
COMPOSITAE
GRAMINEAE
CYPERACEAE
3440
5200
7480
9180
11,400
0
1
2
3
4
5
0 20 40 60 80 100 %

FIGURE 2.4

2.7 CHAPTER 2 -REFERENCES

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CHAPTER 3
STRATIGRAPHY OF THE GLACIAL TIDE LAKE BASIN, SOUTHEASTERN ALBERTA, CANADA

Submitted for publication review. Celina Campbell, John R. Paterson and Ian A. Campbell. Stratigraphy of the glacial Tide Lake basin, southeastern Alberta, Canada.

3.1 INTRODUCTION AND PREVIOUS WORK

Approximately 50% of Alberta is covered by proglacial lake deposits associated with late Wisconsinan, Laurentide ice sheet decay (Quigley 1980). The lakes formed as a result of ice impoundment of glacial meltwater and flow from the re-established river systems (Horberg 1954; St. Onge 1972; Stalker 1973; Quigley 1980; Shetsen 1987; Vreeken 1989; Paterson 1996). As the ice sheet decayed in a general northeastward direction, lakes which were formerly partially ice-bordered decanted into lower level lakes further east (for example Teller 1987). Several non-definitive recessional ice-frontal positions have been proposed based partly on the presence of lacustrine deposits, lake floor elevations, and the heights of inlet and outlet channels (St.-Onge 1972; Christiansen 1979; Clayton and Moran 1982; Shetsen 1984; Dyke and Prest 1987; Teller 1987; Klassen 1994; Paterson 1996). Christiansen (1979) suggested that deglaciation occurred in the study area 18,500-16,700 cal yr B.P. (15,500-14,000 ^{14}C yr B.P.), Clayton and Moran (1982) 13,750-13,150 cal yr B.P. (11,700-11,300 ^{14}C yr B.P.), Vreeken (1989) 13,160-13,060 cal yr B.P. (11,200 ^{14}C yr B.P.) and Evans and Campbell (1992) 20,210-19,430 cal yr B.P. (16,790±270 ^{14}C yr B.P.; AECV-681C). Except where noted, all dates are calibrated to cal yr B.P. [before A.D. 1955] using CALIB Rev 3.0.3 [Stuiver and Reimer 1993]. The date range reported here is that which yields a 100% probability when one standard deviation is used. Because the timing of deglaciation remains controversial, bracketing the timing of these lakes is essential in assessing temporally-related issues such as rates of postglacial landscape evolution, models of Quaternary climate change, ice dynamics, floral and faunal migration and megafaunal extinction.

Preglacial, glacial, deglacial, and postglacial events in southeastern Alberta have been documented by Evans and Campbell (1992). In 31 stratigraphic logs from Dinosaur Provincial Park and surrounding plains, they identified seven major and five minor lithofacies associations (LFA). LFA 1 records preglacial valley deposits in braided fluvial floodplains. LFA 1 grades into LFA 2, recording the formation of an ice-dammed lake as Laurentide ice advanced into the region. Ice rafted shield clasts were deposited at the top of LFA 2. The contact between LFA 2 and LFA 3 is erosional. LFA 3 is a grey till, recording ice movement from the northeast. LFA 3 grades into LFA 4. LFA 4 is described as supraglacially derived diamictons and outwash. The contact between LFA 4 and LFA 5 is sharp. LFA 5 records a period of subglacial moulding and then cavity filling by fluvial debris flow deposits. The contact between LFA 5 and LFA 6 is sharp. LFA 6 is a heterogeneous diamicton recording lodgement and melt-out of glacier ice. LFA's 7a-7e chronicle postglacial glaciolacustrine, fluvial, pond and aeolian depositional environments.

On the basis of channel inlet and outlet elevations, stratigraphic exposures, and the distribution of lacustrine related deposits, the evolution of proglacial lakes in southern Alberta, from the 1100 m (all elevations are in meters above sea level) maximum elevation of Glacial Lake MacLeod to the final 680 m level of Glacial Lake Empress is established (Figure 1). Glacial Lakes Drumheller, Gleichen and Lethbridge drained south through the Etzikom Coulee to the Missouri drainage. Etzikom Coulee was abandoned when the level dropped below the Lethbridge moraine divide, at an elevation of 915 m. Glacial Lake Drumheller outflow ceased through the Strathmore channel at 945 m, and, presumably, discharged instead through the Crowfoot channel until it lowered to 915 m. Flow was then diverted eastwards, into the Bassano-Tilley basin. Glacial Lake Gleichen, which had flowed through the McGregor Lake channel, started to discharge eastward into the interlinked Glacial Lake Bassano-Tilley through its outlet at 860 m. Flow likely continued until the channel deepened to 850 m.

With further ice recession, Glacial Lake Lethbridge lowered and expanded to form Glacial Lake Taber, controlled by the outlet through Chin Coulee at 915 m to 846 m. Drainage then flowed through Forty Mile Coulee until 792 m then through the valley of the South Saskatchewan, becoming channelized at 760 m. Glacial Lake Medicine Hat formed as outlet controlled water levels dropped to roughly 760 m, and Glacial Lake Empress formed to the north at roughly the same elevation.

The glacial lake which formed in the Bassano-Tilley basin is of particular interest to this research; its elevations ranged from a possible maximum of 915 m until final drainage at 690-700 m. Glacial Lake Bassano-Tilley filled after the abandonment of the southward-flow into the Missouri drainage basin system at 915 m and the beginning of generally eastward flowing drainage within Alberta. While Glacial Lake Bassano-Tilley existed, it acted as the base level for all discharge from the western margin of the Laurentide ice sheet from Grande Prairie (>1,000 km to the northwest) to the Cordilleran areas now drained by the North Saskatchewan, Red Deer, and Bow rivers. The height of the lake was regulated both by ice to the north and the east, and the regional topography.

The Bassano-Tilley basin underwent two major episodes of inflow. During the earlier Crawling Valley stage, silt sized sediments were deposited from discharge out of Glacial Lake Drumheller through Crawling Valley. During this stage, Glacial Lake Bassano-Tilley lowered from 851 m to 798 m (levels based on lacustrine deposit elevations). Final rapid drainage of Glacial Lake Drumheller, moved coarser, sand sized sediments during the Red Deer Delta stage into the basin. During the Red Deer Delta stage lake levels in Glacial Lake Bassano-Tilley lowered from *ca.* 813 m to *ca.* >750 m (Paterson, 1996). As ice retreat exposed progressively lower outlets, Glacial Lake Bassano-Tilley eventually emptied with the deepening of the Red Deer valley at 690 m (the Broad Valley stage in Bryan *et al.* 1987).

This paper reports sedimentological, geochemical and geochronological results from Tide Lake. It occupied a basin which may have developed either as a marginal embayment of Glacial Lake Bassano-Tilley, or as a separate, ice-surrounded, small proglacial lake, or as a wholly or partially subglacial lake, or any combination of these various possibilities.

Two Optically Stimulated Luminescence (OSL) dates on what are identified as proglacial lake Bassano-Tilley sediments were obtained. One was dated at 26,000±5,000

OSL yr B.P. This date may well relate to proglacial lake deposition during the late Wisconsinan Laurentide ice advance phase (see LFA 2 in Evans and Campbell 1992). The other was described simply as "early Holocene" (Huntley 1995, pers. comm.). These dates suggest that neither one provides well defined chronologic control for lake formation in the basin (Paterson 1996). The core from Tide Lake provides better local details of the preglacial, glacial, deglacial and postglacial geomorphic development of the basin, and by extension and correlation with previous studies, the plains of southeastern Alberta.

3.2 LOCATION

Tide Lake (25 km^2) is centered at latitude 50° 33' N and longitude 111° 20' W (Figure 2), in the shortgrass prairie of semiarid southern Alberta. The regional bedrock consists of poorly consolidated, horizontally bedded, sandstone and mudstone dominated Upper Cretaceous Judith River Group and Bearpaw Formation.The basin lies within the 730 m bedrock topography contour at the head of a tributary to the preglacial Lethbridge Valley (Figure 3) (Alberta Research Council 1970; Pawlowicz and Fenton 1995). Tide Lake is presently a playa located at a mean altitude of *ca.* 747 m a.s.l. Although the Tide Lake basin is closed below the 754 m a.s.l. contour, there is rarely standing water in the basin. The surface area of the lake floor and its catchment area is *ca.* 495 km^2. Almost all of the major input channels have been dammed to provide ponds for stock watering, and to prevent lake flooding on land leased by gas and oil companies. The core was collected at a well-site (744 m elevation), in Section 20, Township 18, Range 10 West of the Fourth Meridian, on the lease of Beau Canada Exploration Limited.

3.3 METHODS

A truck-mounted, 0.05 m diameter, hollow stem auger extracted 8.1 m of core from a 20.4 m hole. Drilling stopped after 0.8 m of dark grey, fissile, weakly cemented, bentonitic shale Bearpaw Formation bedrock (Caldwell and North 1977) was retrieved. The incomplete nature of the core appears to be due largely to the extreme stiffness of much of the sediment. This caused premature plugging of the core barrel, and, in some cases, collapse of the plastic tube liners. Nevertheless, the core contains excellent samples of depositional sequences, from the present playa surface to just below the bedrock contact.

The core was contained in plastic sleeves, frozen and split lengthwise. The core was described using a variant of the Eyles *et al.* (1983) code for sediment description. It was cut into contiguous 1 cm slices; subsamples of these were used for analysis of the glass:crystal ratio (200 clastic grain count with a petrographic microscope). No tephra beds were identified. Samples were then homogenized into 10 cm segments using a rubber-tipped mortar and pestle, and stored at room temperature in a plastic bag. Grain size analysis was performed using both hydrometers and sieves on each 10 cm homogenized sample (Rutter 1995). Mean grain size was determined after Folk and Ward (1957). For each of the 81 samples, loss on ignition and total carbonates (Dean 1974) were determined. Bulk authigenic geochemistry (Na and Al) was determined by ICP-AES after Malo (1977); this has been shown by Malo (1977) to be an efficient extractant of trace metals and oxide coatings, causing minimal degradation of clays. Duplicate samples were prepared and analyzed for grain size, loss on ignition and total carbonates, and geochemistry every 20 cm. These replicate analyses indicate that precision of the data reported here is approximately ± 2.5% for grain size, ± 2% for loss on ignition and total carbonates, and ± 2% for geochemistry.

3.4 TIDE LAKE SECTION: DESCRIPTION AND INTERPRETATION

The core penetrated 0.72 m of Bearpaw Formation bedrock at a contact *ca.* 20 m below the playa floor. This depth corresponds with the depth to bedrock (*ca.* 730 m a.s.l.) interpolated from Alberta Research Council's (1970) bedrock map (Figure 4). Five lithozones (LZ) above the unweathered bedrock were distinguished based on the combined results of all the data sets, and visually delimited on the basis of trends seen in Figure 5. The term lithozone is used here as "an informal term to indicate a body of strata that is unified in a general way by lithologic features but for which there is insufficient need or information to justify its designation as a formal unit" (Bates and Jackson 1987:385).

3.4.1 LZ-1: WEATHERED AND GLACIALLY DISTURBED BEDROCK

3.4.1.1 DESCRIPTION

LZ-1 consists of 0.53 m of the dark grey (2.5Y N4/4), fissile, Bearpaw Formation shale which grades vertically into 0.19 m of light grey (2.5Y N7/0), structureless, puffy, weathered shale which contains a large, fractured, dark grey (2.5Y N4/4) shale clast. LZ-1 pales upwards (Figure 3.6). The upper contact with LZ-2 is sharp and erosional. There appear to be no non-Bearpaw clasts in LZ-1. A sample recovered from within the drill head contained isolated coal fragments.

3.4.1.2 INTERPRETATION

The gradational contact between the weathered (LZ-1) and unweathered bedrock indicates that the bedrock and LZ-1 are genetically related. LZ-1 is interpreted as weathered Bearpaw Formation bedrock that has been glacially overridden and disturbed as indicated by the incorporation of a sheared and fragmented clast of unweathered Bearpaw shale. This may have occurred when glacial ice froze undisturbed bedrock, LZ-1 and LZ-2 preglacial sediments to its base, which resulted in the incorporation of unweathered Bearpaw shale into LZ-1. Alternatively, as LZ-1 is clearly a weathered surface, the fragmented clast may have been incorporated by preglacial cryoturbation. It is also possible that ice overrode LZ-1, incorporated the unweathered Bearpaw shale clast into LZ-1, sheared off the top of LZ-1, and then subsequently LZ-2 was deposited.

3.4.2 LZ-2: FINELY LAMINATED FINE SILT AND CLAY

3.4.2.1 DESCRIPTION

LZ-2 is composed of 0.7 m of light brownish grey (2.5Y 6/2), finely laminated, sand, fine silt, and clay. A thin 0.01 m thick, finely laminated, light olive brown (2.5Y 5/4) wedge with sharp upper and lower contacts occurs near the base of this lithozone (Figure 3.7). LZ-2 appears devoid of shield clasts and contains highly oxidized, microscopic-sized plant fragments. LZ-2 is dominated by very fine silt, containing approximately 33% clay sized particles and 20% sand sized particles, probably derived from a local bedrock source. It is high in Na and low in Al. Loss on ignition measured 5% organic material and less than 1% carbonates.

3.4.2.2 INTERPRETATION

LZ-2 is interpreted as a preglacial, Empress Formation, fluvial deposit, as suggested by the fine bedding structures, relatively high organic content, the site's location in the head of a preglacial valley, and the lack of shield clasts. However, the sample is very small, so the absence of shield clasts can not be taken as definitive. The fine grained nature of the sediment suggests that LZ-2 was deposited in a low energy environment, perhaps as overbank deposits or in waning flow conditions. Alternatively, it is possible that LZ-2 is a rip-up rather than *in situ*.

3.4.3 LZ-3: CROSS-STRATIFIED SANDS AND SILTS

3.4.3.1 DESCRIPTION

LZ-3 occurs 19.09-18.29 m below the modern surface. There is a sharp contact with LZ-2 below and the contact with LZ-4 is missing (between retrieved core segments). LZ-3 consists of a generally olive brown (2.5Y 4/4) to dark greyish brown (2.5Y N4/2), interbedded, ripple, planar and trough cross-bedded series of sands and silts, with many erosional contacts. Shield clasts appear to be introduced for the first time in the stratigraphic section and occur throughout, often as dropstones, within the fine grained sediments of LZ-3 where they have deformed the underlying sediment. Abundant rip-up clasts are dominantly clay aggregates and therefore bias the grain size analysis towards the fines (Figure 3.8). LZ-3 shows a general fining upwards sequence. The concentrations of Na, Al, organics and carbonates are generally moderate to low. Loss on ignition measured a decrease in organics from 5% in LZ-2 to 2.5% in LZ-3, and 1% carbonates.

3.4.3.2 INTERPRETATION

LZ-3 is interpreted as a moderate to high-energy stream deposit possibly formed in a subglacial cavity, or in a proglacial lake dominated by current-lain silts. LZ-3 may represent the initial stage of deglacial drainage infilling of the Tide Lake basin. The presence of shield clasts implies that the deposit was formed during or subsequent to late Wisconsinan Laurentide glaciation. The abundance of erosional contacts records very dynamic conditions. The fining up sequence suggests a decrease in energy. The low values for Al imply that much of the clay-sized material may be rock flour derived from the local bedrock rather than clay minerals.

3.4.4 LZ-4: INTERBEDDED LAMINATED AND CROSS-BEDDED SILTS WITH RIP UP CLASTS, DROPSTONES, LOADING STRUCTURES AND RHYTHMITES

3.4.4.1 DESCRIPTION

LZ-4 is located 17.66-3.25 m below the present surface and consists of generally olive brown (2.5Y 4/4) to dark greyish brown (2.5Y 4/2) sediments with abundant granular sized dropstones (including shield clasts) and rip-up clasts throughout. LZ-4 consists of interbedded laminated silts with scattered stones which deform the underlying laminations (Figure 3.9), massive poorly sorted fine grained diamictons with occasional load structures, poorly developed laminations, and cross laminations with high angles of climb (Figure 3.10). There are multiple erosional contacts principally at the bases of the diamict units (Figure 3.11). The fines display rhythmic bedding and are often graded. There are five distinct rhythmically graded bands near the top of LZ-4, each approximately 0.05 m in thickness (Figure 3.12). Each of these rhythmites shows a sandy base with granules, and grades upwards to clays. There are finely bedded internal laminations within these rhythmites. The mean grain size shows a coarsening upwards from 17.6-11.5 m and a fining upwards from very fine silt to clay from *ca.* 11.5-3.25 m. Na and Al rise rapidly in this unit. Loss on ignition indicates 3.2-3.9% organic and 1.6-1.9% carbonate content. A small, fish vertebra was found near the top of this unit. An A.M.S. date on wood from near the base of this unit (16.96 m below the playa surface) is 22,870-21,880 cal yr B.P. (18,860 ± 500 ^{14}C yr B.P., TO-5947, calibrated after a linear interpolation of Bard *et al.* 1990).

3.4.4.2 INTERPRETATION

LZ-4 records deposition in a lacustrine environment. The variability in grain size and sedimentary structures suggests highly dynamic conditions, with an increase in energy from 17.6-11.5 m, and a decrease in energy from 11.5-3.25 m. The fining upwards sequence may document either a reduction in sediment input or lake deepening. The lake may have been ice marginal or subglacial.

It is possible that all of LZ-4 was deposited by supraglacial, englacial and subglacial meltwater decanting off the nearby icesheet onto a delta in a proglacial lake, with debris-flows interbedding with more complacent lacustrine or deltaic deposits. The low values for Al suggest the clay-sized particles may be dominantly rock flour. The graded rhythmites from *ca.* 5.0-4.75 m may be true annual varves. The internal laminations may represent short-duration events or turbidites (although the lack of basal erosion and the presence of internal laminations argue against this). The occurrence of a fish vertebra at 3.5 m and the relatively high organic content suggest that at least part of the lake was subaerially exposed.

Alternatively, with the exception of the top of the lacustrine sequence where the fish vertebra was found, LZ-4 may have been deposited in a subglacial lake. Evidence in favour of a subglacial hypothesis include its location in a preglacial topographic hollow. The basin is surrounded by moraine, and a possible beaded esker to the northwest of the lake may have fed a subglacial Tide Lake. The above interpretation would remain unchanged except that the deposition would be subglacial rather than subaerial; it would still have likely been close to the ice margin.The subglacial sediments could have been eroded by ice recoupling, which could be represented by one of the erosional contacts. Above this contact lie proglacial lacustrine sediments. Alternatively, LZ-4 may have formed as a small, high energy subglacial cavity. There may of course have been a floating ice shelf on a lake at the margin of the glacier. With only one core available for analysis, it is not possible at this time to distinguish between these hypotheses.

The wood sample is presumed to have been incorporated during meltwater scouring from preglacial sediments, and to thus predate glaciation. The A.M.S. date of 22,870-21,880 cal yr B.P. provides a maximum date for glaciation of this area.

3.4.5 LZ-5: MASSIVE PED-RICH SILT

3.4.5.1 DESCRIPTION

LZ-5 is 3.25-0 m below the modern surface. The sediment is dark grey (2.5Y 4/0), extremely sticky, massive, and is composed of ~0.05 m sized peds There are abundant roots, plant, and insect macrofossils. The contact between LZ-4 and LZ-5 is gradational.

The mean clastic fraction is 2% sand, 74% silt and 24% clay sized grains. The top 0.2 m show an upwards increase in sand and clay and a decrease in silt. Loss on ignition shows relatively high organic (4-9%) and carbonate (0.8-2.2%) contents. Bulk geochemistry indicates relatively high amounts of Na and Al with Na declining in the top 0.4 m.

3.4.5.2 INTERPRETATION

LZ-5 records deposition in an evaporitic lake environment. The high values for Al suggest an abundance of clay minerals in the clay-size fraction. The high values of Na may relate simply to exchangeable Na-rich clays, but given the abundance of saline seeps and playas in the region today, most likely reflect the presence of evaporitic salts. The top 0.2 m of LZ-5 records a pronounced increase in grain size (Figure 5) related to an increase in very fine sand and coarse silt. This material is possibly a product of increased late Holocene aeolian deposits mixing with the fines associated with playa deposition.

3.5 CHRONOLOGY

The preglacial wood date in LZ-4 of 22,870-21,880 cal yr B.P provides a maximum date for glaciation of this area. A date of 20,210-19,430 cal yr B.P. (16,790 $\pm$ 270 ^{14}C yr B.P., AECV-681C), from mammoth bone at Dinosaur Provincial Park, constrains the date of deglaciation (Evans and Campbell 1992, 1995; C. Campbell and Campbell 1997). Therefore, LZ-1, LZ-2, and possibly LZ-3, occurred prior to 22,870-21,880 cal yr B.P.

Based on the absence of any of the regional tephra beds which provide important chronostratigraphic markers, a chronologic outline can be developed.The absence of the Glacier Peak tephra in the proglacial lake sequence (LZ-4) suggests that its deposition was prior to 13,460-12,975 cal yr B.P. (11,300$\pm$230 ^{14}C yr B.P.; Mehringer *et al.* 1977) but after 20,210-19,430 cal yr B.P. The absence of the Mazama tephra bed implies that LZ-5 began to deposit after 7891-7478 cal yr B.P. (7015-6830 ^{14}C yr B.P.; Bacon 1983). This suggests that there was possibly a major depositional hiatus or erosional episode between *ca.* $\geq$ 13,460-12,975 to 7015-6,830 cal yr B.P.

3.6 DISCUSSION AND REGIONAL CORRELATIONS

Prior to late Wisconsinan glaciation, and the deposition of the Empress Formation, the paleo-Lethbridge Valley was stripped down to Bearpaw Formation bedrock. This underwent weathering, followed by deposition of preglacial LZ-2 deposits. These were then glacially overridden and disturbed as indicated by the incorporation of unweathered and sheared Bearpaw Formation shale clasts into the weathered LZ-1. If LZ-2 is preglacial, it may be contemporaneous with Evans and Campbell's (1992) LFA-1, which is considered to be $\geq$ mid-Wisconsinan in age.

The preglacial wood date in LZ-3 of 22,870-21,880 cal yr B.P provides a maximum date for glaciation of this area. The date of 21,210-19,430 cal yr B.P. constrains the date of deglaciation (Evans and Campbell 1992, 1995). This is consistent with a late Wisconsinan ice maximum advance some 350 km further south into Montana at 21,650-21,310 cal yr B.P. (18,000 yr B.P., Dyke and Prest 1987). A simple uniform model, with no period of ice margin stability at glacial maximum or readvances, would suggest a minimum average rate of both ice advance and recession of at least 200-290 m cal yr^{-1}. If these dates are correct, the southern edge of the Laurentide ice sheet advanced, disintegrated, and disappeared from northern Montana and southeastern Alberta very rapidly, possibly in as little as *ca.* 1670 cal yr. Such a rate of advance and retreat is greater by nearly an order of magnitude than the 50 m ^{14}C yr^{-1} rate previously proposed for this ice margin by Dyke and Prest (1987), and much less than the <2 km yr^{-1} retreat rate proposed by Clayton *et al.* (1985).

LZ-3 is possibly contemporaneous with Evans and Campbell's (1992) LFA 6. But, on the basis of a single core, caution should be given to interpreting LZ-3 as having been deposited as a subglacial cavity fill deposit. It would, perhaps be simpler to interpret it as a proglacial delta / mudflow deposit. If this is the case, then there is no actual subglacial material present in this core, which may pose a difficulty with the absence of the expected till. Even if it is a subglacial cavity fill deposit, till from either prior to or after the cavity fill deposit would be expected. Therefore, the subglacial cavity fill deposit explanation does not reduce the difficulty of interpretation.

Lack of any clearly identifiable till deposits at Tide Lake suggests the possibility that glacial sediments were stripped away prior to terminoglacial fluvial / lacustrine activity. Basal meltwater or subglacial floods may have removed any till deposited prior to the deposition of LZ-3. In this context, it is important to note that this site is adjacent to the proposed eastern path of the Livingstone Lake megaflood event (Shaw and Kvill 1984; Rains *et al.* 1993; Shaw *et al.* 1996), which has been invoked as an explanation for the thin to absent till cover over much of this region.

LZ-4 may be contemporaneous with Evans and Campbell's (1992) LFA-7a proglacial lake deposits. The lake may then be terminoglacial and possibly ice-marginal, as indicated by the abundant debris flows. The upper portion of the proglacial lake deposit, where rhythmites occur as part of a general fining upwards sequence, is indicative of the gradual retreat of the ice front away from the core site and consequent calming of the water. The drainage of the lake shortly thereafter implies the removal of the ice to a position where it was no longer able to dam the lake (Bryan *et al.* 1987). In this case, the lake could possibly have existed over a maximum time span from *ca.* 20,210-19,430 to 13,460-12,975 cal yr B.P. Given the evident rapidity of regional deglaciation, however, the lake was almost certainly very short lived. There is the possibility, discussed above, that a portion of this LZ-4 was deposited subglacially. Even if this is the case, the fish vertebra found near the top of LZ-4 suggests a subaerial environment for at least the later stage of deposition.

Tide Lake shares divides with Glacial Lake Bassano-Tilley. The elevations of these divides show that Tide Lake could have been connected to Glacial Lake Bassano-Tilley only when the elevation of Glacial Lake Bassano was above the 754 m level (the sill between Glacial Lakes Tilley and Tide). This connection occurred during both the Crawling Valley (*ca.* 851-798 m) and the Red Deer Delta (*ca.* 813-750 m) stages of Glacial Lake Bassano-Tilley. This fall in lake level required a certain amount of time to occur, during which Glacial Lake Bassano expanded eastward, across the basin, as well as southward. During its eastward expansion, the Tide Lake basin would likely have been inundated, since it is located at a lower elevation and west of the Suffield Moraine which acts as the boundary constraining the maximum eastwards extent of Glacial Lake

Bassano. Tide Lake cannot have been connected to Glacial Lake Bassano-Tilley during the final Red Deer Delta stage when the water fell below 754 m. Prior to this period Tide Lake was likely an embayment of Glacial Lake Bassano-Tilley.

Alternatively, LZ-4 may have been deposited subglacially or under an ice shelf. The Tide Lake basin occupies a topographic hollow which existed prior to glaciation as it was part of the preglacial Lethbridge Valley (Alberta Research Council 1970) and is surrounded by glacial moraine (Shetsen 1987). From ~17.66-12.19 m the core contains poorly sorted matrix, load structures, polymodal grain size distributions and lack of preferred clast orientation indicating that these deposits could be flow or melt out deposits (Brodzikowski and van Loon 1991). The lack of graded bedding from the base of LZ-4 to ~5.25 m could also possibly be the result of subglacial lacustrine development (Brodzikowski and van Loon 1991).

After Tide Lake decanted northwards into the Red Deer drainage there appears to have been a major depositional hiatus between ≥ 13,460-12,975 to 7,015-6,830 cal yr B.P. indicated by the lack of any regional tephra beds. Although there is a gradational rather than erosional contact between LZ-4 and LZ-5, there is nevertheless a hiatus of several thousand years; the gradational contact is the result of extensive bioturbation by roots. This period corresponds with a major decrease in regional humidity possibly due to maximum seasonality of insolation (Berger 1978; Schweger and Hickman 1989; Campbell and Campbell 1997). Sometime after 7015-6830 cal yr B.P., the lake began to fill with periodically deposited sediment. Schweger and Hickman (1989) document an increase in humidity north and southeast of the study area at this time. LZ-5 is interpreted as being contemporaneous with Evans and Campbell's (1992) LFA 7d which records the infilling of local ponded depressions on the prairie surface.

Although earlier aeolian events did occur in the region (for example Bryan *et al.* 1987; Vreeken 1989), the lack of earlier evidence of aeolian deposits suggests that either Tide Lake was not a suitable site for deposition or that it acted mainly as a deflation surface during these periods. The increase in very fine sand and coarse silt recorded in the top 0.2 m of LZ-5 thus indicates: (1) an increase in the late Holocene aeolian activity; and / or (2) that the lake floor, while dominantly dry, became moist / vegetated enough to trap aeolian grains.

3.7 CONCLUSIONS

The findings of this study can be summarized as follows:

(1) The possible preglacial LZ-1 is apparently alluvial in origin; this is consistent with the presence of preglacial channels in this area.

(2) The maximum date for ice advance in the area was 22,870-21,880 cal yr B.P.

(3) The deglacial date of 21,210-19,430 cal yr B.P., from nearby Dinosaur Provincial Park, suggests a minimum average rate of ice movement (for both advance and recession from glacial maximum in Montana to Dinosaur Provincial Park) of at least 200-290 m cal yr^{-1}.

(4) Tide Lake may have been an embayment of Glacial Lake Bassano-Tilley during Paterson's (1996) Crawling Valley and early Red Deer Delta stages, but not during the final Red Deer Delta stage (<747 m).

(5) There is a possibility that part of LZ-4 may have been subglacial or under an ice shelf; this hypothesis needs to be verified by additional core studies.
(6) Given the rapidity of deglaciation, Tide Lake was probably very short-lived sometime during 20,210-19,430 to 13,460-12,975 cal yr B.P.
(7) Once the level of Tide Lake fell below the 754 m divide, the height of its southern outlet, it became dependent on local precipitation within its very small catchment area for infilling.
(8) There appears to have been a major depositional hiatus between 13,460-12,975 to 7891-7478 cal yr B.P. evidenced by the lack of any regional tephra beds, indicating that precipitation was insufficient to fill the basin or that the area was not a suitable site for deposition or that it acted mainly as a deflation surface during these periods.
(9) Tide Lake became a playa sometime after 7015-6830 cal yr B.P., probably due to an increase in regional humidity.
(10) During the late Holocene there has possibly been an increase in aeolian deposition in the basin.

3.8 CHAPTER 3 - FIGURES

FIGURE 3.1

Schematic diagram of the proglacial lake system in southern Alberta during the late Wisconsinan deglaciation showing the generalized lake areas superimposed over the major drainage system.

FIGURE 3.2

Location map of Tide Lake.

FIGURE 3.3

Bedrock topography (after Alberta Research Council 1970).

FIGURE 3.4

Cross-section of bedrock topography along transect A - A' on Figure 3.3.

FIGURE 3.5

Tide Lake stratigraphy: Core description, lithozone, granulometry, geochemistry (ppm) and loss on ignition and total carbonates.

Legend to stratigraphic log: Roots; Clasts; Lamination or bedding; Contorted bedding; Cross-beds; Load structures; Erosional contact; Codes on right are adapted from Eyles *et al.* 1983. C=clay; D=diamict; F=fine-grained; S=sand; m=massive; l=laminated) after F only; l=load structures (except after F); s=stratified; c=contorted; x=cross-bedded; -r=roots; -u=rip up clasts; -h=shield clasts; -d=dropstones. Scale on left indicates non-recovery of core material.

FIGURE 3.6

LZ-1: Light grey (2.5Y N7/0), structureless, puffy, weathered shale which contains a large, fractured, dark grey (2.5Y N4/4) shale clast.

FIGURE 3.7

LZ-2: A thin 0.01 m thick, finely laminated, light olive brown (2.5Y 5/4) wedge with sharp upper and lower contacts occurs near the base of this lithozone

FIGURE 3.8

LZ-3: Rip-up clasts are dominantly clay aggregates.

FIGURE 3.9

LZ-4: Interbedded laminated silts with scattered stones which deform the underlying laminations, massive poorly sorted fine grained diamictons with occasional load structures.

FIGURE 3.10

LZ-4: Poorly developed laminations and cross laminations with high angles of climb.

FIGURE 3.11

LZ-4: Multiple erosional contacts principally at the bases of the diamict units.

FIGURE 3.12

LZ-4: Five distinct rhythmically graded bands near the top, each approximately 0.05 m in thickness. Each of these rhythmites shows a sandy base with granules, and grades upwards to clays. There are finely bedded internal laminations within these rhythmites.

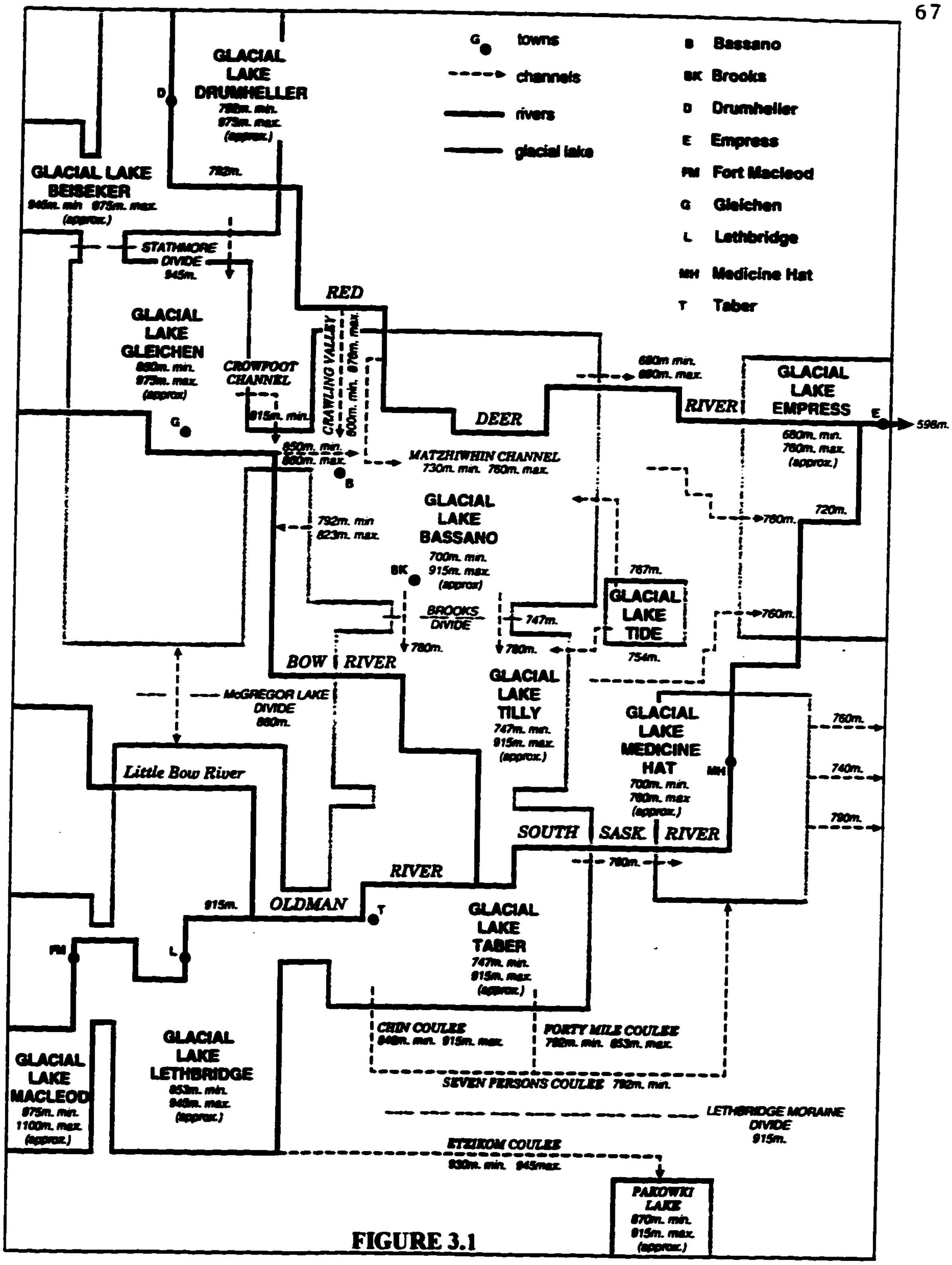

towns
channels
rivers
glacial lake
B Bassano
BK Brooks
D Drumheller
E Empress
FM Fort Macleod
G Gleichen
L Lethbridge
MH Medicine Hat
T Taber
GLACIAL LAKE DRUMHELLER
973m. max.
(approx.)
792m.
GLACIAL LAKE BEISEKER
975m. max.
(approx.)
STATHMORE DIVIDE
945m.
RED
GLACIAL LAKE GLEICHEN
880m. min.
973m. max.
(approx.)
CROWFOOT CHANNEL
915m. min.
CRAWLING VALLEY
800m. min. 876m. max.
DEER
RIVER
680m min.
880m. max.
GLACIAL LAKE EMPRESS
680m. min.
760m. max.
(approx.)
598m.
720m.
760m.
MATZHIWHIN CHANNEL
730m. min. 760m. max.
GLACIAL LAKE BASSANO
700m. min.
915m. max.
(approx.)
792m. min
823m. max.
767m.
GLACIAL LAKE TIDE
754m.
BROOKS DIVIDE
747m.
780m.
BOW RIVER
McGREGOR LAKE DIVIDE
880m.
GLACIAL LAKE TILLY
747m. min.
915m. max.
(approx.)
GLACIAL LAKE MEDICINE HAT
700m. min.
760m. max
(approx.)
760m.
740m.
790m.
Little Bow River
SOUTH SASK. RIVER
780m.
RIVER
915m.
OLDMAN
GLACIAL LAKE TABER
747m. min.
915m. max.
(approx.)
CHIN COULEE
915m. max.
FORTY MILE COULEE
792m. min. 853m. max.
GLACIAL LAKE MACLEOD
975m. min.
1100m. max.
(approx.)
GLACIAL LAKE LETHBRIDGE
853m. min.
945m. max.
(approx.)
SEVEN PERSONS COULEE 792m. min.
LETHBRIDGE MORAINE DIVIDE
915m.
ETZIKOM COULEE
930m. min. 945max.
PAKOWKI LAKE
870m. min.
915m. max.
(approx.)

FIGURE 3.1

FIGURE 3.2

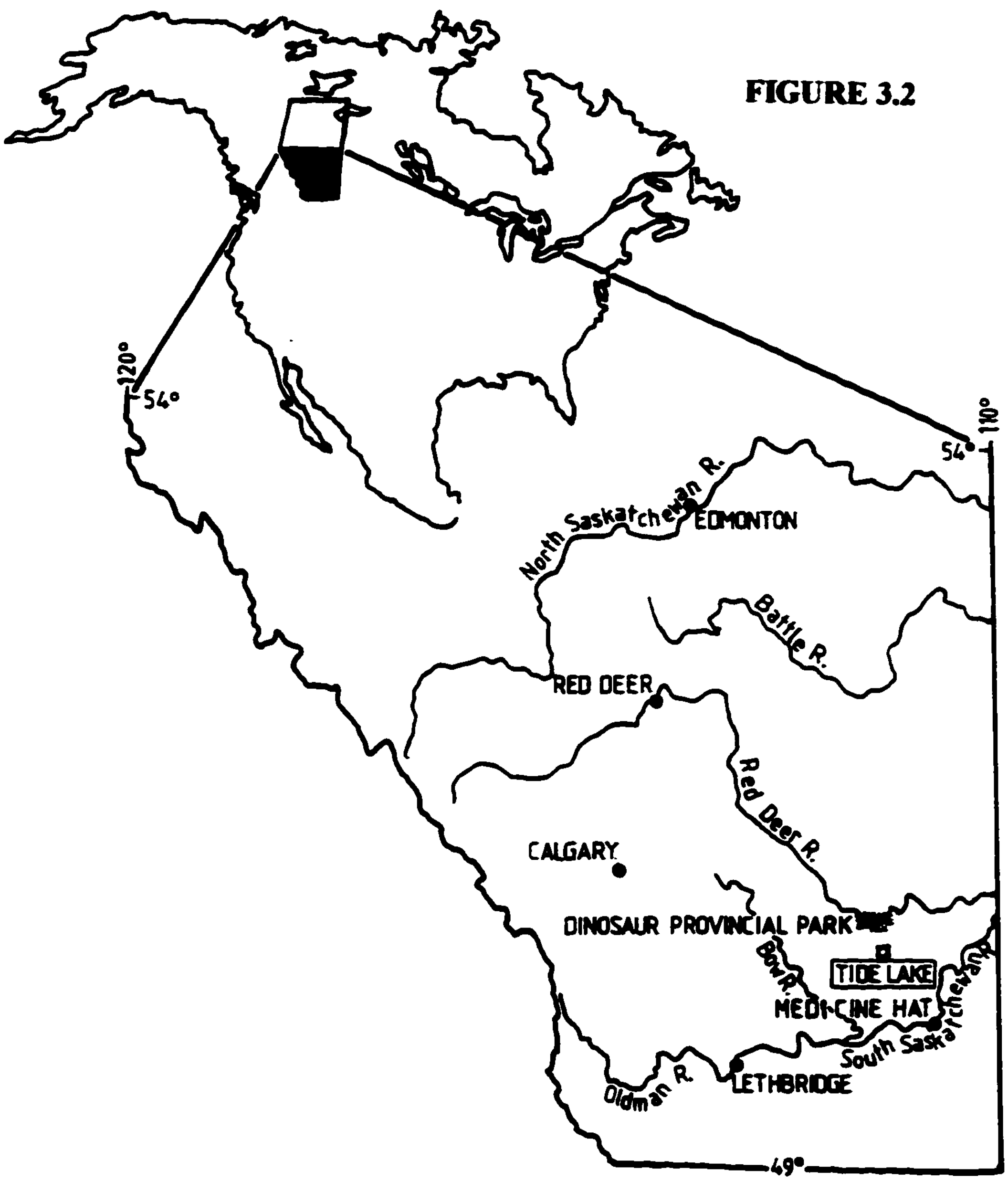

120°
54°
110°
54°
North Saskatchewan R.
EDMONTON
Battle R.
RED DEER
Red Deer R.
CALGARY
DINOSAUR PROVINCIAL PARK
Bow R.
TIDE LAKE
MEDICINE HAT
South Saskatchewan R.
LETHBRIDGE
Oldman R.
49°

FIGURE 3.3

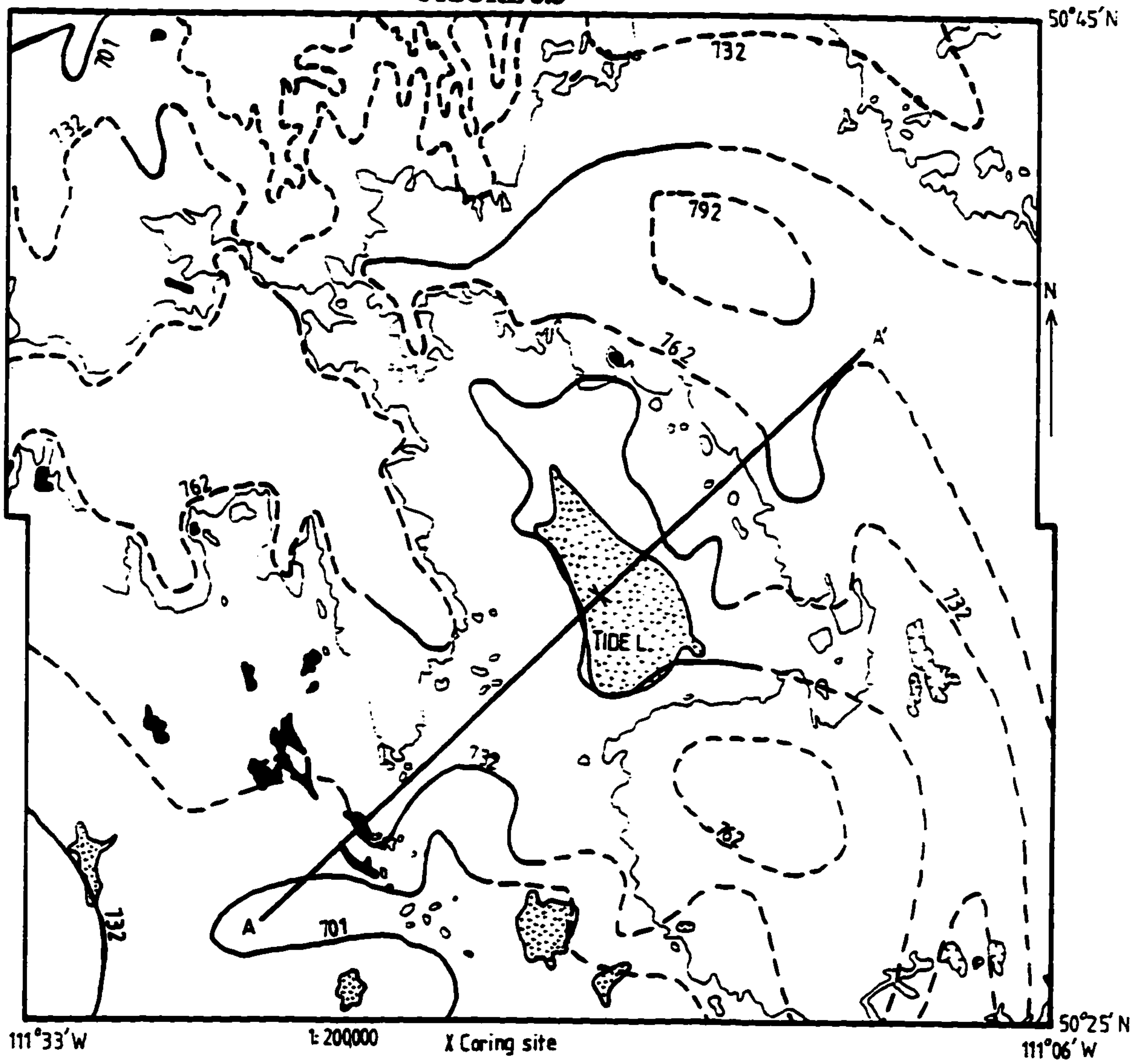

50°45'N
701
732
732
792
N
762
A'
762
TIDE L.
732
732
762
732
A
701
50°25'N
111°33'W
1:200,000
X Coring site
111°06'W

FIGURE 3.4

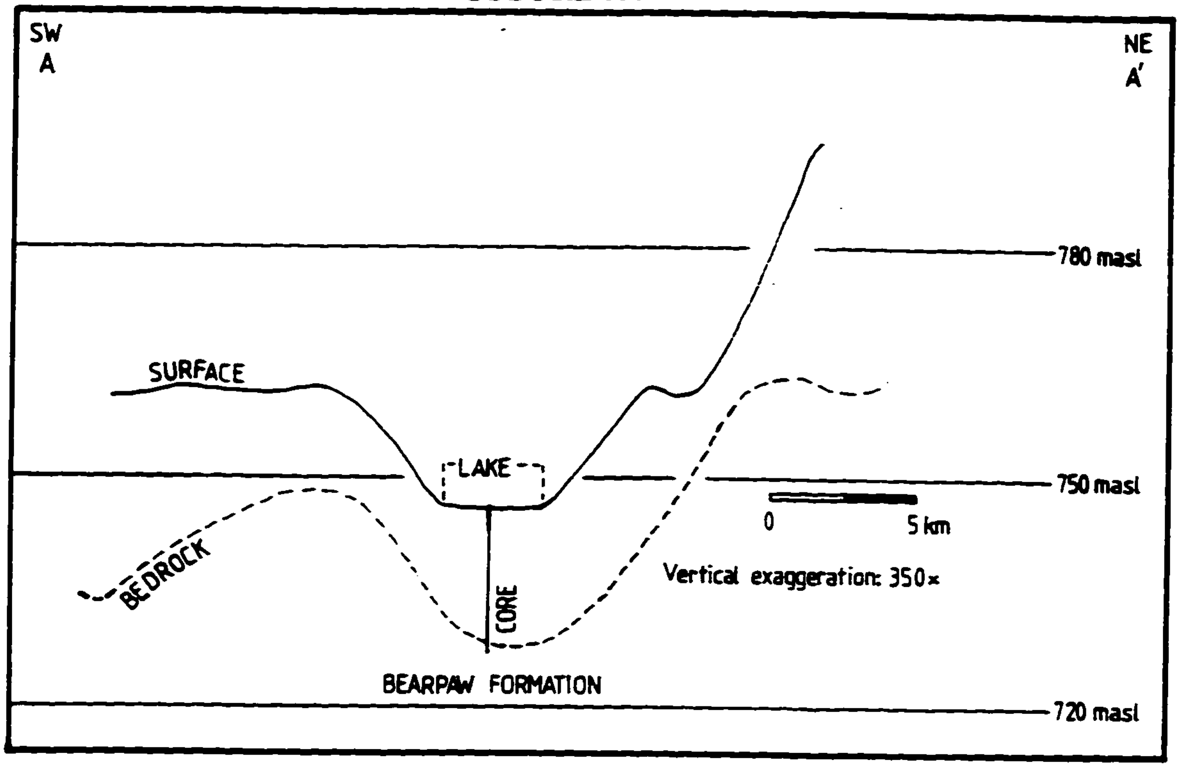

SW
A
NE
A'
780 masl
SURFACE
LAKE
750 masl
0
5 km
BEDROCK
CORE
Vertical exaggeration: 350×
BEARPAW FORMATION
720 masl

FIGURE 3.5

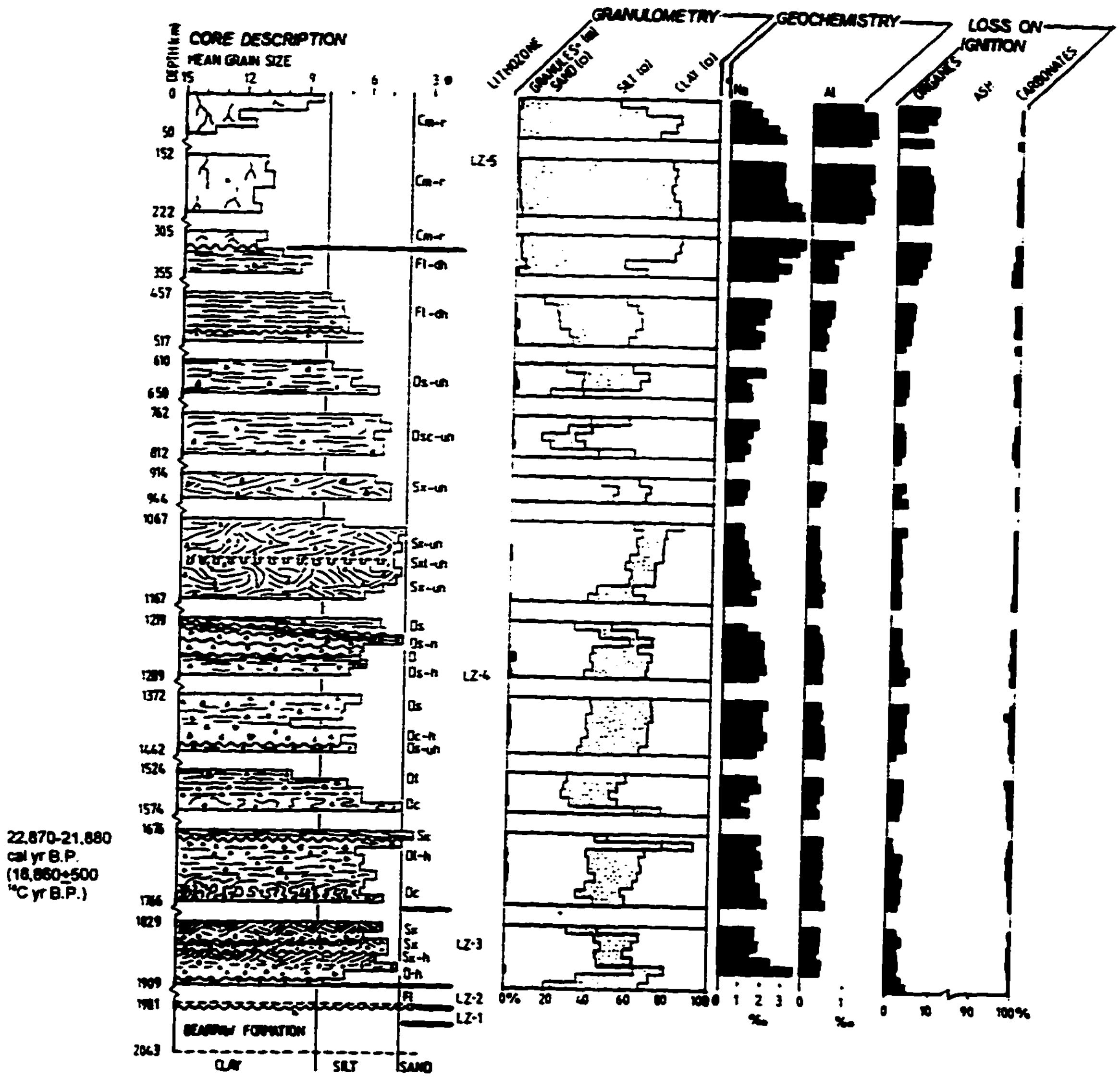

CORE DESCRIPTION
MEAN GRAIN SIZE
DEPTH (cm)
LITHOZONE
GRANULOMETRY
GEOCHEMISTRY
LOSS ON IGNITION
SILT (%)
CLAY (%)
Al
ORGANICS
ASH
CARBONATES
LZ-5
LZ-4
LZ-3
LZ-2
LZ-1
22,870-21,880 cal yr B.P. (18,860±500 14C yr B.P.)
BEAUFORT FORMATION
CLAY
SILT
SAND

FIGURE 3.6

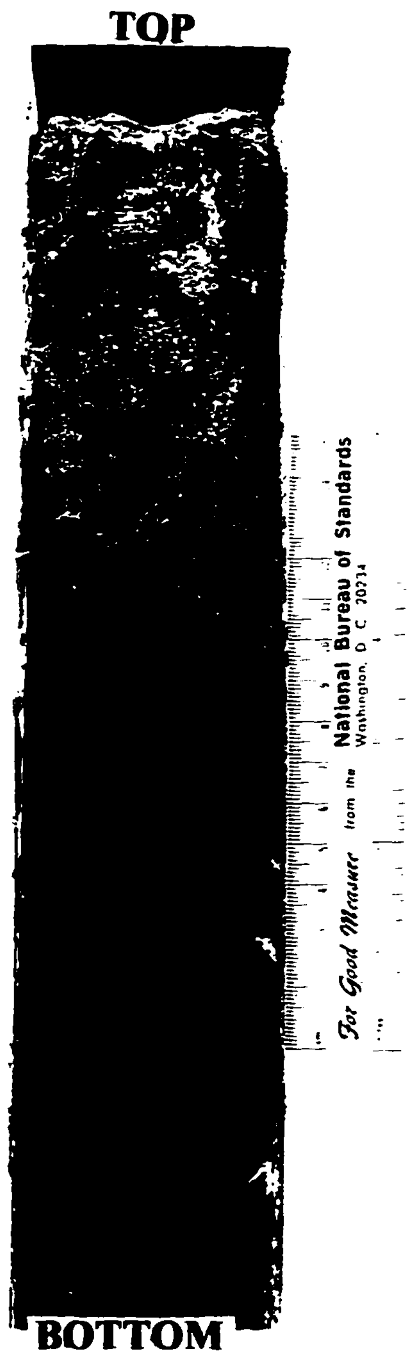

TOP
BOTTOM
For Good Measure from the National Bureau of Standards
Washington, D.C. 20234

FIGURE 3.7

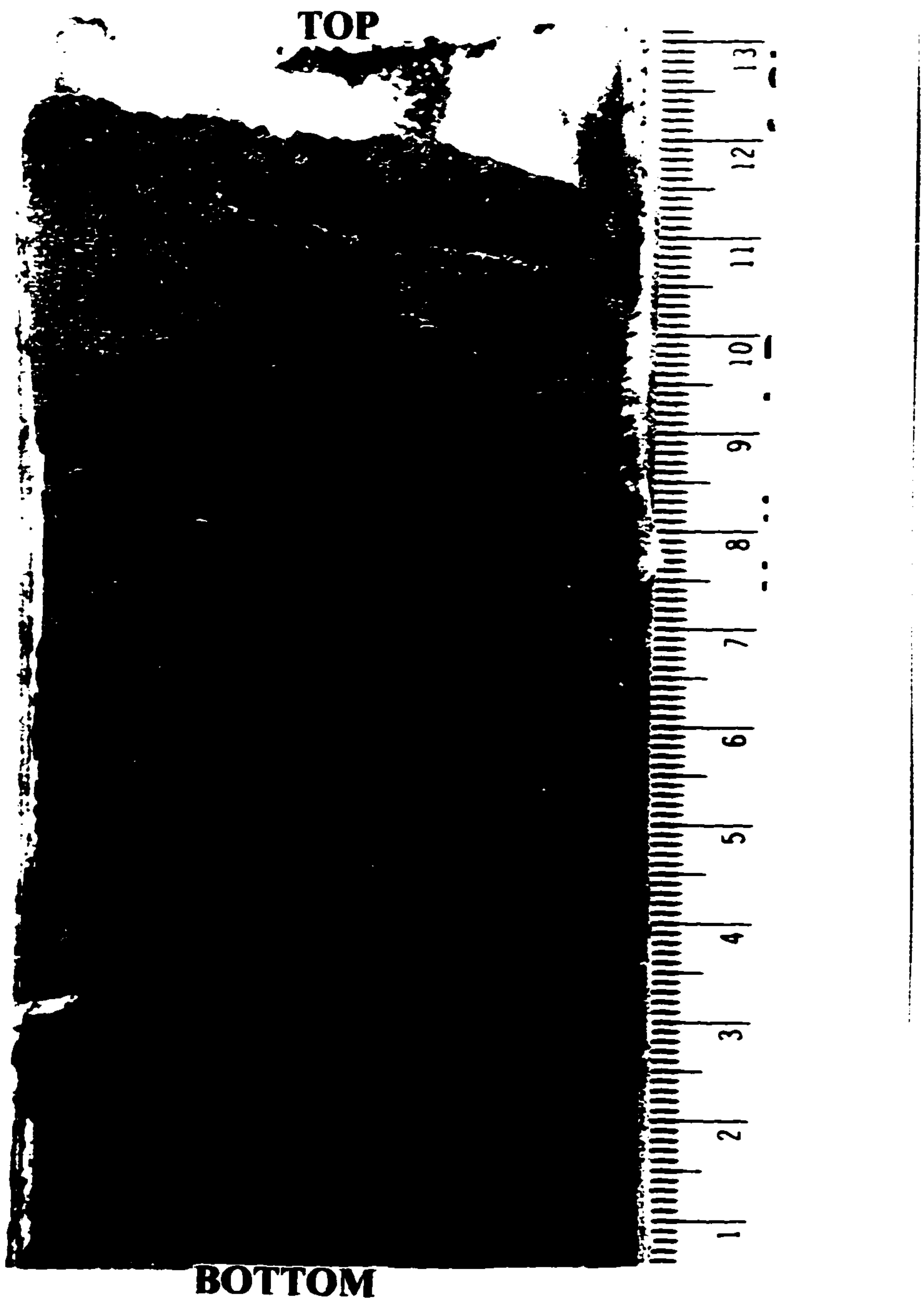


FIGURE 3.8

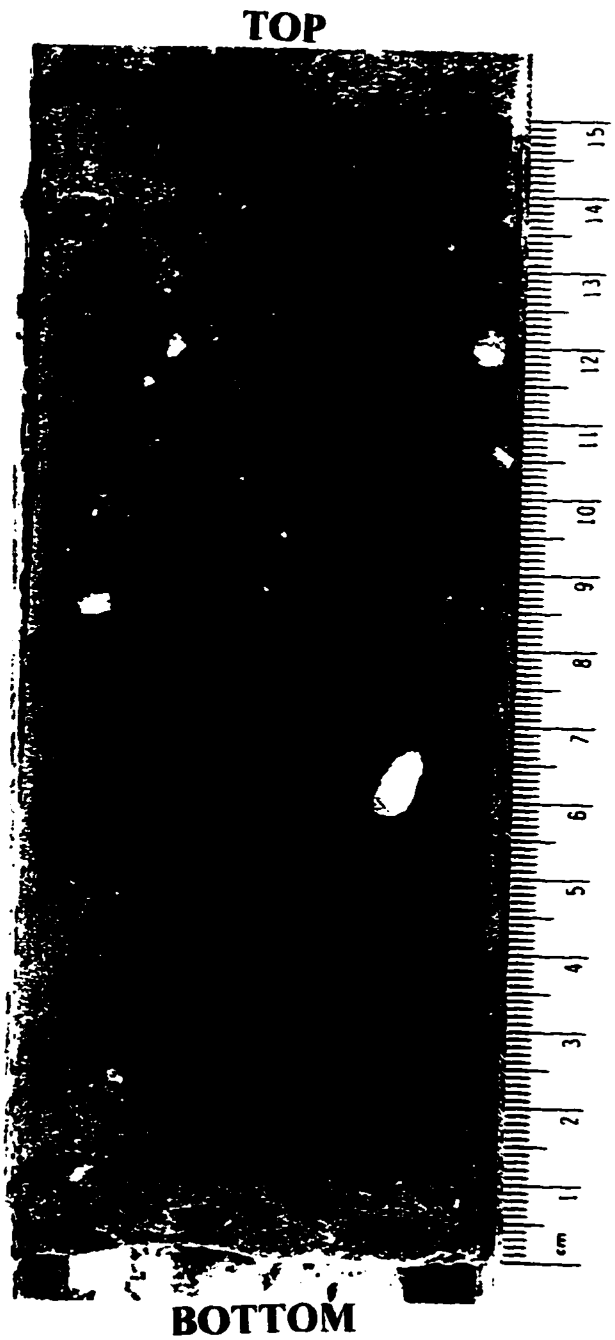


FIGURE 3.9

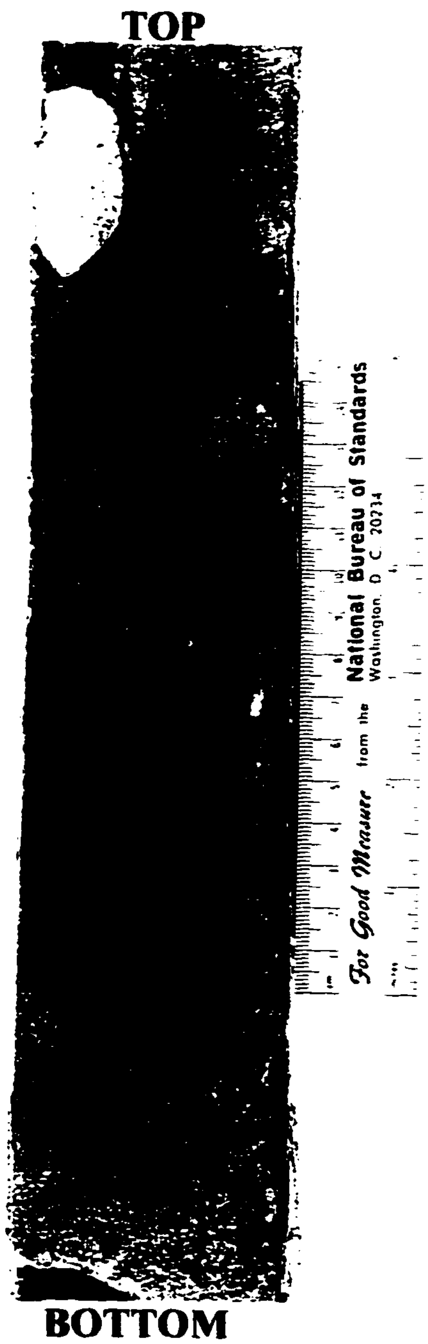

TOP
For Good Measure from the National Bureau of Standards
Washington, D C 20234
BOTTOM

FIGURE 3.10

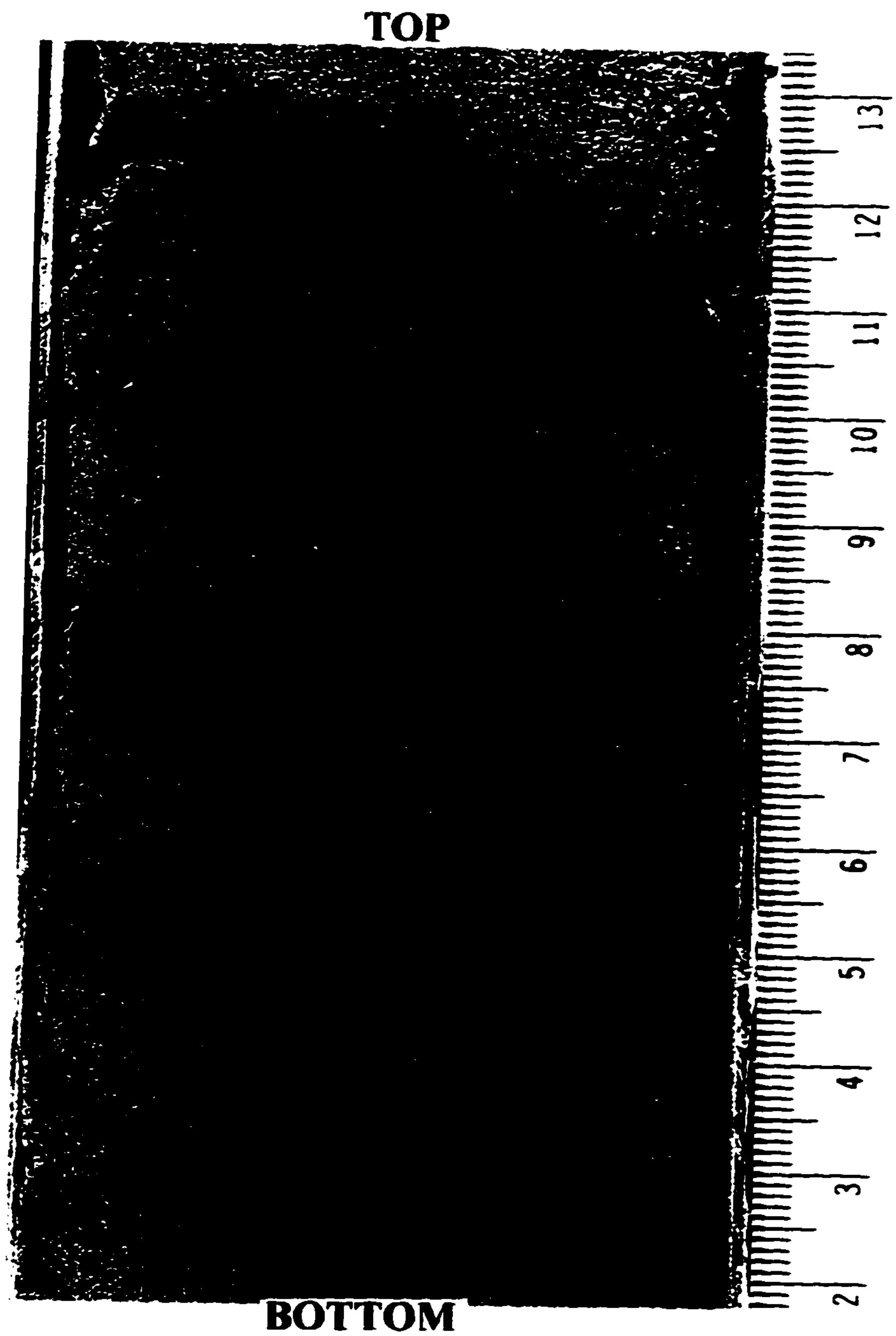


FIGURE 3.11

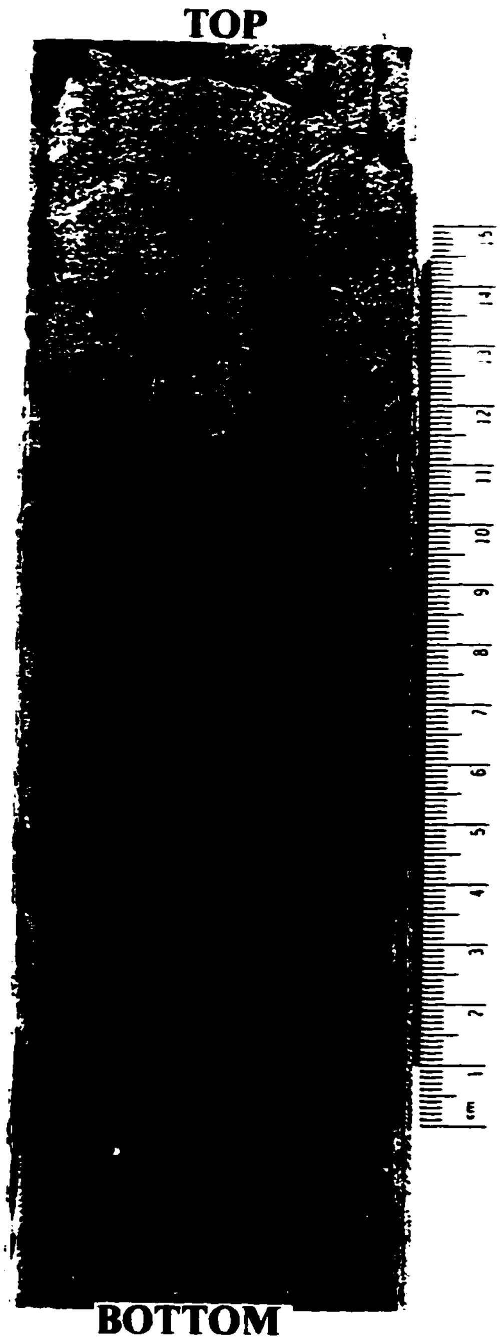
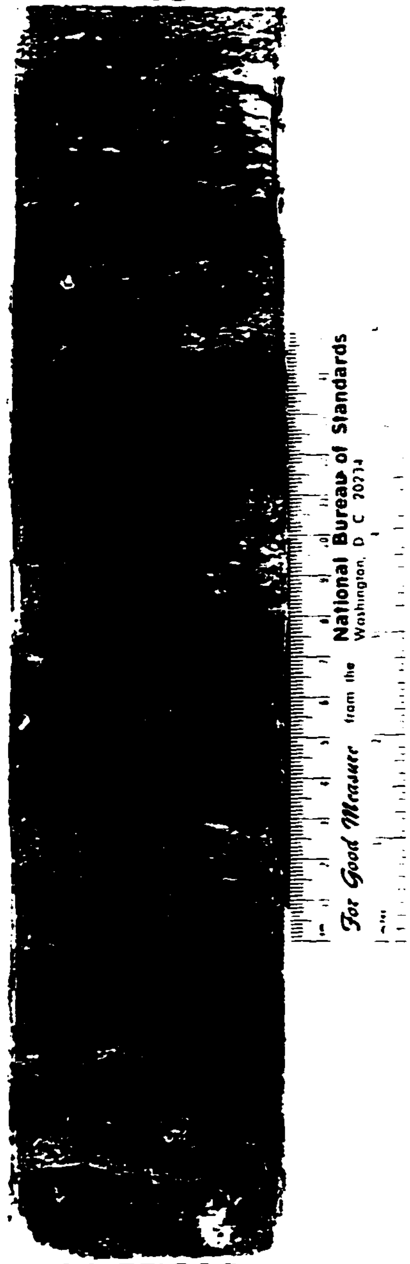


FIGURE 3.12

TOP



BOTTOM

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CHAPTER 4
POSTGLACIAL STRATIGRAPHY OF A FINE-GRAINED ALLUVIAL FAN IN THE RED DEER RIVER VALLEY IN THE PRAIRIES OF ALBERTA, CANADA
4.1 INTRODUCTION

Fine-grained, low angled alluvial fans line the lower Red Deer River valley in southern Alberta, Canada, where they have typically aggraded onto floodplain (e.g pointbar) deposits. The sediment supply to these fans derives from drainage basins which extended into the prairie surface following entrenchment of the Red Deer River by glacial meltwater. Alluvial fan systems are composed of three parts: (1) a drainage basin (zone of erosion); (2) a single confined tributary channel (zone of transport); and (3) an alluvial fan (zone of deposition). The fans form where sediment and water channelled into confined tributaries emerge from valley side catchments onto the valley floor. The change in channel geometry and confinement leads to a decrease in stream power and flow competence, causing sediment deposition and storage (Bull 1977). Sediments derived from aeolian input and river overbank deposits can also contribute to fan aggradation. Fan maintenance and evolution are dependent on the balance between inputs, throughputs and outputs of sediment mass and stream energy in each part of the fan system.

Most of the fans in the lower Red Deer Valley are destroyed within 150-900 years of fan initiation (Seeman 1993), by basal sapping due to downstream pointbar / sweep of meander migrations of around 2.8-3.3 m^{-2} yr^{-1} (Neill 1965; Kondla and Crawford 1971; Seemann 1993), or by piping or headward entrenchment (Seemann 1993). Hence the majority of fans in this region are relatively young. Seemann (1993) calculated that development of the fans occurs on an approximately 1100 year cycle, independent of climate, controlled mainly by river migration. Marken's (1993) study on cottonwoods on the lower Red Deer River affirms Seemann's (1993) conclusions.

The oldest fans identified in the region were studied by O'Hara (1986) and O'Hara and Campbell (1993) in a tributary valley of the Red Deer River. The fans sediments are extremely cohesive due to their high silt and clay content (O'Hara 1986). The layered fan material has been interpreted as sheet wash deposition, formed during fluvial activity over the fan surface. The fans are composed of vesicular layers which are believed to have formed as a result of multiple cycles of wetting and drying which causes the rearrangement of particles, air entrapment, and the formation of vesicles. Fan development was interpreted as having begun soon after valley incision under early Holocene arid conditions (O'Hara 1986). The absence of Mazama tephra deposits (a regional stratigraphic marker) was interpreted as indicating fan deposition ceased prior to *ca.* range 7800-7480 cal yr B.P. (calibrated from Bacon 1983) (O'Hara 1986). Fan deposits are topped by a loess deposit dated nearby in Dinosaur Provincial Park at *ca.* range 5400±800 TL yr B.P. (Bryan *et al.* 1987). Alternatively, the fans may have formed *ca.* 7800-7480 cal yr B.P. to 5400±800 TL yr B.P. All ^{14}C yr B.P. dates used in this paper have been calibrated to cal yr B.P. using CALIB 3.0.3 (Stuiver and Reimer 1993). The date range reported here is that which yields 100 per cent probability when one standard deviation is used.

This paper describes the postglacial evolution of a topographically constrained bajada (coalescing alluvial fans) in the Red Deer River valley, Alberta (Figure 4.1). This area of fan deposition is separated from the river by a large paleochannel gravel bar and has thus been protected from both river migration and headward entrenchment due to river base level changes. The purpose of this paper is to determine the succession of depositional environments observed in a core taken from the fan, and to provide insights into the dynamics of alluvial fan development. These developments are interpreted on the basis of internal structures, Munsell colours, grain size, grain roundness, loss on ignition, charcoal abundance, bulk authigenic geochemistry, mineralogy, and microscopic charcoal.

4.2 STUDY AREA

The study area is located in a chernozem-soil dominated, mixed-grass prairie in the northern Great Plains (Strong and Legget 1992). The area has a midlatitude, semiarid, and continental climate with long cold winters and short warm summers (Table 4.1). Mean annual precipitation is <350 mm yr^{-1} (Table 4.1). Average potential evapotranspiration often exceeds precipitation by >300 mm yr^{-1} (Winter 1989). The study area is designated as BSk (cold winter steppe; I.A. Campbell 1974) in the Köppen Climatic Classification and Sb14 (semiarid; precipitation concentrated in summer; mean annual temperature between 0-10°C when plant growth begins; cold winter) under the Meigs (1952) scheme devised for the UNESCO Arid Lands Research Series. Much of the limited annual precipitation (*ca.* 70-75 per cent) falls during the summer (Table 4.1) and although some falls as short, relatively intense (20-30 mm h^{-1}) convective storms, most precipitation results from prolonged low intensity (1-5 mm h^{-1}) storms (Bryan *et al.* 1988). High intensity precipitation events are rare, although an intensity of 80 mm h^{-1} for six minutes was recorded at the start of a prolonged storm in Dinosaur Provincial Park in 1981 (Bryan *et al.* 1988). The peak recorded precipitation event at the Brooks AHRC climate station (which has the longest record in the region) brought 26 per cent of the mean annual precipitation in one day (Environment Canada 1993). Approximately 35 per cent of the annual precipitation falls as snow, but sublimation and chinook - related melts usually deplete the snow-pack. As a result, spring snowmelt, ranging from late March to early May, varies considerably in magnitude from year to year (Harty 1984).

The regional drainage and flat to undulating prairie topography is governed by the sub-horizontal, poorly consolidated Upper Cretaceous Judith River (formerly Oldman) Group, which formed in shallow water, lagoonal or deltaic environments (Koster 1984; Eberth *et al.* 1996). The lithology of the Judith River Group is dominated by highly erodible, sodic, smectitic mudstones, claystones, siltstones and fine sandstones (Bryan *et al.* 1987). Grain size analysis of Judith River sandstones shows no grains larger than 250 μm (Bryan *et al.* 1984); the clay content of the shales ranges from 30-90 per cent with *ca.* 35 per cent of the clay fraction being <0.3 μm. The dominant clay mineral is montmorillonite, although illite and kaolinite may be present in significant amounts (I.A. Campbell 1987a). Cation adsorption analysis shows that Na and Ca are dominant, and that K, Mg and Al are present (Hodges and Bryan 1982; de Boer and Campbell 1990). Given the regional lithology, "potential sources for the cations are gypsum, calcite and dolomite for calcium and magnesium; the cation exchange complex for sodium and potassium; and the clay mineral lattice for aluminium" (de Boer and Campbell 1990:388). Runoff chemistry is dominated by Na due to Ca from gypsum replacing Na in the cation exchange complex (de Boer and Campbell 1990). The dominance of Na results in the swelling and dispersion of montmorillonitic clays, which is conducive to high erosion rates (Bryan *et al.* 1984). As a result, the valley walls are dominated by spectacular badland topography.

The Judith River Group is overlain by a veneer of Quaternary fluvial, glacial and glaciolacustrine deposits (Evans and Campbell 1992) varying in depth up to 100 m thick (in preglacial valleys) along the Red Deer River valley in the study area (Tokarsky 1986). Since European settlement began in the late 1800s, there has been widespread modification of the prairie vegetation (by damming, dugout construction, irrigation projects, and agricultural practices).

The early postglacial geomorphic evolution of the Red Deer River region is presented in McPherson (1968), Bryan *et al.* (1987), I.A. Campbell and Evans (1990), Evans and Campbell (1992), O'Hara and Campbell (1993), Paterson (1996) and C. Campbell and Campbell (1997). Deglaciation may have begun as early as *ca.* 20,210-19,430 cal yr B.P. in the study region (AECV-681C, calibrated from Evans and Campbell 1992, 1995). Ice disintegration and recession, associated with ice dam failures, drained the proglacial lakes, which decanted, forming extensive meltwater channels (Kehew and Lord 1986; Bryan *et al.* 1987; Teller 1987; Paterson 1996), including that now occupied by the Red Deer River.

In the Dinosaur Provincial Park region, Bryan *et al.* (1987) suggest that as the ice margin disintegrated, meltwater supply to local drainage swiftly abated (possibly accentuated by glacioisostatic rise), causing deep valley incision. Incision occurred in two stages. First, a broad valley stage removed 15-20 m of the prairie surface by meltwater and drainage of a large, ice dammed lake. This broad valley stage was followed by deep incision up to 30-50 m below the top of the present alluvial fill. McPherson (1968) identified, but did not date, gravel deposits associated with this phase of downcutting. The gravel, which lies up to 40 m below the bed of the present Red Deer River. River incision was followed by aggradation (McPherson 1968). Evans and Campbell (1995) demonstrate that some of the valley fill used in McPherson's (1968) interpretation may predate the late Wisconsinan, casting doubt on McPherson's interpretation.

David (1964) and Stalker (1973) suggest that either glacioisostatic rebound or the effects of ice damming to the east raised the Red Deer River's base-level, resulting in river aggradation. The maximum ice thickness in this area was *ca.* 750 m (Rains *et al.* 1990; I.A. Campbell *et al.* 1993). Taking an approximate ice density of 1.0 and an approximate mantle rock density of 3.3 (Walcott 1970), this suggests a maximum isostatic depression of *ca.* 227 m. Analysis of ^{14}C dates indicates that it is unlikely that the ice-load was present long enough for the crust to reach this maximum depression (Chapter 3), so that any postglacial isostatic rebound was almost certainly much less than 227 m. Walcott (1970) suggests total isostatic depression can be estimated at 65.21 m for a 750 m thick ice sheet. Kugler and St.-Onge (1973) suggest that there has been differential isostatic rebound of 40 m per 100 km in southern Saskatchewan and 450 m of total isostatic depression at Eskimo Point, on the western shore of Hudson Bay, where the late Wisconsinan depression would have been greatest.

Isostatic rebound would affect the stream profile, and a river could theoretically incise to maintain its profile as the landscape is lifted. That the Red Deer River has incised by approximately 60 - 80 m, while Walcott (1970) estimated isostatic depression at roughly the same value, may be entirely coincidental, or may relate to the extreme erodability of the valley sediments. This erodability would be a result of both the poorly consolidated pre-Quaternary sediments, and the effectively unconsolidated Quaternary preglacial materials filling the paleovalley which the Red Deer River is in part re-incising. However, as Bryan *et al.* (1987:145) note: "It does not appear necessary to invoke external factors, such as land-level changes, to explain the onset of aggradation, though they may have been involved." Valley aggradation was followed by diminishing discharges and the formation of alluvial fans along the lower Red Deer Valley (McPherson 1968) and its tributaries in Dinosaur Provincial Park (O'Hara and Campbell 1993).

4.3 THE KLASSEN SITE

Originally identified as an abandoned river terrace (McPherson 1968; Berg and McPherson 1972), the Klassen site (unofficial toponym, after the present owner, Paul Klassen) is located on an abandoned paleochannel of the Red Deer River on the north side of the Red Deer River valley (50°, 52' N, 111°, 15' W; Figure 4.1). Located at a mean altitude of 640 m asl, the valley walls rise to the prairie surface *ca.* 709 m asl with a mean gradient of 8° north of the site. At least ten, clearly identifiable, fine grained alluvial fans have coalesced forming a bajada across the site. Granulometric analysis of 21 surface samples shows that there is a general fining from fan apices to toes (Figure 4.2). This finding is consistent with Seemann (1993) who shows that alluvial fans along the Red Deer show a downfan decrease in grain size. Part of the Klassen bajada is bounded to the south (at the distal limits) by a relic mid-channel gravel bar (Figure 4.3). The area of the bar is 405,000 m^2 (measured using an Ushikata X-PLAN360d planimeter on an aerial photograph). The bar appears to somewhat constrain the course of the river; thus most of the fans in the study area have been at least partially preserved from the destruction by the downstream meander migration reported by Seemann (1993). Although there is evidence of river incision in the form of two minor, river terraces on the river side of the bar and truncation of distal fans past either end of the bar, the essentially closed nature of the central part of the bajada has prevented channel shifting from significantly affecting the evolution of this bajada. In addition, possible changes in base level, suggested by headward entrenchment of fan tributary streams elsewhere along the Red Deer, appear not to have affected the fan tributaries behind the bar. This protected part of the bajada is the primary focus of this study.

In the study area, gullies forming the source-basins for the bajada have cut through the *ca.* 5 m thick surficial Quaternary deposits, down into the Judith River bedrock. The gullies have poorly vegetated slopes and well vegetated floors. Bajada source basin area, delimited by aerial photography, is 2.85 X 10^6 m^2. Volumetric loss from the basin is 4.86 X 10^7 m^3 (determined as Area X Mean depth from prairie surface [17.06 m], from the elevations obtained on a 1:50,000 topographic map at 100 random locations within the basin). Thus assuming approximately 13,000 years of subaerial exposure, the mean rate of erosion in the study area is 3.7 X 10^3 m^3 or 1.3 mm yr^{-1}.

Long term process geomorphic studies on the pattern of erosion over a wide range of slopes and lithologies in the Dinosaur Provincial Park badlands just west of the Klassen site show that average denudation rates on open-system 1 m^2 plots are approximately 4 mm yr^{-1}, with a peak annual erosion rate of 13 mm yr^{-1} (I.A. Campbell 1974, 1981, 1982; Bryan and Campbell 1986; I.A. Campbell *et al.* 1993). The higher erosion rates in Dinosaur Provincial Park may be a function of the more barren Dinosaur Provincial Park badland surface. Alternatively, the higher rate of erosion measured in Dinosaur Provincial Park may be a function of the brief period of measurement (10 years). Bryan *et al.* (1984) suggest that denudation rates varied throughout the postglacial, but appear to have been highest during the immediate postglacial period. In Dinosaur Provincial Park, denudation almost ceased after extensive loess deposition *ca.* 5400$\pm$800 TL yr B.P. (Alpha-2070), but rates now appear to be increasing because the loess cover has been largely stripped by streams, piping and tunnel erosion (Bryan *et al.* 1984; O'Hara and Campbell 1993). Thus erosion, and hence deposition rates, at the Klassen site are also likely to have been somewhat variable throughout the postglacial period.

The surface area of the bajada constrained by the bar is 0.66 km^2. The contributing fans have a mean radius of 760 m with a rise of 19 m; their slope is a gentle 1.4°. The volume of the constrained portion of the bajada (slightly less than half the bajada) is calculated at 5.3 X 10^6 m^3 (Area X Mean cross-sectional depth [8 m]); whereas the contributing basin eroded volume is measured at 1.9 X 10^7 m^3, representing a 72 per cent difference. The bajada thus contains a maximum of 28 per cent of the sediment from the gullies. The difference between bajada and contributing basin volumes may be partly due to errors in the estimates of mean bajada thickness and of mean gully depth and to losses prior to channel migration or to sediment loss around the bar.

The surface of the bajada is covered by a distributary, semi-radial pattern of braided streams. These spread out of the fans' apices which are partially obscured by the implementation of centre-pivot well-irrigation farming in 1993 (Figure 4.4). The majority of sediment is deposited on the fans in sheets during spring melt, not during rainfall events (Klassen, personal communication 1995). Klassen's observation agrees with regional process geomorphology studies in Dinosaur Provincial Park, which indicate that grassed surfaces have exceptionally high infiltration capacities (>6 mm h^{-1}, Hodges 1982) and produce runoff only in the most extreme or extended storms or during spring melt when the subsurface is still frozen (Hodges and Bryan 1982; Bryan *et al.* 1988; de Boer 1990). I.A. Campbell (1974) notes that winter frost cycles disaggregate the surface, raising and detaching surface grains and aggregates. These frost-disturbed materials are then eroded and transported down slope during spring and summer runoff. Harty (1984) has noted that snowmelt runoff efficiency is greater than that of rainfall, and is more erosive, due at least in part to the loosening of surficial material by freeze-thaw processes and the inhibition of infiltration by the frozen ground.

Klassen (personal communication 1995) also noted that during the dust storms in the 1980s dust from the prairie surface adjacent to the site blew into the river valley: "the sky was so thick with dust, it blocked out the sun." Research elsewhere in the northern Great Plains suggests that, during the late Holocene, there were many cycles of aeolian activity followed by pedogenesis (Ahlbrandt *et al.* 1983; Vreeken 1986, 1989, 1993; Osterkamp *et al.* 1987; Madole 1994; Wolfe *et al.* 1995); thus it is likely that dust storms similar to those that have occurred historically in the study area also occurred prior to European settlement as indicated in Bryan *et al.* 1987. Figure 4.5 shows the mean cumulative grain size curve of 30 loess deposits from Dinosaur Provincial Park (Bryan *et al.* 1987) compared to that of 33 surface samples obtained from the prairie surface north of the Klassen site (see also Chapter 6). It appears that if the prairie surface is the source for the wind blown dust, as suggested by Klassen (personal communication 1995), then the wind is dominantly eroding the fine sands and silts, leaving behind the coarser and finer materials which are characteristic of the regional prairie surface. Furthermore, while aeolian material is being deflated from the prairie surface, it is being deposited in regional topographic depressions and areas with high surface roughness such as Dinosaur Provincial Park (Bryan *et al.* 1987) and on cliff tops (Vreeken 1993; David 1993), where reduction in wind velocity and decreased turbulence result in aeolian deposition. Where erosion occurs, the prairie surface becomes a lag surface, although accumulations of coarser aeolian material (sand dunes), too heavy for long-distance transport, occur adjacent to major sand dominated source areas (David 1977; Mulira 1986; Shetsen 1987).

Historically, the fans have received sediment from sheetwash (transported by the braided streams which cover the bajada surface), aeolian input, and more recently the unfiltered sprinkler system, which spreads a thin layer of clay and silt-sized sediment onto the site each year (Klassen, personal communication 1995). Based on the regional geomorphic research and the verbal description by the landowner, the bulk of the fan deposits should record sediment deposition by spring melt, infrequent, high-intensity precipitation, and rare aeolian events.

4.4 METHODS

A truck-mounted, five cm diameter, hollow stem auger (Figure 4.6) extracted 15.35 m of core from a 16.07 m hole in the central portion of the bajada, 850 m north of the present Red Deer River. The core was stopped at 16.07 m in coarse gravels just above the present river level (630 m a.s.l.). The core was contained in plastic sleeves, frozen, and was later split lengthwise using a rock saw. One half of the core was then cut into 307 contiguous 5 cm slices. Each 5 cm slice was homogenised using a rubber tipped mortar and pestle, air dired, and stored at room temperature in a plastic bag. The other half of the core was archived in a -15^{0}C freezer. The archived half of the core was described using a variant of the Eyles *et al.* (1983) code for sediment description. Selected sections were photographed. For each of the 307 samples, wet and dry Munsell colours, granulometry, soil texture classification (Hodgson 1976), grain roundness, loss on ignition, per cent charcoal and bulk authigenic geochemistry were determined.

Granulometry was determined using approximately 60 grams of each homogenized sample treated with 25 ml of 30 per cent H_2O_2 at 90°C to remove organics. When oxidation ceased, samples were oven-dried for at least 3 hours in a drying oven at 50°C. Fifty grams of sediment were then weighed and screened at 1 phi intervals using a rotap for 15 minutes for the sediments >300 μm and on a Sedigraph 5100 for the $\leq$300 μm fraction.

Grain roundness was determined after Powers (1953) on 200 random grains per sample (excluding clay and very fine silt). Loss on ignition and $CaC0_3$ content were determined after Dean (1974). Per cent charcoal was determined using the Winkler (1985) method of nitric acid digestion / combustion. Bulk authigenic geochemistry was determined by ICP-AES from a 0.3 N HCl extraction of bulk sediment (Malo 1977) in order to ascertain the concentration of P, which, in the absence of detrital phosphorus-bearing minerals, is an indication of organic productivity. Na, K, and S were also determined to detect possible evaporitic events.

A glass:crystal count of 200 grains per sample, identified with a petrographic microscope, indicated the occurrence of a number of potential tephra beds. Only three of the potential 8-10 glass peaks (Figure 4.7) had sufficiently abundant coarse-grained glass shards to be easily separated for electron microprobe analysis. These were separated from the enclosing sediment by hand panning in distilled water on a large watchglass. Grains other than glass (such as magnetite) were deemed unsuitable for analysis due to the difficulty of ensuring that they were part of the tephra rather than contaminants derived from the enclosing fan sediments; crystalline grains sutured to glass were too rare in the small samples available for useful analysis. Electron microprobe analysis was performed on an ARL-SEMQ electron microprobe fitted with a Tracor Northern 5502 operating system, in the Department of Earth and Atmospheric Sciences at the University of Alberta. It was operated at an accelerating potential of 15.0 kV, a beam current of 10 nanoamperes, and a 10 second integrating time. The beam was rastered for each count, with the target area typically less than 25 μm^2. A ϕ-ρ-z data reduction was used with glass standards USNM 111240/52 (Juan de Fuca Ridge basaltic glass), USNM 72854 (natural rhyolitic glass), and USNM 2213 (synthetic high-Si tektite). Tephra bed designation was determined by Andrei Sarna-Wojcicki (personal communication 1997).

Microscopic charcoal and mineralogy were identified for selected samples from each lithozone (see below). Microscopic charcoal was taxonomically identified from selected samples using a scanning electron microscope, with the aid of reference pictures of microscopic charcoal fragments of regional species. The charcoal appears to preserve microscopic features very well; however, as an atlas for this region has yet to be fully developed, identification is here limited to conifer, deciduous or monocot. The mineralogy of 10 selected, clay sized samples was determined by X-ray diffraction (XRD) on a Rigahu Geigerflex Unit (vertical goniometer) with a cobalt emanating tube at the Department of Earth and Atmospheric Sciences, University of Alberta.

4.5 STRATIGRAPHY

4.5.1 DESCRIPTION

Seven lithozones were distinguished based on the combined results of all the data sets and delimited by eye based on trends seen in Figures 4.7.A-4.7.B and Table 4.2. A lithozone is "an informal term to indicate a body of strata that is unified in a general way by lithologic features but for which there is insufficient need or information to justify its designation as a formal unit" (Bates and Jackson 1987). Figures 4.8-4.15 show examples of some of the structures referred to in the description.

4.5.1.1 LITHOZONE 1

Lithozone (LZ) 1 is 1607- 1587 cm below the modern surface. The auger was stopped within LZ-1 by a 5 cm well-rounded clast, above which there are matrix-supported, generally very well rounded gravels, which include clasts derived from the Canadian Shield. There is a fining upward sequence from gravel to silt loam. Clay content increases rapidly upwards. The coarser sand size fraction is dominated by well rounded grains and the finer silt size fractions are dominated by subrounded grains. Organic matter and charcoal abundance increase upcore. Carbonate abundance decreases upcore.

4.5.1.2 LITHOZONE 2

LZ-2 is 1587- 1122 cm below the modern surface. A broad double peak in glass abundance occurs from 1547-1517 cm. LZ-2 is dominated by massive silty clay loam and clay loam sediments with ped horizons; mud curls occur *ca.* 1402 cm. *Ca.* 1167- 1157 cm, there are a few fine silty laminations, and a red (2.5YR 4/8 - dry Munsell) coloured, organic-rich, sandy horizon with rootlets occurs *ca.* 1142- 1137 cm. Grains are dominantly subrounded. Gastropod shell fragments (too small for identification) were found near the base (*ca.* 1572- 1567 cm). Loss on ignition shows very stable, relatively high, amounts of organic matter and carbonates. Charcoal abundance reaches its highest values in this lithozone. P, Na, and K are stable, although they show a slight peak at *ca.* 1242 cm. Identifiable charcoal consists of fragments displaying bordered pits (conifers, cf. *Picea* spp.) and sieve plates (hardwoods and/or deciduous shrubs). The dominant mineral is quartz with small amounts of microcline, albite, anorthite, dolomite, calcite, gypsum, chromian clinochlore, montmorillonite and illite.

4.5.1.3 LITHOZONE 3

LZ-3 is 1122-897 cm below the modern surface. It consists of fine-grained massive silty clay loam, with horizons dominated by silt loam, and sandy silt loam starting at 1037 cm. Peds occur at *ca.* 1112- 1082 cm, *ca.* 1012 cm, and *ca.* 987-947 cm. Several erosional contacts occur above 1047 cm, as do brief intervals of thin horizontal laminations. A thin sand wedge occurs at *ca.* 1042 cm. The sand content increases upwards through this lithozone. There is a general shift towards greater angularity in grain roundness. Loss on ignition shows a slight increase in carbonates at the expense of organic matter. P is high throughout LZ-3, and there are several pronounced peaks in Na, K and S, all of which reach their highest values in LZ-3. Charcoal abundance decreases from LZ-2. Identifiable charcoal consists of monocots (grasses). The dominant mineral is quartz with small amounts of albite, sanidine, anorthite, dolomite, calcite, gypsum, chromian clinochlore and illite.

4.5.1.4 LITHOZONE 4

LZ-4 is 897-577 cm below the modern surface. Two glass abundance peaks occur *ca.* 702-697 cm and 727-722 cm. LZ-4 is a dominantly fine-grained, massive sediment, with subangular grains. There is a general upwards decrease in clay content. Although ped-rich horizons occur, they are less frequent than in LZ-3. Horizontal laminations are abundant from *ca.* 777-707 cm. Erosional contacts occur throughout LZ-4. Loss on ignition shows a general upwards decrease in carbonate content. There are several peaks in P, Na, K and S. Per cent charcoal values decrease upcore. Identifiable charcoal consists of monocots (grasses). The dominant mineral is quartz with small amounts of albite, anorthite, dolomite, gypsum, clinochlore, illite and montmorillonite.

4.5.1.5 LITHOZONE 5

LZ-5 is 577-320 cm below the modern surface. The clastic fraction is dominated by massive and laminated sandy silty loam. LZ-5 also contains granule sized grains. Grains are mainly subangular; angular and very angular grains reach their maximum abundance in LZ-5. Clay peds are rare, while erosional contacts are frequent. Loss on ignition shows the lowest organic and carbonate content in the core; charcoal abundance reaches its lowest values in the core. P, Na, K and S are all low. A segment from 532-470 cm was lost in coring due to lack of sediment cohesion, suggesting a low clay content. Rootlets and small carbonate concretions occur from 405-380 cm. Black organic rich bands occur between 390-380 cm. An A.M.S. date on grass roots at 355 cm dates to *ca.* 575-764 cal yr B.P. (760 $\pm$ 80 ^{14}C yr B.P., TO-5946. Identifiable charcoal consists of monocots (grasses) and sieve plates (hardwoods and/or deciduous shrubs). The primary mineral is quartz followed by small amounts of microcline, albite, dolomite, calcite, muscovite, clinochlore and magnesioriebeckite.

4.5.1.6 LITHOZONE 6

LZ-6 is 320-95 cm below the modern surface. LZ-6 marks the return to more massive silt, although sandy horizons are still abundant. Ped rich horizons again become frequent and horizontal laminations occur from 235-230 cm. Grey mottles occur from 185-120 cm, and black organic rich bands occur between 150-140 cm. Rootlets and small carbonate concretions occur throughout. LZ-6 marks the beginning of a rapid increase in grain roundness as well as a return to higher values of organics, carbonates, P, Na, and K. S remains low. Charcoal abundance shows a minor increase compared to LZ-5. Identifiable charcoal consists of monocots (grasses) and sieve plates (hardwoods and / or deciduous shrubs).The primary mineral is quartz with small amounts of microcline, anorthite, clinochlore, muscovite and dolomite.

4.5.1.7 LITHOZONE 7

LZ-7 is 95-0 cm below the modern surface. The sediment texture is friable, with red (2.5YR 4/8 - dry Munsell) mottling from 95-55 cm, and a shift in colour from red to weak red (2.5Y 6/2 to 2.5Y 4/2) 60 to 50 cm. The clastic fraction is dominated by silt sized, increasingly well rounded grains upcore. Deposits are massive. At 70-75 cm below the surface there is a relatively coarse-grained interval. Carbonates and organics are moderately high; rootlets and small carbonate concretions occur throughout. P is high, Na is low, K rises to a peak at the surface, and S is low. Charcoal peaks near the surface, and the identifiable charcoal consists of monocots (grasses) with some sieve plates (hardwoods and/or deciduous shrubs). The primary mineral is quartz with small amounts of dolomite, clinochlore, albite, anorthite and illite.

4.5.2 INTERPRETATION

The sediment delivery ratio is the discrepancy between the sediment yield at a particular point in a basin and the actual amount eroded above that point (Roehl 1962). Sediment delivery ratio is not equal across a basin, rather, small areas contribute disproportionate amounts of runoff and sediment (I.A. Campbell 1985). This partial area concept is illustrated by I.A. Campbell (1977) who has shown that at its terminal gauging station downstream of the Klassen site, the Red Deer River basin acquires more than 80 per cent of its total sediment load from less than two per cent of its basin area. Thus in a given basin, sediment, "like the runoff that produces it, is derived principally from spatially limited portions of the basin" (I.A. Campbell 1985:135). Furthermore, theoretically, each environment should have different spatially and temporally limited dominant (although usually not exclusive) partial areas of runoff and sediment supply, throughput, and storage. Thus, dominant partial contributing areas may change markedly through time and space at a given site. While the concept of partial areas has been applied spatially to a wide range of fluvial systems (I.A. Campbell 1985), little attention has been given to the change in partial area properties through time in one place. The sedimentology of the Klassen site is best interpreted in such a framework. Figure 4.16 depicts the following interpretations for each lithozone.

4.5.2.1 LITHOZONE 1

Figure 4.17 shows a cross section of the study area based on the depth of river gravels from the core, well logs (Alberta Environment, personal communication 1996), a field survey, airphoto interpretation and a 1:50,000 topographic map (Energy, Mines, and Resources Canada 1989). The correspondence of the elevation of the fan apex with a pronounced slope break on the south side of the river valley suggests that both north and south channels were occupied simultaneously and that the bar was likely a contemporaneous feature. Alternatively, the break in slope may be an structurally controlled.

LZ-1 (1597- 1587 cm; Figure 16-LZ-1) is a fluvial gravel deposit possibly associated with terminal Late Wisconsinan meltwater. At the base of the Klassen bajada, rounded gravels located in six well-logs between 599-624 m asl were probably contemporaneous with LZ-1 gravels (Figure 4.18) (Alberta Environment 1993). Assuming that the river stage reached 645 m (just above the gravel bar), that the river stretched from bank to bank, and assuming a nearly hydraulically smooth channel (which is reasonable given the likely depth of water relative to the roughness of the bed) with a Manning number of 0.020 the mean velocity of the Red Deer River when the base of LZ-1 was deposited was *ca.* 6.7 m s^{-1} with a discharge *ca.* $3.0 X 10^5$ m^3 s^{-1} (Table 4.3). By comparison, the modern Red Deer River has a mean discharge of $4.3 X 10^1$ m^3 s^{-1}; the peak historic instantaneous discharge was $1.9 X 10^3$ m^3 s^{-1} on June 27 1915 (measured at Red Deer gauging station; Water Survey of Canada, 1992). The Mississippi River has a mean discharge of 2.3 X 10^4 m^3 s^{-1} and the Amazon River 1.6 X 10^5 m^3 s^{-1} (Chorley *et al.* 1984:279). The discharge calculated here for the early postglacial Red Deer River likely occurred during a catastrophic proglacial lake drainage (possibly a jökulhlaup). Waning discharges resulted in channel shifting and abandonment. Following the deposition of LZ-1, channel abandonment created an oxbow lake (LZ-2).

4.5.2.2 LITHOZONE 2

Nemec (1990:3) defines a delta as "... a deposit built by a terrestrial feeder system, typically alluvial, into or against a body of standing water ...." Thus in the case of the Klassen site, as the terrestrial feeder is an alluvial fan, when fan sediments are being deposited into a standing body of water the delta is referred to as an alluvial fan delta. Overlying LZ-1, LZ-2 (1587- 1122 cm; Figure 16-LZ-2) comprises dominantly massive oxbow lake / alluvial fan delta deposits. Peds and mud curls provide evidence of periodic desiccation.

Electron microprobe analysis of the glass from the double peak at 1547- 1517 cm (Table 4.4) indicate that this tephra resembles the Glacier Peak G tephra bed (Sarna-Wojcicki, personal communication 1997), which has an age of *ca.* 12,980-13,460 cal yr B.P. (Mehringer *et al.* 1977). Thus deposition of LZ-1 and the start of LZ-2 deposition occurred prior to this date; therefore deglaciation throughout the Red Deer River basin was largely complete by *ca.* 13, 000 cal yr B.P. The oxbow lake implicit in LZ-2 may have been maintained by any one or combination of four mechanisms: by floods equal to or greater than modern floods; by a river which had not yet incised to its modern depth, by sediment sealing of the oxbow basin maintaining a dominantly meteoric-water pond independent of the river; or by high ground water still recently recharged by glacial meltwater. The first possibility, that there were floods equal to or greater than modern floods, seems plausible as valley glaciers in the mountains were still undergoing advance and recession (Reasoner *et al.* 1994). Historically, a flood caused the river to rise above the level of the present day Klassen house, *ca.* 625 m (Klassen, personal communication 1995). LZ-2 is located 624.3-629 m.a.s.l. The second alternative, that the river had not yet incised to its modern depth, is plausible. Sediment sealing of the oxbow, while likely given the high clay (including the swelling clay montmorillonite) and silt content throughout this lithozone, is unlikely to have been the sole agent given the altitudinal proximity to the modern floodplain and indications of regional aridity at this time (C. Campbell and Campbell 1997). The final possibility, that higher ground water levels were maintained by glacial meltwater, is reasonable but as yet unproven. In all likelihood all or any combination of these processes could have contributed to maintaining a semi-permanent oxbow lake.

The absence of any clearly identifiable alluvial fan deposits in this unit suggests that either the oxbow processes obscure the fan signal or that another source was dominant. A study was conducted to ascertain grain roundness (after Powers 1953) of the different regional sediment sources. Results indicate that Red Deer River overbank deposits are dominated by subrounded grains, loess from Dinosaur Provincial Park by subrounded and subangular grains, Quaternary deposits on the prairie surface by subangular grains and the Judith River Group bedrock sediments are characterised by rounded to well rounded grains (Table 4.5). This suggests that overbank deposits from the Red Deer provided the primary sediment and water source at this time, because the grains in LZ-2 tend to be subrounded. The probable dominance of river sediments is further suggested by the massive structure of the deposit, and the high percentage of clay.

The low variability in grain size, loss on ignition and P, Na and K implies fairly constant dominance of deposition by a single process or else a very stable balance of processes. The presence of dolomite, and particularly of gypsum, suggests the lake was saline and is consistent with periodic desiccation suggested by the mud curls and peds. That this pond periodically dried out suggests that whatever mechanism(s) maintained the lake was not constant during deposition of LZ-2.

Although the abundance of microscopic conifer charcoal may indicate the presence of conifers at the site, it more likely represents input from overbank floods which possibly contributed charcoal obtained from as far west as the foothills region.

The juxtaposition of clay peds, rootlets and rust coloured staining near the top of LZ-2, *ca.* 1142- 1132 cm suggests increasing subaerial exposure, which would be consistent with a reduction in the frequency of overbank floods, due to site aggradation or perhaps due to an increase in climatic aridity and possible river downcutting.

The depression occupied by the oxbow lake may have developed in one of two ways. First, it may have been a pool in the thalweg of the late deglacial channel, and in this way have been essentially a normal oxbow lake. Second, it may have been an impoundment created by the coincidence of the bar, the cliff, and rapidly developing alluvial fans on either side of the core site progressing from rapidly incising gullies on either side of the gully feeding the fan at the core site. These two possibilities are not mutually exclusive. The impoundment may well have started as a pool in the thalweg of the abandoned channel then been maintained by the second mechanism far longer than would have otherwise been the case.

4.5.2.3 LITHOZONE 3

LZ-3 (1122-897 cm; Figure 16-LZ-3) is evaporitic pond sediment, marked by major changes in all sediment characteristics, indicative of a change in dominant process, and consequently in sediment source. Grain size coarsens upwards through LZ-3, and laminations and erosional contacts become more frequent. This likely represents increasing relative importance of alluvial fan sediments and decreasing importance of overbank flood derived sediments. This is supported by increasing angularity of the grains suggesting a change in dominant sediment source. Grains are dominantly subangular, characteristic of the Quaternary deposits on the prairie surface, implying that the gullies which supply the sediment to the alluvial fans were still eroding mainly Quaternary deposits. High levels of P indicate high organic productivity, while the low organic content shown by loss on ignition implies high rates of oxidation of the organic matter. High carbonates shown by loss on ignition combined with gypsum and high values of Na, K and S suggest an evaporitic setting.

The dolomite may be clastic, or, if authigenic dolomite, may reflect a warm arid climate (Vasconcelos *et al.* 1995). Primary dolomite, which has never been chemically synthesized at Earth surface conditions, has been found to form in warm climates in saline water in the presence of certain anaerobic bacteria (Vasconcelos *et al.* 1995). This would be consistent with the other indications of evaporitic conditions in this lithozone, as well as with the high concentrations of S (presumed to be from the gypsum found in this lithozone). This type of environment is presently found in sloughs on the prairie surface (Last 1988).

This change in depositional environment from LZ-2 may have been caused by: (1) sediment aggradation, perhaps in part from coalescence of the bounding fans and encroachment of the middle fan, filling the pond to a level where overbank flooding no longer frequently occurred, or to a level where there was no longer a depression sufficiently deep to support a permanent pond; or (2) river incision or level reduction such that its floods no longer reached the site. In either case, it seems likely that overbank flooding was no longer the dominant source of sediment. Rather, LZ-3 was probably dominated by distal fan deposits in an evaporitic pond. This agrees with the increase in subangular grains evident in this unit.

Process studies in Dinosaur Provincial Park demonstrate that in rainfall events, the first several minutes of rainfall do not produce significant runoff; runoff and attendant sediment transport are greatest on unvegetated pre-wetted slopes (Bryan and Campbell 1980). It is entirely likely that during any runoff event capable of bringing noticeable quantities of sediment to the site, the early stages of runoff would have produced a pond in the depression at the site prior to the onset of significant sediment transport. Hence, the deposits of this lithozone are most likely a mixture of evaporitic pond and alluvial fan delta deposits. Further, the presence of an evaporitic pond suggests a climate in which vegetation on the fan and on gully slopes was very sparse. Therefore it is plausible that the gullies were subject to increased erosion at this time, conceivably explaining the slight increase in grain size. Charcoal consists of monocots suggesting a grassland environment.

4.5.2.4 LITHOZONE 4

LZ-4 (897-577 cm; Figure 16-LZ-4) represents a continuation of the evaporitic pond environment (as indicated by the continued presence of gypsum and dolomite) with frequent alluvial fan delta input events, which both temporarily freshen the pond and deposit finely laminated silts. These laminations are formed of melanocratic (dark coloured mineral dominated) and leucocratic (light coloured mineral dominated) mm-scale couplets. They show fining-upwards within each couplet, with the leucocratic layer being dominantly coarser quartz clasts and the melanocratic layer being more clay-rich. These are likely deposited in standing water as suggested by the continuing high values of P, Na, K and S, and the fine grain size.

The primary source of sediment was the alluvial fan, suggested by the laminations which are characteristic of sheetflood sediments (Bull 1972; O'Hara 1996; O'Hara and Campbell 1993; Seemann 1993). An increase in regional moisture could explain the decreased abundance of evaporitic minerals and the increase in sheetwash. Alternatively, it may be that the aggradation had at this time progressed to the point of nearly eliminating any depression for ponding. The increase in subangular shaped grains and the decrease in clay content suggest a change in dominant sediment source. The most likely explanation is the increased input from alluvial fans incising Quaternary sediments, combined with the decrease in importance of clay-rich overbank deposits.

Electron microprobe analysis was done on the glass from the double peak which occurs at 697-702 and 722-727 cm (Table 4.3). These tephras chemically and physically resemble Mazama tephra, deposited *ca.* 7800- 7480 cal yr B.P. The double Mazama tephra peak is consistent with those recorded elsewhere in Alberta by Zoltai (1989). Microscopic charcoal consists of monocots (grasses) suggestive of a grassland environment.

4.5.2.5 LITHOZONE 5

LZ-5 (577-320 cm; Figure 16-LZ-5) is interpreted as a subaerial alluvial fan environment. LZ-5 deposits are characterized by sandy, subangular, coarsely laminated sheetwash deposits. This unit also contains granular sized grains. The mean particle size is coarser in LZ-5 than in any other lithozone, and organics, carbonates, charcoal, P, Na, K and S reach their lowest values. The combination of coarsening upwards, coarse laminations (interpreted as sheetwash deposition) and the decrease in other parameters suggests that the pond is no longer a near-permanent feature (presumably due to infilling) and that the core site was likely a subaerial alluvial fan. This is further supported by the absence of dolomite and gypsum.

Microscopic charcoal identified as monocots (grasses) and sieve plates (hardwoods and/or deciduous shrubs) suggest an increase in humidity, with vegetation conditions being similar to the present with deciduous shrubs growing along the river or in a sheltered locations.

While it is possible that the increase in regional humidity (C. Campbell and Campbell 1997) was responsible for transporting larger grains, this seems unlikely. The shift from an evaporitic pond to an alluvial fan is delimited by a very sharp boundary. Such a shift most likely indicates a geomorphic threshold crossing, rather than a change in climate.

The shoreline contact of LZ-4 would have concentrated larger grains due to the loss of stream competence at the pond margin. The larger grains in the LZ-4 shoreline in association with the loss of competence imposed by the standing body of water would have effectively acted as a sieve deposit (albeit at a much smaller scale than defined by Hooke 1967), allowing finer grains to be transported further downfan (into the LZ-4 deposits), and concentrating coarser grains at the edge of the water body. Combined with wave-washing of the pond margin, this would have formed a sandy 'shoreline'. Once the pond was no longer active, sheetwash could have transported these larger 'shoreline' grains further downfan.

If this interpretation is correct, sediments in LZ-5 include significant contributions from this 'shoreline', an area much smaller than the total basin. Thus, the coarseness of LZ-5 may not have been related to climatic freshening, but was rather likely a function of a shift in dominant contributing area. This shift in partial area could have been due in part to the cessation of standing water, which in turn was a result of aggradation to the point where ponding events were no longer significant.

A possible aeolian deposit occurs at the top of this lithozone, *ca.* 327-328.5 cm

(Figure 4.8). In the analyses of the 5 cm sample in which it is contained, its characteristics are obscured by the other sediments in the same 5 cm sample.

Evidence of the development of multiple weak paleosols (A / C horizons with carbonate encrustations and nodules) *ca.* 575-760 cal yr B.P. (760 ± 80 ^{14}C yr B.P., TO-5946) in the upper portion of this lithozone, and a gradual decrease in grain size suggest: (1) alternating episodes of landscape stability and instability; (2) depletion of available sediments from the 'shoreline' source area; and, (3) a gradual shift to a new dominant sediment source area.

4.5.2.6 LITHOZONE 6

LZ-6 (320-95 cm; Figure 16-LZ-6) is dominated by fine-grained alluvial fan deposits with multiple, poorly developed, pedogenic horizons. This lithozone is high in herbaceous roots and carbonate concretions, which together with the monocot and some sieve plate charcoal, indicate a grassy environment which may have had deciduous shrubs, again suggestive of conditions similar to the present. Loss on ignition shows an increase in organics; P, Na, and K also increase slightly. This may reflect a more productive biological environment, or perhaps only a slower rate of fan deposition, possibly due to depletion of the shoreline sediments noted in LZ-5, and an increase in a finer grained source material. There may also be downwards leaching of K from fertilizers applied to the soil surface in the last few years.

A trend towards increased grain roundness starts in this lithozone and continues into the next, implying a change in dominant sediment source. There is no marked change in grain size parallelling the change in grain shape through LZ-6 and LZ-7, suggesting more a change in sediment source than in sediment transport process. This may reflect the coalescence of the fans which form the bajada, with this site starting to receive sediments from a different fan, and thus a different partial source area. Aerial photography suggests at least three fans may reach this core site, of which one seems largely blocked by the other two (Figure 4.4).

Figure 4.16 shows a possible scenario that explains the change in grain shape. First, fan deposits reaching this site are derived from a set of very small gullies which trench primarily through Quaternary sediments dominated by relatively angular materials. As the larger gullies to either side entrench more rapidly, capturing an ever increasing fraction of the small gullies' catchment, they also begin to supply an increasing fraction of the sediment arriving at the core site. As these larger gullies are also deeper and are trenching into the relatively round, Judith River Group, rounder grains become more important. Further support in favour of the dominance of a Judith River bedrock source is indicated by the dominance of grains < 250 μm, characteristic of Judith River sandstones (Bryan *et al.* 1984).

This explanation would tend to support the suggestion that the impoundment evidenced in LZ-2 to LZ-4 was at least partially maintained by fans to either side of the site.

4.5.2.7 LITHOZONE 7

LZ-7 (95 cm to surface; Figure 16-LZ-7) is an anthropogenically altered soil composed of coarse fan and possible aeolian deposits. Aeolian deposits (Klassen, personal communication 1995) are likely mixed with the dominant fan deposits, obscuring any possible aeolian signal. Typical modern plough depths range up to 40 cm.

The strong increase of well-rounded grains suggests that the dominant source of sediment is the Judith River bedrock. This can be explained as a continuation and accentuation of the trend started in LZ-6, in which the larger, more deeply bedrock-incising gullies to either side of the main fan-feeding gully directly upslope from the core site, gradually become the dominant sediment source.

Historic farming practices have significantly altered the upper *ca.* 100 cm of sediment, resulting in a friable composition not observed in the rest of the core. Fertilizer applications have undoubtedly affected the chemistry of the sediment. Mottling in the lower part of this lithozone may result from irrigation of the upper portion. The charcoal in this lithozone is dominantly monocots (grasses) with some sieve plates (hardwoods and/or deciduous shrubs), representing modern mixed-grass prairie conditions.

4.6 CHRONOLOGY AND REGIONAL CORRELATIONS

4.6.1 SITE DEVELOPMENT

Unfortunately there is a paucity of early postglacial radiocarbon dates in the region, and the timing of deglaciation has been a topic of significant controversy (Christiansen 1979; Clayton and Moran 1982; Dyke and Prest 1987; Evans and Campbell 1992). LZ-1 was likely created when terminal late Wisconsinan meltwater feeding the Red Deer River fell below *ca.* 624 m a.s.l., abandoning the channel, creating an oxbow lake. In the study area, the modern level of the river is 621 m a.s.l.; therefore the oxbow formed when the level of the river was similar to the present level. This indicates that glacial meltwater had largely ceased flowing in the Red Deer River prior to *ca.* 13,000 cal yr B.P., when the Glacier Peak G tephra bed was deposited. This implies essentially complete deglaciation of the eastern slopes in the headwaters area of the Red Deer River. Vreeken (1989) found Glacier Peak G tephra in loess overlying glaciolacustrine deposits near Lethbridge, suggesting semiarid conditions by this time.

Alternatively, LZ-1 may have been a river terrace, created as a result of the Younger Dryas aged Crowfoot glacial advance in the headwaters *ca.* 13,482-11,223 cal yr B.P. (11,330 $\pm$ 220-10,020 $\pm$ 70 ^{14}C yr B.P.) (Reasoner *et al.* 1994). The oldest dates for river terraces in the major river valleys are 18,260-12,910 cal yr B.P. (Lowden and Blake 1975) on the South Saskatchewan River, and 13,740-11,880 cal yr B.P. for the North Saskatchewan River (Rains and Welch 1988) which, while further north than the study area, may have been similarly affected by increased Younger Dryas aged runoff. While the processes governing river terrace formation are complex (Rains and Welch 1988), it is reasonable to propose that river incision and therefore terrace formation in these major river valleys could have occurred in response to an increase in river discharge and associated channel incision. Effects of glacioisostatic uplift are possibly involved but have not yet been documented.

Thus if deglaciation occurred *ca.* 20,210-15,450 cal yr B.P. (C. Campbell and Campbell 1997), followed by a period in which there was an abundance of surface water, followed by a period of aridity, the occurrence of increased humidity *ca.* 13,482-11,223 cal yr B.P. (Reasoner *et al.* 1994) may have initiated river incision and possibly terrace development.

4.6.2 OXBOW LAKE - ALLUVIAL FAN DELTA

The oxbow lake (LZ-2) occurred *ca.* 13,000 cal yr B.P. during a period believed to have been cooler and moister than present (Harris and Pip 1973; Pennock 1984; Barnosky *et al.* 1987). Moister climatic conditions may have resulted from the effect of continental glaciers still to the east, and abundant surface water on the prairie surface (Beaudoin *et al.* 1996), and moderated by high summer insolation. Alternatively, moister conditions may have resulted from the Younger Dryas aged Crowfoot glacial advance in the headwaters *ca.* 13,482-11,223 cal yr B.P. (11,330 $\pm$ 220-10,020 $\pm$ 70 ^{14}C yr B.P.) (Reasoner *et al.* 1994). At the study site, there may have been an exclusively local cooling effect from the cold meltwater derived from the upstream glaciers. Conifers may have developed in this area under such conditions. If the charcoal at the site is local, the occurrence of abundant conifer charcoal suggests conifer parkland / forest conditions, or perhaps the development of small stands of conifers along the river valleys and other drainage basins.

Paleobotanical research in potholes in southern Saskatchewan and Alberta suggests that following deglaciation, an open forest was followed by perennial wetlands surrounded by aspen in southern Alberta and spruce in southern Saskatchewan (Yansa 1995; Beaudoin *et al.* 1996). Conifer trees do not presently grow in the study area. The nearest natural conifer stands today are near Drumheller (100 km NW); these are small white spruce stands on steep north-facing slopes directly supplied with water from seeps. Recumbent juniper (*Juniper horizontalis*) does occur in the region today, and may have been present at the time of LZ-2.

Alternatively, LZ-2 charcoal may not be local. Charcoal records from the foothills (MacDonald 1989). Given that the sediments in LZ-2 were likely derived primarily from overbank deposits, the charcoal may be reflecting vegetative conditions further upstream.

Aeolian deposits have been identified from this time period in southern Alberta and southern Saskatchewan (David 1993;Vreeken 1993) but there is no indication of aeolian material in LZ-2. There are a number of possible environmental interpretations for aeolian activity. Firstly, the environment was sufficiently arid to prevent significant vegetation development, so that aeolian erosion would be highly effective. Even if the charcoal present in LZ-2 is not local, glacial groundwater recharge probably resulted in large amounts of standing water on the prairie surface, so the surface was probably vegetated. The second is that very strong winds circulating at the ice front (David 1981) were effective despite any vegetation cover. However, at this time, the Laurentide ice front was well to the NE and therefore the argument that there were katabatic winds coming off the ice (David 1981) can not be invoked to explain aeolian activity. The most likely explanation for aeolian activity is that deglaciation provided abundant unconsolidated sediments (lacustrine, deltaic, fluvial) which could be freeze-dried and then deflated by strong winter winds, a phenomenon common in cold environments today (McKenna-Newman 1993; van Djik and Law 1995). At present the region often experiences considerable aeolian activity in the winter as evidenced by very dirty snow accumulations. Regional aeolian activity during the period of LZ-2 deposition was probably a function of winter aeolian activity, not necessarily a reflection of a drier or windier climate than today, in association with a large supply of transportable material from the local area.

The absence of aeolian material in this lithozone may be due to a local absence of aeolian deposition, or simply to the presence of a nearly permanent lake at the site, in which any aeolian input would have been mixed with sediments from other sources, inhibiting recognition. Kettle depressions elsewhere in southern Alberta and Saskatchewan supported ponds at this time (Beaudoin *et al.* 1996), so that a high groundwater table must have been present regionally, and would undoubtedly have contributed to maintaining the oxbow lake in LZ-2.

4.6.3 SALINE POND - ALLUVIAL FAN DELTA

LZ-3 and LZ-4 are believed to have been associated with maximum postglacial aridity from *ca.* 12,000 cal yr B.P. until just after the time of deposition of Mazama tephra, *ca.* range 7480-7890 cal yr B.P. (Bacon 1983). A period of landscape stability from *ca.* 12,000-10,000 cal yr B.P. has been associated with aridity as suggested by the occurrence of the most strongly developed regional paleosols (Lowden *et al.* 1971; Turchenek *et al.* 1974; Waters and Rutter 1984; Valentine *et al.* 1987; David 1993), decreased fluvial erosion and deposition (Waters and Rutter 1984), an environment of non-deposition in formerly active floodplains (Waters and Rutter 1984), alluvial fan accumulation (O'Hara and Campbell 1993), and drying of lake basins in central Alberta (Schweger and Hickman 1989). After 10,000 cal yr B.P. increasing humidity (although still more arid than present) accelerated runoff resulting in terrace development in secondary river valleys (Rains and Welch 1988; I.A. Campbell and Evans 1993) and infilling of local lakes (Sauchyn 1990; Vance and Last 1994, I.A. Campbell and Evans 1990) prior to the deposition of Mazama tephra.

No Mazama tephra has been found in the alluvial fans in Dinosaur Provincial Park, upriver of this site (O'Hara 1986; O'Hara and Campbell 1993). This suggests either that the fans in the Dinosaur Provincial Park badlands were non-depositional and possibly deflationary sites during the early-mid Holocene, or that, like the fans at the Klassen site, they had not yet developed. The Klassen Ranch site, however, did preserve both Mazama tephra beds, perhaps because it was occupied by an evaporitic pond at the time, protecting sediments from deflation.

The evaporitic pond is the clearest climate signal at this site of past climate conditions. Even so, modern conditions, believed to be much more humid, presently maintain evaporitic ponds on the prairie surface (Last 1988); thus the evaporitic pond can only be taken to indicate that the climate at this time was not substantially more humid than it is today. The shift from an oxbow to an evaporitic pond is in keeping with the regional climatic drying, although as noted above, it is not necessary to invoke climate to explain LZ-3 and LZ-4.

4.6.4 ALLUVIAL FAN

Mid-late Holocene climate in southern Alberta is marked by a general increase in regional humidity (C. Campbell and Campbell 1997). This increase was not monotonic, as suggested by a reduction in the rate of tributary valley incisions *ca.* 8000 to 4000 cal yr B.P. (Rains and Welch 1989; I.A. Campbell *et al.* 1993), loess deposition in Dinosaur Provincial Park *ca.* 5400±800 T.L. yr B.P. (Bryan *et al.* 1987), salinity variations at Chappice Lake (Vance *et al.* 1993), and vegetation changes at Harris Lake and Chappice Lake (Sauchyn and Sauchyn 1991; Vance *et al.* 1993; Vance and Wolfe 1996). By *ca.* 4000-3000 cal yr B.P., effective moisture levels were similar to those experienced historically. Late Holocene geomorphic processes are characterised by increased variability manifested in repeated cycles of dune activation and pedogenic intervals (Wolfe *et al.* 1995), renewed valley incision/terrace cutting and stabilization (pedogenic intervals) (Traynor and Campbell 1989; I.A. Campbell and Evans 1993; Rains *et al.* 1994; Barling 1995), fluctuating lake levels (Vance *et al.* 1992), multiple episodes of valley alluviation and incision (O'Hara 1986; O'Hara and Campbell 1993), repeated cycles of loess deposition and pedogenic intervals (Vreeken 1989; 1993), and multiple episodes of landslides in the Cypress Hills (Goulden and Sauchyn 1986; Sauchyn and Lemmen 1996).

At Chappice Lake, Vance and Wolfe (1996) note minor changes in the water balance and suggest a period of peak effective moisture between 2700 and 1000 B.P., followed by a short return to more arid conditions *ca.* 900-1200 A.D. (coeval with the European Medieval Warm Period) and a return to cool and moist conditions *ca.*1450-1850 A.D. (coeval with the European Little Ice Age). The historic environment has been more arid than the previous period, and is reflected in severe historic drought events in the 1880s, 1920s, 1930s, and 1980s A.D. (Vance and Wolfe 1996).

LZ-5, LZ-6 and LZ-7 are marked by a relative absence of overbank or pond deposition. This absence may be taken as an indication of an increase in climatic aridity; however, this is contra indicated by the regional paleoenvironmental record. Fan development is probably the result of aggradation eliminating the depression which had until then supported the pond. LZ-5, LZ-6 and LZ-7 appear to be alluvial fan deposits, modified by pedogenesis and in LZ-7 by anthropogenesis. The increasing number of paleosols in the fan may be indicative of fluctuating climatic conditions enabling fan stability due to increased rainfall resulting in increased vegetation and thus paleosol development.

Aeolian deposition may have occurred throughout these lithozones, but its presence is apparently masked by the fan deposition. Regionally, a major phase of aeolian deposition has been dated as beginning *ca.* 5400±800 T.L. yr B.P. (Bryan *et al.* 1987). However, it is unlikely that the possible aeolian deposit *ca.* 327-328.5 cm dates to this episode as the A.M.S. date at 355 cm of 575-760 cal yr B.P. (760 ± 80 ^{14}C yr B.P., TO-5946) is much younger.

4.6.5 SEDIMENTATION RATES

Rates of sedimentation, calculated using the tephras and the single ^{14}C date for dating control, have changed throughout the core. From 1547-697 cm, the rate of sedimentation was 1.4-1.5 mm yr^{-1}. From 697 cm to 355 cm (*ca.* 764-575 cal yr B.P.) a sediment rate of 0.5 mm yr^{-1} is estimated. From 355 cm to the surface the sedimentation rate increased to 4.6-4.2 mm yr^{-1}.

The sediment accumulation rate in the upper, alluvial fan, portion of the core is similar to those found in late Holocene alluvial fans upstream in Dinosaur Provincial Park (3.8 mm yr $^{-1}$; Seemann 1993). This similarity may relate to the similarities in bedrock and topography subjected to the same climate and hence vegetation cover.

4.7 SUMMARY AND CONCLUSIONS

Changes in postglacial regional climate have been inferred from paleorecords (for example Schweger and Hickman 1989; Sauchyn and Sauchyn 1991; Last and Sauchyn 1993). However, as each lithozone can be understood within the context of changing dominant partial source area development, it is not necessary to invoke fluctuations in climatic controls which would involve the application of more assumptions than required by the partial area explanation. Although at first glance, a fan system like this may appear to be simple enough that climatic interpretations could be made from its changing sediments, the succession of dominant processes and partial areas prevents any such simplistic interpretation. For example, while in the absence of base level changes and tectonism, the basic fan geometry argues that an up-core coarsening should relate to an up-core climatic moistening (since the bajada slope becomes shallower up-core, the grain size should decrease; since it does not, the climate must be getting moister to bring down coarser grains or there was an increase in high intensity storms). The changes brought on by changing dominant source area better explain the same trends.

Despite a number of minor erosional contacts, the Klassen site has effectively been aggrading since river abandonment. Yet, this does not imply that either the rate or the processes of aggradation have been constant through time. Indeed, the site has experienced several different sedimentary environments (river, oxbow lake, evaporitic pond, fan), each with distinct properties.

Succession from one depositional environment to another involves crossing a geomorphic threshold (Schumm 1979) which defines "a boundary between separate systems states" (Bull 1991). In the case of the Klassen site, such crossings appear to be primarily driven by aggradation, resulting in a complex geomorphic response cascade; as illustrated by the drying up of the evaporitic pond (top of LZ-4) which allowed the paleo-evaporitic pond shoreline to migrate in association with younger alluvial fan sediments over the coring site, which enabled more rapid sediment deposition, allowing an increase in landscape instability resulting in minimal paleosol development (LZ-5). Figure 4.19 shows a conceptual model of the impact of aggradation on changes in dominant partial area contributions. As the deposit aggraded, the relative importance of different partial areas sediment sources waxed and waned, although it is likely that more than one source contributed to fan aggradation at any one time. While climate may have affected the timing and sediment yield responses (Brunsden and Thornes 1979), the sequence was controlled by aggradation alone. The sensitivity of this site to climate change is thus overwhelmed by internal processes, or rather, it is dominated by the dynamic nature of fluvial activity in the steeply inclined contributing basin.

Alluvial fan studies have primarily focussed on Quaternary-aged fans in arid, mountainous and most often tectonically active terrain (Harvey 1989; Racjocki and Church [eds.] 1990; Bull, 1961, 1962a, 1962b, 1964a, 1964b, 1968, 1972, 1977, 1991). Most fans that have been well-studied are large, and are dominated by coarse-grained deposits. The Klassen fan is fine-grained, relatively young, and in a tectonically quiet region. Fan evolution models have tended to focus on cyclic-stage oriented and equilibrium paradigms (Lecce 1990). However, as Bull (1991:24) notes: "little is gained by forcing an equilibrium conceptual model on landscapes where change is obvious." Furthermore, the confinement of the Klassen fans at their distal limits, and their development in a setting subject to overbank river deposition, makes this investigation unusual among fan studies.

Structures and grain size analysis alone would not likely have yielded the complex interpretation developed here. The use of multi-proxy data is strongly recommended for determining the succession of fan environments.

Changes in dominant partial source area through time probably occur in many depositional systems, although detecting such changes may be difficult in most settings where the partial areas have very similar properties. In this site, the transitions in sediment source provide a greater contrast between deposits from different partial areas than may be the case in many other systems. Perhaps it is in part the high temporal resolution of this study (an average of 40 years / sample) that allows the separation of partial area contributions in such detail. Given that partial areas do change, interpreting changes in granulometry from a strict climatic or tectonic viewpoint ignores the possible effects of changing partial area properties. As in the case of the Klassen site, change in partial contributing areas appears to be the dominant cause of changes in grain size and shape; furthermore, these changes result not from climate change or tectonism, but simply from inevitable site aggradation.

The findings of this research can be summarised as follows:

(1) The bajada at this site overlies river gravels, oxbow lake, and evaporitic pond sediments.

(2) Glacial meltwater inputs to the Red Deer River were finished prior to *ca.* 13,000 cal yr B.P.

(3) River incision and the creation of a possible river terrace (LZ-1) may have resulted from the Crowfoot Advance (coeval with the Younger Dryas) in the headwaters *ca.* 13,480-11,220 cal yr B.P.

(4) Changing lithozones at this site are a function principally of aggradation, and shifting dominant partial contributing areas.

(5) The transition of dominance by one system to dominance by another may be abrupt (for example the transition between oxbow lake and evaporitic pond at the end of LZ-2) or more gradual (for example the transition between fan sources in LZ-6); in both cases the crossing of a geomorphic threshold is reflected in one or more types of proxy data.

(6) Multiproxy investigation of the Klassen site yielded information on its evolution which would not have been discovered through single-proxy analysis.

(7) Effects of climatic change, while known to have occurred, are not detectable in this deposit, being masked by the changes caused by changing dominant souce area.

(8) Rates of aggradation have changed depending on the type of depositional site.

(9) The sediment accumulation rate in the upper portion of the alluvial fan in the Klassen core is of the same order of magnitude as found in Dinosaur Provincial Park which may

reflect the similarity in bedrock and erosion rates.

4.8 CHAPTER 4 - TABLES

TABLE 4.1

BROOKS AHRC CLIMATE STATION, LATITUDE 50° 33', LONGITUDE 111° 51', ALTITUDE 758M ASL, YEARS OF RECORD 1915-1988

MONTH	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEPT	OCT	NOV	DEC	YEAR
TEMPERATURE (°C)													
Daily Maximum	-6.9	-2.4	3.1	12.2	18.7	23.0	25.9	25.2	18.9	13.6	2.1	-4.6	10.7
Daily Minimum	-18.1	-14.0	-8.5	-2.0	4.0	8.8	10.8	9.7	4.2	-0.9	-9.4	-16.0	-2.6
Daily Mean	-12.5	-8.2	-2.7	5.1	11.4	15.9	18.4	17.5	11.6	6.4	-3.6	-10.2	4.1
Extreme Maximum	17.8	17.2	25.0	31.1	35.6	37.2	40.0	38.9	35.6	31.1	23.5	17.2	
Extreme Minimum	-46.7	-43.9	-37.8	-25.0	-10.0	-2.2	1.7	0.0	-10.6	-24.4	-36.1	-47.2	
Degree Days													
Above 18° C	0.0	0.0	0.0	0.0	2.9	20.7	43.3	37.0	4.5	0.1	0.0	0.0	109
Below 18 ° C	943.4	740.9	642.0	387.4	207.7	82.4	31.7	52.8	196.3	363.3	649.7	877.2	5175
Above 5° C	0.6	1.5	8.4	61.9	202.1	328.3	414.6	387.2	204.8	87.8	10.7	1.1	1709
Below 5° C	394.4	249.3	133.8	17.2	0.1	0.0	0.0	0.0	0.5	14.0	152.4	329.3	1291
Precipitation													
Rainfall (mm)	1.0	0.7	2.3	17.5	38.1	65.5	38.1	36.4	38.3	10.7	2.5	1.0	252.3
Snowfall (cm)	18.3	12.0	15.4	9.84	1.0	0.0	0.0	0.0	0.5	5.3	13.2	18.8	94.2
Precipitation (mm)	18.4	11.9	17.0	27.2	39.1	65.5	38.1	36.4	38.8	16.0	14.9	18.3	341.6
Extreme Daily Rainfall (mm)	4.8	5.1	15.2	29.2	29.2	88.9	36.8	45.5	57.3	20.3	9.9	4.8	
Extreme Daily Snowfall (cm)	16.5	16.5	16.5	21.6	27.9	16.0	9.1	0.0	0.0	17.8	22.9	30.5	
Extreme Daily Ppt. (mm)	16.5	16.5	21.6	34.0	29.2	88.9	36.8	45.5	57.3	22.9	22.9	30.5	
Month End Snow Cover	12	11	N	2	0	0	0	0	0	0	6	13	
Days With													
Maximum Temperature >0°C	10	14	21	28	31	30	31	31	30	29	19	12	286
Measurable Rainfall	*	*	2	7	10	12	10	9	9	5	2	*	68
Measurable Snowfall	6	4	4	2	*	0	0	0	*	1	4	6	28
Measurable Precipitation	6	4	6	8	10	12	10	9	9	6	5	692	
Sunshine (hours)	91.6	114.9	158.3	215.1	266.3	290.2	338.8	302.1	200.9	169.7	105.8	75.1	2328.9

Climate data for Brooks AHRC (Environment Canada 1993). Station with the longest record near the study site.

TABLE 4.2

LITHOZONE CHARACTERISTICS

LZ-1 (Samples 304-307, 1587-1607cm)

Grain Roundness (%)

Well Rounded	*Rounded*	*Sub-Rounded*	*Sub-Angular*	*Angular*	*Very Angular*
26	9	36	25	3	0.6

Loss on Ignition

% Organic Matter	*% Carbonates*	*% Ash*
3	6	92

LZ-2 (Samples 211-303, 1122-1587cm)

Grain Roundness

Well Rounded	*Rounded*	*Sub-Rounded*	*Sub-Angular*	*Angular*	*Very Angular*
2	14	46	34	4	0.6

Loss on Ignition

% Organic Matter	*% Carbonates*	*% Ash*
6	8	86

% Charcoal: 5.92

Chemistry

P (ppm)	*Na (ppm)*	*K (ppm)*	*S (ppt)*
298.9	554.1	390.1	1653.5

Charcoal Identifications

Softwood	*Hardwood*	*Grass*
28	10	0

LZ-3 (Samples 166-210, 897-1122cm)

Grain Roundness (%)

Well Rounded	*Rounded*	*Sub-Rounded*	*Sub-Angular*	*Angular*	*Very Angular*
2	13	35	44	6	0.6

Loss on Ignition

% Organic Matter	*% Carbonates*	*% Ash*
4	9	87

% Charcoal: 3.73

Chemistry

P (ppm)	*Na (ppm)*	*K (ppm)*	*S (ppt)*
607.7	1157.7	650.3	3865.5

Charcoal Identifications

Softwood	*Hardwood*	*Grass*
0	0	18

LZ-4 (Samples 104-165, 577-897cm)

Grain Roundness (%)

Well Rounded	*Rounded*	*Sub-Rounded*	*Sub-Angular*	*Angular*	*Very Angular*
0.7	9	29	49	11	1

Loss on Ignition

% Organic Matter	*% Carbonates*	*% Ash*
4	7	89

% Charcoal: 3.01

Chemistry

P (ppm)	*Na (ppm)*	*K (ppm)*	*S (ppt)*
473.2	670.8	486.1	2118.5

Charcoal Identifications

Softwood	*Hardwood*	*Grass*
0	0	20

LZ-5 (Samples 65-103, 320-577cm)

Grain Roundness (%)

Well Rounded	*Rounded*	*Sub-Rounded*	*Sub-Angular*	*Angular*	*Very Angular*
0.9	10	26	45	16	2

Loss on Ignition

% Organic Matter	*% Carbonates*	*% Ash*
2	6	92

% Charcoal: 2.12

Chemistry

P (ppm)	*Na (ppm)*	*K (ppm)*	*S (ppt)*
278.4	188.7	233.7	454.9

Charcoal Identifications

Softwood	*Hardwood*	*Grass*
0	2	11

LZ-6 (Samples 20-64, 95-320cm)

Grain Roundness (%)

Well Rounded	*Rounded*	*Sub-Rounded*	*Sub-Angular*	*Angular*	*Very Angular*
5	24	38	27	5	0.9

Loss on Ignition

% Organic Matter	*% Carbonates*	*% Ash*
4	6	90

% Charcoal: 2.58

Chemistry

P (ppm)	*Na (ppm)*	*K (ppm)*	*S (ppt)*
354.0	190.4	354.2	334.5

Charcoal Identifications

Softwood	*Hardwood*	*Grass*
1	4	12

LZ-7 (Samples 1-19, 0-95cm)

Grain Roundness (%)

Well Rounded	*Rounded*	*Sub-Rounded*	*Sub-Angular*	*Angular*	*Very Angular*
36	31	19	7	2	0.3

Loss on Ignition

% Organic Matter	*% Carbonates*	*% Ash*
6	5	89

% Charcoal: 3.07

Chemistry

P (ppm)	*Na (ppm)*	*K (ppm)*	*S (ppt)*
342.2	64.5	452.3	141.9

Charcoal Identifications

Softwood	*Hardwood*	*Grass*
0	2	8

TABLE 4.3 KLASSEN SITE: CONDITIONS OF THE RED DEER RIVER JUST PRIOR TO THE OXBOW CUTOFF

Aim: To ascertain conditions of the Red Deer River just prior to oxbow cutoff at the Klassen ranch.

Assumptions: Mid-channel bar, water level was slightly higher than the bar, hydraulically smooth channel

Reference: 1:50,000, 25 ft contour map, Howie 72/L14 (Energy, Mines and Resources, 1989).

Manning Equation:

$$V = \frac{R^{2/3} S^{1/2}}{n}$$

Where:

V = mean velocity

$$R = \text{hydraulic radius} = \frac{\text{cross sectional area}}{\text{wetted perimeter}} \;(\text{width*depth}=2161*20.6923=)\; \frac{44716.0603}{2232.4}$$

=20.030487

S = slope = 0.0003

n = Manning roughness # = 0.02

$$V = \frac{(20.0304875^{2/3})(0.0003^{1/2})}{0.02} = 6.7 \text{ m/s}$$

V = 6.7 m/s

Discharge

Q = VWD

Where:

V = mean velocity =6.7 m/s

W = Width of channel = 2161 m

D = mean depth = 20.6923 m

Q = (6.7)(2161)(20.6923) = 299,597 m^3/s

Q = 299,597 m^3/s

TABLE 4.4

MEAN TEPHRA BED GLASS CHEMISTRY

Sample 128 (n=19) - Identified as Mazama

SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O
74.1	0.439	14.7	2.1	0.1	0.6	1.7	3.6	2.7

Sample 133 (n=18) - Identified as Mazama

SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O
75.6	0.4	14.7	2	0.1	0.4	1.6	2.4	2.8

Sample 302 (n=34) - Identified as Glacier Peak Bed G

SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O
76.7	0.2	12.2	1	0	0.2	1.1	2.2	1.6

TABLE 4.5

PER CENT ROUNDNESS CHARACTERISTICS OF REGIONAL SEDIMENT SOURCES

	% WR	% R	% SR	% SA	% A	% VA
Loess	2	16	41	41	0	0
River Over-bank	1	18	53	26	2	0
Prairie Surface	1	10	25	50	13	1
Judith River Bedrock	52	25	16	6	1	0

Roundness calculated after Powers (1953) on 200 grains > fine silt per sample. WR = well rounded (sample n=5), R= rounded (sample n=5), SR = subrounded (sample n=5), SA = subangular (sample n=5), A = angular (sample n=5), VA = very angular (sample n=5).

4.9 CHAPTER 4 - FIGURES

FIGURE 4.1

Study area. Stippled area = gravel bar, ■ = core site. Scale: 1:50,000.

FIGURE 4.2

Granulometric analysis of 21 surface samples at the Klassen site (grain size per cent <45 μm) shows that there is a general fining from fan apices to toes.

FIGURE 4.3

Klassen site: Photograph of gravel bar, car for scale (Photograph taken by I.A. Campbell).

FIGURE 4.4

The surface of the bajada is covered by a distributary, semi-radial pattern of braided streams which spread out of the fans' apices (Photograph courtesy of Air Photo Services, Government of Alberta, 92-159 AP LN=7 AS4361 20). Scale: 1:31,000.

FIGURE 4.5

Mean cumulative grain size curve of 30 loess deposits from Dinosaur Provincial Park (dark line) compared to that of 33 surface samples obtained from the prairie surface north of the Klassen site (light lines).

FIGURE 4.6

A truck-mounted, five cm diameter, hollow stem auger extracted 15.35 m of core from a 16.07 m hole in the central portion of the bajada (Photograph taken by I.A. Campbell).

FIGURE 4.7

4.7A Klassen site: Stratigraphic log: dates in cal yr B.P., depth, per cent glass, lithozone zonation, grain size, grain shape, loss on ignition, charcoal, chemistry.
4.7B Klassen site: Description of sedimentary structures

FIGURE 4.8

Moderately poorly sorted, matrix-supported, generally very well rounded gravels LZ-1.

FIGURE 4.9

Possible aeolian deposit, cross bedded laminations and massive sandy silt loam LZ-5.

FIGURE 4.10

Black bands, rich in roots and carbonate concretions LZ-6.

FIGURE 4.11

Massive sediments with overlying ped rich horizon LZ-2.

FIGURE 4.12

Fine grained horizontal laminations LZ-4.

FIGURE 4.13

Massive sandy silt loam LZ-5.

FIGURE 4.14

Massive sandy silt loam with laminated horizons separated by an erosional contact LZ-5.

FIGURE 4.15

Massive homogenous sandy silt, possible aeolian deposit LZ-5.

FIGURE 4.16

Hypothesized evolution of the Klassen site (see text for interpretation).

FIGURE 4.17

Cross section of the study area.

FIGURE 4.18

At the base of the Klassen bajada, rounded gravels located in six well-logs between 599-624 m asl were probably contemporaneous with LZ-1 gravels Scale: 1:50,000.

FIGURE 4.19

Conceptual model of deposit evolution showing relative dominance of clastic sediment source partial areas.

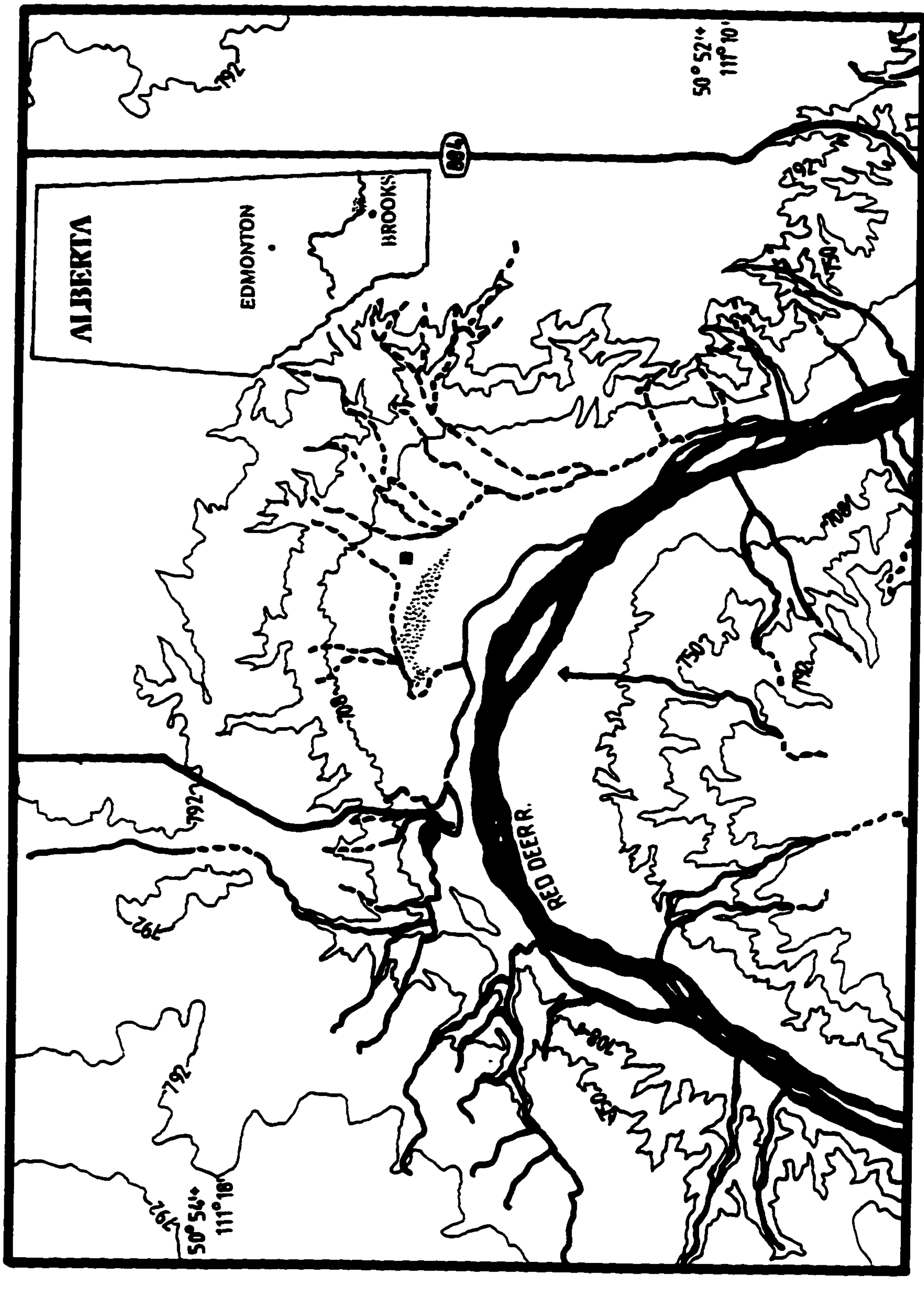

ALBERTA
EDMONTON
BROOKS
884
RED DEER R.
792
750
708
50°52'+
111°10'
50°56'+
111°18'

FIGURE 4.1

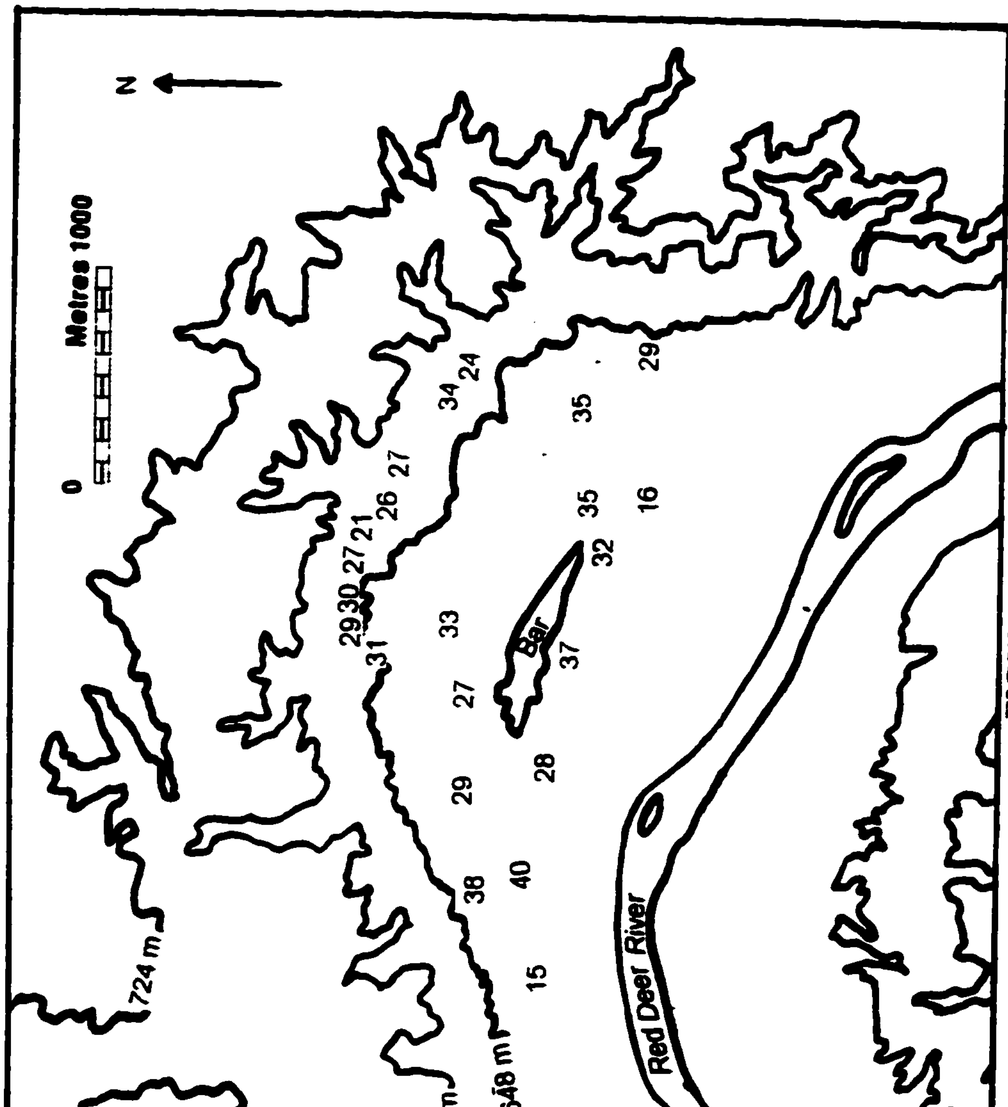


FIGURE 4.2

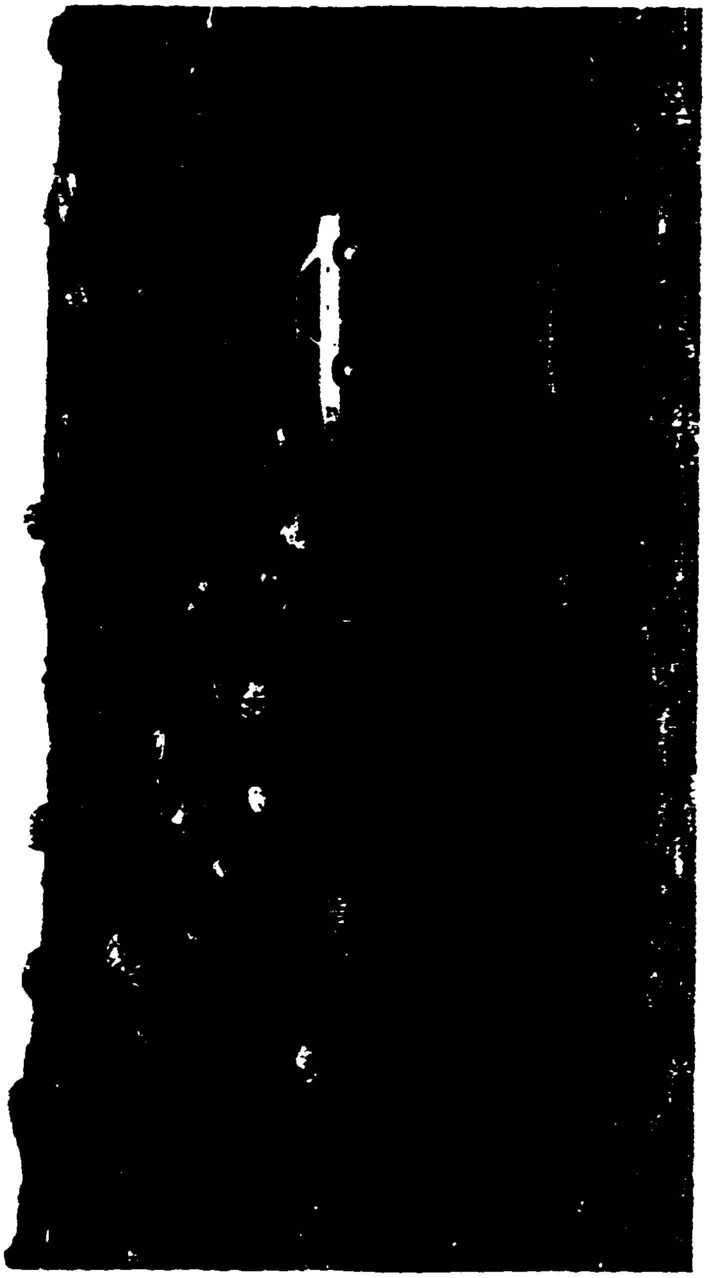


FIGURE 4.3

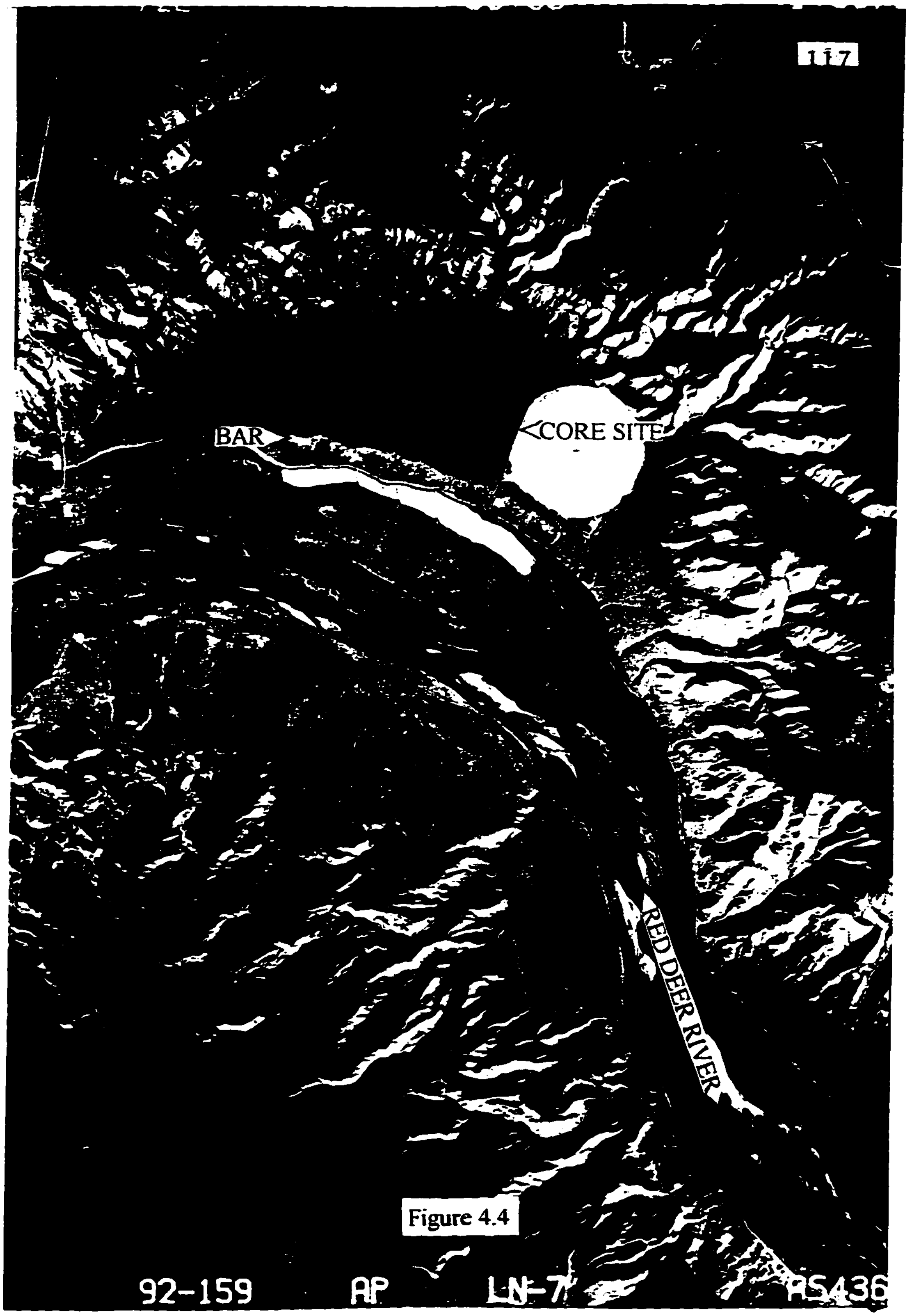


Figure 4.4

FIGURE 4.5

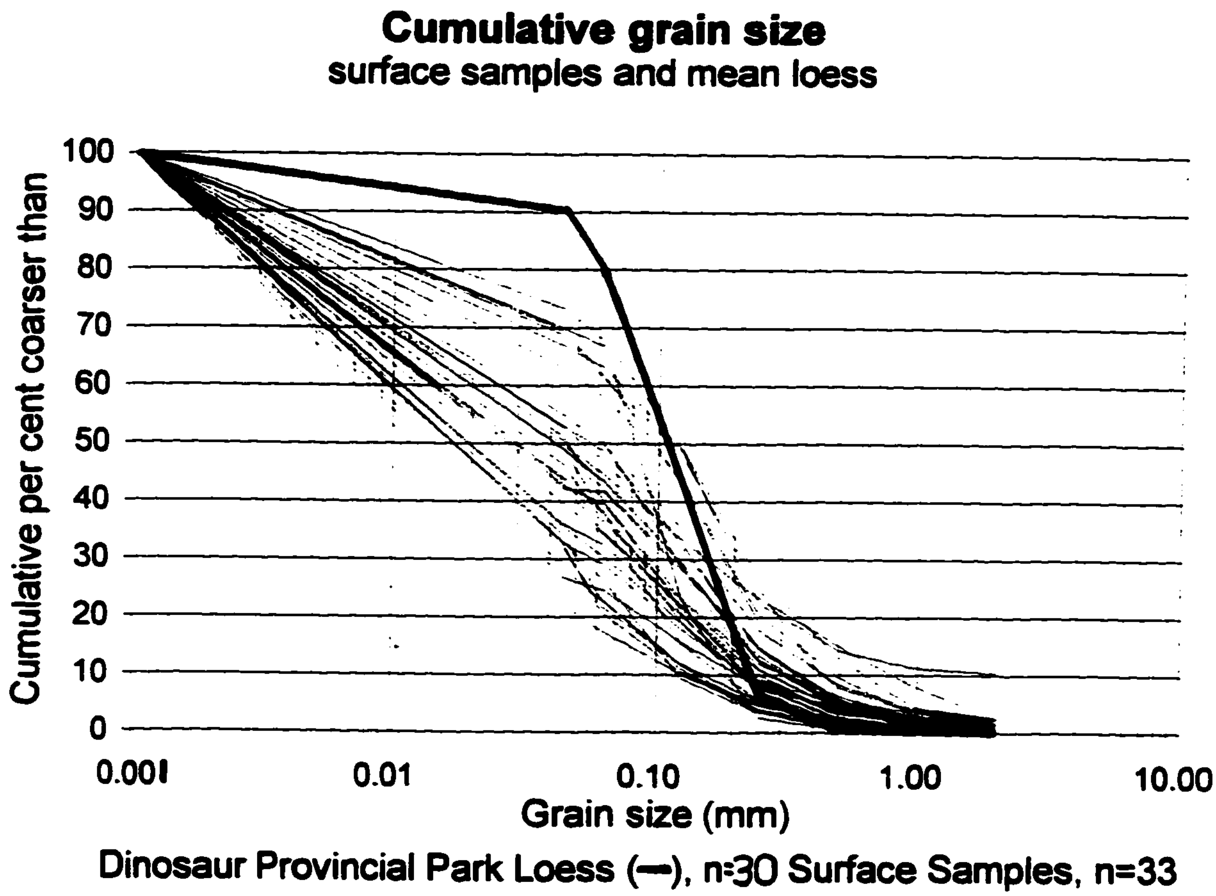
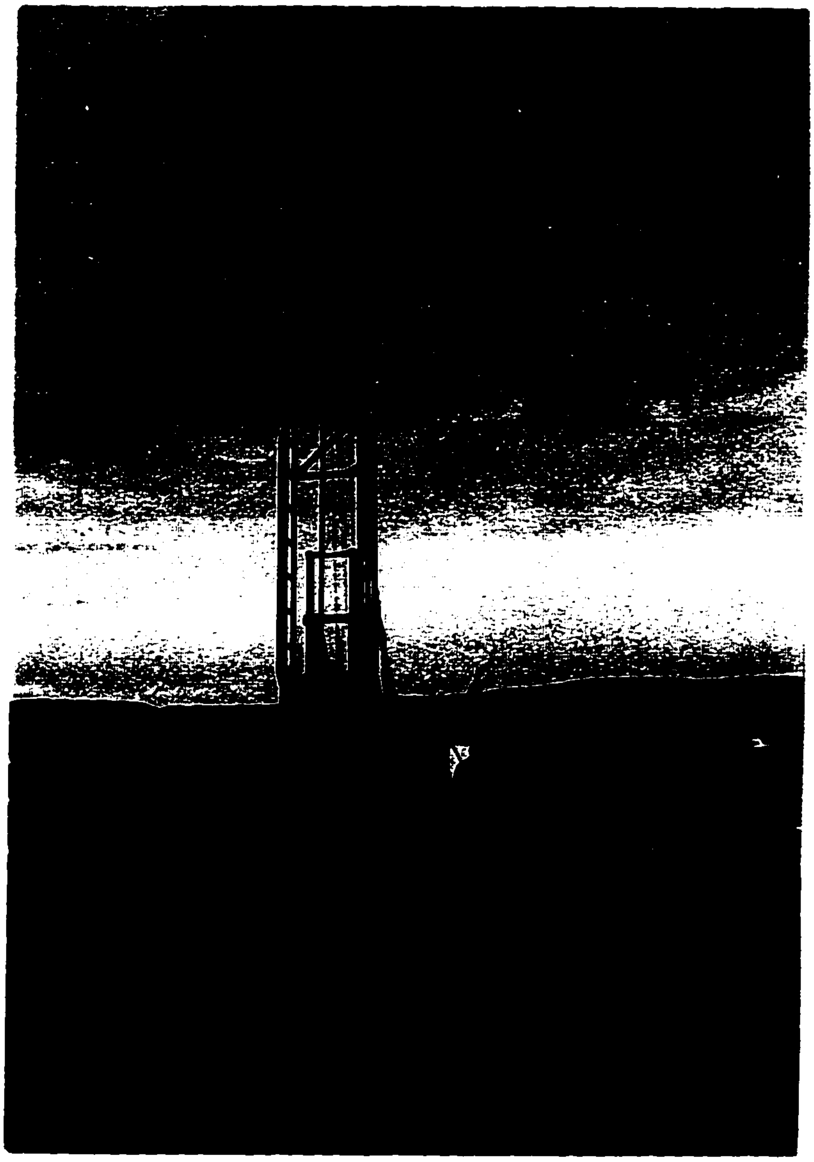
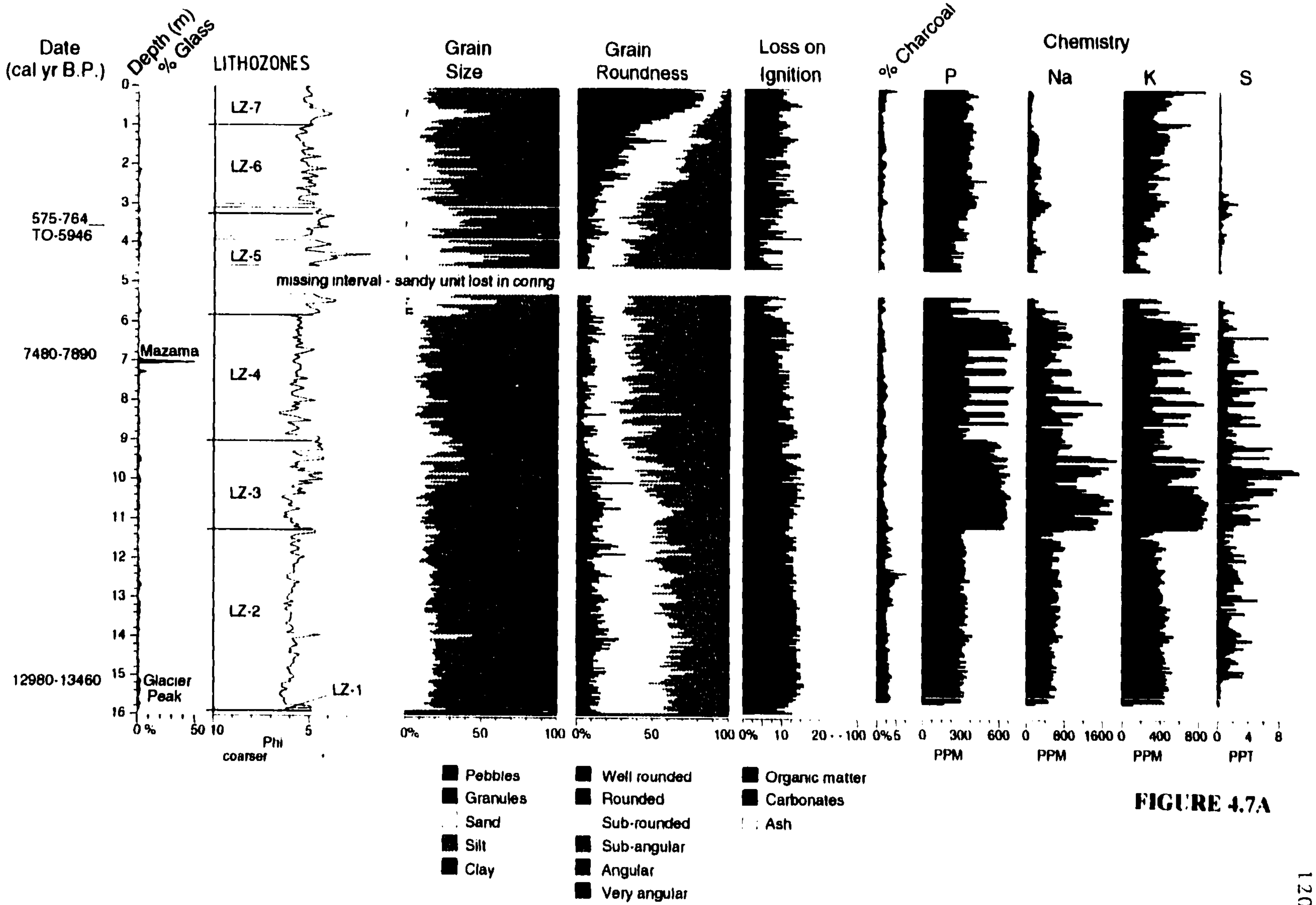

Cumulative grain size
surface samples and mean loess
Cumulative per cent coarser than
100
90
80
70
60
50
40
30
20
10
0
0.001
0.01
0.10
1.00
10.00
Grain size (mm)
Dinosaur Provincial Park Loess (—), n=30 Surface Samples, n=33

FIGURE 4.6





Date (cal yr B.P.)
575-764
TO-5946
7480-7890
12980-13460
Depth (m)
% Glass
Mazama
Glacier Peak
LITHOZONES
LZ-7
LZ-6
LZ-5
LZ-4
LZ-3
LZ-2
LZ-1
Phi
coarser
missing interval - sandy unit lost in coring
Grain Size
Grain Roundness
Loss on Ignition
% Charcoal
Chemistry
P
Na
K
S
PPM
PPT
Pebbles
Granules
Sand
Silt
Clay
Well rounded
Rounded
Sub-rounded
Sub-angular
Angular
Very angular
Organic matter
Carbonates
Ash

FIGURE 4.7A

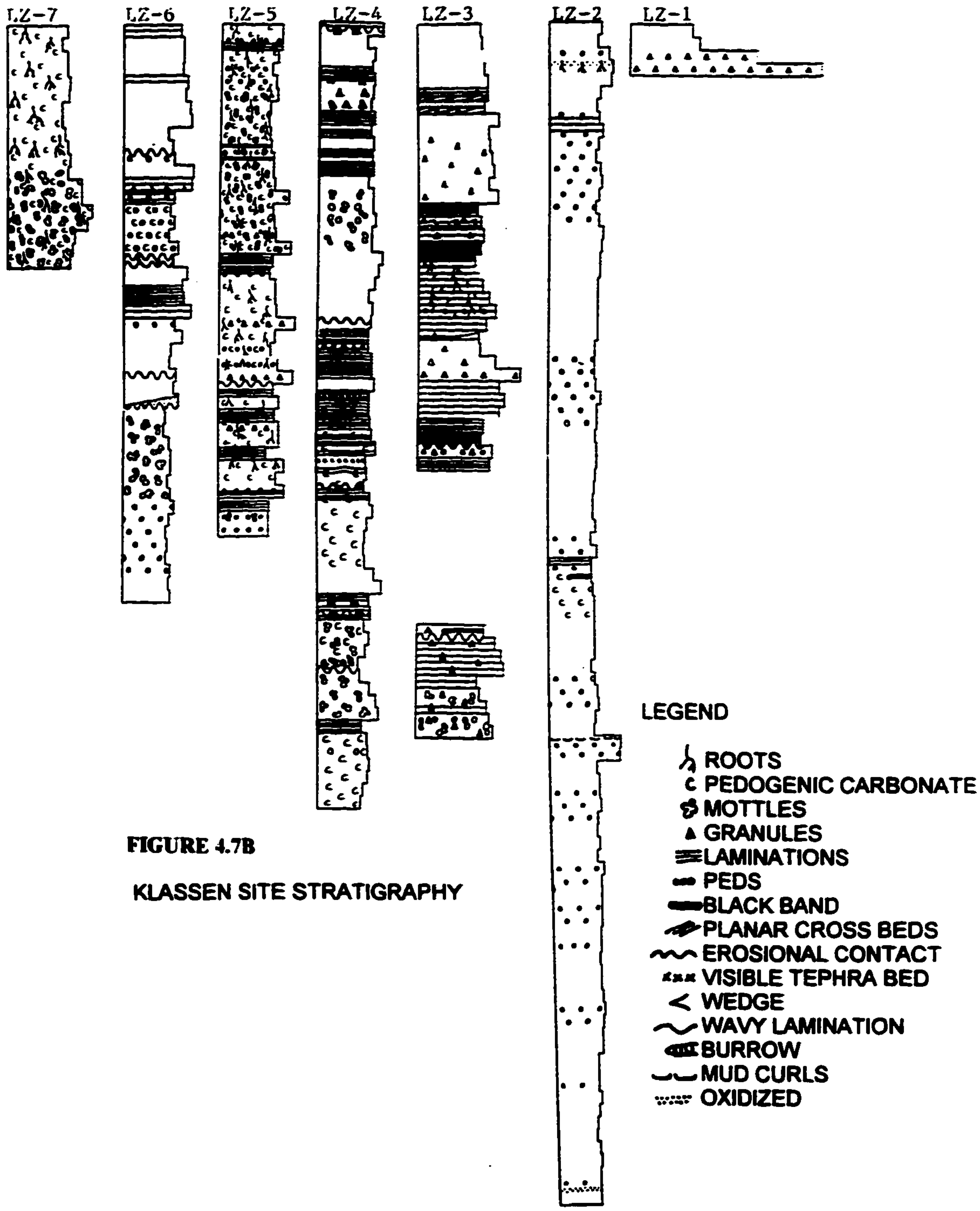


FIGURE 4.7B

KLASSEN SITE STRATIGRAPHY

TOP  BOTTOM

Figure 4.8

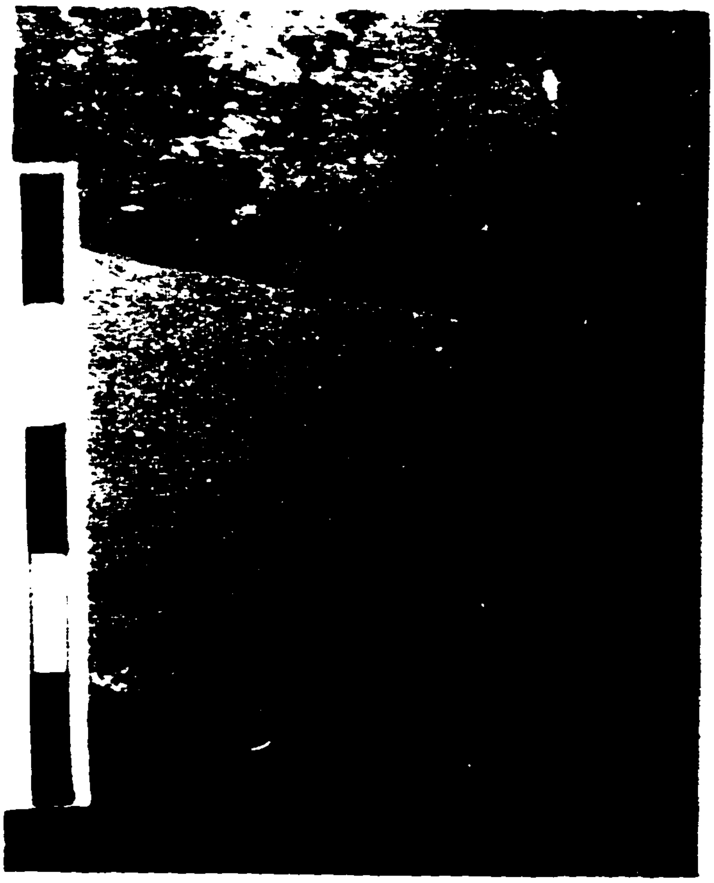
TOP  BOTTOM

Figure 4.9

TOP  BOTTOM

Figure 4.10

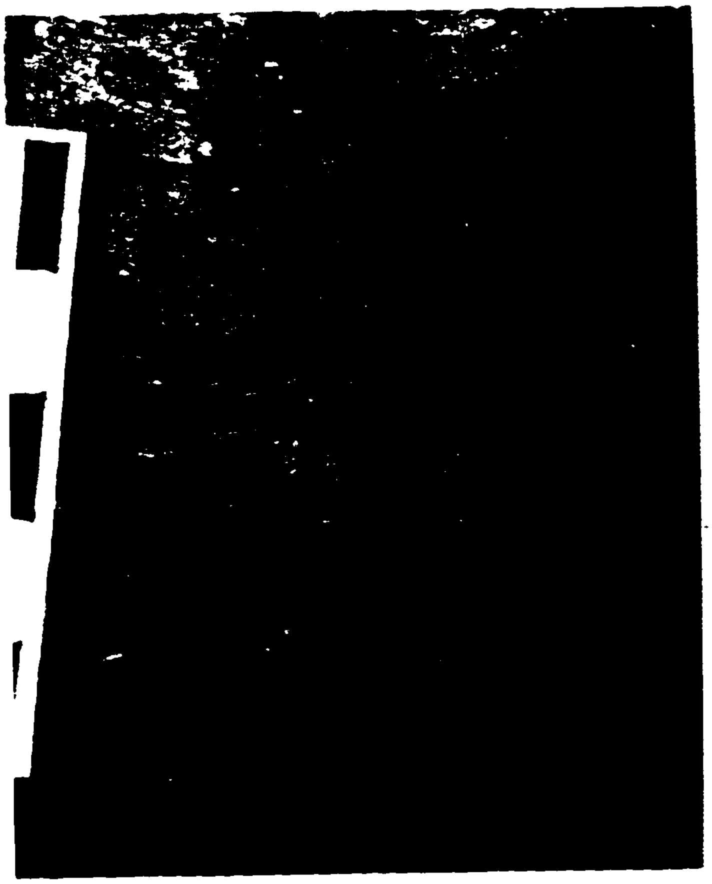
TOP



BOTTOM

Figure 4.11

TOP



BOTTOM

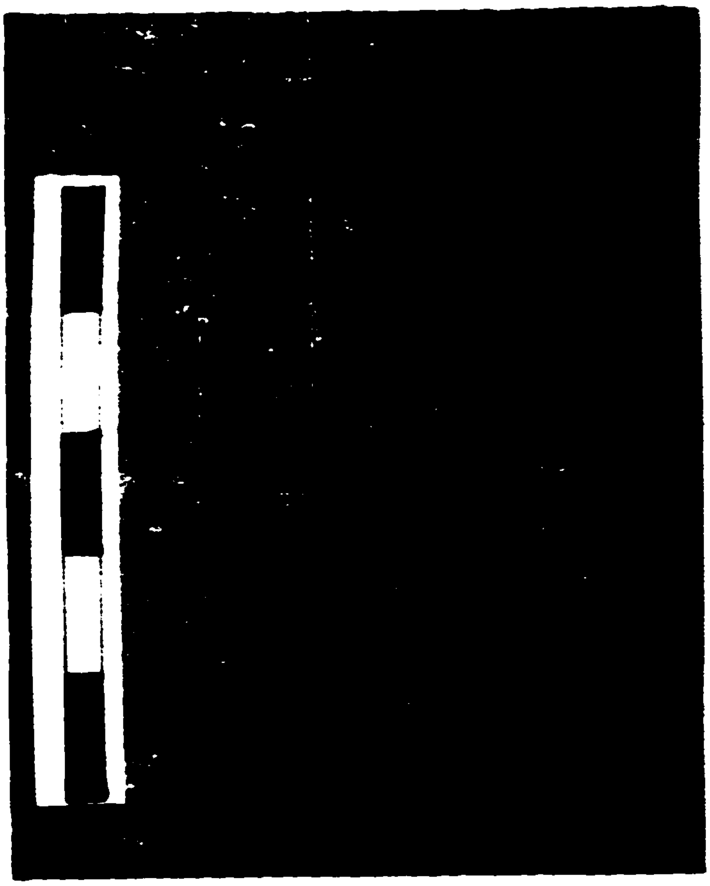
Figure 4.12

TOP

BOTTOM

Figure 4.13

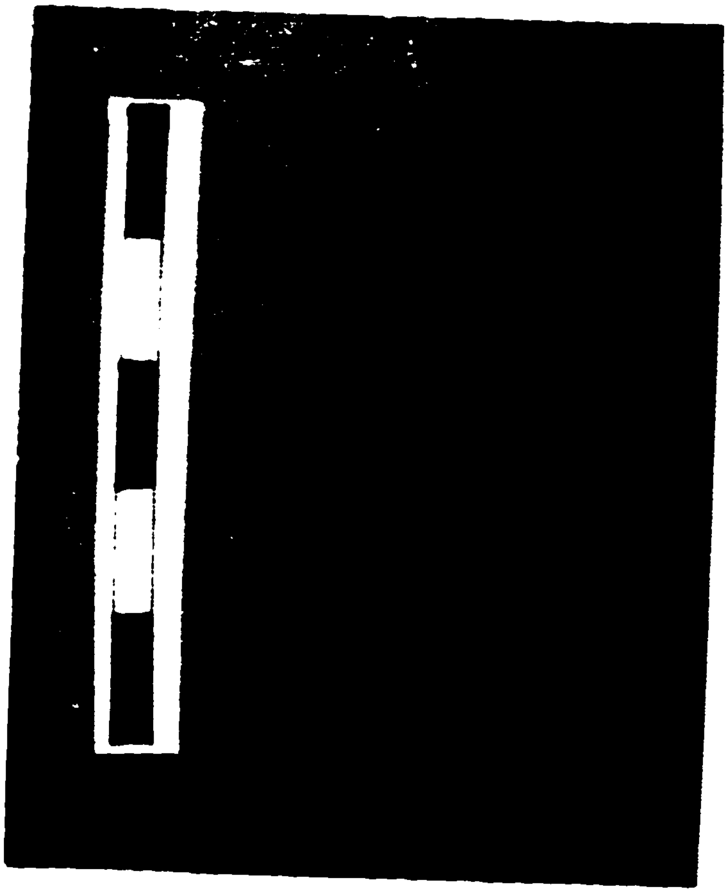
TOP



BOTTOM

Figure 4.14

TOP



BOTTOM

Figure 4.15

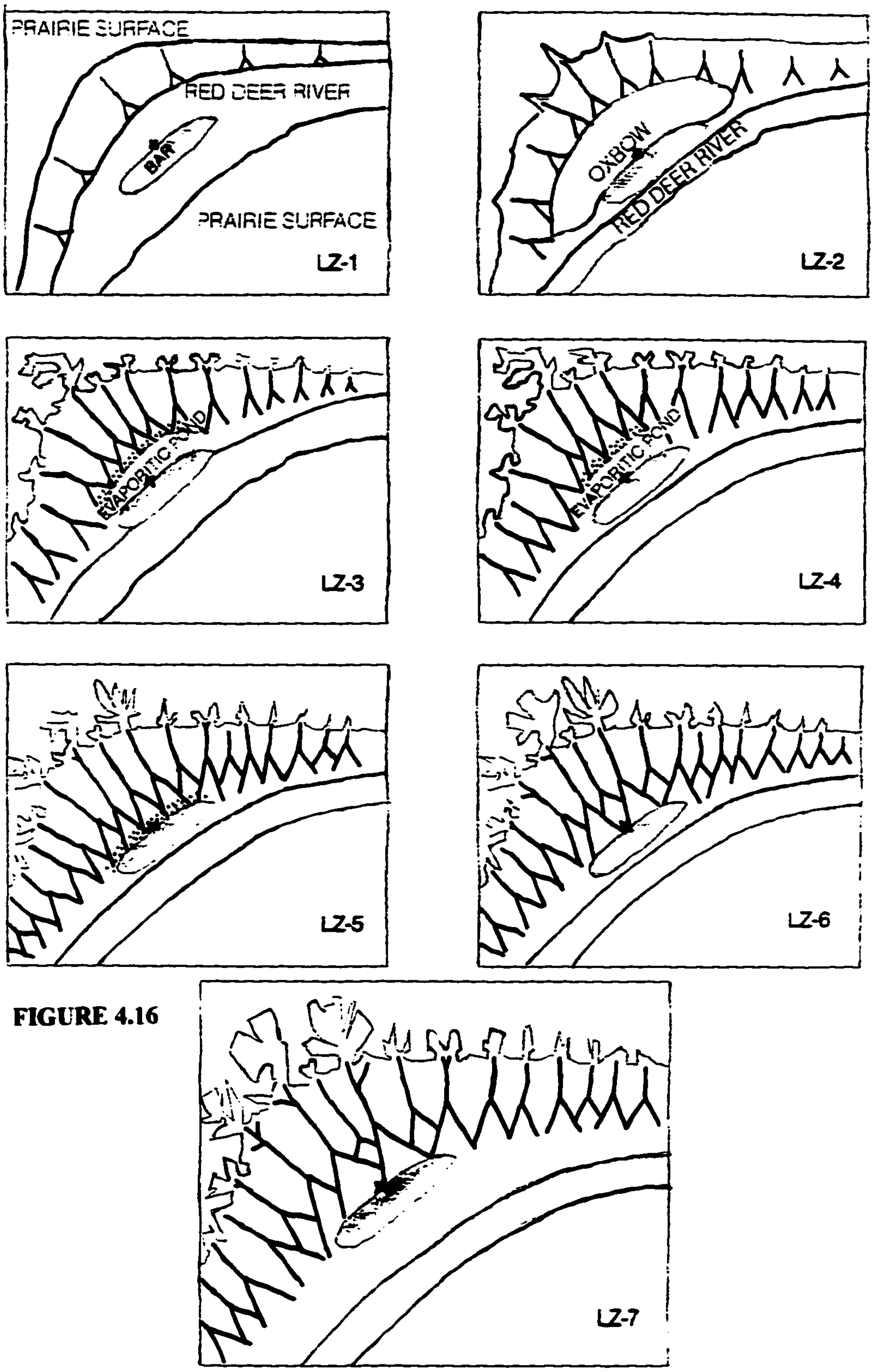

PRAIRIE SURFACE
RED DEER RIVER
BAR
PRAIRIE SURFACE
LZ-1
OXBOW
RED DEER RIVER
LZ-2
EVAPORITIC POND
LZ-3
EVAPORITIC POND
LZ-4
LZ-5
LZ-6
LZ-7

FIGURE 4.16

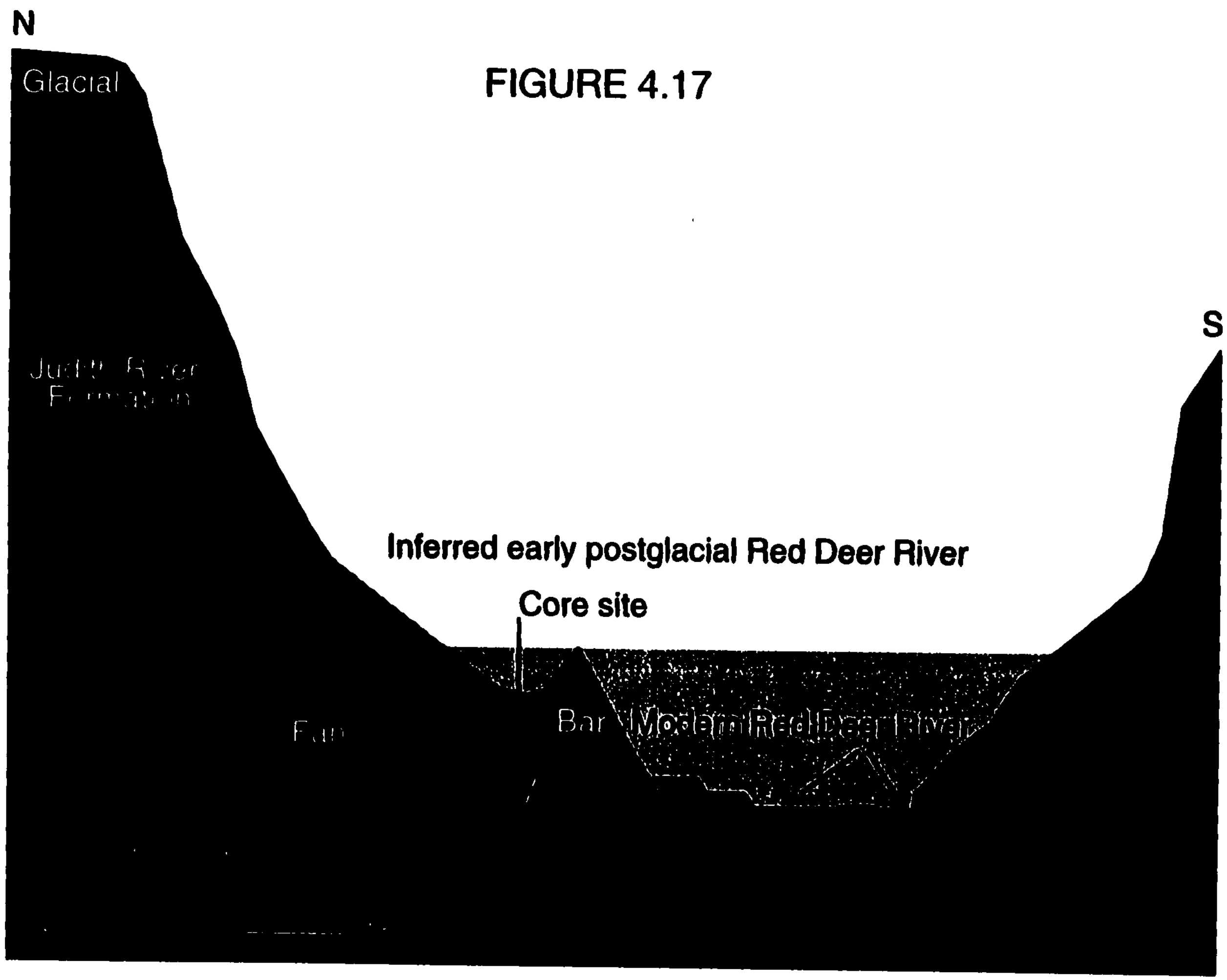

N
S
Glacial
Inferred early postglacial Red Deer River
Core site
Fan
Bar
Modern Red Deer River

FIGURE 4.17

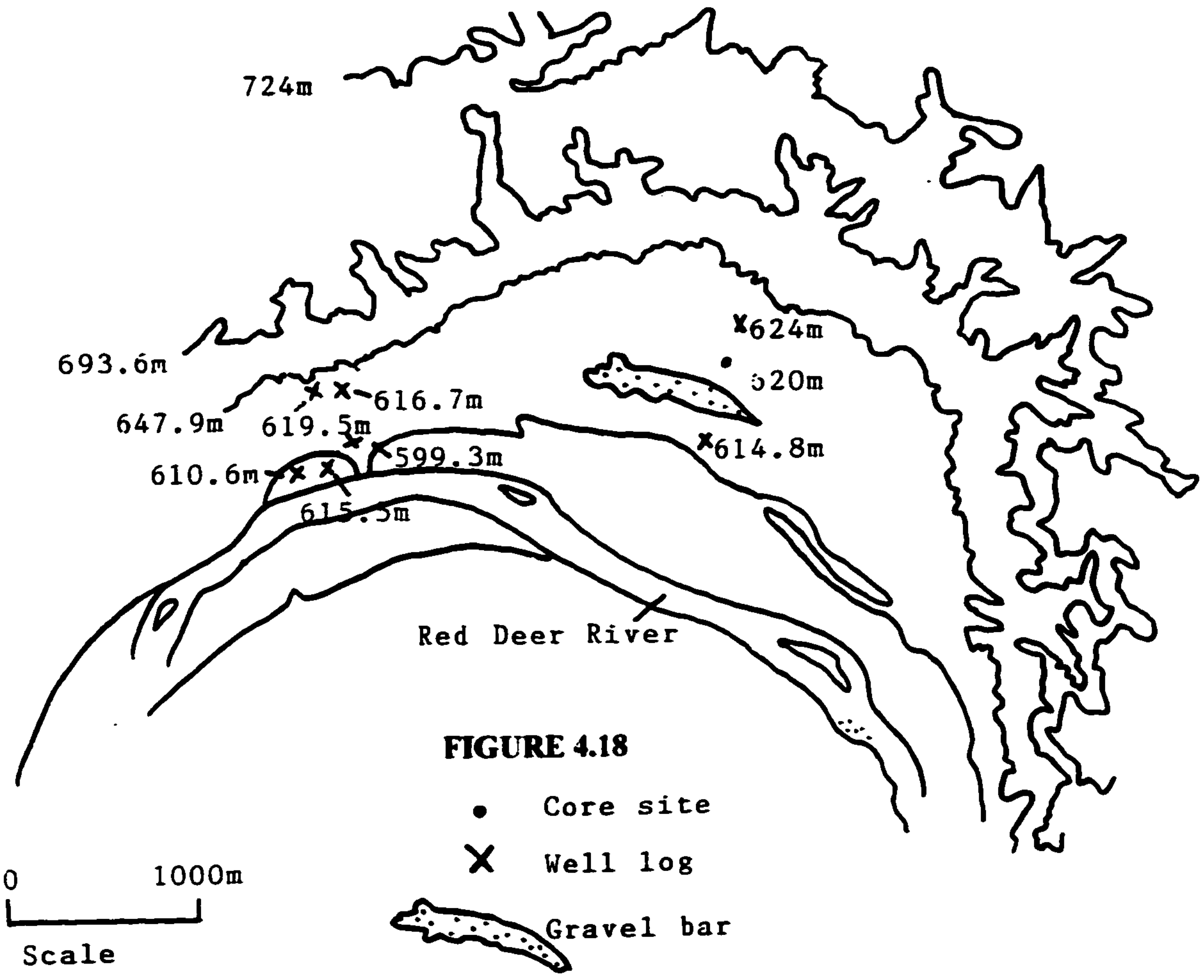
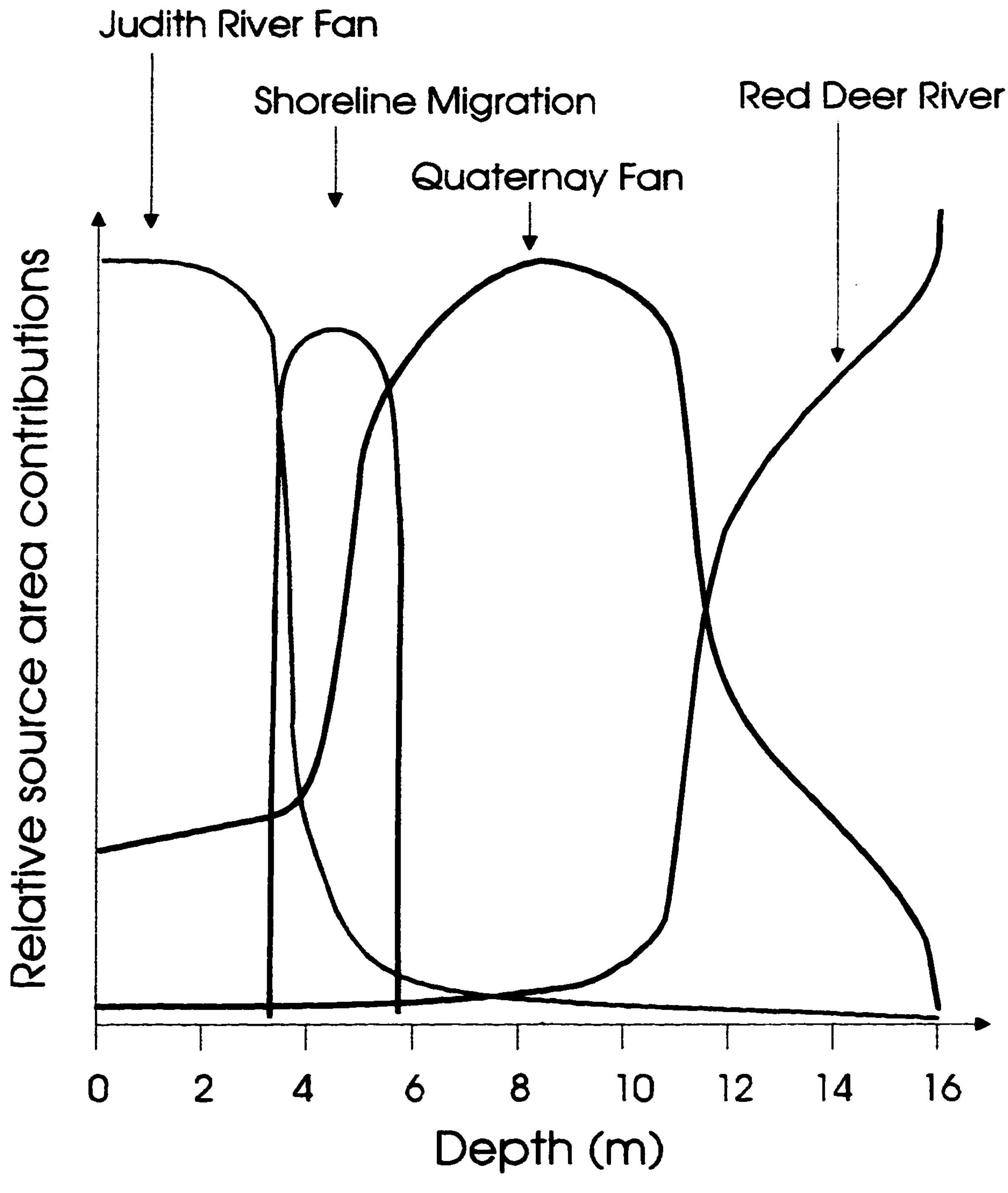

724m
693.6m
647.9m
616.7m
619.5m
610.6m
599.3m
615.5m
624m
620m
614.8m
Red Deer River
Core site
Well log
Gravel bar
0
1000m
Scale

FIGURE 4.18



Judith River Fan
Shoreline Migration
Red Deer River
Quaternay Fan
Relative source area contributions
0
2
4
6
8
10
12
14
16
Depth (m)

FIGURE 4.19

4.10 CHAPTER 4 -REFERENCES

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CHAPTER 5
LATE HOLOCENE STREAM DISCHARGE, TERRACE DEVELOPMENT, AND CLIMATE IN SOUTHERN ALBERTA, CANADA

5.1 INTRODUCTION

Grain size of minerogenic lake sediment has been used as an indicator of water depth and hence paleoclimate, with coarser grains typically interpreted as increasing proximity to shoreline and hence shallower water (Digerfeldt 1986). However, in humid climate lakes where the level is controlled by the outlet sill (for example lakes which are not closed), lake level and area fluctuations are often minimal compared to the fluctuations in closed lakes (C. Campbell *et al.* 1994 - Chapter 2). Changes in the coarseness of sediments in open lakes may be controlled by factors other than the relatively unvarying proximity to shoreline. Other sources of grain size variation include wave influences (which will be minimal in deep lakes), and slumping or turbidity currents bringing coarser shallow water sediments into deeper water (which will be minimized in lakes with broad flat bottoms). If the lake also has steep sides, then the impact of any lake level variations which do occur will be minimized. Further, if the lake is small enough and the inlet streams are weak enough, there may be no strong within-lake currents. In humid region lakes which meet all of these criteria, variations in the grain size of clastic sediments in the middle of the lake will probably reflect mainly past stream discharge.

As stream velocity (a function of discharge) increases, larger sediment grains can be transported in the stream's suspended load. When a stream's sediment load is carried into the still water of a lake, the coarser suspended sediment settles out and is deposited on the lake bed. If the lake is small enough to have a relatively short mean lake water residence time and the outlet stream flows quickly enough, the finer clays will be removed from the lake before they are deposited. Thus, in lakes meeting the above criteria and where other factors are unchanged, changes in the clastic grain size distribution of the sediments may be indicative of fluctuations in outlet stream discharge. The grain-size record of lake sediments may therefore be a proxy of past changes in climate relating to the excess of precipitation over evapotranspiration and other losses, or effective precipitation. The variations in outlet flow will also reflect mean lake water residence time.

Here, analyses of grain size at continuous one cm resolution intervals for Pine Lake, Alberta, Canada for the last 4000 years are presented. If the hypothesis (that grain size at this site reflects climate with coarser grains indicating moister periods) holds true, this should provide the longest continuous non-vegetation based high-resolution record of late Holocene climate available for this region. To determine the validity of the hypothesis, the grain-size record is compared first with known historic climate fluctuations, then with known prehistoric climate fluctuations, and finally against stream terrace development in associated small tributary valleys of the Red Deer River.

5.2 STUDY SITE

Pine Lake (52° 04' N 113° 27' W) is located in a meltwater channel 35 km southeast of Red Deer, midway between Calgary and Edmonton, Alberta (Figure 5.1). It is 3.89 km^2 in surface area, steep-sided, flat-bottomed, 13.2 m deep (maximum), with an estimated volume of 24,087,620 m^3, and drains 150 km^2 of hummocky terrain, mostly by percolation through glacial deposits rather than by direct overland flow (Crosby 1990; Sosiak and Trew 1996). The basin area to lake surface area ratio is 39:1 (Sosiak and Trew 1996). The lake itself is long and narrow, and was probably formed as a subaerial or subglacial meltwater channel carved into the Tertiary Paskapoo Formation. It lies near the northern limit of the Great Plains in the southern Groveland Subregion of the Aspen Parkland (Strong and Leggat 1992). There are two permanent streams and one intermittent stream draining into the lake. One permanent stream (Ghostpine Creek) drains the lake through Three Hills Creek to the Red Deer River.

5.3 METHODS

A 3.75 m long, 7.6 cm diameter Reasoner (Reasoner 1986) core was extracted in the winter of 1992 from the deepest part of Pine Lake along with a freezing sampler (Swain 1973) core of the top 0.71 m of sediment. Two columns from the freezing sampler were subdivided into one cm slices, and each sample placed in a hermetically sealed bottle. A third column was sent frozen intact to Flett Research labs in Winnipeg for ^{210}Pb analysis. A sample of wood charcoal from 3.63 m was dated by A.M.S. at IsoTrace Laboratory, University of Toronto.

The sediment is irregularly laminated, organic, silty clays, with abundant $CaCO_3$ and shell horizons. The laminations may represent annual accumulations, but are too irregular to be useful for dating the sediment; however, their presence demonstrates the relative lack of bioturbation or other postdepositional disturbance of the sediment. The Reasoner core and the freezing core were correlated using a prominent shell horizon and the occurrence of tephra from the 1980 Mt. St. Helens eruption at a depth of seven cm.

Samples of five grams dry weight were treated with 25 ml of 30 % H_2O_2 at 90° C to remove organics. When oxidation ceased, samples were oven dried for at least three hours in a drying oven at 50° C. Grain size analysis was performed on three grams per sample using a Sedigraph 5100, with potassium hexametaphosphate as the dispersing agent, and the expected grain density set at 2.65 (quartz). Duplicate samples were prepared and analyzed for grain size every 20 cm. These replicate analyses indicate precision of the data reported here is approximately $\pm$2.5%. The results are reported using the Wentworth (1922) classification.

5.4 RESULTS

5.4.1 DATING CONTROL

Dating control is provided by ^{210}Pb analysis of the 0.71 m freezing core (I.D. Campbell 1996), the occurrence of the 1980 Mount St. Helens volcanic ash at 0.07 m depth (Table 1), and an A.M.S. radiocarbon date on wood charcoal at 3.63 m of 3540 $\pm$ 160 ^{14}C yr BP (TO-4160; 3620-4080 cal yr BP - calibrated [before A.D. 1955] using CALIB Rev 3.0.3 which assigns probabilities of the calibrated date falling into various intervals, the date range reported here is that which yields a 100% probability when 1σ is used [Stuiver and Reimer 1993]). The A.M.S. date is the average of two normal precision targets and has been corrected for natural and sputtering fractionation to a base of $\delta^{13}C$ = 125‰. Dates for each one cm interval have been interpolated and extrapolated using the power function age(yrs) = 0.16 + 0.70·depth (cm)$^{1.37}$; this function accounts for compaction of the sediment with depth, and provides a better interpolation than would either a straight regression line or "connect the dots" (I.D. Campbell 1996). Using these dates, one cm near the top is approximately one year of sediment accumulation, while near the bottom, one cm is closer to 17 years of accumulation. This has the effect of imposing a running smooth with increasing bandwidth towards the bottom of the core, muting both the frequency and amplitude of short-term variation in grain size towards the bottom.

5.4.2 GRAIN SIZE

Figure 5.2 shows the grain size distribution. The sediment is dominantly fine to very fine silt, with abundant clays, particularly in the lower part of the core. There is very little sand and granules within the sediment. Figure 5.3 shows the standardized median grain size curve, with an inset showing an expansion of the curve for the historic period. There is pronounced variation in the median grain size, which shows troughs in the 1980s, 1960s and in the late 1920s and 1930s, and 1890s as well *ca.* 750-1250 years ago, and prior to *ca.* 2000 years ago.

5.5 INTERPRETATION

5.5.1 GRAIN SIZE

The pronounced declines in the median grain size in the 1980s, 1960s, the late 1920s-1930s, and 1890s correspond with historic drought periods on the Canadian prairies; the 1980s and 1960s drought periods are found locally in the Pine Lake precipitation record (Ahmed 1995). However, the grain size - climate relationship is not a simple increase in the relative abundance of a single fraction of fine grains during dry periods, as might be expected if the change in grain size were due to the increase in aeolian dust, but, rather, is an increase in the fineness of grains during dry periods. For example, during the 1980s trough, there is first a decrease in the relative abundance of fine silt and an increase in very fine silt, than of clay, then a return to very fine silt and eventually (in 1990) to fine silt. Thus, the process responsible for these grain-size variations must be capable of producing increasingly fine sediment with increasingly droughty conditions, whereas aeolian activity would be expected to produce an increase in silt and sand with drier climate.

The most likely explanation for the variations in grain size is variation in stream power at both the inflows and outflow to Pine Lake, and hence in the removal of fine sediments from the lake. The standardized median grain size curve in Figure 5.3, may therefore be taken as an index of lake water residence time. Flow through the lake and discharge down through Ghostpine Creek removes the fine clays; the faster the outflow, the shorter the lake water residence time and the coarser the material removed.

It is possible, that at various times the water level in Pine Lake fell below the outlet sill. At such times, the lake water residence time becomes a function of precipitation and evapotranspiration, rather than streamflow. Also, the movement of water in the lake, which determines the grain size of the sediment, becomes a function of wind rather than of climatic moisture. There is no evidence in this record for such a drop in lake level, but this possibility must be considered, particularly in reference to the driest periods shown in this record. The lake would have to fall at least 0.7 m to close the sill at 888.9 m (Ames 1992), and has not varied by more than +0.5 to -0.6 m during the period 1964-1987 (Crosby 1990). Were the level to fall below the sill, it would have the effect of imposing a lower boundary on the sensitivity of grain size to climate; any values significantly above the lowest median grain size index values may be considered valid proxies of past climate. At the lowest values, the median grain size index can only be considered to indicate a climate "drier than" the moisture level required to keep Pine Lake an open system.

In addition to the historic fluctuations, the median grain size shows fluctuations corresponding with the wet Little Ice Age (Vance *et al.* 1992), with peak discharge *ca.* 330 years ago; the dry Medieval Warm Period (Vance *et al.* 1992; Laird *et al.* 1997), with lowest discharge *ca.* 990 years ago; the wet early Neoglacial (Vance *et al.* 1992), with peak discharge *ca.* 1950 years ago; and preceding these, an unnamed dry interval with two wetter episodes embedded in it, and a slightly moister interval towards the base of the core. Similar interpetations of lacustrine sequences with climatic fluctuations recorded in lacustrine sediments have been described regionally (Vance *et al.* 1992) and globally (with a plethora of associated names), although there are regional variations in direction and magnitude (for example Baerreis and Bryson 1965; Bryson and Murray 1977; Lamb 1977; Cook *et al.* 1991; Mikami [ed.] 1992; Hughes and Diaz 1994; Stine 1994; Vance *et al.* 1992), which is to be expected given the complexity of the global climate system. The detection of events such as the Little Ice Age and Medieval Warm Period in proxy records from every continent (Grove, 1988) implies a global scale forcing mechanism. Stine (1994) has suggested that a shift in the position of or contraction of the polar vortices could account for the Medieval Warm period and other climatic variations could presumably be accounted for in similar fashion.

5.6 TERRACES ON STREAMS TRIBUTARY TO THE RED DEER RIVER

Ghostpine Creek drains Pine Lake, then flows into Three Hills Creek and thence into the Red Deer River near Drumheller. Terraces along Ghostpine Creek have been studied by Rains *et al.* (1994). Terraces have also been studied along Little Sandhill Creek (I.A. Campbell and Evans 1990), Onetree Creek (I.A. Campbell and Evans 1993), and Matzhiwin Creek (Barling 1995), all of which join the Red Deer River further downstream. Figure 5.3 shows the calibrated radiocarbon dates available for those terraces formed in the last 4000 cal years, together with the standardized median grain size curve from Pine Lake.

The T-4 (most recent) terraces dated at Little Sandhill Creek and Onetree Creek, can be seen to have formed within the last *ca.* 600 years. This corresponds with an increase in stream discharge at Pine Lake. The T-3 terraces dated at Little Sandhill Creek, Matzhiwin Creek, and Onetree Creek, formed between *ca.* 1000 and 2000 cal yr BP, corresponding with another period of high discharge at Pine Lake. The T-2 terraces formed sometime prior to this, with dates at Little Sandhill Creek, Ghostpine Creek, and Onetree Creek, ranging from *ca.* 1700 cal yr BP - 3100 cal yr BP. Since these are dates of materials found in deposits on the terrace, and therefore post-dating stream incision and terrace formation, the terraces probably formed towards the earlier end of that range.

There is no indication of terrace formation between *ca.* 1000 cal yr BP - 600 cal yr BP, which is seen as a time of low stream discharge at Pine Lake.

5.7 DISCUSSION AND CONCLUSIONS

Streams will entrench, causing terraces to form, in response to either a relative lowering of base level (which may be due to tectonism, impoundment, or many other factors) or in response to an increase in discharge (Schumm, 1977). There is no indication that base level was significantly lowered during the periods shown here to have been regional terrace forming periods. However, the Pine Lake record does demonstrate that these were periods of increased stream discharge.

Other proxy studies at Pine Lake on the same core (pollen, charcoal, loss on ignition, grain shape, mineralogy and geochemistry) show little response to changing climate, and indeed show little variation except in response to EuroCanadian anthropogenesis (I.D. Campbell, J. McAndrews, W. Last, pers. comm. 1996; also C. Campbell *et al.*, 1995). The stable isotopes do show a response to climate (M. Padden, pers. comm., 1996), but the expense of stable isotope analysis precludes the one cm interval resolution used here. In this site, grain size analysis provides the most sensitive affordable paleoclimatic proxy record. Grain size analysis of this site shows a clear climate signal. The proposed mechanism for this response is that finer grained sediments are removed from the water column by streamflow with increasingly fine sediments being retained and deposited as streamflow declines.

The formation of terraces in secondary streams is regionally synchronous, implying a response to regional climate change. The late Holocene terraces examined here formed during periods of increased stream discharge, and hence increased climatic moisture.

5.8 CHAPTER 5 - TABLE

TABLE 5.1.

CHEMISTRY OF PINE LAKE TEPHRA AND THE TEPHRA IT MOST CLOSELY RESEMBLES, BY ELECTRON MICROPROBE ANALYSIS OF GLASS

Sample	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	Cl	Water by Difference
Pine Lake 7 cm (n=12; this study)											
	74.45	0.38	15.07	2.28	0.03	0.59	2.47	2.69	1.97	0.07	1.83
Standard dev.											
	0.69	0.04	0.39	0.17	0.02	0.06	0.16	1.09	0.06	0.03	0.85
Mt. St. Helens 1980 (Sarna-Wojcicki *et al.* 1981)											
	71.4	0.44	15.1	2.30	0.06	0.49	2.76	4.30	1.94	0.10	0.81

5.9 CHAPTER 5 - FIGURES

FIGURE 5.1

A. Location of Alberta.

B. The Red Deer Drainage basin (stippled area = badlands [modified atter Campbell, 1974]), Threehills Creek = THC, Ghostpine Creek = GPC.

C. Pine Lake.

D. Pine Lake bathymetry (modified after Sosiak and Trew 1996).

E. The lower Red Deer valley.

FIGURE 5.2

Pine Lake grain size distribution. Ash at 7 cm is Mt.St. Helens 1980.

FIGURE 5.3

5.3A Dated river terraces in the study area for the last 4000 cal yr B.P. (Campebll and Evans 1990, 1993; Rains *et al.* 1994; Barling 1995). Black bars represent 100% calibrated date range at 1σ. All dates have been corrected for $\delta^{13}C$ fractionation and calibrated into calendar years (Stuvier and Reimer, 1993).

5.3B Relationship between cal yr B.P. and ^{14}C years (using intercal93.14c, CALIB Rev 3.0.3; Stuvier and Reimer, 1993).

5.3C Standardized median Pine Lake grain size curve (cal yr B.P.), with an inset showing an expansion of the curve for the historic period (yr A.D.).

FIGURE 5.1

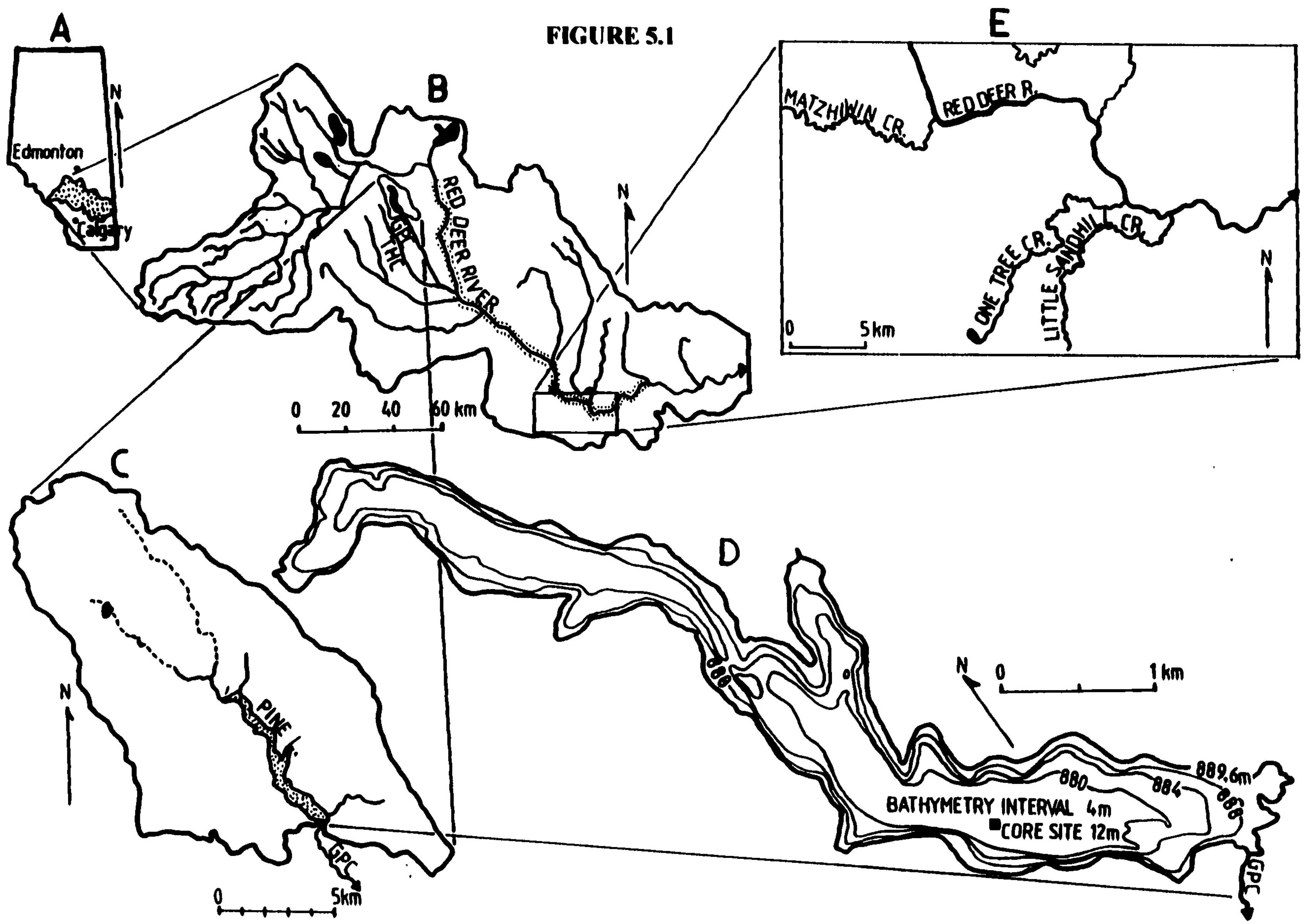

A
N
Edmonton
Calgary
B
RED DEER RIVER
GP
THC
N
0 20 40 60 km
E
MATZHIWIN CR.
RED DEER R.
ONE TREE CR.
LITTLE SANDHILL CR.
N
0 5km
C
N
PINE L.
GPC
0 5km
D
N
0 1km
889.6m
880
884
888
BATHYMETRY INTERVAL 4m
CORE SITE 12m
GPC

FIGURE 5.2

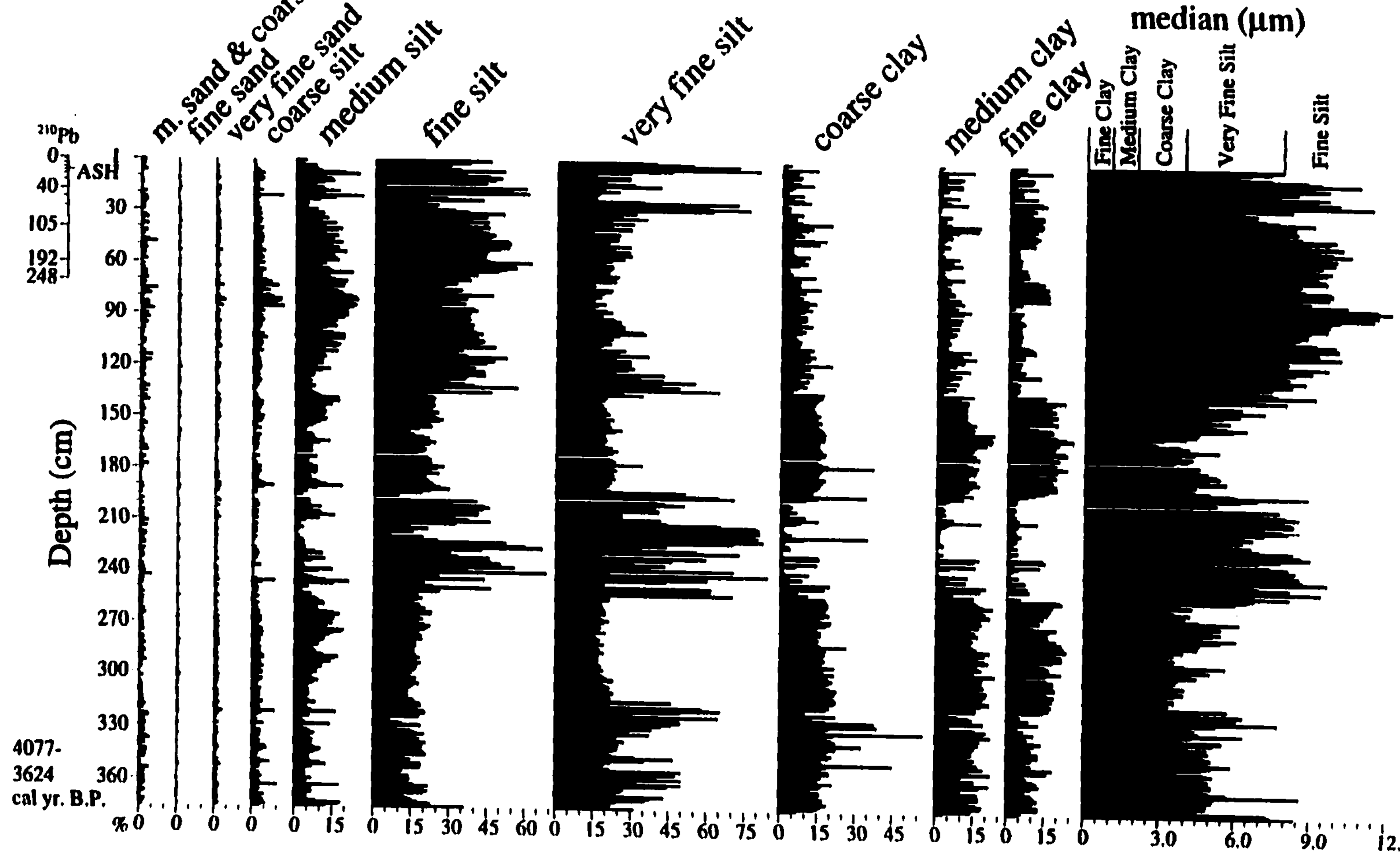
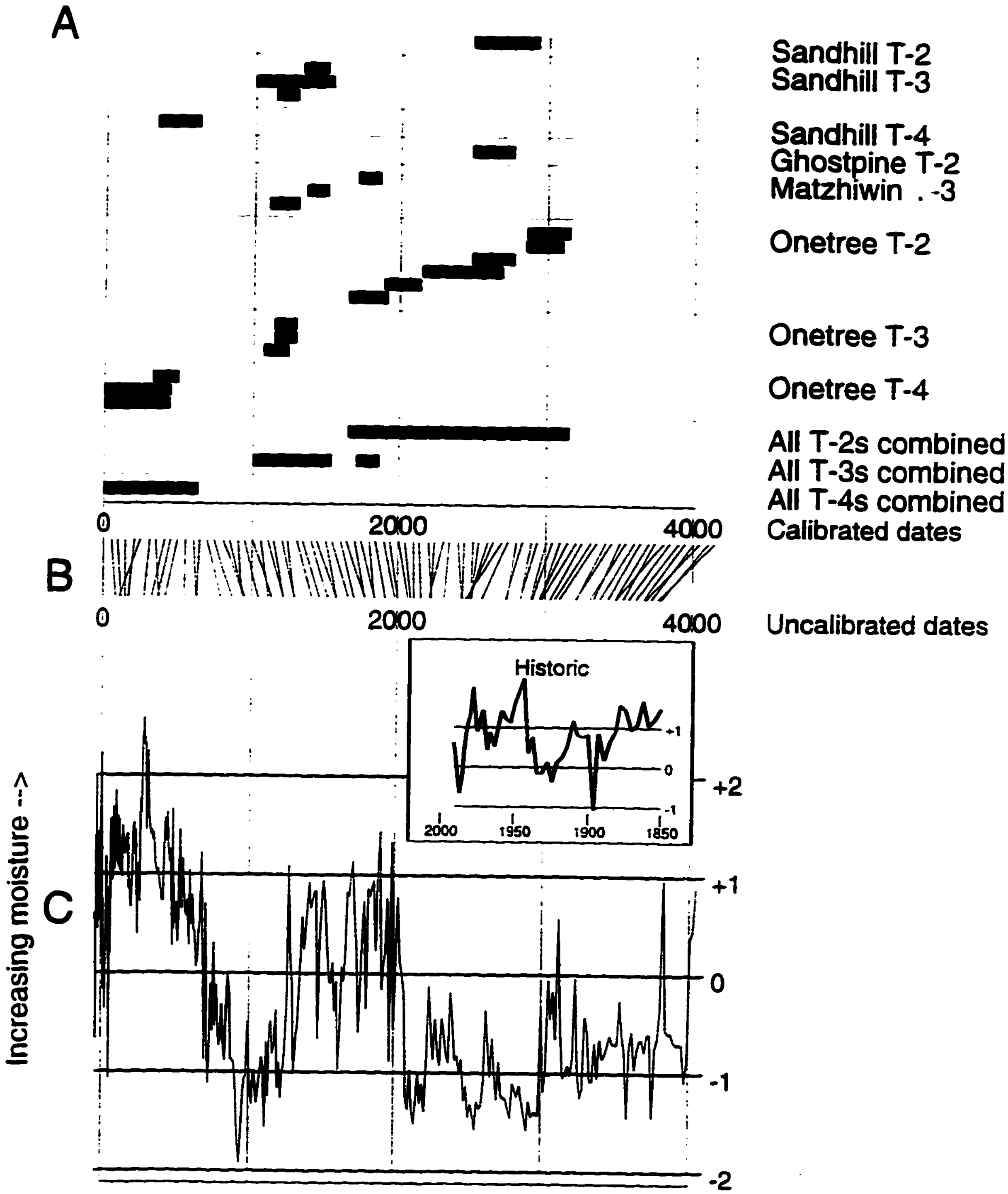

m. sand & coarser
fine sand
very fine sand
coarse silt
medium silt
fine silt
very fine silt
coarse clay
medium clay
fine clay
median (µm)
Fine Clay
Medium Clay
Coarse Clay
Very Fine Silt
Fine Silt
210Pb
0
40
105
192
248
ASH
Depth (cm)
30
60
90
120
150
180
210
240
270
300
330
360
4077-
3624
cal yr. B.P.
%
0
15
30
45
60
75
3.0
6.0
9.0
12.0

FIGURE 5.3


A
Sandhill T-2
Sandhill T-3
Sandhill T-4
Ghostpine T-2
Matzhiwin . -3
Onetree T-2
Onetree T-3
Onetree T-4
All T-2s combined
All T-3s combined
All T-4s combined
0
2000
4000
Calibrated dates
B
0
2000
4000
Uncalibrated dates
Historic
+1
0
-1
2000
1950
1900
1850
C
Increasing moisture -->
+2
+1
0
-1
-2

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CHAPTER 6
AEOLIAN DEPOSITION AND PRESERVATION IN SOUTHEASTERN ALBERTA WITH SPECIAL EMPHASIS ON LOESS DEPOSITION

A modified version of Chapter 6 has been submitted for publication review.
Campbell, C. and Campbell, I.A. Aeolian deposition and preservation in southeastern Alberta with special emphasis on loess deposition.

6.1 INTRODUCTION

The semiarid prairies of the northern Great Plains contain extensive areas of wind deposited sediments. This region is important for agriculture and ranching, and large scale aeolian activity, such as that which occurred during the 1920s, 1930s and 1980s, can have serious socioeconomic consequences (Johnston and Wickenden 1935; Main 1935; Chepil 1945, 1957; Gray 1978; Coote and Pettapiece 1987; Coote and Padbury 1988a, 1988b; Lang and Jones 1988; Wheaton and Chakravarti 1987; Wheaton and Wittrock 1989). As in similar climatic regions elsewhere, this area is extremely sensitive to climatic fluctuations. A more detailed understanding of the nature of the postglacial aeolian geomorphic response to climatic change may help in assessing the probable consequences of future responses to climatic change and variability.

The relationship between climate and aeolian mobilization, deposition, and stabilization in the mid-high latitudes is a complex one because changes in aeolian activity as recorded in wind-blown deposits may reflect the effects of processes which are not necessarily exclusively a function of climate. These processes involve both dependent and independent variables and include: fire, geology, morphology, hydrology, pedology, vegetation, animal and human influences - all of which are temporally and spatially interdependent (David 1977, 1981, in press; Seppälä 1981; Wright 1984; Filion 1984; Sweet 1992). Field and simulation studies on boundary layer and surface controls (*e.g.* wind shear velocity, sediment texture, pore water, snow, ice, vegetation, and frozen and wet surfaces) in niveo-aeolian conditions suggest a remarkably more complex system of interactions than in hot-arid settings (Nickling 1978, 1984, 1988; McKenna Neuman 1989, 1990a, 1990b, 1993; McKenna Neuman and Nickling 1989).

Across the northern Great Plains, evidence of episodic late Holocene aeolian activity has been interpreted from sedimentary sequences in sand dunes and loess (including cliff-top loam) deposits (*e.g.* David 1970, 1972, 1977, 1993, in press; Ahlbrandt *et al.* 1983; Jorstad *et al.* 1986; Forman *et al.* 1992; Vreeken 1993; Madole 1994). Madole (1994) notes that the study region is climatically near the geomorphic threshold at which widespread aeolian transport takes place and that relatively slight variations in average surface temperature and annual precipitation are required to destabilize the surface. David (in press) indicates that this aeolian activity is both temporally and spatially highly variable. In the Canadian prairies, deflation may be maximal from late winter through early summer, when agricultural fields are unvegetated, winds are strong, and normally stable soil aggregates have been comminuted by freeze-thaw action and desiccation (Wheaton 1984; McKenna Neuman 1993). However, as McKenna Neuman (1993) notes, there is a paucity of research on detailed, quantitative studies on niveo-aeolian processes in Canada. Furthermore, the effects of the highly variable continental climate on aeolian processes are poorly understood.

Despite the widespread occurrence of aeolian deposits, little is known about the history and processes of postglacial aeolian activity in southeastern Alberta (David, in press). The correlation and interpretation of aeolian chronologies are hampered by a sparse data-base, the lack of any widespread temporal correlation between aeolian sections (*e.g.* Vreeken 1989, 1993; David 1993; Wolfe *et al.* 1995), a relatively low resolution Holocene climate change record, and an incomplete understanding of how climate variations in mid - high latitudes affects aeolian activity. Here, a composite paleohydrology curve is compared with the stratigraphic, spatial and temporal distribution of aeolian sediments across a west-east transect in southeastern Alberta, in order to assess the relationships between geomorphic setting, climate, and aeolian activity. Moreover, it will be shown that profound differences exist between the conditions applicable to aeolian sands and those relevant to loess deposits. Finally, a conceptual model is proposed for a moisture-driven relationship between loess mobilization, deposition and preservation.

6.2 STUDY AREA

Located in the Northern Great Plains, southeastern Alberta (Figure 6.1) is a chernozem soil-dominated mixed-grass prairie (Strong and Leggat 1992). The area has a midlatitude, semiarid, continental climate with long cold winters and short warm summers (Environment Canada 1993). Potential evapotranspiration often exceeds precipitation by >300 mm yr^{-1} (Winter 1989). Historically, the area has been subjected to recurrent droughts associated with the development of stable high pressure ridges displacing cyclonic tracks, moist air masses, and fronts northward (Dey and Chakravarti 1976). Early and mid-Holocene arid phases, however, may have been caused by different synoptic patterns (Vance 1984). Southeastern Alberta is an area which is particularly favourable to aeolian erosion and deposition; there is an abundance of poorly consolidated fine grained material, precipitation is low, vegetation cover is sparse, and the extensive, relatively flat surface encourages persistent, strong, winds (Odynsky 1958; David 1977; Mulira 1986). Mean annual windspeeds surpass 20 km hr^{-1} across the region with considerable fetch over the essentially flat, largely treeless, terrain (Walmsley and Morris 1993). Strong westerly and southwesterly winds prevail, and in the winter and early spring, chinook winds > 90 km/hr have been recorded (Longley 1972). In combination, these conditions strongly encourage aeolian mobilization and transport.

Based on a calibration of radiocarbon dates of geomorphic events, the glacial and postglacial period can be divided into four major episodes (Chapter 1, Chapter 3): (1) glaciation *ca.* 22,870-21,870 to 20,210-19,430 cal yr B.P.; (2) deglaciation *ca.* 20,210 cal yr B.P. - 12,000 cal yr B.P.; (3) landscape stability *ca.* 12,000-10,000 cal yr B.P.; and (4) landscape instability *ca.* 10,000 cal yr B.P. to the present. Evidence of regional aeolian activity is restricted to these last two periods. All radiocarbon dates used in this paper have been calibrated after Stuiver and Reimer 1993; the range reported here is that which yields a 100% probability with 1σ analytic uncertainty.

6.3 REGIONAL PALEOHYDROLOGY

Figure 6.2 shows a composite synthetic paleohydrology curve for the study area for the last 10,000 cal yr B.P. The bulk of the curve is based on the palynological and mineralogical evidence from Harris Lake, in the Cypress Hills of southwestern Saskatchewan (Sauchyn 1990; Sauchyn and Sauchyn 1991; Last and Sauchyn 1993). Climatic interpretation from the Chappice Lake sedimentary record (Vance *et al.* 1993) is incorporated to provide greater detail in the last 8000 cal yr B.P. Because the Chappice Lake curve is largely based on indicators of salinity, which can be expected to show a long-term increase in an unchanging arid or semiarid climate, the Chappice Lake curve is adjusted by eye, to fit the moistening trend of the Harris Lake curve. The most recent 4000+ years of the record are interpreted on the basis of the grain size record in the sediments of Pine Lake (Chapter 5).

This composite synthetic curve is consistent with the geomorphic (C. Campbell and Campbell 1997) and palynological (Schweger and Hickman 1989; Sauchyn and Sauchyn 1991; Vance *et al.* 1993) records of the region. It shows a long-term moistening trend throughout the Holocene within which multiple events of different magnitudes of wet and dry conditions are superimposed. The most recent wet period corresponds with the Little Ice Age *ca.* 330 years ago, and the preceding dry period with the Medieval Warm Period *ca.* 990 years ago (Vance and Wolfe 1996). This general pattern is also in agreement with the theoretical insolation curve derived from orbital cycles (Berger 1978; Schweger and Hickman 1989; C. Campbell and Campbell 1997).

6.4 REGIONAL AEOLIAN GEOMORPHOLOGY

6.4.1 SAND DUNES

Sand dunes in southern Alberta are dominantly "wet-sand" (David 1979, in press) parabolic dunes with small blowout areas. The dunes are composed primarily of unconsolidated silts to medium size sand grains (*ca.* 39-500 μm) (David 1977, in press; Mulira 1986; Wolfe *et al.* 1995). Odynsky (1958), David (1977), Mulira (1986) and Shetsen (1987) mapped the distribution of large dune deposits in southern Alberta (Figure 6.1). There appears to be a close association between dune fields and glaciolacustrine deltaic deposits throughout southern Alberta.

Muhs and Holliday (1995) developed a conceptual model of aeolian activity in response to climate change in the Great Plains based on an index of dune mobility (Lancaster 1988) and tree-ring records. They suggested that increased dune mobilization in the late Holocene is a function of droughts which, when accompanied by periods of higher than normal mean temperatures, greatly lowers the precipitation to evapotranspiration ratio, reducing vegetation cover, and resulting in dune mobilization. A similar model developed by Wolfe *et al.* (1995), suggests that increased aeolian mobilization in the Great Sand Hills of Saskatchewan was associated with increased average aridity, and that periods of dune stability (indicated by paleosol development) are associated with periods of increased humidity. However, according to McKenna Neuman (1993), warm climate-based wind erosion models do not apply in temperatures below 0°C, possibly rendering such models inapplicable to this study area if peak aeolian activity occurs in late winter as suggested by Wheaton (1984).

David (in press) suggests that the three conditions required for aeolian activity are: a supply of favourably sized material, sufficiently strong winds, and a locally sparse to absent vegetation cover. Based on stratigraphic evidence, David (1981) proposed that the Cree Lake parabolic dunes in northern Saskatchewan were initially formed by dry katabatic winds coming off the receding Laurentide ice sheet. It is reasonable to suppose that dune genesis in southern Alberta began under similar conditions when vast expanses of unvegetated, poorly consolidated, fine grained deposits were exposed to deflation (Odynsky 1958; David 1993; Vreeken 1993), *ca.* 20,210-19,430 cal yr B.P. (Evans and Campbell 1992, 1995) and that dunes and related deposits (sandsheets etc.) formed throughout the postglacial period whenever and wherever conditions permitted. Optically Stimulated Luminescence (OSL) dates on dune deposits in southern Saskatchewan (*ca.* 1980s, 1940s, <200-600 B.P., <2600 B.P.; Wolfe *et al.* 1995) show episodic mobilization and stabilization throughout the late Holocene. This finding agrees with dates on dune activity elsewhere in the Northern Great Plains. For example, Madole (1994) shows that extensive areas of aeolian sands along the valley of the South Platte River, Colorado, were mobilized up to three times within the last 1000 years, and stabilized prior to the historic period.

6.4.2 LOESS

Loess is a terrestrial windblown silt deposit (Pye 1987). The loess in southern Alberta (for example Catto 1983; Bryan *et al.* 1987; Vreeken 1989, 1993) is dominantly 'sandy loess' as the sand component generally exceeds 20% (Pye 1987). Globally, loess is associated with highly variable climatic conditions, ranging from arid to semiarid conditions in both hot and cold climates (for example Pye 1987; Middleton 1989). In the Great Plains loess is primarily formed from sediments derived from outwash deposits associated with late Wisconsinan glaciation (Smalley 1972; Pye 1987; Shetsen 1987; Vreeken 1989). Suspended fines, blown from these exposed sediments, are trapped by vegetation and topographic obstacles. Typically particle size and deposit thickness decrease systematically with distance from the source (Ruhe 1969; Catto 1983; Pye 1987), resulting in aeolian differentiation, which is a function of particle segregation by fall velocity during downwind transport. Catto (1983) found that cliff top loess in the Cypress Hills, Saskatchewan, decreased in abundance (from 3.27 to 0 m in thickness) and coarseness (coarse to fine silt), in a wedge shaped distribution with distance from the escarpment across 9 km.

There is evidence of multiple cycles of loess mobilization during the late Pleistocene and Holocene in the Northern Great Plains (Souster *et al.* 1977; Colton 1978; Benedict 1981, 1985; Bryant *et al.* 1981; Jorstad *et al.* 1986; Pennock and Vreeken 1986; Vreeken 1986, 1993; Madole 1994). However, while loess is apparently widespread it is neither areally nor volumetrically as significant in Canada as it is in the central United States (Thorp and Smith 1952; Krinitzsky and Burnbull 1967). Several loess deposits have been identified in southeastern Alberta (Figure 6.1; Thorp and Smith 1952; Jungerius 1966; Westgate 1968; Catto 1983; O'Hara 1986; Vreeken 1986, 1989, 1990, 1993; Bryan *et al.* 1987; O'Hara and Campbell 1993). Where studied in detail, it appears that loess deposits in southern Alberta preferentially accumulate on cliff tops and in topographic depressions (David 1972; Catto 1983; Vreeken 1989, 1993).

Vreeken (1986, 1989, 1990, 1993) has shown that deposition of post late Wisconsinan loess began in southern Alberta before the deposition of the Glacier Peak tephra bed *ca.*13,460-12,980 cal yr B.P. (after Mehringer *et al.* 1977). Vreeken distinguished up to six paleosol / loess sequences prior to the deposition of the Mazama tephra bed (*ca.* 7800-7480 cal yr B.P.; after Bacon 1983) and up to five sequences after the deposition of the Mazama tephra bed in the Lethbridge area. However O'Hara (1986), Bryan *et al.* (1987), and O'Hara and Campbell (1993) found only a single, but widespread, mid-Holocene loess deposit in Dinosaur Provincial Park. Because the frequency and size of suitable source areas is fairly uniform across southern Alberta this northeastwards decrease in the number of loess deposits (from 11 to one) is interpreted here as reflecting increasingly unfavourable geomorphic conditions for loess mobilization and deposition (*e.g.* lower moisture, decreased chinook activity, and lower density of vegetation).

6.5 AEOLIAN DISTRIBUTION, GRANULOMETRY AND AGES OF AEOLIAN DEPOSITS: DESCRIPTION AND INTERPRETATION

The area described here consists of three parts: (1) the Red Deer delta / Duchess sand dune field, (2) the Dinosaur Provincial Park badlands, and (3) the prairie surface (Figure 6.1D).

6.5.1 DUCHESS DUNE FIELD

6.5.1.1 DESCRIPTION

The dominant aeolian sediment source in the study area is almost certainly associated with the ancestral Red Deer River delta (Shetsen 1987), which formed when Glacial Lake Drumheller decanted into Glacial Lake Bassano during the late Wisconsinan. The delta sediments are dominantly silt sized (Paterson 1996). The sand dunes of the Duchess Dune field have formed on the downwind distal limits of the deltaic deposits.The dune area (Figure 6.1D) is comprised of three lobes of parabolic and blow-out sequences of aeolian deposits covering a total of *ca.* 400 km^2. These dunes are mostly of silt to medium sized sand grains forming deposits up to 7m thick; the interdune corridors are predominantly ice-contact glaciolacustrine and glaciofluvial deposits up to 25 m thick (Shetsen 1987). Bryan *et al.* (1987) believed that the dunes formed shortly after deglaciation at *ca.* 12,600 yr B.P.

Exposures in 17 dunes along widely separated transects in the Duchess Dune field were described (Figure 6.3), and selected sequences were dated by five OSL and one AMS date (Table 1). All the dune depositional sequences contain evidence of multiple events of dune stabilization with, in one exposure, $\leq$ 15 paleosols. None of the exposures contained visible tephra beds.

6.5.1.2 INTERPRETATION

Calibrated Accelerator Mass Spectrometry (AMS) and OSL dates on samples of dune deposits, and the absence of the regional Mazama tephra bed (*ca.* 7800-7480 cal yr B.P.) from the Duchess dunes, supports the interpretation that these dunes have been mobilized repeatedly throughout the late Holocene. If the late Holocene dunes are composed of redeposited sediments from older sand dunes, then conditions prior to the late Holocene may have been unfavourable for long term dune stabilization.

Because the OSL and AMS values date deposition of aeolian sand which has not been remobilized since the dated deposition, the deposits can be interpreted as indicating terminal conditions for dune mobilization. When compared with the regional hydrology curve (Figure 6.2), there is no apparent association between dune stabilization and periods of either wetter or drier conditions during what is generally a relatively moister late Holocene. It can be posited that, in addition to aridity, other variables must exist in order to result in dune mobilization; these include, for example, abrupt periods of severe drought (David 1982), ground breaking by fire (Filion 1984), or bioturbation (faunal or anthropogenic). While aridity is therefore believed to be required (*e.g.* Muhs and Holliday 1995; Wolfe *et al.* 1995), it may not be, of itself, sufficient to cause aeolian mobilization, nor may it be the sole cause of dune activity. All the dates are limited to the relatively moist late Holocene which suggests that such climatic conditions favour dune stabilization, and that any evidence of prior dune mobilization has been obliterated, or is at a greater depth.

6.5.2 DINOSAUR PROVINCIAL PARK LOESS

6.5.2.1 DESCRIPTION

Badlands fringe the Red Deer River for over 300 km, forming an area of about 800 km^2

(I.A. Campbell 1974). The Dinosaur Provincial Park badlands are exceptionally well developed and well studied. The area is dominated by fine grained, poorly consolidated Upper Cretaceous Judith River Group (dominant bedrock) sediments, which are characterised by highly erodible sodic, smectitic mudstones, claystones, siltstones, and fine sandstones (Bryan *et al.* 1988).

The badlands formed following rapid postglacial (late Wisconsinan) down-cutting by the Red Deer River (Bryan *et al.* 1987). Bryan *et al.* (1987) recognized four postglacial surfaces in Dinosaur Provincial Park. The upper two surfaces resulted from spillway development associated with deglaciation. Of the lower surfaces, surface 3 is associated with locally generated runoff, and surface 4 is related to the Red Deer River floodplain (Figure 6.4).

A structureless, homogeneous horizon of fine sand and silt (dominated by very fine sand [625-125 μm]) with an abundant mesh of fine herbaceous roots, caps many remnants of surfaces 1-3. It is interpreted as loess deposited over a period of a few hundred years (Bryan *et al.* 1987; O'Hara and Campbell 1993). A basal date of 5400 $\pm$ 800 TL yr B.P. (Alpha-2070) was obtained from a loess deposit (Bryan *et al.* 1987).

For this study grain size was determined from 30 samples of loess collected from different aged surfaces within Dinosaur Provincial Park, to develop a regional loess standard. Grain size analysis was performed after Rutter (1995). For each sample, at least 100 grams were sieved through a 2000 μm (-1 ϕ) sieve. Approximately 60 grams of each homogenized sample were treated with 25 ml of 30 % H_2O_2 at 90° C to remove organics. When oxidation ceased, samples were oven-dried for at least 3 hours in a drying oven at 50°C. Fifty grams of the organic-free sample were dispersed for at least 8 hours in 125 ml of 4% sodium hexametaphosphate solution. The sample and solution were homogenized in a blender for 15 minutes. The suspension was then wet sieved through a 4.5 ϕ screen until the water ran clear (about 15 minutes), to remove the clays and fine silts from the sample, as they formed peds up to coarse sand size during dry and wet sieving, biasing the results. The fraction > 4.5 ϕ was retained and dried in an oven for at least 3 hours at 50°C. It was then weighed and sieved through a series of seven sieves: 2000 μm (-1 ϕ), 1000 μm (0 ϕ), 500 μm (1 ϕ), 250 μm (2 ϕ), 125 μm (3 ϕ), 625 μm (4 ϕ) and 45 μm (4.5 ϕ). The sieve results were plotted as cumulative curves. There was minimal grain size difference between all 30 samples. Figure 6.5 shows the cumulative combined curve of all the loess deposits. This standard shows that approximately 70% of grains are very fine sand - fine sand (625-250 μm), with a low clay content. This grain size distribution accords with the analysis of Bryan et al. (1987) on their loess samples.

6.5.2.2 INTERPRETATION

In southeastern Alberta, there are very few sites on the prairie surface that contain tephra beds. However, the widely separated regional nature of the tephra occurrence (*e.g.* Christiansen 1961; David 1970; Westgate *et al.* 1970, 1977; Westgate 1975; Waters and Rutter 1984; Vreeken 1986, 1989, 1990, 1993; Sauchyn and Sauchyn 1991; Vance *et al.* 1992, 1993; Last and Sauchyn 1993) argues against either a source or transport limitation explanation for its absence. Glacier Peak and Mazama tephra deposition should, in theory, have been ubiquitous across the prairie surface. The widespread absence of tephra-bearing sediments implies that the limitation is, therefore, not due to an inadequate source of material, but is rather one reflecting a limitation on the deposition and / or preservation of the deposits. Similarly, historic (1920s, 1930s) aeolian activity would have been expected to produce a blanket of windblown soil deposits in the region;

evidence of such deposits is conspicuously absent. The thickest loess sections (Vreeken 1986, 1989, 1990, 1993) in southeastern Alberta *are* generally associated with tephra bed deposits, indicating that the sites where tephras are found are also suitable for loess sedimentation and preservation. In addition, tephra beds are also commonly associated with well developed paleosols, indicative of stable surfaces.

Loess deposition in this region thus appears to be limited not by supply of silt to fine sand sized material, but by the lack of conditions which were suitable for loess deposition and preservation. The single occurrence of loess in Dinosaur Provincial Park implies that the conditions required to deposit and preserve loess are not easily met in this region.

Although sediments are presently being mobilized by wind in Dinosaur Provincial Park, these sediments are not forming loess deposits. Goudie (1983) found that maximum dust storm activity occurs in climates with between 100 and 200 mm yr^{-1} rainfall. The Dinosaur Provincial Park area today receives approximately 250 mm yr^{-1} rainfall, and an additional 95 mm yr^{-1} as snow (record from nearby Brooks climate station, Environment Canada 1993), which does not have the same effect as rainfall in inhibiting dust storms. The modern climate near Dinosaur Provincial Park is clearly only slightly too moist for maximum loess mobilization. Figure 6.2 indicates that the mid-Holocene dated loess from Dinosaur Provincial Park (Figure 6.5) coincides with a relatively humid episode during a generally arid period, possibly consistent with 100-200 mm yr^{-1} of precipitation. That loess was not deposited in significant quantities either before or since that time implies that the climate of Dinosaur Provincial Park has not remained within the critical precipitation range long enough for widespread loess deposition.

6.5.3 PRAIRIE SURFACE

6.5.3.1 DESCRIPTION

Shetsen's (1987) map of the surficial geology of the region shows that the prairie surface is covered by Quaternary glacially-derived sediments, dominantly glaciolacustrine and till, which are characterized by highly variable grain size compositions (Kjearsgaard, 1988; Kjearsgaard *et al.* 1983). Because no comprehensive study of aeolian materials has been conducted in this region it was necessary to assess the extent of aeolian coverage on the prairie surface.

Ninety-three samples were collected for grain size analysis (Figure 6.1). The samples were collected at *ca.* two km spacing along selected roads. The organic soil layer was removed and each sample taken from just below the root layer. Grain size analysis was performed using the method described above.

Based on a comparison of the grain size distribution of the surface samples to the regional standard loess curve, only three of the deposits analysed here fall within the loess size parameters (Figure 6.6). While there is no apparent association between major geomorphic feature (e.g., moraine vs. glacial lake) and loess deposition, the samples identified as potential loess were recovered from moist, topographic lows where moisture conditions are expected to be higher than on the topographic highs.

6.5.3.2 INTERPRETATION

Only three per cent of the prairie surface surveyed is covered by deposits which may have loess characteristics. The distribution of samples makes it unlikely that the apparent absence of loess is due to sampling bias. While it might be argued that using roads automatically biases sampling in favour of certain geomorphic settings (roads tend to avoid wetlands and cliffs, for example) the general pattern of roads in this region follows the township and range grid with very little deviation, ensuring that a representative sampling was collected from all geomorphic settings although not necessarily in direct proportion to their landscape coverage.

The non-loess surface samples are not depleted in very fine sand and coarse silts. This suggests that the soils which form much of the prairie surface may be largely non-deflational, possibly due to the high clay content, which acts to bind the surficial sediments. The paucity of aeolian deposits also suggests that the prairie surface is largely a transportational surface (minimal erosion or deposition), perhaps due to its relatively flat and dry nature, rather than being a major depositional site.

The distribution of surface samples indicates that loess is more likely to occur selectively in topographic depressions (such as prairie sloughs) rather than uniformly across the prairie surface. Aeolian deposition is due to air-transported particles being caught in the laminar sublayer flow, deposition in the lee / windward slopes of topographic obstacles due to flow divergence and reduction of drag, funnelling effects, and capture by moister depression surfaces (Pye 1987). The spatial distribution and variation of thickness of deposits of loess can be very complex. Gossens (1988) demonstrates that loess sedimentation can be highly variable even within a very small catchment area. Based on >2000 field points, in a hilly and rather topographically complex 40 km^2 test area in the vicinity of Leuven, Belgium, Gossens (1988) found that loess sedimentation occurred differentially over very short distances, ranging in depths between <1 - >8 m. Scale model simulations of loess deposition indicate that topography (and its impact on aerodynamical characteristics of the wind) is the primary factor determining its spatial distribution (Gossens 1988). While material and climatic conditions suitable to loess deposition existed at least once during the postglacial in Dinosaur Provincial Park, the lack of suitable depositional / erosional sites on the prairie surface appears to have restricted aeolian mobilization / deposition.

6.6 DISCUSSION AND CONCLUSIONS

Ager (1993) has proposed that the stratigraphic record is more gaps than records of depositional events, which raises important implications in the study of aeolian deposits, between most dates (*e.g.* paleosols) are of periods dominated by a lack of aeolian activity rather than of mobility. Radiocarbon dates of a paleosol, or OSL or TL dates of the aeolian deposit indicate periods of non-movement of the sediment which is being dated. In attempting to use aeolian deposits as a proxy for wind-dominated paleoclimate it is most often a cessation of aeolian activity which is being dated, not aeolian mobilization.

The occurrence in this area of sand dune deposits dating only to the late Holocene is most parsimoniously explained as a result not of dune mobility being confined to the late Holocene, but rather of repeated mobilization expunging prior episodes of dune deposition from the stratigraphic record. Before the late Holocene dunes were likely completely remobilised. The loess record, confined as it is to the middle Holocene, requires a more elaborate explanation.

Loess size material should be deflated most easily in the driest climate, and to decrease in mobility as climate becomes wetter whereas the ability of a given site to trap loess will increase with moisture. The combination of these two functions can be expected to produce a curve of loess deposition which shows a clear maximum at some moisture value which is neither too dry for deposition and trapping, nor yet too moist for deflation and mobilization (*ca.* 100-200 mm mean annual precipitation) (Figure 6.7).

A further complication to this conceptual model would be the addition of a curve representing post-depositional erosion by water. This curve would not affect loess deposition, but would affect its preservation in sites where water erosion occurs or is likely. Such a curve would be sigmoid to show the slow increase in erosion by water at very low moisture levels, an erosional threshold at moderate levels, and an asymptotic increase at high moisture levels. The resulting combination of the three curves produces not only an optimum moisture level for loess deposition and preservation, but it implies that a substantial excess of moisture results in a net loss of loess because the rate of fluvial erosion of previously deposited loess exceeds deposition of new loess.

This conceptual model can be applied to Dinosaur Provincial Park to explain the occurrence of loess dating only to the mid-Holocene. Prior to that time, it may have been too dry for loess deposition - the sites which are presently depositional may have been areas of continued transport rather than final deposition, lacking the vegetation cover or moisture to effectively bind the loess. After the mid-Holocene, the area became moister, and may have become at times too moist for loess deflation and mobilization, and at times too wet (erosion) for loess preservation. This hypothesis is supported by the observation that the loess cover in Dinosaur Provincial Park is presently being stripped off by runoff erosion (Bryan *et al.* 1987; O'Hara and Campbell 1993). Alternatively, further southwest, at a higher elevation in the Lethbridge area, conditions favourable for loess deposition have occurred throughout much of the postglacial (Vreeken 1989). Such a transition suggests that some loess-related geomorphic threshold condition exists between the Dinosaur Provincial Park and Lethbridge region.

The pattern of aeolian deposits over a transect of three distinct geomorphic situations indicates that in the areas studied:

(1) Aeolian dunes developed on or adjacent to the fluvioglacial deltaic source area, and have not migrated substantially; such a pattern seems common in Alberta (David 1977; Shetsen 1987).

(2) There is no evidence of early Holocene aeolian mobilization / deposition in the Duchess dune field. If dunes formed prior to the late Holocene, subsequent re-remobilisation has obscured and erased evidence of their genesis. The preservation of evidence of dune mobility in the late Holocene may be a consequence of increased moisture causing soil formation and partial dune stabilization.

(3) Mobilization of aeolian sand during the late Holocene has been highly localized and episodic, likely responding to variable site-specific deflational controls - as it is today.

(4) Dune mobilization occurs during periods of regional aridity, but not all periods of aridity caused dune mobilization. While aridity is a necessary precondition for dune mobilization, a further destabilizing event - such as fire or bioturbation - may be required to trigger movement. Thus, while aridity will predispose dunes in the region to mobilize, the actual mobilization may be site-specific, depending on the occurrence and intensity of a local destabilizing event.

(5) Conditions suitable for loess stabilization occurred only once since deglaciation in this area.

(6) Loess stabilization occurs in large topographic depressions such as Dinosaur Provincial Park, in sites on the prairie surface where it is moister, and on cliff tops in Dinosaur Provincial Park where it protected from remobilisation and erosion due to airflow (boundary layer) separation.

(7) The prairie surface is dominantly a transportation zone of non-erosion and non-deposition due to well-bonded, clay-rich deposits which are not generally conducive to aeolian deflation except when severely disturbed.
(8) Aeolian erosion and deposition are spatially heterogenous, as functions of suitable sources of material, topography, and geomorphic setting.
(9) The study area has been subject to repeated cycles of dune mobilization / deposition in the late Holocene. It is likely that a relatively minor decrease in moisture could set the stage for remobilisation of dunes but dune migration will be very limited.

TABLE 1a AMS DATE FROM THE DUCHESS DUNE FIELD (date determined by Isotrace Laboratory, University of Toronto)

Sample	Material	δ ^{13}C (°/oo)	AMS age $\pm$ 1σ ^{14}C yr B.P	Calibrated using CALIB Rev 3.03 date range reported here is that which yields 100% probability when 1σ is used (Stuiver and Reimer 1993).
TO-5478	Bison cervical vertebra in paleosol #5 NWII 3E	-25 °/oo	840 $\pm$ 100	668-898

TABLE 1b OSL DATES FROM THE DUCHESS DUNE FIELD (date determined by D. Huntley, Simon Fraser University)

Sample I.D.) (0X$_{(0)}$)	^{40}K Whole Sample	^{40}K Surroundings %	^{238}U μg/g	^{232}Th* μg/g	H_20 Δ	Depth Below Surface (m)	Grain Size μm	Dose Rate Gy/ka	D_{eq} Gy %	Age (OSL yr B.P.)
IC2-01	1.26	1.31	0.90	2.8 $\pm$ 0.03	0.05	1.5	180-250	2.58$\pm$0.12	2.14$\pm$0.06	830$\pm$45
IC3-01	1.43	1.30	0.99	(3.0)	0.05	1.5	180-250	2.75$\pm$0.13	0.56$\pm$0.04	203$\pm$17
IC4-01	1.46	1.41	1.20	(3.6)	0.02	2	180-250	2.94$\pm$0.13	10.8$\pm$0.8	3700$\pm$300
IC5-01	1.46	1.49	1.22	(3.6)	0.06	3	90-125	2.63$\pm$0.13	6.1$\pm$0.2	2300$\pm$150
IC5-02	1.43	1.42	1.02	(3.0)	0.08	0.08	180-250	2.66$\pm$0.13	0.62$\pm$0.05	233$\pm$22
			All $\pm$ 0.04							

Five dunes samples (IC2-01 - IC5-01) were collected for analysis. Table 6.1B lists the OSL age determinations (Huntley pers. comm. 1996). Optical dating is a radiation-exposure-based luminescence dating method.The dating signal within a mineral grain is a function of the length of time that the sample has been subjected to the weak flux of environmental nuclear radiation and cosmic rays since its zeroing by daylight during deposition. The longer the exposure to radiation, the larger the signal. The age is represented by the equation:

$$\text{age} = \frac{\text{equivalent dose}}{\text{dose rate}}$$

The equivalent dose is estimated from laboratory measurements comprising the dating signal and corresponds to the dose of nuclear radiation responsible for producing the luminescence signal; the SI unit of dose is the gray (Gy). The dose rate is the rate at which the sample assimilates energy from the radiation in its local vicinity; this includes, alpha, beta, and gamma radiation from ^{40}K, ^{238}U and ^{232}Th, and their daughter products, together with a small cosmic-ray contribution (Aitken 1992).

CHAPTER 6 - FIGURE CAPTIONS

FIGURE 6.1

A. Study area.

B. Loess distribution in southeastern Alberta (after Thorp and Smith 1952; Vreeken 1986, 1989, 1990, 1993; Bryan *et al.* 1987; Shetsen 1987).

C. Dune distribution in southeastern Alberta (after David 1977; Shetsen 1987).

D. Surface sample locations. Highlighted areas = possible prairie loess deposit.

FIGURE 6.2

A. Summer and winter insolation values for 50°N latitude for the last 10,000 years (Berger 1978).

B. Relationship between cal yr B.P and ^{14}C yr B.P.(produced using file intcal93. 14c, CALIB Rev 3.0.3 [Stuiver and Reimer 1993]).

C. Composite paleohydrology curve.

D. Harris Lake (curve derived and calibrated from Sauchyn and Sauchyn 1991; Last and Sauchyn 1993).

E. Chappice Lake (calibrated after Vance *et al.* 1993).

F. Pine Lake grain size curve (Chapter 5).

G. Dated postglacial aeolian deposits in southeastern Alberta. AMS and OSL dates on dune deposits (1. ALPHA-2070 (Bryan *et al.* 1987); 2. TO-5478; 3. IC2-01; 4. IC3-01; 5. IC4-01, 6. IC5-01, IC5-02 [see Table 6.1]).

FIGURE 6.3

Duchess Dune Field: 17 stratigraphic logs, demonstrate a large number of paleosols (see Figure 6.1 for location of Duchess Dune Field).

FIGURE 6.4

4A Composite diagram showing topographic relationships between various surfaces and loess in Dinosaur Provincial Park (modified after Bryan et al. 1987). **4B** Picture of loess deposit in Dinosaur Provincial Park.

FIGURE 6.5

Cummulative curve of 30 samples of loess from Dinosaur Provincial Park.

FIGURE 6.6

Presumed loess samples from prairie surface based on shape of the curve and grain size characteristics.

FIGURE 6.7

Conceptual model used as an explanation for aeolian deposition.

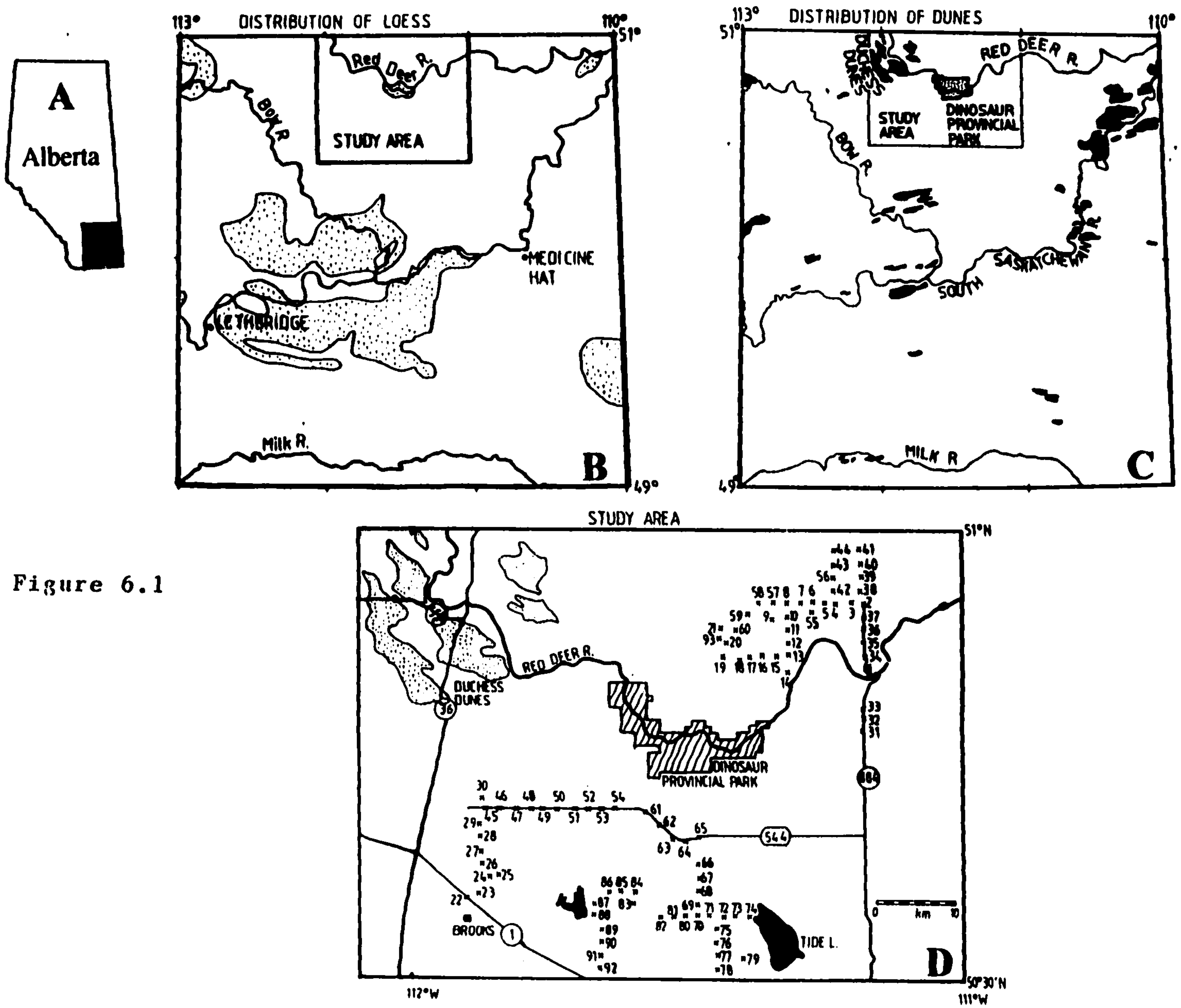
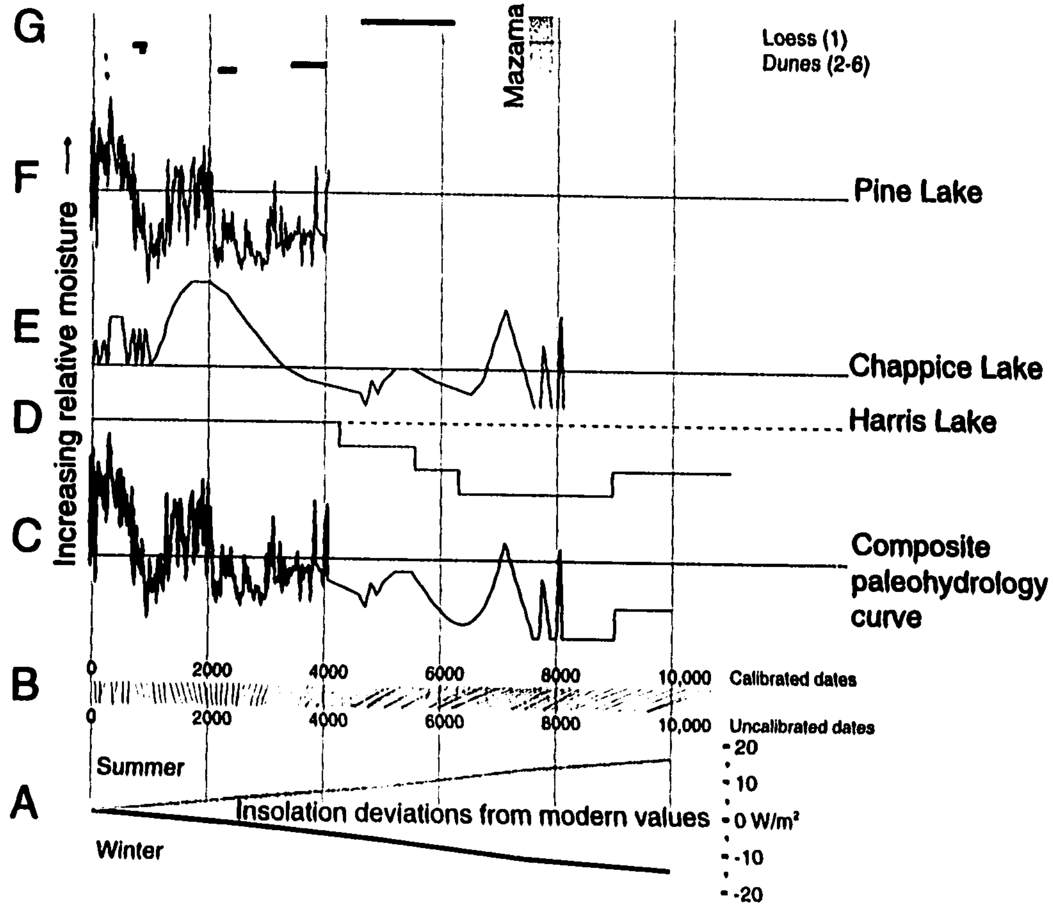

A
Alberta
DISTRIBUTION OF LOESS
113°
110°
51°
49°
Red Deer R.
STUDY AREA
BOW R.
MEDICINE HAT
LETHBRIDGE
Milk R.
B
DISTRIBUTION OF DUNES
RED DEER R.
DUCHESS DUNES
STUDY AREA
DINOSAUR PROVINCIAL PARK
BOW R.
SOUTH SASKATCHEWAN R.
MILK R
C
STUDY AREA
51°N
RED DEER R.
DUCHESS DUNES
DINOSAUR PROVINCIAL PARK
BROOKS
TIDE L.
km
0
10
D
112°W
111°W
50°30'N

Figure 6.1

FIGURE 6.2



G
F
E
D
C
B
A
Increasing relative moisture
Mazama
Loess (1)
Dunes (2-6)
Pine Lake
Chappice Lake
Harris Lake
Composite paleohydrology curve
0
2000
4000
6000
8000
10,000
Calibrated dates
Uncalibrated dates
Summer
Winter
Insolation deviations from modern values
20
10
0 W/m²
-10
-20

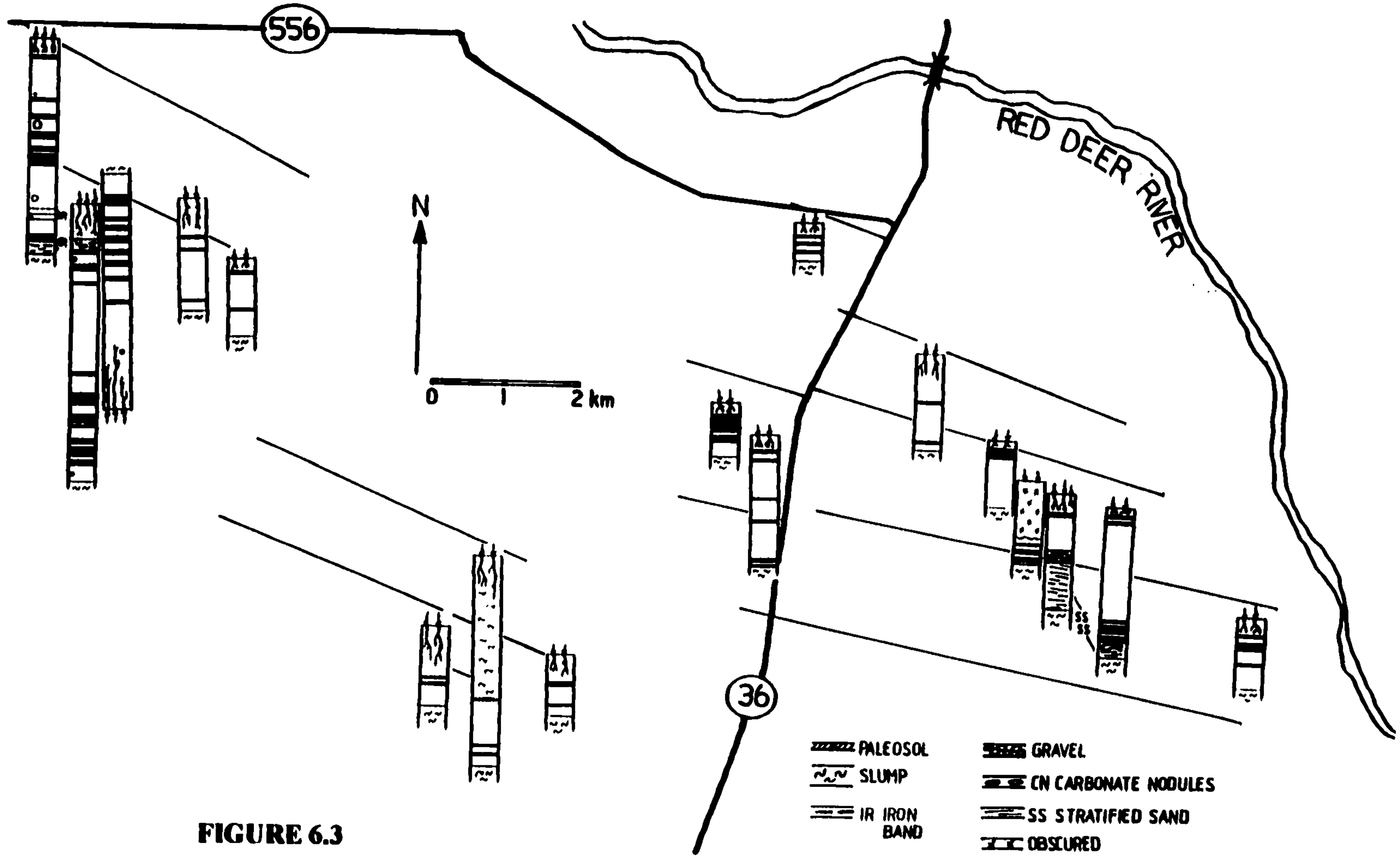
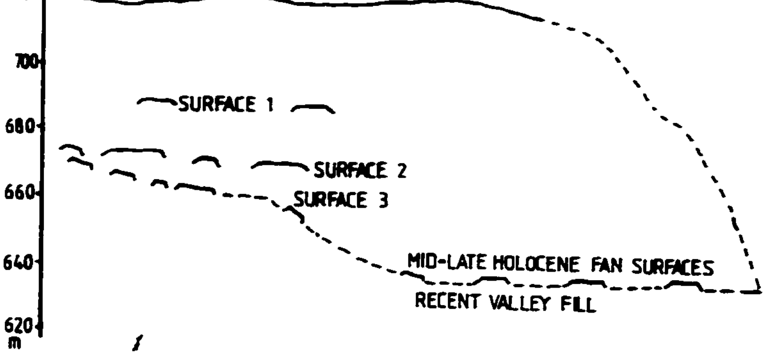

556
RED DEER RIVER
N
0
1
2 km
36
PALEOSOL
SLUMP
IR IRON BAND
GRAVEL
CN CARBONATE NODULES
SS STRATIFIED SAND
OBSCURED
SS
SS

FIGURE 6.3

Figure 6.4


PRAIRIE SURFACE
720
700
680
660
640
620
m
SURFACE 1
SURFACE 2
SURFACE 3
MID-LATE HOLOCENE FAN SURFACES
RECENT VALLEY FILL

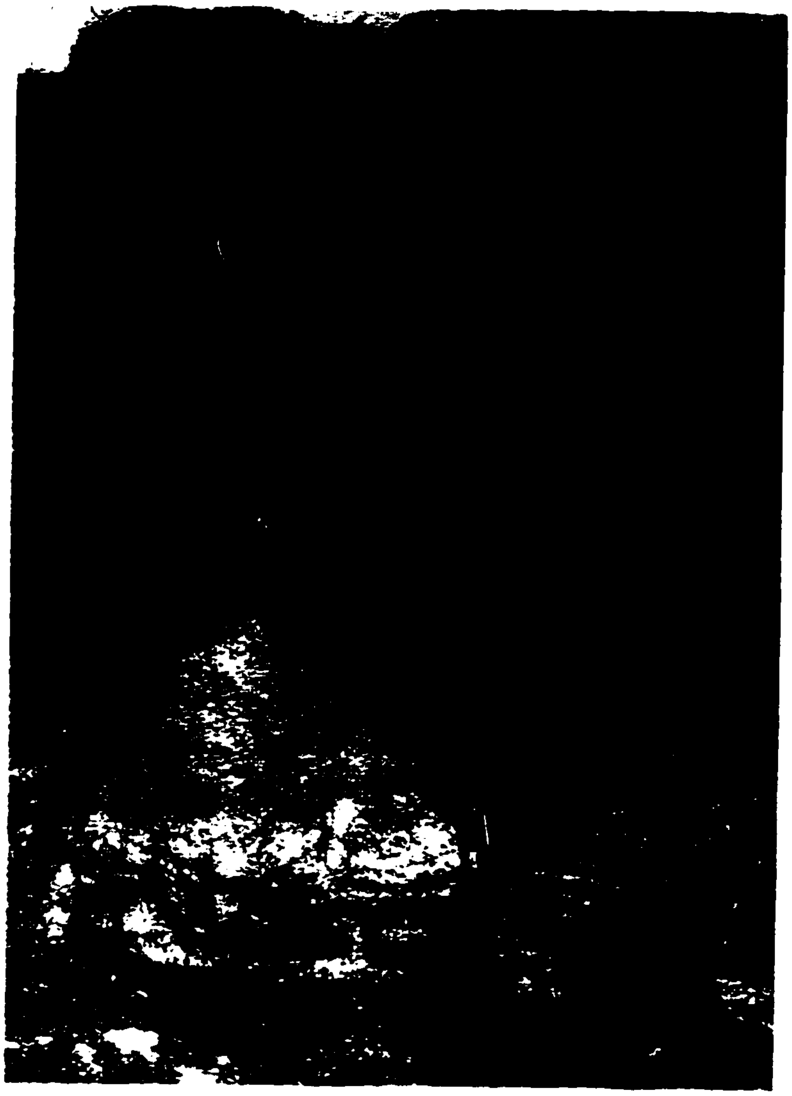


FIGURE 6.5

STANDARD LOESS - CUMULATIVE CURVE

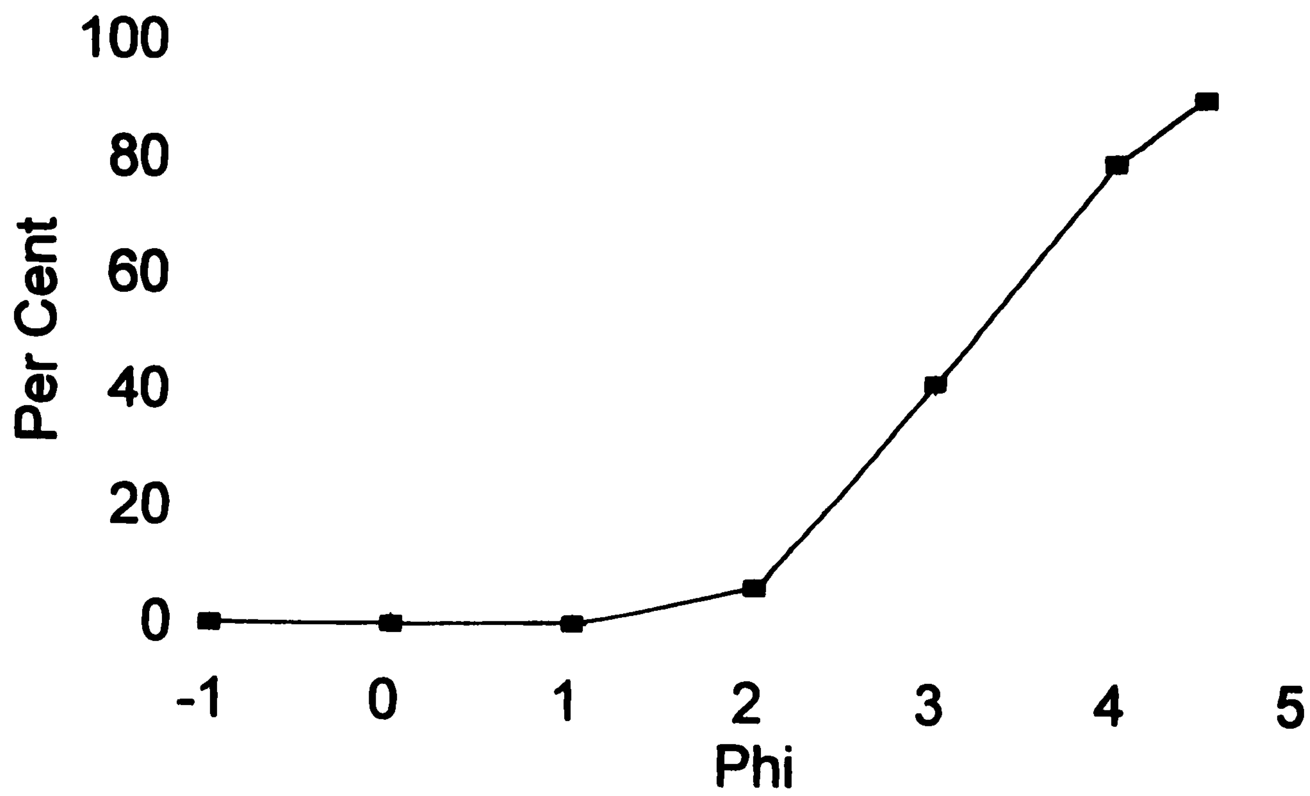


FIGURE 6.6

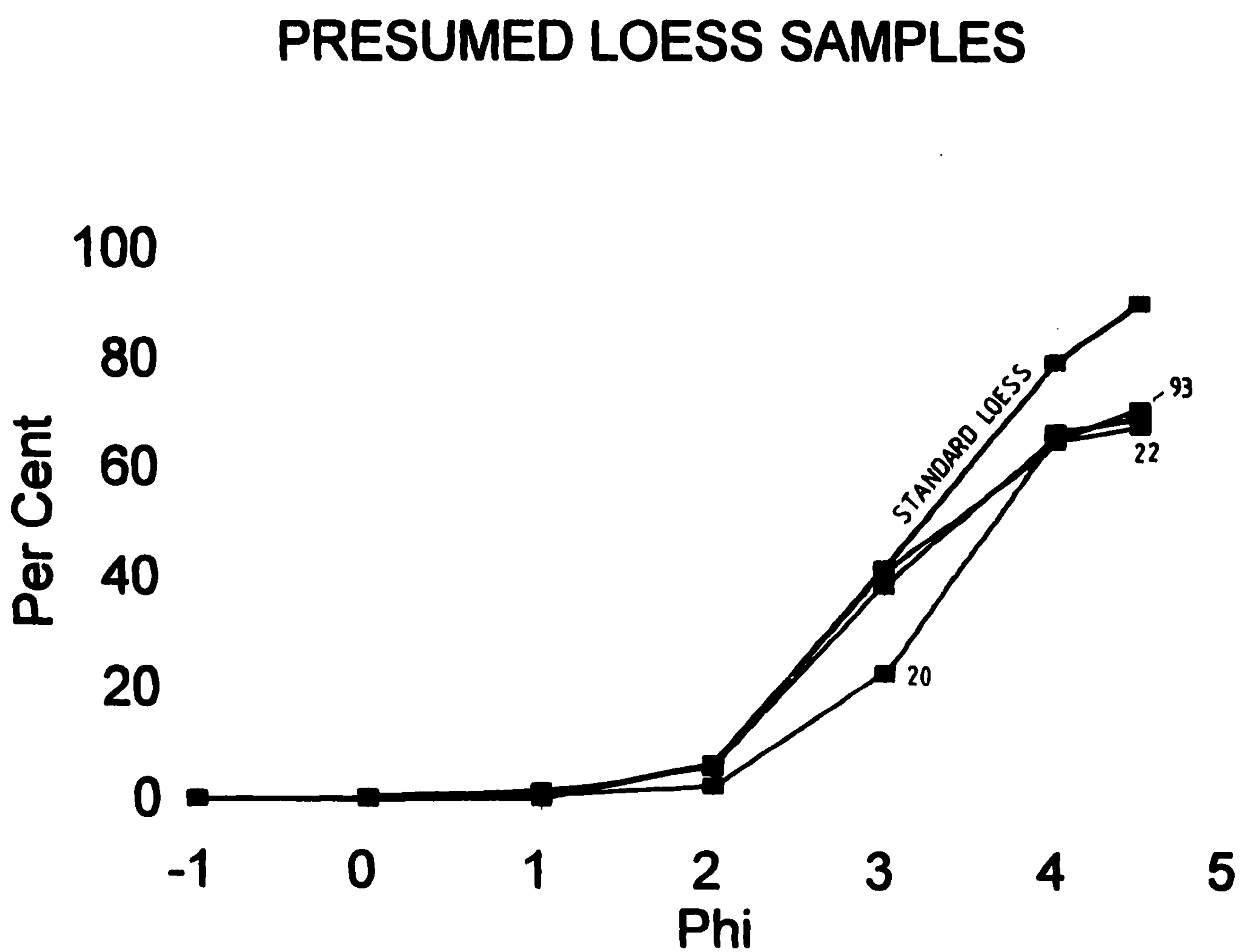
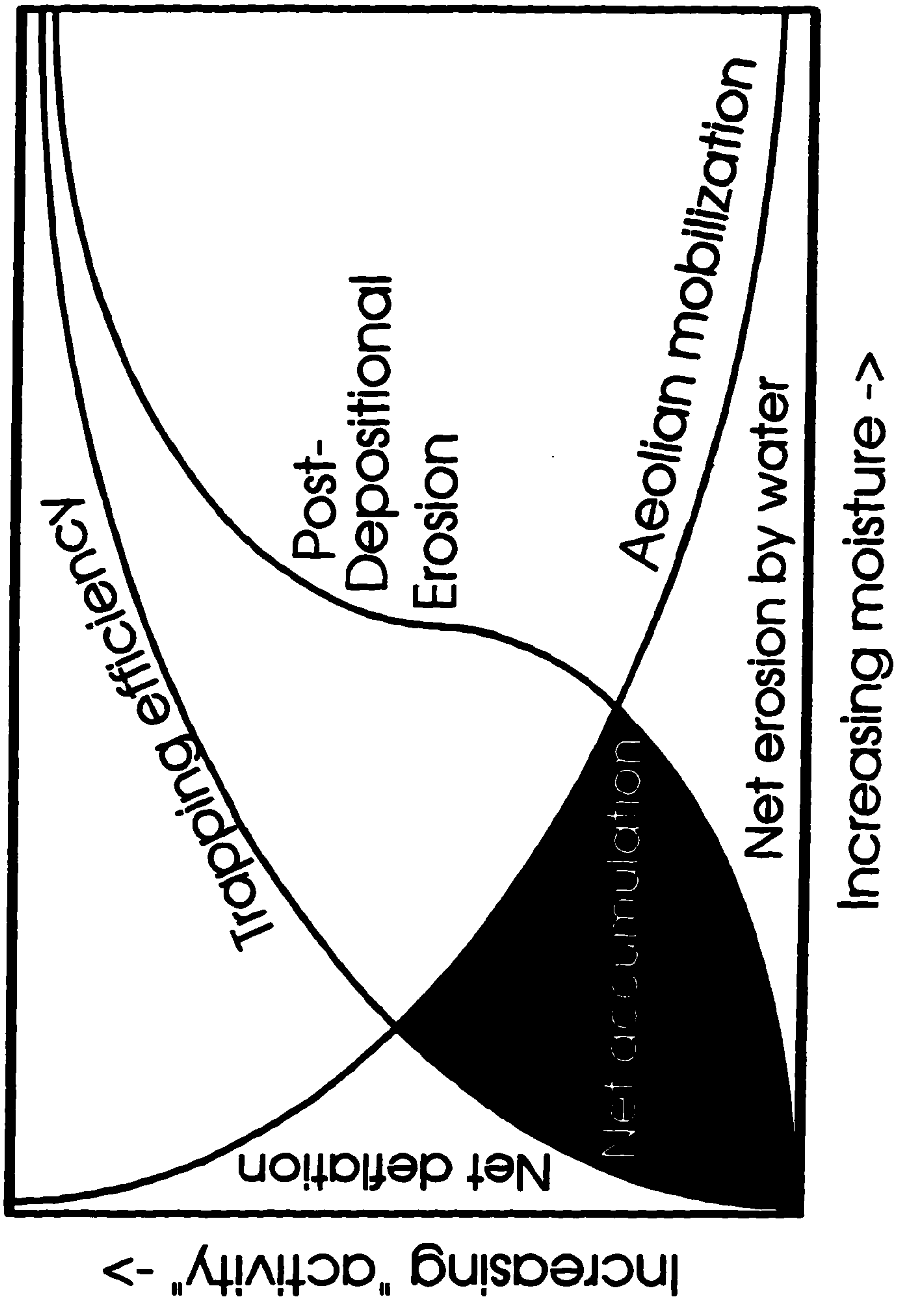

PRESUMED LOESS SAMPLES
100
80
60
40
20
0
Per Cent
-1
0
1
2
3
4
5
Phi
STANDARD LOESS
93
22
20

FIGURE 6.7


Trapping efficiency
Post-
Depositional
Erosion
Aeolian mobilization
Net erosion by water
Net accumulation
Net deflation
Increasing moisture ->
Increasing "activity" ->

6.7 CHAPTER 6 - REFERENCES

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CHAPTER 7

7.1 CONCLUSIONS

This thesis tested the hypothesis that the tempo and pattern of geomorphic changes in southeastern Alberta over the postglacial period reflects the nature and intensity of climatic forcing as expressed principally through hydrological response. In Chapter 1 calibration of postglacial radiocarbon dates shows an association between geomorphic events and seasonality of insolation in this region. The late Pleistocene and Holocene was divided into four major episodes: glaciation, 22,870-21,880 to 20,000 cal yr B.P.; deglaciation, *ca.* 20,000-12,000 cal yr B.P.; landscape stability, *ca.* 12,000-10,000 cal yr B.P.; and landscape instability, *ca.* 10,000-0 cal yr B.P. Insolation is presumed to drive geomorphic activity in this semiarid region through changes in hydrology. Chapters 2-6 provide additional information which refines the above chronology and allowing for the development of a relative paleohydrology curve.

Figure 7.1 shows the sites used to develop Figure 7.2. Figure 7.2 shows a paleohydrology curve for last 20,000 cal yr B.P. developed from data from the literature and this thesis. The date for glacial advance is based on the A.M.S. date on preglacial wood of 22,870-21,880 cal yr B.P. from Tide Lake (TO-5946). A date of 20,210-19,430 cal yr B.P. (AECV-681C), from mammoth bone, constrains the date of deglaciation (Evans and Campbell 1992 1995). Ice disintegration and recession drained the proglacial lakes, forming extensive meltwater channels (Kehew and Lord 1986; Bryan *et al.* 1987; Teller 1987; Paterson 1996). The period of proglacial lakes associated with deglaciation is indicated by a wedge in Figure 2 which spans the period of dates on proglacial lake sediments in the study area (Chapter 1).

Following regional deglaciation, it is most likely that ground and surface water was recharged, as indicated by pond deposits over glacial lake deposits, *ca.* 18,960-17,280 cal yr B.P. (Christiansen 1979), and continuous development of peat in a hummock *ca.* 17,330-17,050 to 12,220-11,230 cal yr B.P. (Klassen 1993). During this period, seasonality of insolation was increasing, suggesting increasing aridity. However, the occurrence of both peat and paleosols indicate a period with more available moisture than at present, perhaps due to groundwater recharge by glacial meltwater.

In the western Cordillera, the source area for the major rivers in the area, increased moisture accompanied the Crowfoot Advance (coeval with the Younger Dryas) *ca.* 13,482-11,223 cal yr B.P. (calibrated after Reasoner *et al.* 1994). This period was significantly moister than the times either just before or after it. Since Crowfoot moraines are at approximately the same elevation as the Little Ice Age moraines (Reasoner *et al.* 1994), this period is tentatively assigned a similar maximum climatic moisture. This period also saw the incision of major river valleys in Alberta, as evidenced by dates of material in the uppermost terraces. While it may be that the rivers were responding to changes in baselevel, the simultaneity of incision events argues against this. Similarly, isostatic rebound would be expected to cause metachronic rather than synchronic incisions across such a broad region. The most parsimonious explanation for such simultaneous incision events is an increase in the regional water supply due to the Younger Dryas climatic fluctuation.

Peak aridity followed the Crowfoot Advance. This arid period was associated with maximum seasonality of insolation, and characterized by the most strongly developed regional paleosols, decreased fluvial erosional and depositional activity, a period of non-deposition in formerly active floodplains, alluvial fan accumulation, and the drying of lake basins (Chapters 1, 3 and 4). Interestingly, no lake deposits in the region have yet been dated to this period, which is likely indicative of an arid interval.

From 10,430-5600 cal yr B.P., regional lakes began to fill (Sauchyn 1990; Vance and Last 1994; I.A. Campbell and Evans 1990 1992), reflecting an increase in regional moisture availability. Coeval with lake filling, upper level terraces formed in tributary stream valleys.

The rest of the curve is derived from data from three lake sites (see Chapter 6). The palynological evidence from Harris Lake, in the Cypress Hills of southwestern Saskatchewan (Sauchyn and Sauchyn 1991; Last and Sauchyn 1993) provides an 11,000 cal yr record. Paleosalinity interpretation of the Chappice Lake sedimentary and palynological record (Vance *et al.* 1993) provides greater detail in the last 8000+ cal yr B.P. Because lake salinity can be expected to show a long-term increase even in an unchanging arid or semiarid climate, the Chappice Lake curve is fit to the trend of the Harris Lake curve. The most recent 4000+ cal yr B.P. of the record are taken from the grain size record in the sediments of Pine Lake, which reflects changes in stream discharge (Chapter 5). The variations in the last 4000+ cal yr B.P. of the paleohydrology curve correspond to the dates of stream incision as recorded in stream terraces (Chapter 5).

This composite synthetic curve, which is of necessity a subjective blend of data given the disparate nature of the evidence is consistent with the geomorphic (Campbell and Campbell 1997) and palynological (Schweger and Hickman 1989; Sauchyn and Sauchyn 1991; Vance *et al.* 1993) records of the region. It shows maximum aridity immediately following the Crowfoot Advance (Younger Dryas), followed by a moistening trend throughout the Holocene. While most authors have identified the period from 9,000-8,000 cal yr B.P. (e.g. Schweger and Hickman 1989) as the driest postglacial interval, the absence of lake deposits from the earlier, aridity maximum has generally prevented its recognition. Multiple wet / dry cycles are superimposed on the Holocene moistening trend; the most recent wet period corresponds with the Little Ice Age *ca.* 300 years ago, and the preceding dry period with the Medieval Warm Period *ca.* 1000 years ago (Vance and Wolfe 1996). The reconstruction is also in broad agreement with the theoretical insolation curve derived from orbital cycles (Berger 1978; Schweger and Hickman 1989; Chapter 1).

The paleohydrology curve has variable temporal resolution, and in many parts is somewhat *ad hoc*, but nevertheless shows the major trends and episodes of departure from them. Several of these events are found globally, for example the Little Ice Age (Grove 1988) and the Younger Dryas (Peteet 1995). This suggests that the forcing mechanisms producing some of the climatic events in this record operated at a global scale. It further suggests that many Holocene climatic fluctuations thought to occur in only one region may in fact have occurred globally, and will likely be revealed as global events as more data from different methods in different regions becomes available.

The paleohydrology curve is an improvement to the model developed in Chapter 1 as it incorporates the new data contained in this thesis, and indicates a much more complex pattern of postglacial geomorphic response to climatic change in southeastern Alberta than originally indicated in Chapter 1. While as suggested in Chapter 1, insolation may drive geomorphic activity at a gross scale in this semiarid region, relatively small scale variations in humidity, such as the Younger Dryas (Chapter 4) and the Little Ice Age (Chapter 6) are overprinted on the general insolation based trend. The hypothesis tested in this thesis, that the tempo and pattern of geomorphic changes in southern Alberta over the postglacial period reflects the nature and intensity of climatic forcing as expressed principally through hydrological response, appears to be most probable in light of the evidence presented here. However, great caution must be used in interpreting geomorphic records as demonstrated in the Klassen site whose evolution is interpreted as a function of changes in internal variables. In addition to the climatic signal inherent in some of the records analysed in this thesis, a number of additional points are raised.

First is the issue of gaps. Gaps in sedimentary records are indicative of either periods of inactivity or erosion. From any given site record, it may be impossible to discern which process resulted in the gap; however, when regarded in a regional context, a gap in a record may potentially be informative. For example, the absence of Glacier Peak or Mazama tephra beds in Tide Lake indicate that the site was non-depositional during the times that these tephras were deposited, or was subsequently erosional. That Tide Lake was in the plume of these tephras is demonstrated by their presence in the nearby Klassen site. Given the paleohydrological context and the stratigraphy of the site, it does not seem unreasonable to propose that it was dry during the early Holocene, and therefore a poor site for the preservation of tephras.

Gaps in the sedimentary record are not always confined between two periods of deposition. For example, in the regional chronology, the basal dates of lakes in the area are used to define a period of increasing moisture in the paleohydrology curve. Thus while the occurrence of a gap may be tentatively taken to indicate certain types of conditions, the cessation of a gap is much clearer in its interpretation.

Gaps are not only temporal, but also spatial. In Chapter 6, the coverage of the prairie surface is found to be highly discontinuous. So just as temporal gaps in the sedimentary record may carry useful information, so may spatial gaps - the occurrence of aeolian materials in greater abundance in Dinosaur Provincial Park than on the surrounding prairie surface yields valuable insight into the conditions required for loess deposition and preservation.

A second theme is the importance of internal dynamics in some geomorphic systems. At the Klassen site in particular (Chapter 4), there are great variations in sedimentation, which are attributed to internal processes at the site, overwhelming any climatic signal that might be present. Nevertheless, the basal date - the date of initiation of the system, complete with its internal dynamics - may be informative. Thus while great caution must be exercised at all times, it may be that interpretations of the data relating to system initiations can be more readily related to external forcing factors than can data relating to system transitions.

A third theme, related to the second, is the importance of thresholds in geomorphology. Since geomorphic activity is a function of disequilibria, thresholds gain enormous importance. Most systems embody a variety of processes and subsystems, and thus a variety of thresholds. When external conditions cross one of these thresholds, a system response is triggered. Since each system response is a function of the particular elements comprising the system, each site has the capacity to respond differently, or not at all, to a given external stimulus.

A fourth, and perhaps most important point, is the importance of patterns. Because of individualistic responses of sites and systems, which are in turn due largely to their internal dynamics, a single datum - such as the date of filling of a given lake - can be invested with only very limited meaning. If, however, a spatial and temporal pattern of lake fillings can be found, then the pattern may confirm changes which are not readily extracted from an individual site. This is in essence a population issue - while an individual site may behave very differently from the majority of sites, the majority will still follow a trend, according to the law of averages. A regional synthesis provides the optimum approach for modelling regional trends.

7.2 CHAPTER 7 - FIGURE CAPTION

FIGURE 7.1

Study area showing locations of dated sites shown in Figure 7.2. The area of Dinosaur Provincial Park is indicated by a circle.

Glacial Advance: 1. TO-5947 (Chapter 3); **Proglacial Lake**: 2. AECV-681C (Evans and Campell 1992); 3. GSC-4675 (Vreeken 1989); 4. TO-694 (Vreeken 1989); **Collapsed Pond Deposits Over**: Glacial Lake Deposits: 5. S-300B (Christiansen 1979); 6. S-300A (Christiansen 1979); **Crowfoot Advance**: 7. CAM 3063, CAM 3177, CAM 3065 (Reasoner *et al.* 1994); **Primary River (Meltwater Channel) Terraces**: 8. S-2385 (Rains and Welch 1988); 9 .GSC-805 (Lowden and Blake 1968); 10. S-1923, 11. S-1706, 12. S-1799 (Rains and Welch 1988); 13. Glacier Peak Tephra (Chapter 4); **Secondary River Terraces**: 14. GSC-236 (In: Waters & Rutter 1984 - Dyke *et al.* 1965); 15. S-1798, 16. S-1926, 17. S-1787 (Rains and Welch 1988); 18. TO-1829 (Rains *et al.* 1994); 19. S-1789 (Rains and Welch 1988); 20. GSC-1735 (Lowden and Blake 1975); 21. S-1788, 22. S-2387, 23. S-1800, 24. S-2392, 25. S-1797, 26. S-1795, 27. S-2389, 28. S-1794, 29. S-1796, 30. S-2388 (Rains and Welch 1988); 31. AECV-1276C (I.A. Campbell and Evans 1993); 32. AECV-904C (I.A. Campbell and Evans 1993); 33. AECV-684C (Traynor and Campbell 1989); 34. AECV-911C (I.A. Campbell and Evans 1993); 35. AECV-9440 (Rains *et al.* 1994); 36. AECV-906C (I.A. Campbell and Evans 1993); 37. S-2390 (Rains and Welch 1988); 38. AECV-905C (I.A. Campbell and Evans 1993); 39. S-1783 (Rains and Welch 1988); 40. AECV-913C (I.A. Campbell and Evans 1993); 41. TO-4577 (Barling 1995); 42. S-1785 (Rains and Welch 1988); 43. AECV-682C (Traynor and Campbell 1989); 44. TO-4579 (Barling 1995); 45. AECV-604C (I.A. Campbell and Evans 1990); 46. TO-4576 (Barling 1995); 47. AECV-914C, 48. AECV-80 (I.A. Campbell and Evans 1993); 49. AECV-683C (Traynor and Campbell 1989); 50. AECV-912C (I.A. Campbell and Evans 1993); 51. S-1793, 52. S-1786, 53. S-1790, 54. S-1784, 55. S-1791 (Rains and Welch 1988); 56 AECV-680C (C. Campbell and Campbell 1997); 57. AECV-903C (I.A. Campbell and Evans 1993); 58. S-1792 (Rains and Welch 1988); 59. AECV-1278C, 60. AECV-1277C (I.A. Campbell and Evans 1993); **Basal Lake Dates / Lacustrine-Ponding Events**: 61. S-2908 (Sauchyn 1990); 62. Clearwater Lake (Vance and Last 1994); 63. TO-959 (Campbell and Evans 1992) (in a series of lacustrine-ponding events); 64. Chappice Lake (Vance and Last 1994); 65. AECV-605.C (I.A. Campbell and Evans 1990) (in a series of lacustrine-ponding events); 66. Elkwater Lake (Vance and Last 1994).

FIGURE 7.2

A. Summer and winter insolation values for 50°N latitude for the last 20,000 years (Berger 1978).

B. Relationship between cal yr B.P and ^{14}C yr B.P.(produced using file intcal93. ^{14}C, CALIB Rev 3.0.3 [Stiver and Reamer 1993]).

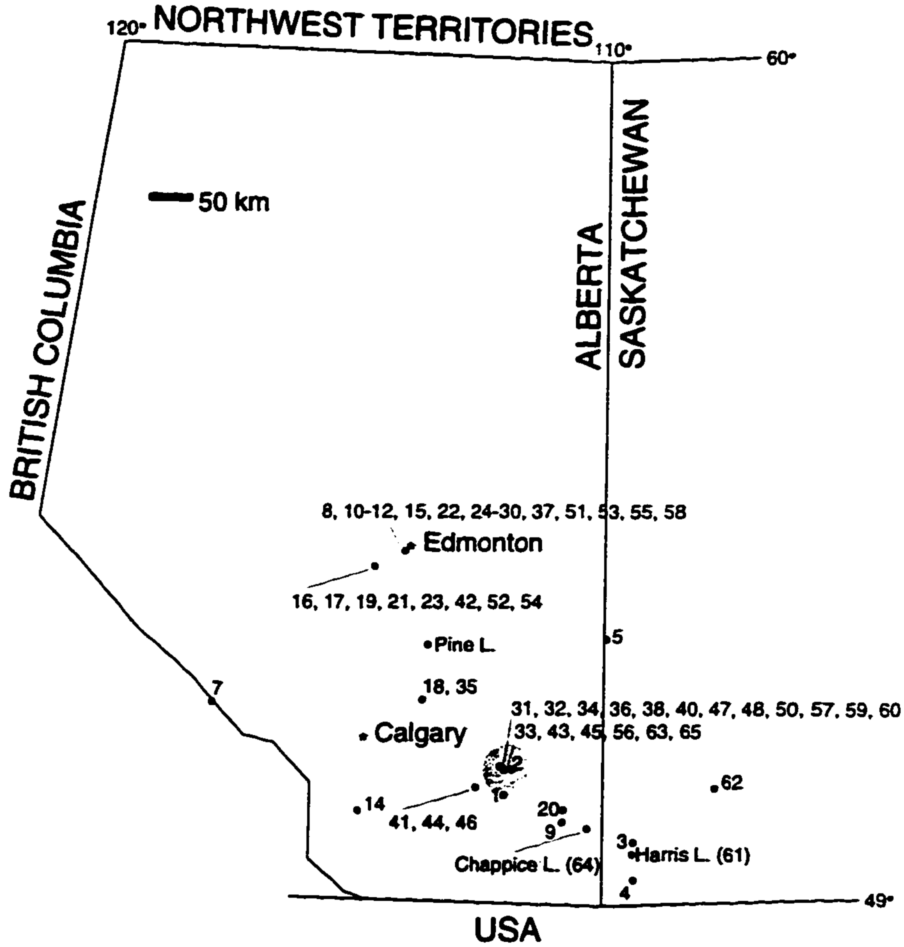
C. Composite paleohydrology curve.

D. Harris Lake (curve interpreted and calibrated from Sauchyn and Sauchyn 1991; Last and Sauchyn 1993).

E. Chappice Lake (calibrated after Vance *et al.* 1992 1993).

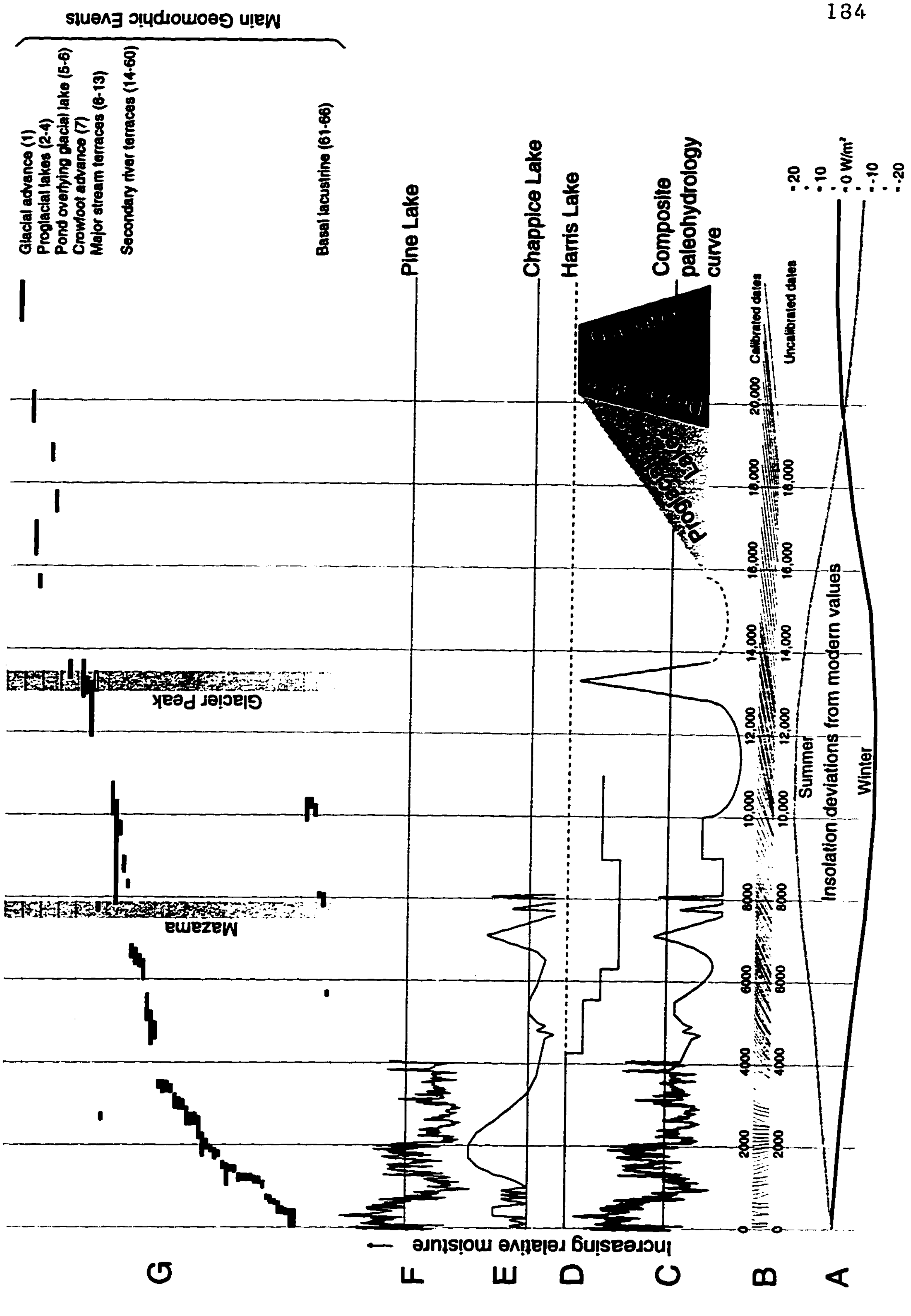
F. Pine Lake grain size curve (Chapter 6).
G. Dated postglacial hydrologic deposits in southeastern Alberta and adjacent areas (all dates in calendar years). Bars represent the 100% calibrated date range at 1σ using CALIB Rev 3.0.3 (Stuiver and Reimer 1993). Gray bars indicate uncorrected for $\delta^{13}C$ fractionation are indicated by a gray line. The white bar refers to a basal tephra date. The numbers in brackets (right hand column) refer to site numbers on Figure 7.1.

FIGURE 7.1



120°
NORTHWEST TERRITORIES
110°
60°
50 km
BRITISH COLUMBIA
ALBERTA
SASKATCHEWAN
8, 10-12, 15, 22, 24-30, 37, 51, 53, 55, 58
Edmonton
16, 17, 19, 21, 23, 42, 52, 54
Pine L.
5
7
18, 35
31, 32, 34, 36, 38, 40, 47, 48, 50, 57, 59, 60
33, 43, 45, 56, 63, 65
Calgary
2
1
62
14
41, 44, 46
20
9
3
Harris L. (61)
Chappice L. (64)
4
49°
USA

FIGURE 7.2


Main Geomorphic Events
Glacial advance (1)
Proglacial lakes (2-4)
Pond overlying glacial lake (5-6)
Crowfoot advance (7)
Major stream terraces (8-13)
Secondary river terraces (14-60)
Basal lacustrine (61-66)
Glacier Peak
Mazama
Pine Lake
Chappice Lake
Harris Lake
Composite paleohydrology curve
Proglacial Lakes
Increasing relative moisture
Calibrated dates
Uncalibrated dates
Summer
Winter
Insolation deviations from modern values
20
10
0 W/m²
-10
-20
0
2000
4000
6000
8000
10,000
12,000
14,000
16,000
18,000
20,000
G
F
E
D
C
B
A

7.3 CHAPTER 7 - REFERENCES

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