LANDSCAPE EVOLUTION
AT
WANUSKewan HERITAGE PARK,
SASKATOON, SASKATCHEWAN

ARIGAL HERN DURE
1597
Landscape Evolution at Wanuskewin Heritage Park, 
Saskatoon Saskatchewan.

A Thesis Submitted to the College of 
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in Partial Fulfilment of the Requirements 
for the Degree of Master of Science 
in the Department of Geography 
University of Saskatchewan 
Saskatoon

Abigail Keren Burt 
Summer, 1997

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Abstract

Regional deglaciation of the study site (52° 13’ N, 106° 35’ W) and drainage of glacial Lake Saskatchewan occurred by ca. 10.5 ka BP. Channelisation of flow into the South Saskatchewan River and a gradual drop in base level caused small tributary stream valleys, such as the Opimihaw Creek valley, to be incised through surficial sediments and underlying till formations. Large stream discharges associated with a cool, moist post-glacial climate contributed to the erosion of the landscape.

The following physiographic elements were observed in the study area: the till plain; alluvial terraces; mass movement landforms; the modern stream channel. The alluvial terraces located in the valley bottom are the focus of this study.

Prior to ca. 4.6 ka BP, a combination of a rise in base level and gradual climate change led to the crossing of a geomorphic threshold, shifting the stream from an incising to an aggrading system. Slope wash resulting from a reduction in vegetation cover, combined with variable precipitation and stream discharge, led to initial rapid channel aggradation. Sedimentation rates gradually declined as the environment became increasingly moist leading to a more constant stream discharge, a denser vegetation cover and reduced slope wash.

Excavations at several valley bottom alluvial terraces reveal five sedimentary facies: facies 1, vertical accretion sediments; facies 2, proximal channel sediments; facies 3, alluvium and/or colluvium; facies 4, fluvial sands; facies 5, channel gravels. The
generally fining upwards sequence from coarse-grain fluvial channel deposits (facies 3 - 5) to finer-grained proximal channel and vertical accretion sediments (facies 1 and 2) records the migration of the stream channel across its floodplain throughout the aggradation phase. Repeated pedogenesis and human occupation indicate periodic subaerial exposure of the floodplain. Downstream sites are generally dominated by finer-grained facies and are characterised by more rapid sedimentation rates than upstream sites.

Within the last ca 0.1 to 0.2 ka BP, a geomorphic threshold was crossed and the stream incised its floodplain. The proximity of the South Saskatchewan river to alluvial terraces at the mouth of the creek indicates the creek was likely responding to a drop in base level. Currently, there is very little flow in the creek, due in part to several beaver dams along its length.
Acknowledgements

I would like to thank my supervisor, Dr. A. E. Aitken, his support and guidance throughout the process of researching and writing this thesis. I would also like to thank the members of my advisory committee, Dr. E. Walker and Dr. J. McConnell, for their guidance and Dr. D. Sauchyn for his comments. Enthusiastic and knowledgeable field assistance was provided by G. Slemko.

Very special thanks are due to the management and staff of Wanuskewin Heritage Park and to the WIHI Board of Elders for their co-operation during my field season.

This project could not have been completed without the support I received from Grant, my parents, and my brother. Thank you for your patience and understanding.

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Chapter 1

Introduction

Over the years, two different, yet related, approaches to the interpretation of palaeolandsapes have arisen. The first approach is that of the sedimentologist and stratigrapher who examines the sediment record for evidence of erosion and deposition related to changes in energy within a system. This approach has the advantage of the ability to observe changes in the landscape over long periods of time (Miall, 1990; Schumm, 1981). The second approach is that of the geomorphologist who examines modern landforms, and processes of erosion and deposition acting on them, to gain a better understanding of the role of changing environmental conditions on landscape evolution. Geomorphologists have the advantage of being able to observe and measure processes currently operating under a wide range of climatic and geological conditions.

Schumm (1981) explains the advantages of adopting an integrated approach in studying fluvial systems;

the observations of the sedimentologist should aid the geomorphologist in the interpretation of landscape evolution because the record lies in the sedimentary deposits...a fuller understanding of the erosional and transportation processes acting on the landscape will provide a sound basis for the interpretation of fluvial and fluvially-influenced sedimentary deposits (Schumm, 1981, p. 19).
Numerous studies integrating the Holocene geomorphology and sedimentology of a variety of sites have been conducted in recent decades. Much of this work has been aimed at understanding processes operating in the arid to semi-arid American Southwest (Baker et al., 1977; Boison and Patton, 1985; Bull, 1991; Leopold et al., 1966; Patton and Boison, 1986; Schumm and Parker, 1973), humid regions in Europe and eastern North America (Karrow, 1986; Rose et al., 1980; Scully and Arnold, 1981) or mountainous regions (Bull, 1991). There has also been research directed at understanding change over shorter time periods of the last couple of centuries (Patton and Schumm, 1981; Schumm and Lichty, 1963; Womack and Schumm, 1977).

Little work has been conducted on the history of fluvial systems in the Canadian prairies. Research has been done on the Assiniboine river system in southern Manitoba (Nielsen et al., 1993) and part of the North Saskatchewan River basin in Alberta (Rains and Welch, 1988). Work conducted in the non-glacial Cypress Hills has examined the relationship between the paleoclimate record and landscape stability (Last and Sauchyn, 1993; Sauchyn, 1990).

Since the 1960's, there has been an increasing data base of Holocene paleoenvironmental information acquired from across Canada and the United States. A better understanding of the nature of environmental change has enabled researchers to speculate on possible driving forces behind the observed changes within the landscapes under consideration. Recently, numerous studies of the paleoclimate record have been conducted in conjunction with the Palliser Triangle project in southern Alberta,
Saskatchewan and Manitoba (Lemmen et al., 1993). This has helped to focus attention on Holocene environmental change across the southern Canadian prairies.

This study is focused on the interpretation of post-glacial landscape evolution in a small stream valley located in Wanuskewin Heritage Park in south-central Saskatchewan. This site was selected for study for two reasons. Rains and Welch (1988) argue that large river systems probably reflect “very complex associations of past controls, processes and responses” while smaller tributary basins would likely respond to short-term, low magnitude environmental controls. Given the time period and resources available to most researchers, these smaller basins could be more feasibly subjected to the necessary intense investigation.

The second reason is that early recognition of the cultural significance of the site means that it has been protected from major modification or disturbance during modern times. Wanuskewin, a Cree word meaning ‘seeking peace of mind’ was home to Plains Indians for over 5000 years. Occupation of the valley ended in the 1870's when treaties were signed and First Nations people were moved to reserves. In 1982 the land was purchased by the Meewasin Valley Authority for future development as a park. By 1984, the site had became the first pre-contact archaeological resource to be brought under the protection of the Saskatchewan Heritage Act. The importance of the site was recognised nationally in 1987 when Queen Elizabeth unveiled a plaque declaring the valley a National Historic site. The park was opened to the public in 1992.

An archaeological inventory conducted by Dr. E. Walker in 1982 - 83 revealed 19 individual pre-contact sites within the 116 hectare park. These include summer and
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An archaeological inventory conducted by Dr. E. Walker in 1982 - 83 revealed 19 individual pre-contact sites within the 116 hectare park. These include summer and
winter campsites, buffalo jumps, processing sites and a medicine wheel. The investigation of these sites is an important step in formulating a better understanding of Northern Plains cultural developments as well as paleoenvironmental change in this region.

1.1 Objectives of Study

There are three primary objectives of this study. The first objective is to produce a detailed inventory of landforms within the study site, locating each landform on an air photo. The second objective is to examine the stratigraphy of the landforms and relate this information to depositional and erosional processes operating within the study site over time. The third objective is to establish a chronologically controlled model of landscape evolution with emphasis on the response of landforms to Holocene environmental change. This will be the first model of landscape change produced in the region forming a starting point for future studies. In addition, it will act as a framework of physical change for use by park resource managers and archaeologists.
Chapter 2

Study Site

2.1 Location

The study site incorporates the lower reaches of the Opimihaw Creek valley and surrounding uplands within the boundaries of Wanuskewin Heritage Park. The park is located north of the Saskatoon city limits at 52°13’N, 106°35’W (Fig. 2.1). The stream valley is oriented on an NE-SW axis, draining south into the South Saskatchewan River. The head of the stream is found within the Hudson Bay slough, a paleochannel of the South Saskatchewan River.

The study area is located within the Saskatchewan Rivers Plain region of the Saskatchewan Plain, Southern Interior Plains of Canada (Fig. 2.2). The Saskatchewan Plains are bounded by the Canadian Shield to the north, the Manitoba Escarpment to the east, the Missouri Coteau to the west and the Canada/USA border to the south (Klassen, 1989). The Southern Interior Plains rise towards the west and south-west. The Manitoba Escarpment rises 300 m from the Manitoba Plain up to the Saskatchewan Plain. The Missouri Coteau rises a further 100 m to the Alberta Plain. Elevations within the Saskatchewan Plain range from 400 to 800 m above sea level (a.s.l.) with several upland areas including Moose Mountain, the Touchwood Hills and the Allen Hills (Klassen,
Figure 2.1: Location of the study site, north of the Saskatoon city limits.
Elevations within the study site range from 460 m a.s.l. at the bottom of the valley to 495 m a.s.l. on the adjacent uplands.

2.2 Geology

The bedrock geology of Saskatchewan is dominated by the Precambrian Shield in the north and Cretaceous shales, siltstones and sandstones in the central and southern regions of the province (Fig. 2.3) (Klassen, 1989). These are separated by a band of Palaeozoic limestones, dolomites, sandstones and shales. Tertiary siltstones and sandstones form the Wood Mountain and Cypress Hills uplands.
Interbedded gravel, sand, silt and clay deposits of glacial, fluvial, lacustrine and colluvial origin lay between bedrock and Pleistocene glacial tills (Table 2.1). These deposits, named the Empress Group, range in thickness from 6 to 55 m and have been subdivided into a lower preglacial unit and upper proglacial unit based on sand and gravel lithologies (Whitaker and Christiansen, 1972). The lower unit only occurs in
Table 2.1: Lithostratigraphy and chronostratigraphy of central Saskatchewan. Sources: Christiansen, 1968b, 1970, 1992; Sauer and Christiansen, 1991; Skwara Woolf, 1980; Whitaker and Christiansen, 1972;

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<tr>
<td></td>
<td>Early Wisconsinan</td>
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<td></td>
<td>Sangamon</td>
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<td></td>
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<td>Oldman Formation</td>
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preglacial river valleys (Fig. 2.4). The upper unit occurs both in preglacial valleys and on bedrock uplands (Christiansen, 1992). The Empress Group has been estimated as late Tertiary to Quaternary in age (Whitaker and Christiansen, 1972).

Pleistocene sediments in Saskatchewan are divided into two groups - the Sutherland Group and the overlying Saskatoon Group - on the basis of carbonate content, texture, stratigraphic position, the presence of weathering zones, and engineering properties (Christiansen, 1992). These parameters are also used to differentiate between formations within each group. Calcite and dolomite contents are generally lower in the Sutherland group than in the Saskatoon Group. Each formation is composed of basal proglacial sediments, till, upper proglacial sediments and postglacial sediments.
Figure 2.4: Preglacial drainage valleys across the Southern Interior Plains of Canada. Adapted from Klassen, 1989.

The Sutherland group, ranging in thickness from 17 m to 71 m, is subdivided into three formations. The Mennon Formation overlies the Empress group, followed by the Dundurn Formation, and finally the Warman Formation (Table 2.1). The tills of the Sutherland group are pre-Illinoian (early to mid-Pleistocene) in age. Three nonglacial stages (the Wellsch, Wascana, and Red Cliff) have been noted in localised type sections across the Canadian Prairies, but have not been correlated with specific formations in the Sutherland group (Stalker, 1976). The earliest (Wellsch) nonglacial stage followed an earlier Quaternary glaciation informally referred to as the “PW event” at 2 Ma (Fenton, 1984; Storer, 1989).
The Saskatoon Group is divided into two distinct formations; the Floral Formation and the overlying Battleford Formation. The Floral Formation ranges from 1 m to 70 m in thickness and consists of tills interbedded with sand and gravel (Christiansen, 1968b). The till records at least two separate glacial advances. These are termed the upper till and lower till as there is not sufficient compositional differences for the tills to be designated as formal stratigraphic units. The Floral Formation tills are hard, jointed and stained with iron and manganese where oxidised (Christiansen, 1968). The tills are separated by the nonglacial Riddell Member consisting of fossiliferous sand (Skwara Woolf, 1980). The lower till, associated with the Dunmore glaciation, is Illinoian in age and the Riddell Member, associated with the Osler nonglacial, is Sangamonian in age (Fenton et al., 1984; Fulton, 1984; Klassen, 1989). The upper till was deposited during the Burke Lake glaciation of the early Wisconsinan.

The Battleford Formation, ranging from 0 to 50 m in thickness, lies directly below Holocene surficial deposits. It is composed of soft till with little or no iron or manganese staining. The formation is further subdivided into an upper and lower till. Klassen and Vreeken (1987) suggest that the lower unit is lodgement till whereas the upper unit is ablation till based on compaction. The Battleford Formation exhibits a sandy loam to loam texture, generally containing more sand than the underlying Floral Formation (Christiansen, 1970). Based on radiocarbon dated carbonaceous material from the Watino nonglacial stratified sediments between the Floral and Battleford Formations, the Battleford Formation was deposited during the Late Wisconsinan Lostwood Glaciation which reached its maximum ca. 23 ka BP (Fenton, 1984).
The postglacial surficial deposits are composed of "deglacial lacustrine, outwash and ice-contact sediments and postglacial alluvium, colluvium, aeolian and landslide deposits" (Christiansen, 1992, p 1776). Glacio-lacustrine sediments deposited in glacial Lake Saskatchewan are the most significant, spatially, in the Saskatoon area. The surficial sediments within the study site are classified as bouldery moraine with surficial sandy glacio-lacustrine sediments west of the valley (Christiansen, 1970). The contact between the surficial deposits and the underlying Battleford Formation is gradational and conformable as a result of mixing as stagnant ice melted (Christiansen, 1992).

2.3 Modern Climate and Hydrology

Saskatoon has a continental climate with long cold winters, short warm summers and a mean annual precipitation of 350 mm. The winter period extends from early November to early April (Maybank and Bergsteinsson, 1970). Continental arctic air masses from the north, characterised by cold, stable air, are dominant but occasionally warmer Pacific air will penetrate into the region bringing warmer temperatures (Hare and Thomas, 1974; Phillips, 1990). The coldest temperatures occur in January and February with mean daily temperatures highs ranging from -10°C to -15°C, mean daily lows ranging from -20°C to -25°C and record lows from -45°C to -50°C (Hare and Thomas, 1974). Precipitation is low from October through to March inclusive, with monthly totals ranging from 0 mm to 55 mm and monthly means of 17 mm to 19 mm (Hare and Thomas, 1974). Most winter precipitation occurs as snow with continuous snow cover developing by early November. The snow cover is generally greatest at the end of February and has melted by the end of March to early April (Longley, 1972).
The dominant air masses change in the summer. The continental arctic air is replaced by milder maritime polar air masses. The majority of precipitation comes from Pacific air masses, although occasionally moist air from the Gulf of Mexico will cause increased precipitation (Hare and Thomas, 1974). Temperatures increase during the months of March, April and May although precipitation levels do not greatly increase until late April. By the beginning of May, rain is the dominant form of precipitation. On average, 60% of the annual precipitation occurs between April and August with June (57 mm) as the wettest month. The warmest temperatures occur in June, July and August with mean daily maximums of 22°C - 26°C, mean daily lows of 8°C to 12°C and maximums of 38°C to 40°C. Saskatoon has on average 2450 hours of sunshine a year (Phillips, 1990).

In most years there is a moisture deficit across southern Saskatchewan (Phillips, 1990). This may be attributed to high temperatures, low annual precipitation and low relative humidity due in part to the rain-shadow effect of the western mountain ranges. In the Saskatoon area, the mean annual runoff is estimated to be less than 25 mm (Hare and Thomas, 1974).

The source of the Opimihaw Creek is the Hudson Bay slough located north of the study site. There are no perennial tributary streams within the study site. Additional flow comes from groundwater, as observed at several small springs located along the valley margins. The spring snow melt is an important source of moisture within the study site. The Opimihaw Creek basin drains southward into the South Saskatchewan River.
2.4 Modern Soils and Vegetation

Saskatoon is located within the zone of Dark Brown Chernozemic soils. Two soil association complexes have been identified within the study site (Acton and Ellis, 1978; Ellis and Stonehouse, 1970). The soil complex to the east of the stream valley includes the Weyburn and Asquith associations. These are dominantly Orthic Dark Brown Chernozems with significant Calcareous Dark Brown Chernozems and Gleysols (Acton and Ellis, 1978). The parent material is glacial till with a moderately to very stony, fine sandy texture. To the west of the stream valley the soil complex includes the Bradwell and Weyburn associations. These are Orthic Dark Brown Chernozems with significant Eluviated Dark Brown Chernozems, Calcareous Dark Brown Chernozems and Orthic Regosols. The parent material is defined as slightly to moderately stony glacio-lacustrine deposits with till 1-2 m below the surface.

Temperature and precipitation are the most important factors controlling the nature of the vegetation cover and soil development (Brady, 1990). Parent material affects the soil texture and the types of clay minerals present within the soil, which in turn influence soil drainage. Topography influences erosion and acts as an important modifier of climatic effects and vegetation by regulating moisture availability. Time is also an important component as it takes many hundreds of years for Chernozems to develop fully. At sites where pedogenic processes have not had time to fully modify the parent material, Brunisols and Regosols dominate. In many cases a particular soil may be out of phase with its current climate and/or vegetation zone.
The study site is located within the Aspen Parkland, a transitional zone between the mixed-grass prairie to the south and transitional mixed forest to the north (Archibold and Wilson, 1980). The two primary plant communities within the aspen parklands are forest and grasslands. The grassland association in the Saskatoon area is referred to as the *Festuca* grassland. Three grass species, *Festuca scabrella* (rough fescue), *Stipa spartea* (porcupine grass) and *Koeleria cristata* (June grass), dominate the basal cover. Nearly 70% of the non-herbaceous ground cover is made up of *Solidago glaberrima* (low goldenrod), *Artemisia frigida* (pasture sage), *Anemone patens* (prairie crocus), *Antennaria microphylla* (small everlasting), *Phlox hoodii* (moss phlox) and *Cerastium arvense* (field chickweed). Significant shrubs are *Symphoricarpos occidentalis* (western snowberry) and *Elaeagnus commutata* (wolf willow). The forested regions of the parkland are dominated by *Populus tremuloides* (aspen poplar). The important forest communities in the Saskatoon area are the aspen poplar community, the maple, elm and ash flood-plain community and the willow community.
Chapter 3

Methods

3.1 Field Methods

A preliminary survey of geomorphic features was conducted noting their location on a topographic map and air photos of the study site. The features identified for further consideration were: the till plain; terrace remnants; alluvial terraces; mass movement landforms on slopes; and the modern stream channel. Sites for excavation and auguring were selected on the basis of geomorphic importance and permission from park authorities. Previous archaeological work, and therefore this study, was focused on the alluvial terraces near the valley-bottom.

An engineering level was used to survey along the length of the valley, focusing on elevation changes across and between alluvial terraces. The position and elevation of each excavation and auger hole were recorded and a detailed survey of the risers and treads of one set of terracettes completed. A theodolite was used to record the location and elevations of terrace remnants located near the tops of slopes. The elevations are reported relative to a datum located on the Newo Asiniak site alluvial terrace which is central to the study site. This is a more meaningful reference point than sea level as it emphasis small elevation differences.
The sediments composing the alluvial terraces and colluvial slopes were examined through a combination of excavations and auger holes. The lithostratigraphy of two alluvial terraces and two slope sites were examined by re-excavating old archaeological dig sites. There is an on-going archaeological excavation at one additional alluvial terrace site. Slit trenches were excavated to examine the internal structure of an alluvial terrace - colluvial slope contact and a terracette. A hand auger, capable of extracting a 0.05 m diameter core in 0.2 m sections to a maximum depth of 2 m, was used in locations not previously excavated. Large numbers of boulders and stones in the colluvium limited the use of the auger essentially to the alluvial terraces, although a small number of short cores were obtained from the slopes.

The re-excavated pits, trenches and each section of core were logged noting the sediment texture and colour, the type and depth of boundaries, organic and cultural horizons and structure. Texture was described in the field using the ‘feel method’. Samples were taken where the thickness of units permitted. The pits and trenches yielded the most reliable data as unit boundaries were often distorted during auguring. Problems were also encountered with material dropping into the hole, from both the walls and auger bottom, and small sample sizes. Bone material was recovered whenever possible for radiocarbon dating.

3.2 Laboratory Methods

Laboratory analyses were performed in order to describe and classify selected sediment samples into distinct sedimentary facies thus enabling a comparison of the sediments within and between individual profiles. The laboratory analyses conducted
during this study were the determination of the percentage of organic and inorganic carbon by loss on ignition and particle size analysis by sieving and pipetting.

3.2.1 Sample Selection

Each of the re-excavated archaeological sites, the trenches and several auger holes were selected for detailed analysis where field descriptions of lithology indicated that changes in the nature and/or rate of sedimentation may have occurred. Field data on sediment texture, colour and organic matter content were utilised to develop a preliminary set of sedimentary facies represented in the profiles. The number of samples selected for further analysis was based on the relative importance of each facies within the profile. Of the 650 samples taken, 219 were selected for further analysis.

3.2.2 Organic and Inorganic Carbon Level Determination

Percentages of organic and inorganic carbon were determined using a modified loss on ignition procedure from Dean (1974) and Scott (1991). Approximately 5 g of air dried sample was weighed to 0.0001 g then placed in a 550°C muffle furnace for one hour, cooled in a dissector and weighed. The sample was then placed in a 1000°C muffle furnace for one hour, cooled and weighed. A detailed procedure and calculations are presented in Appendix A.

The loss on ignition method was used as several samples could be analysed simultaneously with minimal operator time. The method has the additional advantage of using standard laboratory equipment. The accuracy and precision of this method compares favourably with percent organic carbon determined chromatographically and percent inorganic carbonates determined by acid neutralization, atomic absorption and
total carbon analysers (Dean, 1974). Clay will lose some lattice water in the 550 - 1 000°C temperature range resulting in a higher inorganic carbonate estimate than is actually present within the sediment. It has been determined, however, that the error is proportional to the amount of clay within the sample; pure clay samples may experience up to 5% weight loss (Dean, 1974).

3.2.3 Particle Size Analysis

Two methods were used for particle size analysis. The gravel and sand fractions were analysed according to standard sieving procedures and the silt and clay fractions according to standard pipette procedures (Day, 1965; Kunze, 1965; Lewis and McConchie, 1994; Scott, 1991) (Appendix A). Each sample was pre-treated with hydrogen peroxide to remove organic material. The sample was not treated to remove carbonate cementation as the parent material (till) for the sediments within the study site is carbonate rich. Removal of carbonate cement would, therefore, remove a portion of the sediments to be analysed. Careful visual inspection of the sediments samples revealed that only one sample contained significant cementation.

Sediments were classified according to the Wentworth scale where each grain size class is half the interval of the previous class. When the grain size data are plotted, this produces a skewed distribution. The distribution of grain sizes within a sediment tends to become more symmetrical when the logarithm of the particle diameter is used instead of the particle diameter itself (Inman, 1952; Krumbein, 1938; Otto, 1939). The phi (\(\phi\)) notation developed by Krumbein (1936) converts the diameter in millimetres of a particle into a logarithmic scale by the formula
\[ \phi = -\log_2 \xi \] (3.1)

where \( \xi \) = the diameter of the particle in millimetres.

When a near symmetrical distribution is used, measures of central tendency, dispersion and shape will be more significant. The diameters and corresponding phi notations used in this study are presented in Table 3.1.

Table 3.1: Diameter in millimetres (\( \xi \)) and corresponding phi (\( \phi \)) units of class limits used in particle size analysis.

<table>
<thead>
<tr>
<th>( \xi )</th>
<th>( \phi )</th>
<th>( \xi )</th>
<th>( \phi )</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.000</td>
<td>-1.0</td>
<td>0.125</td>
<td>3.0</td>
</tr>
<tr>
<td>1.400</td>
<td>-0.5</td>
<td>0.090</td>
<td>3.5</td>
</tr>
<tr>
<td>1.000</td>
<td>0</td>
<td>0.063</td>
<td>4.0</td>
</tr>
<tr>
<td>0.710</td>
<td>0.5</td>
<td>0.020</td>
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</tr>
<tr>
<td>0.500</td>
<td>1.0</td>
<td>0.005</td>
<td>7.5</td>
</tr>
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<td>1.5</td>
<td>0.002</td>
<td>9.0</td>
</tr>
<tr>
<td>0.250</td>
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</tr>
<tr>
<td>0.180</td>
<td>2.5</td>
<td></td>
<td></td>
</tr>
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</table>

3.2.4 Radiocarbon Date Material

Bone samples from four units were selected for radiocarbon dating. Samples from near the base of the Amisk and Newo Asiniak sites were submitted to constrain the beginning of colluviation and channel aggradation, respectively. One sample was submitted from the Redtail site in an effort to correlate the current study with previous archaeological work.

Samples were cleaned with distilled water and a soft nylon brush, then air dried. Each sample was sealed in a sample bag and refrigerated at 4°C prior to submission for analysis. Three of the four samples were too small for bulk dating and were, therefore, submitted to the Isotrace Laboratory, University of Toronto, for AMS dating. The
remaining sample was submitted to the Geological Survey of Canada Radiocarbon Laboratory for standard radiocarbon dating.

3.3 Data Analysis

Once particle size analysis had been completed, the relationship between the distributions of sediment sizes and possible depositional environments were examined. The best method for this analysis has been the subject of considerable debate for some time. Both graphical methods (directly comparing the shape of curves) and numerical methods (comparing calculated parameters) have been used although numerical methods are currently favoured. The parameters calculated are a measure of central tendency (average grain size), dispersion (standard deviation), and shape (skewness and kurtosis) of the distribution. The dispersion is commonly translated into degree of sorting within the sediment sample (Folk, 1968). The results are then compared in order to group sediments into facies.

Graphic probability plots combine graphic representation of the data with calculated parameters. A cumulative frequency curve of the grain size data is plotted onto probability paper using the phi interval as the independent variable as described by Folk (1968). The use of a probability scale straightens out the S-shape of the frequency curve providing greater accuracy to interpolations between known points (Glaister and Nelson, 1974). The expansion of the tail portion of the curve allows greater accuracy reading the graph and improved extrapolations beyond the analysed portion of the curve. The curve is then used both for subjective shape comparisons and to determine the phi intervals of various percentiles used in the numerical calculations. Several formulae have
been developed to determine each statistical parameter from graphic plots. The most common set of formulae were developed by Folk and Ward (1957) and are used in this study (Appendix B).

This method has the advantage that irregularly spaced phi intervals can be plotted; this is important in the silt and clay fraction of the sediment. In most grain size analysis, the fine fractions often remain unanalysed resulting in an “open tail” (Lewis and McConchie, 1994). This method easily accommodates open-ended distributions as the unanalysed portion of the sediment is extrapolated instead of being lumped into a single size class.

3.4 Significance of Cultural Material to Current Study

The study site has been periodically occupied by humans for over five thousand years. This habitation has left behind a wealth of material which is slowly being uncovered and studied by archaeologists. Detailed studies have been conducted at three sites within the valley; the Amisk site (Amundson, 1986), the Newo Asiniak site (Kelly, 1986) and the Redtail site (Ramsay, 1992). Two additional sites, the Thundercloud and Tipperary Creek sites, are currently under investigation. Numerous sites, located during an archaeological inventory by Dr. E. Walker, have yet to be examined.

The rich cultural history is a valuable tool in the interpretation of physical changes. A series of 38 radiocarbon dates have been obtained from alluvial terrace cultural horizons, constraining the period of valley aggradation (Appendix C). Relative dates based on tool traditions are useful where absolute dates have not yet been determined, such as at the Thundercloud site.
3.5 Radiocarbon Dating and Dates

Accurate methods of dating are required for establishing chronological control when attempting to reconstruct past environments. For example, peat layers, both above and below glacial tills, are dated to provide an age bracket when glaciers would have covered the land (Grootes, 1978). Dating sets of paired terraces and valley-fill deposits provides an indication of phases of fluvial aggradation and incision (Rains and Welch, 1988). Dates from paleosols have been used to determine periods of slope stability (Klassen and Vreeken, 1987).

There are several isotopes that are currently used for radiometric dating, but radiocarbon dating is most applicable for the reconstruction of Quaternary events in Canada. Radiocarbon dating can be performed on a wide variety of materials such as wood, peat, charcoal, soils, shells, bone, organic detritus, organic-rich sediments and gyttja with varying degrees of reliability (Klassen and Vreeken, 1985).

Three isotopes of carbon exist naturally, $^{12}\text{C}$, $^{13}\text{C}$ and $^{14}\text{C}$. $^{12}\text{C}$ and $^{13}\text{C}$ are stable with $^{12}\text{C}$ being the most abundant form. The unstable isotope, $^{14}\text{C}$, is produced when neutrons bombard atmospheric nitrogen in the upper atmosphere. The $^{14}\text{C}$ is oxidised into $^{14}\text{CO}_2$ and gradually mixes with stable forms of CO$_2$ in the lower atmosphere. $^{14}\text{C}$ breaks down into its stable daughter product, $^{14}\text{N}$, and weak beta particles are emitted. The $^{14}\text{CO}_2$ is taken into plants and animals through photosynthesis and respiration in the same way as CO$_2$. As long as the organism is alive, $^{14}\text{C}$ is constantly replaced reflecting the levels that are found within the atmosphere. When the organism dies, this exchange is no longer maintained and $^{14}\text{C}$ levels decline.
For standard radiocarbon dating, the amount of $^{14}$C in an organism is determined by measuring the beta-particle activity in a sample as a number of disintegrations per minute. Alternatively, ions may be produced, accelerated, detected and counted in accelerator mass spectrometer (AMS) dating (Stuiver, 1978B). This has the advantage of requiring a much smaller sample size than standard dating techniques. In both methods, the count will depend on how much of the $^{14}$C has already decayed and can, therefore, be converted into a fairly accurate age estimation using the isotopes half-life. This was originally calculated by Libby as 5 568±30 years and later adjusted to 5 730±43 years (Bowen, 1978). Radiocarbon dates are only a statement of a Poisson Probability Curve and are, therefore, reported with ± one standard deviation (Stuiver and Polach, 1977). The base year for reporting radiocarbon dates is 1950.

The levels of $^{14}$CO$_2$ in the atmosphere have fluctuated with time. These fluctuations are related to volcanic emissions, temporal fluctuations in the cosmic rays bombarding the upper atmosphere and more recently, the combustion of fossil fuels and nuclear testing (Goetz, 1990; Vogel, 1983). Calibration curves have been developed for recent times using techniques such as dendrochronology and varve dating (Pearson et al., 1977; Stuiver, 1978A). Checks have also been made by comparing results of radiocarbon and uranium series dating on marine phosphates, thermoluminescence and magnetic dating (Burnett and Kim, 1986; Stuiver, 1978A). Recent adjustments to CALIB (a computer calibration program) allow corrections to 18 400 BP (corrected to 22 000 BP) (Stuiver and Reimer, 1993).
Contamination of samples is a major source of error in radiocarbon date determinations. Although contamination from dust or cleaning reagents may occur in the laboratory, most contamination occurs in the field, usually in the form of the incorporation of either old or young carbon. The degree of error can generally be reduced by removing the outer layers of the sample that would have come in closest contact with the contaminant.

Carbonates secreted by aquatic plants and animals are sensitive to contamination from old carbon in the form of hard water (Aravena et al., 1992), particularly in regions with carbonate-rich bedrock or glacial deposits such as the Saskatoon area. The carbon enters the ground water system and is incorporated into the organism, giving an age that is too old. Old carbon may also be incorporated into samples from trace amounts of fine-fraction Palaeocene lignite and Cretaceous black shale. Humic acids, organic decay products, fresh calcium carbonate and the penetration of modern roots will contaminate soil and peat samples, generating dates which are too young (Goudie, 1983).

Clayton and Moran (1982) claim that the potential for contamination with older carbon is large enough that all dates except those obtained from wood should be disregarded. Klassen (1983) and Jackson (1983) criticise the rejection of all non-wood dates, contending that contamination is not as widespread as Clayton and Moran suggest. They argue that peat, charcoal and bone samples are reliable but that dates on organic silts and clays should be treated with caution. It should be noted that the porous nature of both wood and charcoal make them susceptible to the absorption of dissolved organic substances (Godwin, 1969). The continuous input of organic material into soils
consistently results in dates younger than the true age of the soil. The degree of error is controlled primarily by the rate of organic carbon cycling in the soil (Wang et al., 1996). Buried soils have an additional source of error, particularly in high rainfall regions, from the downward translocation of water-soluble acids from more modern soils. At some point, the soil will reach a steady-state, after which radiocarbon dating will provide no indication of the age of the soil.

Time zero, the point at which decay actually began, is an important concept in radiometric dating that is often overlooked. Dating a piece of bone found in a fluvial deposit will not necessarily give the age of the deposit as the bone may have been very old before it was incorporated into the deposit. For this reason, dates should be considered as maximum ages for a particular deposit.
Chapter 4

Literature Review

4.1 Deglaciation

Research concerning glacial advances and retreats during the late Wisconsin has focused on regional mapping of maximum ice margins. Landforms (such as ice-thrust and ridged moraine, end moraines, ice-marginal channels and spillways), the stratigraphy of glacial lake basins, drainage patterns and radiocarbon dates facilitate the reconstruction of ice margins (Christiansen, 1979). Little work has focused on identifying when specific sites were deglaciated making it difficult to establish the exact timing of the deglaciation of the Saskatoon area.

The late Wisconsinan glacier retreated from the southwest to the northeast of the province following the slope of the land. Nine phases of deglaciation have been proposed for Saskatchewan indicating the positions of ice margins, proglacial lakes and major drainage systems (Christiansen, 1979). The ice-margin positions likely represent zones of stagnating ice. Phases five through seven record the deglaciation of the Saskatoon area (Fig. 4.1). Meltwater from rivers and glacial lakes flowed through various spillways into Glacial Lake Agassiz throughout these phases (Table 4.1).
Figure 4.1: Nine-Phase model of deglaciation in Saskatchewan. Phases five through seven indicate the pattern of deglaciation in the Saskatoon area. Source: Adapted from Christiansen, 1979.
Table 4.1: Active drainage patterns in Saskatchewan during phases five through seven of the late Wisconsin deglaciation. These drainage patterns are a critical component of ice margin definition. Source: Christiansen, 1979.

<table>
<thead>
<tr>
<th>Phase</th>
<th>Channels</th>
<th>Spillways</th>
<th>Extraglacial Rivers</th>
<th>Glacial Lakes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Five</td>
<td>Battle</td>
<td>Blackstrap, Lewis, Last Mountain, S. Saskatchewan, Qu’Appelle, Assiniboine, Cutarm</td>
<td>S. Saskatchewan, Oldman, Red Deer</td>
<td>Saskatchewan, Last Mountain, Saltcoats, Agassiz</td>
</tr>
<tr>
<td>Six</td>
<td>Wakaw</td>
<td>Vermilion, N. Saskatchewan, Watrous, Last Mountain, Qu’Appelle, Assiniboine</td>
<td>S. Saskatchewan, Oldman, Red Deer</td>
<td>Saskatchewan, Melfort, Agassiz</td>
</tr>
<tr>
<td>Seven</td>
<td>Shell Brook, Sturgeon, Spruce</td>
<td>N. Saskatchewan, Assiniboine</td>
<td>S. Saskatchewan, Oldman, Red Deer</td>
<td>Saskatchewan, Agassiz</td>
</tr>
</tbody>
</table>

According to Christiansen’s model, the Saskatoon region was still under ice during phase five ca. 12.5 ka BP. No dateable material was recovered by Christiansen relating to this margin. The phase four ice margin, radiocarbon dated ca. 14 ka BP based on bone, gyttja, organic detritus and carbonaceous silt, is the maximum age for the phase five margin (14 670 ± 240, S-300A; 15 850 ± 225, S-300A; 14 300 ± 320, GSC-1369; 14 040 ± 465, S-685; 13 900 ± 240, I-3476; Appendix C). By phase six, ca. 12 ka BP, the ice front had retreated north of Saskatoon and the North and South Saskatchewan River valleys and surrounding areas were flooded by Glacial Lake Saskatchewan (Mott and Christiansen, 1981). Chronological control of this margin is based on wood found in
alluvium in the Qu’Appelle spillway marking the abandonment of the spillway (12 025 ± 205, S-553). By phase seven, ca. 11.5 ka BP, Lake Saskatchewan had retreated north, following the glacier margin, and the Saskatoon area was free of ice and water. Carbonaceous material from beneath till deposited in the Ladder Valley by a readvance during phase seven was dated at 11 610 ± 450 BP (GX-2254). Glacial Lake Saskatchewan drained shortly after as the glacier retreated down the Manitoba Escarpment, further constraining the phase seven margin age estimate. Organic sand from the base of a pond in the lake basin was used to date the drainage of the lake (11 560 ± 640, GSC-648).

The ice margins proposed by Christiansen have been accepted, but the chronology attached to the margins is generally rejected as being too early. Concern over the chronology focuses on the use of unreliable dates from organic detritus and carbonaceous material (e.g. the phase seven margin). The wood date from the Qu’Appelle alluvium (12 025 ± 205, S-553) is called into question as no evidence is presented to support the contention that it was deposited after abandonment. It has been reinterpreted as having been deposited at the beginning of melt-water flow during phase four (Clayton and Moran, 1982; Klassen, 1989). In addition, several ice margins do not have any radiocarbon dates associated with them. An additional source of error is that Christiansen does not consider the glacial chronology to the south and east of the province, in particular the history of Glacial Lake Agassiz (Teller et al., 1980). Defining the period of establishment of Glacial Lake Agassiz is important as Christiansen proposes that meltwater drained into it from phase four onwards.
The date attached to the phase four margin is likely too old as it conflicts with a group of wood dates from the Bemis Moraine in Iowa (13 820 ± 400, W-513; 13 900 ± 400, I-1268; 13 910 ± 400, I-517; 14 200 ± 500, I-1402; 14 470 ± 400, W-512; 14 700 ± 400, W-153) which establish that ice was present within the Lake Agassiz basin at the same time that Christiansen states meltwater is draining into it (Clayton and Moran, 1982; Teller and Fenton, 1980; Teller et al., 1980). It is postulated that subsequent to this, eight readvances of the Red River-Des Moines Lobe occurred, so that ice was in Iowa and South Dakota as late as ca. 12.3 ka BP, retreating into the Red River Valley, and forming Glacial Lake Agassiz ca. 11.7 ka BP (Clayton and Moran, 1982; Teller, 1987). This is clearly much later than Christiansens' ca. 14 ka BP (phase four) estimate for the establishment of the lake. Each succeeding phase of deglaciation must now be placed later.

The trend in estimating the ages of the ice margins has been to place them later and later in time (Table 4.2). Clayton and Moran (1982) propose that although the Saskatoon area was still under ice ca. 11.3 ka BP (phase four ice margin), deglaciation progressed rapidly so that the ice had reached Christiansen's phase seven margin ca. 11.1 ka BP (Table 4.2). Teller (1987) places the phase seven ice margin ca. 11 ka BP, but it is not clear whether Glacial Lake Saskatchewan had drained from the region at this time. Klassen (1989) places the phase four margin ca. 12 ka BP based on wood, peat and freshwater shell dates in Manitoba and Saskatchewan (12 025 ± 205, S-553; 12 100 ± 160, GSC-1319; 12 630 ± 80, TO-216). The model is in close agreement with those of Clayton and Moran (1982) and Teller (1987) in its estimate of the retreat from
Table 4.2: Summary of chronologies for the deglaciation of the Saskatoon area. Note the general trend toward later dates for more recent research. Dates are ka BP. Sources: Christiansen, 1979; Clayton and Moran, 1982; Kehew and Teller, 1994; Klassen, 1989; Teller, 1987; Thorleifson, 1996.

<table>
<thead>
<tr>
<th>Source</th>
<th>Phase Four</th>
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<td>---</td>
<td>11</td>
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</tbody>
</table>

Saskatoon, placing the ice near Christiansen’s phase eight margin ca. 11 ka BP based on wood dates from the North Saskatchewan River and Assiniboine River alluvium (10 690 ± 190, GSC-677; 10 840 ± 355, S-2077; 10 875 ± 660, S-1374). Current investigations focusing on Glacial Lake Agassiz fluctuations place Christiansen’s margins even later (Kehew and Teller, 1994; Thorleifson, 1996). A generalised phase five margin was placed at ca. 11 ka BP by Kehew and Teller (1994). Thorleifson places the phase four ice margin ca. 11.1 ka BP, retreating north of Saskatoon (phase six) ca. 10.9 ka BP, with the region free of ice and water ca. 10.4 ka BP (phase seven). This date is based on 440 varves situated below a distinctive red clay marker at Dryden, Ontario. The chronology includes research conducted on various meltwater channels, glacial lakes and the various outlet channels used by Lake Agassiz throughout its history.

Establishing a reliable date for the phase seven margin is important because it is the minimum age for the drainage of Glacial Lake Saskatchewan which, in turn, is the maximum age for the incision of the Opimihaw Creek. The regional models of deglaciation that have been developed for the Southern Interior Plains region indicate that
deglaciation of the Saskatoon area would have occurred by ca. 11 ka BP, with the area fully drained as late as ca. 10.4 ka BP.

4.2 Glacial Lake Saskatchewan and the Development of the South Saskatchewan River Drainage System

Glacial Lake Saskatchewan (originally named Glacial Lake Saskatoon) was formed by glacial melt-water and water draining from southern Saskatchewan and Alberta through the North and South Saskatchewan River systems. Large sandy deltas formed where these rivers entered the lake. It is estimated that the maximum water level of the lake in the Saskatoon area was about 550 m a.s.l. (Christiansen, 1970). As the ice and lake migrated north, the water level dropped to 510 m a.s.l., gradually becoming concentrated in a broad channel covering the modern river channel, the Hudson’s Bay slough area and the Saskatoon terrace (Cherry, 1962; Christiansen, 1970). Levels dropped to approximately 495 m a.s.l., forming the Saskatoon terrace, when Glacial Lake Saskatchewan drained into Glacial Lake Agassiz. Aeolian activity modified the sandy delta sediments, depositing silt and sand at Beaver Creek and on the Saskatoon Terrace (Chapter 4.3.1; Christiansen, 1970; Turchenek et al., 1974). A profile of the South Saskatchewan River bed (Cherry 1962) indicates that there is between 4.5 and 6 m of aggraded river deposits including gravel, sand and pebbly clays beneath the South Saskatchewan River. This profile indicates that the river incised to 460 m a.s.l. before aggrading (Cherry, 1962). Little attempt was made by Cherry to account for these changes in the fluvial regime beyond the suggestion of variations in local base level and isostatic rebound.
4.3 Holocene Climatic Change

Over the past few decades researchers have developed global circulation models that attempt to describe the substantial changes in the earth’s climate during and after the last glaciation. Changes in the radiative regime and/or continental relief were able to drive the changes in global circulation responsible for the primary climatic variations throughout the Holocene (COHMAP, 1988). The proxy record, however, suggests a more complex response than is generally modelled.

Between 10 ka BP and the present, the timing of the perihelion of the earth’s orbit has shifted from July to January. During this same period, the tilt of the earth’s axis has shifted from 24.5° to 23.5°. These factors combined to increase solar radiation at the top of the earth’s atmosphere during the summer by 8% and decrease it during the winter by 8% (COHMAP, 1988). The period of greatest seasonality would have lasted ca. 12 to 9 ka BP (Bryson, 1985; Webb et al., 1987). The result of this increased seasonality on continental interiors would have been to increase temperatures during the summer and decrease them during the winter. There is general consensus in the published literature that the timing of the Hypsithermal lags behind the period of maximum solar radiation. Globally, the period of maximum aridity has been placed between 8 ka and 3 ka BP, indicating a great deal of regional variation in climate (Lamb, 1971). The change to maximum aridity was time transgressive occurring earlier in western plains regions than in eastern Canada (Vance et al., 1995). The continuing presence of the Laurentide ice sheet in central and eastern Canada effectively delayed the Hypsithermal period of maximum
aridity until after 6 ka BP (Bryson, 1985; COHMAP, 1988; Ritchie, 1983). Conditions gradually became cooler and moister after the Hypsithermal interval.

A variety of paleoclimatic indicators will be examined in the following sections including paleosols, plant macrofossils and pollen, landscape stability and lake level fluctuations. The discussion is limited to a brief outline of the technique and the results of relevant studies based on the original author’s interpretations. Studies were selected from the central region of the Southern Interior Plains in an effort to minimise variability due to the time transgressive nature of Holocene climate change. There is, however, both a general lack of published studies and considerable variation in the nature of climatic change between the few studies that have been conducted.

4.3.1 Paleosols

Studies in Saskatchewan have examined buried paleosols; soils which formed on past landscapes and have subsequently been buried (Valentine and Dalrymple, 1976). They are used as stratigraphic markers and in the interpretation of past environments through identification of soil types and knowledge of pedogenic processes (Rutter, 1969; Turchenek et al., 1974). Organic material within the soil may be used for radiocarbon dating providing chronological control for stratigraphic studies. Relative dating techniques are based on soil properties that change over time such as the composition of organic matter and the degree of profile development (Evans, 1985).

There are several sources of error in the interpretation of paleosols. The nature of soil development varies significantly across a small area based on landscape position (Pennock et al., 1987) This is due primarily to moisture variability and its influence on
vegetation growth and profile development. Soils also vary according to the initial parent material and textural differences within a landscape. Different interpretations of former climates may occur depending on where a particular profile is located topographically. Misinterpretations may be avoided, however, if the catenary position of the paleosol is examined (Pennock and Vreeken, 1986). Changes that occur in the soil after burial such as continued decomposition of organic material, and the movement of carbonates and clay particles is another significant source of error (Catt, 1987; Valentine and Dalrymple, 1976).

Two postglacial paleosols occurring on river terraces adjacent to the South Saskatchewan river were investigated at Batoche, Saskatoon and Beaver Creek (Turchenek et al., 1974). The older paleosol at Batoche, an Orthic Black Chernozem, indicates soil development under a moister climate than present conditions. Black Chernozems currently develop in cool to cold, subhumid environments under grassland or grassland-forest communities (Agriculture Canada Expert Committee on Soil Survey, 1987). Mottling of the soil was interpreted as an indication of poor drainage possibly due to the river level being close to the terraces during soil formation.

Burial of this soil by aeolian deposits ca. 8 ka BP (8 100 ± 120, S-234) suggests a change to more arid conditions at this time. This conclusion is based on the assumption that soil formation indicates an extended period of landscape stability, hence radiocarbon dates on paleosols relate to the deposition of overlying materials rather than deposition of underlying sediments (Wang et al., 1996).
The younger paleosol, a Regosol or weakly developed Chernozem, indicates a
decrease in aeolian activity whereas its burial marks a return to more arid conditions ca.
7 ka BP (7 070 ± 115, S-445). Regosols are weakly developed soils that lack B horizon
development. They form under a wide range of climates and vegetation often as a result of recent deposition or unstable conditions (Agriculture Canada Expert Committee on

At the Saskatoon site only the older paleosol, a weakly developed Regosol, was intact. The similarity between the A horizon and C horizons of this paleosol, due to youthfulness of the sediment was interpreted as a cumulic soil (Turchenek et al., 1974). The paleosol was buried in aeolian sands at about the same time as the Batoche paleosol (8 160 ± 150, S-296).

The older paleosol at Beaver Creek exhibits a similar morphology to the Saskatoon paleosol, but is radiocarbon dated 2 ka earlier (9 940 ± 160, S-442), indicating earlier burial by aeolian sediments. The younger Beaver Creek paleosol, a Regosol, was also buried by aeolian deposits dated ca. 7.6 ka BP (7 640 ± 150, S-443).

4.3.2 Palynology and Plant Macrofossils

Palynology is the study of organic-walled microfossils, primarily pollen and spores
(MacDonald, 1990). Pollen and spores are most commonly obtained from lake cores where anaerobic conditions aid preservation. They are easily extracted from the sediment and can be identified down to the family, genus, or species level. Reconstructing the former vegetation cover of a region using palynology is based on the assumption that the proportion of pollen grains for each species within a sediment sample is representative of
the number of parent plants, reflecting the composition of vegetation at a particular site. The composition of the vegetation cover may change over time and these changes will be recorded in the pollen spectra. A researcher may use the climatic limits of one or more indicator species to determine paleoclimates (MacDonald, 1990). Alternatively, multiple regression analysis may be performed on pollen percentages to calibrate modern distributions with climate variables such as temperature or rainfall. This information may then be applied to the fossil pollen record to obtain paleoclimatic information (Arigo et al., 1986). This is not a simple process as in many cases modern analogues do not exist. Even where analogues are used, the match may only be for one or two climatic variables such as temperature or precipitation and not for climate as a whole (Bryson, 1985).

The pollen record is regional in nature, particularly for larger lakes (MacDonald, 1990). Small lakes and ponds are more useful as they contain the greatest proportion of local material. The dispersal mechanisms of plants, productivity and sorting by both wind and water influence the accumulation of material at a site biasing the representation of the local flora in the pollen record which contributes to incorrect interpretations (Warner, 1990). Reworking of older material may also cause errors.

Palynology has been one of the most widely used paleoclimatic indicators, although recent studies are tending to place a greater emphasis on the investigation of plant macrofossils (Vance et al., 1995). Plant macrofossils are the reproductive and vegetative plant parts including fruits, seeds, wood, leaves, buds and mosses that are visible to the unaided eye (Wasylikowa, 1986). Plant macrofossil indicator species may be used for paleoenvironmental reconstruction in a similar fashion to pollen. They are also
subject to many of the same considerations of error. The primary advantages of plant macrofossils are that they are more easily identified to the species level and provide a more accurate picture of local flora (Wasylikowa, 1986).

A series of four lakes in central Saskatchewan have been examined for their sediment and pollen record (Mott, 1967, 1973). Spruce-dominated pollen assemblages suggest that a closed spruce forest invaded central Saskatchewan following the retreat of glacial ice. Spruce gradually declined and herbs invaded open upland areas. An open grassland vegetation assemblage may have spread northward as far as the Prince Albert area north of Saskatoon. The herbaceous taxa subsequently declined in favour of birch, alder and willow, indicating the development of open forest conditions. South of Saskatoon, grasslands dominated after the decline of the initial spruce forest. High percentages of Chenopodiinea and Ambrosieae and low percentages of Artemisia (sagebrush) indicate dry conditions. These trends were then reversed indicating a return to a moister climate.

Radiocarbon-dated lake sediments (8 520 ± 170, GSC-643, Cycloid Lake; 6 000 ± 170, GSC-1335, Cycloid Lake; 10 260 ± 170, GSC-647, Waskesiu; 11 560 ± 640, GSC-648, Prince Albert; 7 590 ± 220, GSC-1506, Clearwater Lake; 9 310 ± 150, GSC-1506, Clearwater Lake) provide basal dates for the lake sites. Dates obtained for the organic and inorganic carbonate fraction of the same sample indicate considerable error. The value of this study is further limited by the lack of chronological control for the vegetation changes outlined above.
Plant macrofossils preserved within a small kettle depression (Andrews Site) on the Missouri Coteau southeast of Moose Jaw record environmental changes between 10.2 and 5.8 ka BP (Yansa, 1995, 1996). An early post-glacial spruce woodland, dated 10.3-10.2 ka (10.2 ± 0.14 ka BP, ASA-D95-2; 10.2 ± 0.08 ka BP, GSC-5882; 10.23 ± 0.14 ka BP, ASA-D95-1), and subsequent deciduous parkland and pond environment indicate moist conditions extending to at least 8.8 ka (8 790 ± 140, TO-5019). Conditions subsequently became drier with peak aridity ca. 7.7 ka BP (7 670 ± 80, TO-4780) recorded by the presence of saline plants and charcoal. After ca. 7.7 ka BP, the environment became moister and a semi-permanent, calcareous slough formed. This slough was in existence to at least ca. 5.8 ka BP (5 770 ± 80, TO-5018). The moist conditions during this period have been attributed in part to variations in local groundwater discharge. The end of the record corresponds to drier conditions when the slough became ephemeral.

Martins Slough, 14 km west of Saskatoon, was used in a palynological study by Mott and Christiansen (1981). They note an abundance of herb and grass assemblages indicating a treeless landscape ca. 11 ka BP. This was replaced by an early *Picea* (spruce) dominated assemblage ca. 10.5 ka BP. The prevalence of herbaceous taxa indicate an open forest. This forest subsequently declined and a grassland assemblage of *Artemisia* (sagebrush) and other herbs developed ca. 10 ka BP which persisted into modern times.

Chronological control is provided by three radiocarbon dates (8 350 ± 200, S-1197; 10 240 ± 250, S-1198; and 11 070 ± 245, S-1199) obtained from lake sediments.
Organic sediment does not yield reliable dates so while the pollen analysis may be accurate, the timing attached to the events is suspect. When viewed in terms of current deglaciation theories which place deglaciation and drainage of Glacial Lake Saskatchewan ca. 10.4 ka BP, it is evident that the Martins slough chronology is too early.

A continuous record of Holocene environmental change has been obtained from cores extracted from Harris Lake in the Cypress Hills (Last and Sauchyn, 1993; Sauchyn, 1990; Sauchyn and Sauchyn, 1991). The studies performed on the core integrate sediment analysis, including mineralogy and lithostratigraphy with palynology and diatoms (Wilson et al., 1997). The pollen record indicates the establishment of a *Populus* (poplar) - grassland - shrub vegetation complex by 9 ka BP (9 120 ± 250 BP, S-2908; Sauchyn and Sauchyn, 1991).

*Populus* and aquatic species declined throughout the period ending ca 7.7 ka BP while herbs increased, indicating a slow transition to grassland. The climate during this period was drier and warmer than the present climate (Sauchyn, 1990). Low sedimentation rates indicate the landscape was fairly stable. Sedimentation of clastic materials derived from aeolian and fluvial erosion dominate the core although chemical precipitation increased during this period. Sections of the core indicating magnesium/calcium ratios greater than 10 are attributed to short periods of saline to hypersaline conditions (Last and Sauchyn, 1993). The absence of low magnesium calcite indicates that the magnesium-rich ground water was dominant over surface runoff and direct precipitation as a source of water in the basin throughout the Holocene.
A shift to high salinity waters and salt tolerant plants marks the increasingly warm and dry conditions of the Hypsithermal interval from 7.7 ka to 5 ka BP (Mazama ash dated 6.8 ka BP; 5 120 ± 60 BP, TO-1055; Bacon, 1983; Sauchyn and Sauchyn, 1991). Rates of clastic sedimentation increased possibly due to a loss of ground cover leading to more rapid aeolian and fluvial erosion.

Increased levels of coniferous and aquatic species ca. 5 to 3 ka BP (5 120 ± 60, TO-1055; 3 450 ± 50 BP, TO-1054), suggests a climatic deterioration with a return to cooler and moister conditions. Rotational landsliding became the dominant geomorphic process contributing to maximum sedimentation rates in the core. The high sedimentation rates subsequently declined reflecting a decrease in the magnitude and frequency of landslides as the landscape adjusted to a new equilibrium (Sauchyn, 1990). Last and Sauchyn (1993) further hypothesise that a shift from clastic to chemical sedimentation ca. 4 ka BP may mark a diversion of a major portion of the fluvial system out of Harris Lake. This diversion may have been the result of a landslide located west of the Harris Lake drainage basin. Modern plant taxa were established by 3.2 ka BP indicating a shift to modern semi-arid climatic conditions (Sauchyn and Sauchyn, 1991).

The late Holocene history of the Saskatoon area is further refined by the investigation of Waldsea Lake, 100 km east of the city. The lithostratigraphy of the site was investigated in conjunction with the pollen record (Last and Schweyen, 1985; Schweyen and Last, 1983). The changes in the nature and rate of sedimentation are attributed to variations in climate. An ephemeral, hypersaline lake existed ca. 4.5 - 4 ka BP. Hypersaline algal mud flats were produced on-shore at the same time that organic,
salt-rich muds were produced within the lake (Schweyen and Last, 1983). The presence of a hard, dense crust containing alkaline earth carbonates formed by evaporation of subsurface brines, carbonate sands and grits and rounded gypsum grains indicate exposure, desiccation and reworking of sediments on the lake shore where the sediments were not normally inundated with lake brines (Last and Schweyen, 1985). The lack of bedded salts during this phase suggests the domination of clastic inflow from sheetwash, wind and ephemeral streamflow. Vegetation mats, dated ca. 4 ka BP (3 970 ± 90, Beta-6892), within the shallow lake deposits were likely derived from basin-margin vegetation. The pollen record suggests the presence of an aspen parkland during this period.

The increasing abundance of pine and spruce pollen in lake sediments ca. 4 - 2.8 ka BP indicate a transition to a cooler, moister climate. The lake became deeper by 3 ka (2 920 ± 70, Beta-6508) in response to the cooler, wetter trend at the end of the Hypsithermal interval. Thin laminae of relatively pure aragonite irregularly spaced within the deep water sediments indicate periods of increased streamflow or precipitation (Last and Schweyen, 1985).

A short return to warmer climates, ca. 2.8 - 2.2 ka BP (2 340 ± 70, Beta-6507) was associated with the expansion of grass and herb assemblages and lowered water levels. By 2 ka, high water levels were regained and vegetation assemblages were established (Last and Schweyen, 1985). These conditions have prevailed into modern times with the exception of a minor lowering of lake levels about 700 years ago (no radiocarbon date available). Evidence for this is the deposition of coarser shoreline/nearshore sediments further out into the basin. High water levels were
interpreted as representing cool, moist conditions while low lake levels were interpreted as representing warm, dry conditions.

4.3.3 Peatlands

Studies of peat development are confined to boreal forest regions north of the study site. In the grassland and parkland regions of the province marshes and shallow ponds, which do not develop peat, are prevalent (Kuhry et al., 1992). In addition, extreme variations in water table depths inhibit the development of fens within the prairies (Zoltai and Vitt, 1990).

The accumulation of significant peat deposits within the southern boreal forest has been closely linked to the onset of cooler and moister conditions after 6 ka BP (Kuhry et al., 1992; Zoltai and Vitt, 1990). It is estimated that the modern distribution of peat-producing fens and bogs was attained ca. 3.5 - 2 ka BP. Prior to 6 ka, the development of peat deposits was inhibited by the warm, dry conditions of the early to middle Holocene. By comparing the modern and 6 ka BP fen distribution limits, Zoltai and Vitt, (1990) estimate that growing degree days were 6 to 20% higher and precipitation was 19% lower. This would have resulted in an environment 17 to 29% more arid than the present.

4.3.4 Composite Model

A variety of proxy data indicate that the climate on the northern Great Plains has changed considerably during the Late Pleistocene and Holocene (Fig. 4.2). Although different proxy climate indicators reflect change in different climate parameters, they
Fig. 4.2: Comparison of timing of Holocene moist and arid periods from studies in central Saskatchewan. A general picture of environmental change is presented based on dates and interpretations within each source. Suspect dates are discussed within the body of the text. Note that the category "Modern" includes both modern distributions and stated modern climates. Sources: Kuhry et al., 1992 (Peatlands); Last and Sauchyn, 1993 (Harris Lake); Last and Schweyen, 1985 (Waldsea Lake); Mott and Christiansen, 1981 (Martins Slough); Sauchyn, 1990 (Harris Lake); Sauchyn and Sauchyn, 1991 (Harris Lake); Schweyen and Last, 1983 (Waldsea Lake); Turchenek et al., 1974 (Batoche Paleosol, Saskatoon Site Paleosol, Beaver Creek Paleosol); Vance, 1986; Vance et al., 1995 (Composite Model); Yansa, 1995 (Andrews Site); 1996 (Andrews Site); Zoltai and Vitt, 1990 (Peatlands).
generally reveal similar broad trends. The invasion of a spruce forest ca. 10 ka suggests early post-glacial cool, moist conditions (Mott, 1973). The climate then became warmer and drier (Hypsithermal) as indicated by the invasion of grassland species (Anderson et al., 1989). Ritchie (1976) places the start of the Hypsithermal interval ca. 10 ka, although other investigations indicate that this estimation is too early. The shift to arid conditions was placed at ca. 8 ka BP by Turchenek (1974), based on paleosol studies of the Batoche terraces north of Saskatoon.

A reasonable picture of climate and environmental conditions on the prairies ca. 6 ka BP has been developed from pollen, macrofossil, lake levels, paleosols and tree line studies (Vance et al., 1995). These studies illustrate several broad trends. The Hypsithermal interval of peak aridity had passed by 6 ka BP but conditions remained warm and dry. The grassland - parkland boundary was located north of the modern day boundary. Lake levels were lower and reduced vegetation cover allowed increased aeolian activity. Subsequent to 6 ka BP, climatic conditions gradually shifted to a cooler and moister environment. This shift was time transgressive across the prairies occurring earlier in Alberta (ca 6 - 5 ka BP) than Manitoba (ca. 4 - 3 ka BP). Studies in Saskatchewan suggest moister climates were prevalent by ca. 4 ka BP.

Quantitative estimates of the climatic shift have been developed from pollen transfer functions and treeline studies. Despite contradictions and uncertainty inherent in the techniques, it is suggested that mean annual temperatures were 0.5 to 1.5 °C higher than current values and summer temperatures 0.5 to 3.0°C higher (Vance, 1986; Vance et al., 1995). The mean annual precipitation may have been reduced by as much as 65 mm
and the growing season precipitation by 50 mm. Peatland studies suggest conditions were between 10 and 29 % more arid than at present (Zoltai and Vitt, 1990).

Late Holocene climate changes are not as well documented in the current literature. Lake level studies east of Saskatoon reveal considerable fluctuation with shallow, saline conditions possibly corresponding to dry conditions prior to ca. 4 ka and 2.5 ka BP (Last and Schweyen, 1985). Higher levels, responding to moister conditions, occurred ca. 4 - 3 ka BP and ca. 2 ka BP.

Recently, the calibration and reinterpretation of the radiocarbon data base in Southeast Alberta has indicated that maximum landscape stability, corresponding to the period of maximum aridity, occurred much earlier than previous estimates ca. 12 - 10 ca. ka BP (Campbell and Campbell, 1997). This is followed by an increasingly moist period extending to after the deposition of Mazama ash ca. 7 ca ka BP. A return to arid conditions is suggested ca. 8 - 4 ca ka BP before a final shift to modern conditions beginning ca. 4 ca ka BP. Allowing for variation due to the use of calibrated versus non-calibrated dates, this represents a major departure from current thought. The calibration of radiocarbon from a wider area and the rejection of unreliable dates may change other chronologies.

4.4 Stream Channel Patterns

Channel pattern refers to the shape that is observed in plan form over long stretches of total stream length. There are five categories of channels: straight, sinuous, meandering, braided and anastomosing, as well as gradational patterns between each of these categories (Fig. 4.3). Straight, sinuous and meandering channels are usually
Figure 4.3: Alluvial channel patterns. A) bedload channels, B) mixed-load channels, c) suspended-load channels. Straight patterns are represented by 1, 6, 11; meandering patterns by 7, 8, 12, 13; braided patterns by 3, 4, 5, 9, 10; anastomosed patterns by 14 and sinuous patterns by 2. Source: Miall, 1992 after Schumm, 1981.

differentiated by their sinuosity index. This is defined as the ratio of stream length (measured along the channel centre) to valley length (measured along the centre of the valley) (Ritter, 1986). Braided and anastomosing channels are characterised by multiple channels.

Experimental flume studies and field observations indicate that channel pattern changes between straight, meandering and braided forms occur at critical threshold values of stream power, gradient and sediment load (Fig. 4.4; Schumm, 1981). Each of
Figure 4.4: Relationship between slope and sinuosity during experimental testing. Sediment load, stream power and velocity increase with slope. Source: Schumm, 1981, 24.

the fourteen channel patterns illustrated in Figure 4.3 may be related to these thresholds (Fig. 4.5). Within each grouping (bed-load channels, mixed-load channels, suspended-load channels) straight and low sinuosity channels are maintained until a critical threshold value of valley slope, stream power and/or sediment load is crossed (Schumm, 1979). Beyond the threshold, sinuosity increases through sinuous and meandering patterns. Continued increases in stream slope, power and load cause a second threshold to be crossed at which time the channel pattern switches to braided forms.

Streams that are close to threshold values of valley slope, stream power or sediment load will be more likely to experience changes in channel pattern. Small changes in controlling variables, such as discharge, within a stream basin may cause dramatic changes in both channel pattern and the ability of the channel to transport sediment (Schumm, 1981). Streams which are not close to threshold values will be less likely to shift between patterns. Channel patterns and their main controls will be
Figure 4.5: Relative effect of valley slope, stream power and/or sediment load on river pattern thresholds for: A) bed-load channels; B) mixed-load channels; and C) suspended-load channels. Numbers indicate the locations of the channel patterns illustrated in figure 4.2. Source: Schumm, 1981.

discussed in the remainder of this section. Geomorphic thresholds will be considered under the heading of river terraces (Chapter 4.6).

**Straight and Sinuous Channels**

Straight channels have a sinuosity index less than 1.05 (Morisawa, 1985). They generally consist of a single channel with alternate bars, pools, riffles and a meandering thalweg. The thalweg is the line of maximum water depth and velocity which connects pools in the channel. Riffles are shallow zones between pools which trap coarse sediment. Erosion leads to minor channel widening and incision. Streams rarely maintain a straight channel for long distances, usually shifting to sinuous and meandering forms.
Sinuous channels are an intermediate form between straight and meandering with a sinuosity index of 1.05 to 1.15.

**Meandering Channels**

Meandering channels also contain pools, riffles and bars, but have a much higher sinuosity index (>1.15) (Morisawa, 1985). Pools are located at bend apices with riffles between the pools. Point bars and riffles often merge together, eventually becoming indistinguishable (Reid and Frostick, 1994). A clay-rich suspended sediment load, localised bank erosion within generally cohesive sediments, a minimum gradient of 0.4% and helicoidal flow within the channel are necessary for meanders to form (Reid and Frostick, 1994; Summerfield, 1991). Stream flow erodes sediment from the outside of meander bends creating cutbanks, depositing it downstream on point bars. Channel migration resulting in meander cutoffs are distinctive features of this fluvial system. The dominant depositional forms are point bars which build up laterally across the channel and in a downstream direction.

**Braided Channels**

Braided streams have two or more channels, separated by bars and small islands. The bars and islands are not stable channel features and the individual channels and the thalweg shift continuously. Braided streams are usually shallow and wide with a sinuosity index greater than 1.3 (Morisawa, 1985). Individual channels within the stream complex often have higher sinuosity indexes (Leopold et al., 1964). Braiding has been attributed to many factors, the most important of which are: 1) easily eroded bank material enabling channel shifting; 2) large volumes of coarse sediment transported as bedload (>
11% bedload necessitating shallow, rapid flow); 3) rapid and frequent variations in discharge preventing vegetation growth on the bars and 4) a steep channel gradient (Schumm, 1981). Experimental and analytical studies have shown that the number of braids in a channel not only increase with stream power, but also with increasing rates of sediment transport (Knighton, 1984). Channel widening, particularly around bars and islands, is the dominant erosive action. Deposition includes channel aggradation and the formation of mid-channel bars and sandflats (Morisawa, 1985). Braided streams may be further subdivided into pebbly and sandy channels (Collinson, 1986). Often both will occur within a single stream, with pebbly forms in the upstream reaches, grading into sandy forms in the downstream reaches.

**Anastomosing Channels**

Anastomosing streams incorporate many features of both braided and meandering channels. Anastomosing stream systems are composed of

an interconnected network of low-gradient, relatively deep and narrow, straight to sinuous channels with stable banks composed of fine-grained sediment (silt/clay) and vegetation...separating the channels are floodplains consisting of vegetated islands, natural levees, and wetlands. (Smith and Smith, 1980, 157-158)

Unlike braided streams, bank stability, associated with cohesive fine-grained sediments, is necessary to slow channel migration and erosion of the vegetated islands. A rapidly aggrading system with a dominantly fine-grained sediment load and a low channel gradient are also critical for anastomosing to occur. Peat bogs, backswamps and floodplains predominate in an anastomosing system: channels, levees and crevasse splays cover a smaller area (10 - 40%) (Smith and Putnam, 1980). Anastomosing streams are
dominated by vertical accretion producing stringer-like bodies of coarse sediment rather than lateral accretion such as point bars (Smith and Smith, 1980).

It is unlikely that the study site valley would have been wide enough to accommodate an anastomosing system. This stream type will, therefore, not be considered further.

4.5 Fluvial Facies Models

A facies is a sediment unit with a distinctive group of characteristics including lithology, physical and biological structures, composition and/or texture within a unit of sediment or rock that differ as a group from those above or below (Hamblin, 1991). The sedimentary facies associated with meandering and braided streams, and their floodplains, will be discussed in the following section. These facies are important as they will form the basis for the interpretation of former fluvial environments within the study site. Seventeen distinct fluvial facies are commonly recognised (Table 4.3).

Meandering Stream Facies Model

The meandering stream facies model proposed by Walker and Cant (1984) indicates the spatial relationship between channel meanders, point bars and meander cutoffs (Fig. 4.6). The channel floor is represented by a lag deposit (Gm and Gt) as seen in Figure 4.7. Lateral accretion deposits are deposited over this channel lag by point bar migration. Trough and planar cross-bedding (St, Sp), created by migration of dunes on the floor of the channel, form low on the point bar. As the point bar migrates, the water will become shallower and sediment will fine upwards. Ripple cross-laminations (Sr) will be preserved in the sediment. Horizontal laminations (Sh) may be
Table 4.3: Classification of fluvial facies according to texture and internal sedimentary structure. Facies codes will be used throughout the text. Primary Sources: Miall, 1978B; 1992. Secondary Sources: Allen, 1985; Blatt et al., 1991; Collinson and Thompson, 1982; Leeder, 1982; Tucker, 1991; Visher, 1972; Walker and Cant, 1984.

<table>
<thead>
<tr>
<th>Facies Code</th>
<th>Facies</th>
<th>Sedimentary Structures</th>
<th>Interpretation and Explanatory Comments</th>
</tr>
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</table>
| Gms         | massive, matrix supported gravel | graded | debris flow deposits  
- indicated by simultaneous deposition of clasts and matrix |
| Gm          | massive or crudely bedded gravel | horizontal bedding, imbrication | longitudinal bars, channel lag deposits  
- imbrication produced by current in bedload transport |
| Gt          | gravel, stratified | trough cross-beds | minor channel fills  
- trough cross-beds produced by migration of dunes  
- water depth over the crest approximately equals the dune height |
| Gp          | gravel, stratified | planar cross-beds | longitudinal bars, deltaic growths from older bar remnants  
- planar cross-beds are produced from the downcurrent migration of dunes |
| St          | sand, medium to very coarse, may be pebbly | solitary or grouped trough cross-beds | dunes  
- individual cross-beds produced if there is no net sedimentation  
- sets of cross-beds produced under conditions of net sedimentation  
- generally deposited in deeper water, often near the base of point bars or cross-channel bars |
| Sp          | sand, medium to very coarse, may be pebbly | solitary or grouped planar cross-beds | bars and waves  
- deposited by straight-crested dunes |
| Sr          | sand, very coarse to fine | ripple cross-lamination | ripples  
- may indicate deposition in shallow water such as near the tops of bars  
- tabular and trough cross-laminations undifferentiated  
- rapid deposition may produce |
<table>
<thead>
<tr>
<th>Code</th>
<th>Description</th>
<th>Characteristic Features</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sh</td>
<td>sand, very fine to very coarse, may be pebbly</td>
<td>horizontal lamination produced by continuous movement of sediment - no apparent water turbulence - laminae deposited by a single event</td>
</tr>
<tr>
<td>Sl</td>
<td>sand, very fine to very coarse, may be pebbly</td>
<td>low angle (&lt; 10°) cross-beds scour fills, washed-out dunes, antidunes - low-angle cross-sets caused by sandwave migration, typically in braided streams</td>
</tr>
<tr>
<td>Se</td>
<td>erosion scours with intraclasts</td>
<td>crude cross-bedding scour fills</td>
</tr>
<tr>
<td>Ss</td>
<td>sand, fine to very coarse, may be pebbly</td>
<td>broad, shallow scour scour fills</td>
</tr>
<tr>
<td>Fl</td>
<td>sand, silt, mud deposits</td>
<td>fine lamination, very small ripples overbank or waning flood - indicated by fine grain size</td>
</tr>
<tr>
<td>Fsc</td>
<td>silt, mud</td>
<td>laminated to massive backswamp deposit - massive deposits indicate continual, steady deposition from suspension or later destruction of laminations - laminations differentiated on the basis of grain size, colour</td>
</tr>
<tr>
<td>Fcf</td>
<td>mud</td>
<td>massive, may contain fresh water molluscs backswamp pond deposits - steady deposition from suspension</td>
</tr>
<tr>
<td>Fm</td>
<td>mud, silt</td>
<td>massive, desiccation cracks overbank or drape deposits, drying after flood events - steady deposition from suspension</td>
</tr>
<tr>
<td>C</td>
<td>coal, carbonaceous mud</td>
<td>plant, mud films swamp deposits - may exhibit high organic contents</td>
</tr>
<tr>
<td>P</td>
<td>carbonate</td>
<td>pedogenic features paleosol</td>
</tr>
</tbody>
</table>
Figure 4.6: Block diagram indicating the morphological elements of a meandering river system. Source: Walker and Cant, 1984.

Interbedded with the main lateral accretion facies indicating higher flow regime events (Walker and Cant, 1984).

Coarser textured vertical accretion sediments (Fl) are deposited near the channel where floodwaters have a higher velocity (Fig. 4.7). Further from the channel, stagnant floodwaters will result in the deposition of fine, massive facies (Fsc, Fcf). Desiccation cracks and vegetation may be evident as the flood waters dry out resulting in Fm and C facies. Depending on the position of the subsequent (or current) channel, channel lag (Gm) or levéé facies (Sr) may be found on top of either the lateral accretion or vertical accretion facies. Levéé deposits would indicate the outside of a meander bend whereas channel lag and point bar deposits would indicate the inside of the bend.
An important component of the meandering stream model, facilitating differentiation from braided stream deposits, is the identification of meander cut-offs in the sediment record. There are two categories of cut-offs; chute cut-offs and neck cut-offs (Fig. 4.8; Walker and Cant, 1984). The facies sequence consists of channel lag (Gm), trough cross-bedding (St, Sp) and ripple cross lamination (Sr) capped with fine-grained, vertical accretion facies (Fl, Fsc, Fcf, Fm). In chute cutoffs, old swales are reoccupied and the main channel will gradually be abandoned (Walker and Cant, 1984). This slow abandonment is recorded by the deposition of ripple cross-laminated sand (Sr)
Figure 4.8: Abandonment of meander loops by chute cut-off and neck cut-off. ACT. - active river  AB. - gradual abandonment  V. A. - vertical-accretion deposits. Stippled area at neck cut-off indicates sudden abandonment and sealing off by deposition of sand plugs. Source: Walker and Cant, 1984.

reflecting a decrease in water discharge through the former channel. During neck cut-offs, the meander loop is quickly abandoned and sealed. Vertical accretion facies, deposited primarily during flood events, dominate the sequence.

Over time, flood deposits form natural levees which confine and gradually raise the stream above its floodplain. A catastrophic levee break may cause the river to avulse, forming a new channel (Walker and Cant, 1984). This will leave a distinctive pattern of long, lateral accretion sand facies bounded by fine-grained vertical accretion facies in the sediment record (Fig. 4.9).

**Braided Stream Facies Model**

The braided stream facies model is not as well developed as the model for meandering streams. Fewer studies have been conducted on ancient and modern braided fluvial systems. Separate facies models have been developed for sandy braided
Figure 4.9: A) Block diagram of flood plain aggradation of a meandering river system. Shoestring sands are surrounded by vertical accretion silt and clay. B) Block diagram of sandy braided system. Vertical accretion may occur during flood stage, but the deposits are rarely preserved. Source: Walker and Cant, 1984.

Sandy braided streams consist of four main elements; channels, bars, sand flats and vegetated islands (Cant and Walker, 1978). Sand Flats (Fig. 4.10, #1) generally develop out of emergent nuclei (higher areas which are exposed during low flow;
Figure 4.10: Block diagram showing elements of a braided sandy river. Stippled areas are exposed, all other areas are underwater. Source: Walker and Cant, 1984.

Fig. 4.10, #3) on cross-channel bars (Fig. 4.10, #4). Unlike meandering streams, three facies sequences are common, depending on the location within the stream.

The first facies sequence represents a sand flat depositional environment (Fig. 4.10A, 4.11). The base of the sequence is a channel lag facies (Gm) overlain by poorly-defined, then well-defined trough cross-bedding (Se, St) deposited by migrating dunes on the channel floor (Miall, 1992). Solitary planar cross-beds, 0.2 m - 5 m thick, indicate the formation of cross-channel bars (Sp) (Walker and Cant, 1984). Multiple planar cross-beds (Sp) are the next facies, formed by emerging nuclei during a series of flood and low level stages. The actual sand flat is indicated by sets of parallel laminations deposited during floods as well as thin trough cross-bedding (St) and ripple
cross lamination (Sr) related to decreasing water depth and flow velocity (Cant and Walker, 1978). These facies are overlain by thin vertical accretion facies (Fl, Fm, Fsc).

Channel depositional environments (Fig. 4.10 C, 4.11) preserve a different facies sequence. The bottom channel lag and trough cross-bed facies (Gm, Se, St) are similar to the sand flat facies sequence. Unlike the sandflat sequence, however, large trough cross-bedding continues throughout much of the sequence. Cross-channel bars, evident in thick solitary planar cross-beds (Sp), may or may not occur in these sequences. Near the top of the sequence, the trough cross-bedding becomes thinner (Cant and Walker, 1978). Planar cross-beds and ripple cross-laminations (Sp, Sr) indicate shallowing water and a reduction in flow velocity as the channel fills. The top of the sequence may include
massive and laminated vertical accretion facies (Fl, Fsc, Fcf, Fm) deposited as the channel is abandoned.

The third sequence (Fig. 4.10 B, 4.11) describes a depositional environment with both sandflat and channel influences. In this particular example, a sand flat started to develop, but was subsequently scoured by a minor channel. The sand flat then re-established itself, continuing to the top of the sequence.

Gravelly, braided streams are common in modern paraglacial and alpine environments (Rust and Koster, 1984). The term paraglacial refers to processes that are "directly conditioned by glaciation . . . both proglacial processes and those occurring around and within the margins of a former glacier that are the direct result of the earlier presence of ice" (Church and Ryder, 1972). In proximal settings, horizontally bedded, imbricated gravels are the most common facies (Gm) (Fig. 4.12). These facies may be massive where beds are thick and the texture of the gravel is uniform. Clast-supported gravel suggests deposition by a high-energy aqueous flow that keeps the finer particles in suspension while stratified sand matrix-supported gravels are deposited by low energy aqueous transport (Rust and Koster, 1984). Planar cross-stratified gravels (Gp) may also occur but are found much less often (Rust and Koster, 1984). Fine-grained sediments (sand and/or clay drapes) are deposited from suspension during waning floods. Scour surfaces within the facies sequence represent fluctuations in discharge.

Distal gravelly braided deposits are different from those described above (Fig. 4.13). The bottom of the sequence consists of trough cross-stratified, clast-supported gravel above an erosional base (Gt) (Rust and Koster, 1984). Multiple sets of this facies
Figure 4.12: Depositional models for proximal gravelly braided river deposits. A) Modern deposition: facies Gm is strongly dominant due to shallow flow in a semiarid or paraglacial environment. B) Ancient deposition: Gm is still dominant, but Gp is more prominent because the bar/channel relief is greater. Source: Rust and Koster, 1984.

Fine upwards to horizontally-bedded gravels (Gm) and finally trough cross-stratified sands (St). This occurs in response to gradually shallowing water and a decrease in flow velocity over bars and in the channels. Migration of active channels is probably important in this process. Fine-grained muds (Fm) and organic-rich sediments are deposited as sections become inactive.

Both fining-upward and coarsening-upward units may be present in the sediment record. Fining-upward sequences are attributed to channel abandonment or, when coupled with low-angle cross-bedding, to lateral accretion of sediment (Collinson,
Coarsening-upward units may result from the gradual re-activation of an individual channel or the progradation of a mid-channel or lateral bar.

The differentiation of sedimentary facies based primarily on texture and sedimentary structures is not without complications. Prior to the 1960’s, fine-grained deposits that did not have recognisable channel forms were classified as lacustrine deposits (Collinson, 1986). More recently, it has been recognised that fine-grained fluvial deposits may have a massive structure. Unfortunately, the increasing knowledge of the complexity of fluvial systems “makes the application of simple models less and less satisfactory and leads to uncertainty and ambiguity in interpreting...ancient” fluvial
systems (Collinson, 1986, p 20). For example, it is now known that the distinctive architecture of a point bar deposit may be found in braided streams not just meandering streams as originally postulated (Miall, 1992).

In summary, the strongest evidence for meandering stream patterns is the preservation of lateral accretion, fining-upward deposits (Fig. 4.9). As a general rule, meandering stream deposits are finer in texture than braided stream deposits (Collinson, 1986). Fine-grained vertical accretion deposits in braided systems are rare; they are deposited infrequently and are rarely preserved (Walker and Cant, 1984). The lack of bounding clay deposits results in channels that are much wider than those found in meandering systems. An additional factor which may be used to distinguish between the two patterns is the identification of cross-channel bars in braided stream facies (Cant and Walker, 1978).

4.6 Terrace Formation

River terraces are remnants of abandoned floodplains preserved as benches along valley walls. They are composed of two distinct sections; a steeply sloping riser which faces the river and a generally flat tread. Terraces are produced by channel aggradation or lateral erosion followed by vertical erosion (incision), as the river system adjusts to changes in channel gradient, stream discharge or sediment load.

River terraces have been examined for a variety of purposes such as: 1) the reconstruction of longitudinal profiles for paleohydrological reconstructions; 2) the reconstruction of the geomorphic history of drainage basins; 3) the establishment of temporal stratigraphic correlation; 4) evidence of recent tectonism; and 5) the
interpretation of the effects of climate and sediment load changes on fluvial systems (Germanoski and Harvey, 1993).

Two broad classifications of terraces are used alone or in combination. The first classification system is based on the material underlying the terraces. Strath, or rock-defended, terraces are cut into bedrock. In contrast, alluvial or valley-plain terraces are composed entirely of alluvial material (Leopold et al., 1964). In the second classification system, terraces are described as either erosional or depositional. Erosional terraces have a generally uniform layer of alluvium deposited on top an erosional surface (Fig. 4.14; Morisawa, 1985). Alluvial deposition and erosion of the underlying surface occur concurrently, such as during point bar formation and meander migration. The terrace scarp is formed by lateral erosion. In contrast, depositional terraces have an irregular underlying surface resulting in varying thicknesses of alluvium. This suggests that alluvial deposition occurs after the underlying surface has been eroded.

In central Saskatchewan, bedrock surfaces are deeply buried by glacial tills and surficial sediments so that strath terraces are not expected to have formed during the Holocene. For this reason, the following discussion will focus exclusively on alluvial terraces.

Repeated cycles of alluvial deposition and incision may occur within a valley. Sediments deposited subsequent to fluvial incision will either be inset within the trench cut by the previous period of incision or be of sufficient volume to overflow the trench forming overlapping deposits (Fig. 4.15) (Leopold et al., 1964). Inset fills are a product
Figure 4.14: Erosional terraces with a flat, underlying surface formed by concurrent erosion and deposition. Depositional terraces with irregular underlying surfaces formed by deposition after erosion of the underlying surfaces. Adapted from Morisawa, 1985.

Figure 4.15: Comparison of inset and overlapping valley fills. Adapted from Leopold et al., 1964.
of a long-term lowering of stream level whereas overlapping fills cause a gradual build up of sediment burying former terraces. Terraces formed during each cycle of channel aggradation and incision may be evident in the case of inset fills. Richards (1982) notes that different gradients along the course of a river may cause oscillations between inset and overlapping fills making it difficult to trace individual terrace surfaces. Similar surface configurations of terraces may be produced regardless of the number of cycles of incision and degradation (Fig. 4.16).

Terraces formed by rapid incision are often paired with surfaces preserved at the same elevation on both sides of the valley. These may be used as evidence of sudden and intermittent changes in the fluvial system. Unpaired terraces do not have matching surfaces on opposing sides of the valley. They are formed by slow gradual incision that has a significant lateral component and do not necessarily indicate changes in environments or the fluvial system (Rains and Welch, 1988; Summerfield, 1991). This type of incision may occur as the stream system attempts to reach base level.

**Channel Aggradation and Incision**

It has been proposed that channel erosion and aggradation are episodic; periods of stability are separated by unstable periods of adjustments. Instability and adjustment to a new equilibrium form occur when geomorphic thresholds are crossed (Schumm et al., 1984). A geomorphic threshold may be defined as a threshold, extrinsic or intrinsic, of landform stability. Extrinsic thresholds relate to forces or processes external to the river channel such as climatic change, changes in basin runoff or tectonic movements (Summerfield, 1991). In these cases, an external mechanism is necessary for the stream
Figure 4.16: Examples of valley cross sections showing some possible stratigraphic relations in valley alluvium. Note that within each terrace configuration (A, B and C) one or more alluvial fills result in the same surficial form. Source: Leopold et al., 1964, 460.

to cross a threshold. Intrinsic thresholds are crossed without any change in external conditions such as "natural channel shift, steepening of the valley floor through deposition and natural meander cut-offs" (Schumm et al., 1984). External changes are not necessary for a threshold to be crossed; erosional and depositional changes are viewed as inherent in a fluvial system.

The complex responses exhibited by many fluvial systems may be a product of a combination of extrinsic and intrinsic controls. Generally, a complex response occurs in the final phase of episodic erosion, as the system approaches a new equilibrium (Schumm, 1977; Schumm and Brakenridge, 1987). Once the initial environmental change has occurred, oscillations between aggradation and incision may be initiated by crossing intrinsic thresholds without further changes in the environment (Schumm and Parker,
For example, an initial fall in base level causes channel incision and an increase in channel gradient which moves progressively upstream through knickpoint retreat. The sediment supply to downstream reaches is increased causing channel aggradation. The upstream progression of channel aggradation reduces the channel gradient and hence the sediment supply to downstream reaches. The reduced sediment supply initiates a second cycle with renewed incision in the downstream reaches. This type of response to an environmental change means that unequal numbers of terraces may form in upstream and downstream reaches, making correlation difficult (Germanoski and Harvey, 1993).

In many situations it may be difficult to distinguish between terraces formed due to extrinsic controls and those due to intrinsic controls. Womack and Schumm (1977), in their classic Douglas Creek study, suggest that intrinsic controls be used to explain terraces formed too quickly to be explained by "base-level, climatic and land-use changes" (Womack and Schumm, 1977, p 72). The logical extension of this idea is that terraces formed when there is no detectable tectonic, isostatic or climatic change may be attributed to intrinsic controls. Intrinsic controls could also be useful in explaining differences in terrace formation between small, local drainage basins.

Channel aggradation will occur when the production of debris exceeds the amount of debris that can be removed through transportation (Leopold et al., 1964). The capacity of a stream to carry sediment may be exceeded for a variety of reasons including increased sediment production, decreased discharge and a reduction in channel gradient. For example, active gully incision in up-stream reaches may cause aggradation in the lower reaches of a small stream channel (Schumm, 1981). Specific factors leading to
variations in sediment production, discharge and channel gradient will be discussed below.

Incision generally occurs through the headward migration of a knickpoint. Transportation of sediment and flattening of the slope are at a maximum at the knickpoint (Leopold et al., 1964). A stream may incise for a variety of reasons, the most common of which are outlined in Table 4.4 (Schumm et al., 1984). Factors decreasing the erosional resistance of channel sediments increase susceptibility to headward erosion. These factors are significant in the initiation of gullies and channels. Factors increasing the power of a stream increase both the competency and capacity of the stream.

Table 4.4: Mechanisms for channel incision including decreased erosional resistance and increased erosional forces. Source: Schumm et al., 1984, 12.

<table>
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<tr>
<th>Decreased erosional resistance</th>
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<tr>
<td>i) decreased vegetation cover.</td>
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<tr>
<td>ii) surface disturbance causing decreased sediment/soil permeability and cohesion.</td>
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<tr>
<th>Increased erosional forces</th>
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<tr>
<td>i) constriction of flow such as by vegetation.</td>
</tr>
<tr>
<td>ii) steepening of gradient and energy slope by deposition of sediment, channelisation and meander cut-offs, and base-level lowering</td>
</tr>
<tr>
<td>iii) increased discharge and flood peaks.</td>
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<tr>
<td>iv) decrease of sediment load.</td>
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Climatic changes may lead to periods of aggradation and incision due to fluctuations in discharge and sediment load. The relationship is complex as discharge and sediment loads are responding to more than variations in precipitation. Other factors which may be significant are

"alterations in the density and type of vegetation cover, changes in the rates of evapotranspiration related both to temperature and vegetation characteristics, changes in the intensity of precipitation and its distribution throughout the year and changes in . . . the contribution of snow-melt to runoff." (Summerfield, 1991, p 359)

There is a lag between a climate change and the formation of terraces and other landforms as the quantity and distribution of precipitation and vegetation cover are altered followed by a shift in the basin discharge and sediment load. Finally, there will be a period of adjustment as the river and channel reach a new equilibrium, creating the landforms.

The relationship between climate change and channel aggradation and incision is complex, depending in part on the effective precipitation (precipitation minus evapotranspiration) in a region. In arid to semiarid regions with limited vegetation, an increase in precipitation and runoff would increase the sediment load of a stream leading to aggradation (Langbein and Schumm, 1958). This relationship between environment change and the stream profile continues until a threshold is reached. Any further increase in precipitation results in an increase in the vegetation cover, thus protecting the basin from erosion (Fig. 4.17; Knox, 1983).

In more humid regions an increase in precipitation would be expected to increase the vegetation cover, decreasing sediment yield. Increased precipitation would also
Figure 4.17: The relationship between mean annual sediment yield and mean annual effective precipitation where the mean annual temperature is 10°C. Source: Knox, 1983.

Promote incision through an increase in stream discharge resulting in an increase in stream capacity and competency (Williams et al., 1993). Dry periods lead to aggradation as vegetation dies back, increasing slope wash and the sediment load carried by the stream. The reduction in precipitation also causes stream discharge, and therefore stream capacity and competency, to decrease.

Changes in climate represent only one type of threshold which may affect a fluvial system. Fluctuations in base level are another important external geomorphic threshold (Schumm and Brakenridge, 1987). Base level is the elevation at which the mouth of the channel joins with a major body of water. A rise or drop in base level will result in changes to the channel gradient initiating incision or aggradation. The specific response of a stream depends on many factors including the rate, amount and direction of change in base level, river character and the rate of sediment supply. Many natural systems will
adjust to smaller variations in base level by changes to the channel pattern (Chapter 4.3, Figures 4.4 and 4.5) as well as width, depth and roughness. Slow rates of change in base level may also be accommodated in this fashion (Schumm, 1993). Greater amounts of base level lowering will increase the gradient of a stream leading to incision. The increase in stream gradient will gradually progress upstream by knickpoint retreat. A rise in base level will reduce the stream gradient causing aggradation in the lower reaches of the channel that gradually propagate upstream. The morphology of the valley containing the stream is another important variable. Streams flowing in wide valleys or on plains will be able to increase in sinuosity in response to lowering whereas a confined stream would be more likely to incise.

In most cases, changes in the ultimate base level would not be expected to extend throughout an entire drainage basin. Short term changes in local base levels by processes such as river capture, isostatic rebound, local tectonics and disturbances in drainage patterns such as fluctuating levels in ice dammed lakes due to glacier retreats, would be expected to have a far greater impact on terrace formation at the local scale (Richards, 1982).

A critical component of any terrace study is the mapping, correlation and interpretation of terrace surfaces. Correlations will be most reliable where a former floodplain was continuous throughout a valley and subsequent incision limited in nature, producing numerous terraces at a uniform height throughout the valley. This ideal set of circumstances is, however, the exception rather than the rule.
There are many sources of error in terrace studies, particularly where surface form is used without an investigation of the subsurface stratigraphy (Frye and Leonard, 1954). Error may occur in the initial stages of an investigation by the misidentification of terrace surfaces such as mapping terraces that don’t actually exist. Landslide deposits near valley walls may result in a scarp completely unrelated to terrace formation. Scarps and treads formed by differential erosion of different valley fills after abandonment of a floodplain may be misidentified as being formed by separate incision events (Frye and Leonard, 1954).

Significant error may be encountered in the correlation of terraces by altitude matching (Leopold et al., 1964). This would be most likely to occur at sites with numerous terraces or where the number of terraces changes along the length of a particular valley. Terraces formed due to the complex response mechanism will not be correlated throughout the length of the valley as aggradation and incision may occur simultaneously in different reaches of the stream (Germanoski and Harvey, 1993). Even if these problems did not exist, there is considerable scope for error in the determination of terrace elevations. Terraces commonly slope toward the channel that formed them. In some cases, this slope will be steep enough for the difference between the front and back terrace height to exceed the height of the terrace above the floodplain. This is often accentuated by erosion and mass movements depositing material on the terrace from higher up the valley wall. The colluvial material may become interfingered with alluvium during aggradation as illustrated in Figure 4.18. In some cases, colluvium may actually be dominate within the terrace sediment. Dissection of the terrace surface by
Figure 4.18: Cross section of valley fill illustrating the inter-fingering of alluvium and colluvium at valley margins. Source: Leopold et al., 1964.

small channels and the diversity of the initial floodplain, such as meander scrolls, cut-offs and levees, may also be a problem (Cotton, 1940). Leopold et al. (1964) suggest several stratigraphic and physiographic criteria for the correlation of terraces including:

- stratigraphic discontinuities between terrace fills, differences in particle size and sorting and in sedimentary structures, fossil fauna and flora, artefacts ranging from Indian pottery to tobacco tins, buried soils or paleosols, and frost features...(and) physiographic relations to other landforms. (Leopold et al., 1964, p 467).

Differential terrace preservation, and terraces responding to different scales of change are an additional complication. The preservation of terraces varies depending on two characteristics of the fluvial system: 1) the magnitude and frequency of flood events and 2) the relative importance of colluvial versus fluvial processes (Green and McGregor, 1987). Short-term, high magnitude stream floods will have a greater impact on smaller basins than larger basins. The relative magnitude of change will be greater in small basins so that geomorphic thresholds will be crossed more frequently. Long-term preservation of terraces is less likely to occur in these small basins as many localised, short-term terraces are formed. The second factor involves the interaction of terrace formation and
destruction. Higher order streams and downstream reaches generally have wider floodplains. Slope processes acting to destroy terraces are confined to the valley walls so there is a greater potential for the preservation of the wider, downstream terraces. These factors combine to produce a complex, long-term terrace record in higher order stream valleys and local, short-term terraces in smaller, low order stream valleys.

4.7 Terracette Formation

Terracettes have been referred to as an "enigmatic phenomena" in geomorphology (Vincent and Clarke, 1976). Throughout the twentieth century, many different theories as to how they formed have been postulated. Terracettes are networks of shallow, grass-covered steps, which usually form on steep slopes (Summerfield, 1991). Anderson (1972) proposed a classification with two broad categories; normal terracettes and tear terracettes. Normal terracettes form on slopes less than 30° in areas of permanent pasture and mineral soils (Anderson, 1972; Selby, 1982). They have long, wide, low angle treads that are generally parallel to each other and normal to the slope. The risers have a continuous vegetation cover and exhibit slopes similar to the hillslope angle. Tear terracettes have narrow and slightly steeper treads that are short and rarely parallel. The risers are bare and are much steeper and narrower than those of normal terracettes. Tear terracettes are more likely to form on slopes greater than 30° (Anderson, 1972; Selby, 1982).

A third category, grazing-step terracettes, has been proposed by Higgins (1982). The treads are narrow, often bare of vegetation and have a very gentle slope, approaching
In contrast, the risers are generally vegetated. These terracettes may form very rapidly, possibly in a few months (Parsons, 1988).

A few of the more plausible mechanisms of terrace formation suggested over the years are briefly outlined in the remainder of this chapter. Terracette formation has yet to be resolved satisfactorily in the literature. This situation suggests that a variety of mechanisms may lead to their formation.

Theories based on solifluction and/or gelifluction work under the basic assumptions that terracettes have been observed in regions currently experiencing gelifluction and terracettes located in regions not currently experiencing gelifluction are fossil features. Some theories propose that creep, involving the faster movement of vegetation and the upper A horizon of soils, may be the dominant mechanism. Alternatively, small cracks in the ground may be enlarged by frost action. The separated sections would then be tilted by soil creep, forming both the riser and tread of the terracette.

It has been noted that “many of the structural features of mass creep (are) analogous to those found in terracettes” (Vincent and Clarke, 1976, p 69). This has led to suggestions that variations in the rate of soil creep is the dominant process in the formation of terracettes (Carson and Kirkby, 1972; Davies, 1969, Summerfield, 1991). Terracettes have been observed in locations where it is unlikely that they could have formed via solifluction.

Some researchers have found terracette formation to be correlated with the thickness of soil and debris on the hillslope. Where the soil is shallow, the stabilising
effect of vegetation will be increased. Large scale mass movements such as landslides will not occur, instead small scale movements will dominate leading to the formation of terracettes (Carson and Kirkby, 1972). Poorly consolidated material also promoted the formation of terracettes. Scheidegger (1984) suggests that variations in the thickness of material may cause the movement of material to be impeded. The resulting humps would then develop into terracettes. Since the 1920’s, it has been suggested that terracettes may help to stabilise slopes thus preventing further movement downslope. During the 1960’s, research indicated that slopes up to 43° were stabilised by terracettes (Vincent and Clarke, 1976).

For many years, the formation of terracettes by animal treading has been a popular theory. The grazing-step terracettes described by Higgins (1982) were probably initiated by the movements of animals. Many theories view animal disturbances as accelerating and enhancing other processes instead of as an independent mechanism (Young and Saunders, 1986). Selby (1982) states that “animal treading is almost certainly primarily responsible for continuous terracette systems which may have been built upon natural forms developed beneath forest and by irregular soil creep”. He further hypothesises that once the terracettes have begun to form, they will receive greater inputs of animal excrement, which promotes plant growth. This would help to stabilise the edges of the terracettes. Still other researchers use the regularity of terracettes to discredit theories of formation involving animals (Parsons, 1988).
Chapter 5
Sedimentary Facies and Depositional Environments
Within Wanuskewin Heritage Park

5.1 Sedimentary Facies

A facies is defined as "a distinctive group of characteristics within part of a rock body (such as composition, grain size, or fossil assemblages) that differ as a group from those found elsewhere in the same rock unit" (Hamblin, 1991, p 359). These characteristics result from the nature of sedimentation.

Several criteria were used in the development of five sedimentary facies within the study site, the most important of which are grain size distribution, organic carbon content and colour. These physical characteristics are readily assessed in field situations. The grain size classification developed by Folk (1954) was used to classify the texture of sediment samples. The shape of cumulative frequency curves and statistical parameters, including mean grain size and standard deviation (particle sorting), are also used in defining the facies although these elements are generally viewed as more important in the interpretation of the facies with respect to the reconstruction of depositional environments. Sedimentary structures were not used as the observed sediment units were massive.
There are two primary methods of investigating the shape of grain size distribution curves. The first assumes the shape of the curve to be influenced by the importance of the processes of traction, saltation and suspension in the transport and deposition of sediment. According to this method, each curve consists of two or more straight line segments which represent the different modes of sediment transport (Fisher, 1969; Glaister and Nelson, 1974). The number of segments, the slope of the segments and position of intercepts vary systematically with the relative importance of each process so that different depositional environments yield a different shaped curve (Fig. 5.1).

The braided stream deposit exhibits a large traction, or bedload, component indicated by the lowermost line segment. The aeolian sediments are dominated by a saltation load component. The slope of the line segments indicates that the dune deposits are better sorted than the braided stream deposits. The dune deposit does not contain a traction load segment as wind is not an effective way to transport gravel.

![Diagram of probability curves](image)

**Figure 5.1:** Typical probability curves: A) braided stream deposit; B) dune deposit. Note variation in number, length and slope of line segments. Adapted from Glaister and Nelson, 1974.
The second method considers the shape of the curve to be affected by the relative proportions of one or more populations of particles exhibiting a log-normal grain size distribution. These include a gravel population (mean -3.5φ to -2φ, standard deviation 0.7φ to 1.5φ), a sand population, (mean 1.5φ to 4φ, standard deviation 0.4φ to 1.0φ), and a clay population (mean 7.0φ to 9.0φ, standard deviation 2.0φ to 3.0φ) (Spencer, 1963). Sediment with a unimodal grain size distribution will result in a near-normal frequency curve. In a bimodal distribution, the shape of the curve depends on the proportion of each grain population present within the sample. The greater the proportion of a population, the closer the curve is to an idealised curve for that population (Fig. 5.2).

Figure 5.2: Theoretical probability curves for three sediments composed of different proportions of sand and clay populations: A) 90% sand and 10% clay; B) 50% sand and 50% clay; C) 10% sand and 90% clay. Frequency curves of a pure sand and a pure clay are also shown. Adapted from Spencer, 1963.
These two methods are not contradictory as the gravel population may be seen as corresponding to the traction load, the sand population to the saltation load and the clay population to the suspension load. It should be noted, however, that the suspension load will generally include fine sands, as well as the silt and clay fractions of a sediment. In each case, it is the relationship between the different particle size classes that is of greatest importance not the specific particle size. These studies are useful as they demonstrate that sediment samples with similar shaped curves are more likely to have been deposited in environments with similar energy than are sediment units with different shaped curves. Grain size envelopes, constructed by plotting the end members of the group of curves, are used to aid the interpretation of samples from unknown environments (Fig. 5.3).

The statistical parameters of mean and standard deviation may be compared to aid in the identification of depositional environments (Folk, 1966; Folk and Ward, 1957). Generally, coarser sediments are found in high-energy environments and finer sediments are found in low-energy environments. Standard deviation provides a quantitative measure of the degree of sorting within a sediment sample. Sorting also provides an indication of the energy within an environment with better-sorted sediments occurring in higher-energy environments. Of the sediments present within the study site, till deposits are expected to be poorly sorted and aeolian deposits to be well sorted. Researchers have failed to agree on a meaningful geological interpretation for skewness and kurtosis so these statistical parameters are largely ignored in the published research literature.
Figure 5.3: Grain size envelope for a theoretical ‘sand’ facies constructed by plotting the end members of the facies.

Studies based on the interpretation of depositional environments from statistical parameters have met with only limited success. There are several possible reasons for this. The texture of the parent material will affect the texture of the sediment deposit under consideration as a coarser parent material will yield a coarser deposit. Proximity to a sediment source is another important factor. Sediments which have been transported over greater distances will tend to be finer-grained and better sorted than sediments transported over shorter distances. Problems with interpretations may arise if sediments from different sites are compared without knowing how far each has been transported.
Post-depositional alteration of ancient sediments introduces further error into any interpretations. In a study of the reliability of environmental identifications using cumulative weight percentage and moment measures, it was found that the probability of error increases with an increase in the number of possible depositional environments (Tucker and Voucher, 1980). When used in conjunction with other parameters such as sedimentary structures, bioturbation and position within a landscape, the examination of statistical parameters does, however, facilitate the interpretation of depositional environments.

5.2 Facies Descriptions

Field observations and laboratory analyses indicate the presence of five sedimentary facies within the study site, two of which are subdivided into organic-rich and organic-poor sub-facies. Grain size envelopes have been plotted for the five sedimentary facies within the study site (Fig. 5.4). There is considerable overlap in the grain size envelopes, however, the degree of sorting serves to distinguish the various facies. Facies 1 is fine-grained and very poorly sorted, as demonstrated by the gentle slope of the envelope lines. Facies 2 and 3 exhibit the greatest degree of overlap. These facies are distinguished on the basis of gravel content, with facies 2 containing less than 5% gravel and facies 3 containing more than 5% gravel. This gives facies 2 a generally finer texture. Facies 5 sediments contain a very high percentage of gravel, setting them apart from the rest of the facies. Likewise, the well-sorted nature of facies 4 is distinctive.
Figure 5.4: Grain size envelopes for the five sedimentary facies identified for this study site.

A summary of average textural composition, mean grain size and sorting, as well as organic and inorganic carbon contents, of the various lithofacies is presented in Table 5.1. This table highlights the variations between individual facies and sub-facies.

5.2.1 Facies One

Facies one, characterised by the absence of gravel, is divided into two subfacies differentiated on the basis of texture and organic carbon content (Fig. 5.5). Sub-facies 1A in finer-grained than sub-facies 1B. Sub-facies 1A exhibits diverse textures ranging
Table 5.1: Average textural composition, mean grain size, sorting, organic carbon and inorganic carbon for each sedimentary facies within the study site. Total measured thickness refers to the total combined unit thickness of each facies in the analysed sections, cores and trenches.

<table>
<thead>
<tr>
<th></th>
<th>Facies 1</th>
<th>Facies 2</th>
<th>Facies 3</th>
<th>Facies 4</th>
<th>Facies 5</th>
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<td>A</td>
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<td>22.66</td>
<td>16.73</td>
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<td>3.22</td>
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<td>8.74</td>
</tr>
<tr>
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<td>1.02</td>
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<td>12.68</td>
<td>3.635</td>
</tr>
<tr>
<td>Thickness (m)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Figure 5.5: Bivariate plot of mean grain size versus standard deviation for facies 1A and facies 1B demonstrating general textural differences between the two facies. Note that facies 1B tends to be coarser-grained than facies 1A.
from clay through mud, silt, sandy mud and sandy silt to slightly gravelly mud as defined by Folk (1954). The structure of the facies is massive. Organic carbon content varies from 4.36% - 22.7% with an average of 7.84% and inorganic carbon content varies from 2.44% - 13.91% with an average of 10.16%. The thickness of the sampled units ranges from 0.04 m to 0.78 m with an average thickness of 0.27 m and a total measured thickness of 2.06 m. Dry Munsell colour ranges from black, dark brown, very dark grey, very dark greyish brown, dark olive brown, olive brown to light olive brown.

Sub-facies 1B is composed of sediments that exhibit muddy sand and sandy silt textures as defined by Folk (1954). The structure of the facies is massive. Organic carbon content varies from 1.94% - 4.26% with an average of 3.11% and inorganic carbon content varies from 1.24% - 12.94% with an average of 6.64%. The thickness of the sampled units ranges from 0.09 m to 0.68 m with an average thickness of 0.26 m and a total measured thickness of 1.02 m. Dry Munsell colour ranges from very dark grey, dark olive brown to olive brown.

5.2.2 Facies Two

Facies two, characterised by less than 5 percent gravel, also contains two sub-facies differentiated primarily on the basis of organic carbon content (Fig. 5.6, Fig. 5.7). Although there is considerable overlap in texture, facies 2B is generally coarser and more poorly sorted than facies 2A.

Sub-facies 2A is composed of sediments that exhibit slightly gravelly, sandy mud and slightly gravelly muddy sand textures as defined by Folk (1954). The structure of the facies is massive. Organic carbon content varies from 3.48% - 15.19% with an
Figure 5.6: Bivariate plot of mean grain size versus standard deviation for facies 2A and facies 2B demonstrating general textural characteristics of the two facies. Facies 2B tends toward a coarser mean grain size and has a greater range of sorting than facies 2A.

average of 5.23% and inorganic carbon content varies from 0.84% - 26.76% with an average of 5.27%. The thickness of the sampled units ranges from 0.01 m to 0.42 m with an average thickness of 0.14 m and a total measured thickness of 4.19 m. Dry Munsell colour ranges from black, very dark brown, dark brown, very dark greyish brown to very dark grey.

Sub-facies 2B is composed of sediments that exhibit slightly gravelly sandy mud and slightly gravelly muddy sand textures as defined by Folk (1954). The structure of the facies is massive. Organic carbon content varies from 0.58% - 3.35% with an average of 1.9% and inorganic carbon content varies from 0.08% - 33.49% with an average of 11.04%. The thickness of the sampled units ranges from 0.01 m to 1.58 m with an
Figure 5.7: Examples of sedimentary facies found within the study site. The uppermost unit is organic-rich facies 2A. This is underlain by 0.05 m of organic-poor facies 2B. The lower unit is coarser, classified as facies 3.
average thickness of 0.33 m and a total measured thickness of 12.68 m. Dry Munsell colour ranges from dark brown, brown, olive grey, olive brown, light olive brown, light brownish grey to light grey.

5.2.3 Facies Three

Facies three is composed of sediments that exhibit gravelly mud and gravelly muddy sand textures as defined by Folk (1954). Facies 3 is composed of coarser sediments than facies 2 (Fig. 5.7, Fig. 5.8). The structure of the facies is massive. Organic carbon content varies from 0.04% - 5.79% with an average of 1.83% and inorganic carbon content varies from 1.19% - 24.73% with an average of 8.74%. The thickness of the sampled units ranges from 0.02 m to 0.45 m with an average thickness of 0.18 m and a total measured thickness of 3.64 m. Dry Munsell colour ranges from brown, olive brown, light yellowish brown to light olive brown.

5.2.4 Facies Four

Facies four is composed of sediments that exhibit sand, muddy sand, slightly gravelly sand and slightly gravelly muddy sand textures as defined by Folk (1954) (Fig. 5.9). The structure of the sand facies is massive. Organic carbon content varies from 0.38% - 1.36% with an average of 0.7% and inorganic carbon content varies from 0.93% - 10.45% with an average of 6.3%. The thickness of the sampled units ranges from 0.04 m to 0.84 m with an average thickness of 0.19 m and a total measured thickness of 2.06 m. Dry Munsell colour ranges from brown, light olive brown to light yellowish brown.
5.1.5 Facies Five

Facies five is composed of sediments that exhibit muddy, sandy gravel and gravelly muddy sand textures as defined by Folk (1954) (Fig. 5.9). The structure of the facies is massive. Organic carbon content varies from 0.52% - 0.92% with an average of 0.76% and inorganic carbon content varies from 20.92% - 32.81% with an average of 28.47%. The thickness of the sampled units ranges from 0.09 m to 0.3 m with an average thickness of 0.13 m and a total measured thickness of 1.13 m. Dry Munsell colour ranges from brown, olive brown, to light olive brown.
Figure 5.9: Examples of sedimentary facies found within the study site. The upper and lower units are facies 5 gravel separated by facies 4 sand.
5.3 Facies Interpretations and Depositional Environments

There are four depositional environments represented within the study site; glacial, fluvial, colluvial and aeolian. Sediment samples are assigned to each of these environments on the basis of facies characteristics and stratigraphic position within the valley. The shape of grain size distribution curves, mean grain size and particle sorting are used to refine the interpretation of depositional environments. The facies defined in the preceding section are not confined to a single depositional environment; a particular facies may have two or more possible depositional environments. The massive structure of the sediment units observed complicates the interpretation of depositional environments.

Facies 1 is found entirely at valley bottom locations making a strong argument for a fluvial origin. The texture and structure is consistent with the vertical accretion facies described by Miall (1978B, 1992) (Chapter 4.5). The high proportion of silt and clay and the absence of gravel within this facies indicates that these sediments were associated with deposition within meander cut-offs and/or overbank sedimentation during flood events. The organic-rich sediments of facies 1A indicates deposition on a floodplain followed by stability and soil formation. The sediments would then be buried by subsequent flood events. The organic-poor units of facies 1B may have been deposited in either a meander cut-off or on a floodplain. The wide area over which these sediments are found suggests a floodplain location except where surface depressions indicate meander scars. The increased percentage of sand within this subfacies suggest deposition closer to the stream channel than the facies 1A sediments. Cultural layers found within
many facies 1A and facies 1B units indicates subaerial exposure and therefore a floodplain location.

Facies 2A and 2B sediments are found in both valley bottom and slope locations, suggesting two or more depositional environments. The position of the units within the landscape is important in determining depositional environments as there is no separation on the basis of structure, the shape of cumulative frequency curves, mean grain size or standard deviation (Fig. 5.10).

On the alluvial terraces, the organic-rich facies 2A sediments tend to be located near the tops of sections and are often associated with cultural layers. These factors point to a floodplain location. The facies 2A sediments are coarser-grained than the vertical accretion sediments of facies 1 suggesting a location proximal to the stream channel (Fig. 5.11). In many of the fluvial sections, the organic-poor facies 2B sediments are interbedded with sand and gravel units or directly overlie the sand and gravel suggesting a channel location. These sediments may be related to deposition on bars during lateral aggradation. Alternatively, facies 2B sediments may have been deposited on a floodplain, closer to the channel than the facies 2A sediments (Fig. 5.12). Cultural layers are located within several facies 2B units making a channel location for these sediments unlikely. The exact location of deposition cannot be stated with certainty due to the lack of structure within the facies.

Facies 2A and 2B deposits located high up on the slopes would have been deposited by a combination of mass movement and slope wash. The organic-rich sediments are likely the result of soil formation on a stable surface followed by burial.
Figure 5.10: Bivariate plot of mean grain size versus standard deviation for colluvial and fluvial sediments, Facies 2A and 2B. Identification of an unknown depositional environment could not be made on the basis of mean grain size and standard deviation. Each sample was identified as colluvial or fluvial on the basis of position within the landscape. Samples that could have been deposited by either mechanism were not included in this figure.

due to renewed slope activity. Unlike the facies 2 sediments found in valley bottom locations, many of the slope units contain occasional cobbles and small boulders.

Facies 3 is found in both alluvial terrace and slope locations. A bivariate plot of mean versus standard deviation of samples from known terrace, slope and till locations indicates a reasonable degree of separation (Fig. 5.13). The samples obtained from terrace locations have a coarser mean grain size and are slightly better sorted than the slope samples. The till samples are similar in mean grain size and sorting to the slope samples.
Figure 5.11: Bivariate plot of mean grain size versus standard deviation for facies 1 and facies 2 demonstrating general textural differences between the two facies. Facies 2 tends to be coarser-grained than facies 1.

Figure 5.12: Sketch indicating the position of deposition of facies 2A and facies 2B sediments.
Figure 5.13: Bivariate plot of mean grain size versus standard deviation of facies 3 sediment samples from slope, terrace and till locations. Unlike the facies 2 sediments, it would be possible to separate the slope and terrace sediments. Each sample was identified as slope or terrace on the basis of position within the landscape. The till samples were obtained from outside the main valley (chapter 6.1). Samples that could not be categorised with confidence were not included in this figure.

The facies 3 units found in slope locations were deposited by mass movements and/or slope wash. There is a higher concentration of gravel in this facies than the facies 2 slope units previously described. This suggests an intensification of processes, causing the concentration of gravel as silt and clay is removed. Facies 3 sediments are represented in all colluvial sections.

The sediments in facies 3 do not fit into the fluvial facies model developed by Miall (1978B, 1992) as easily as the other facies within the study site (Chapter 4.5; Table 4.3). The percentage of gravel is high enough that it is unlikely that these are overbank deposits, but not high enough for this to be a channel lag. Two cultural layers at the
Newo Asiniak site are located within a facies 3 unit which also makes a channel lag explanation unlikely.

Mass movement processes may have introduced coarser sediment from the slopes. Sediment deposited in the channel or on a point bar could then be reworked by the stream leaving a coarser and better sorted sediment. Alternatively, an increase in stream discharge may have resulted in both transportation and deposition of coarser sediment.

The well sorted sands of facies 4 deposits are found both within alluvial terraces and near the top of the valley slopes. It is unlikely that either colluvium or till would be as well sorted, leaving a fluvial or aeolian origin as the most likely mechanism of deposition. The sand units within the alluvial terraces and other valley bottom locations are most likely channel sands. Overbank sediments, containing a higher percentage of silt and clay, would not be as well sorted. The lack of organic carbon enrichment, such as would be expected in floodplain sediments, further strengthens this argument. The massive structure of the facies makes it impossible to determine whether the sands are associated with point bars or the main channel bottom.

The sand units found on the upper colluvial slopes suggest the reworking of the original sandy, glacio-fluvial deposits by wind. Several samples were obtained for comparison from dune deposits within the Beaver Creek Conservation Area, located several kilometres south of Saskatoon. These additional samples were included in a bivariate plot of mean versus standard deviation for facies 4 (Fig. 5.14). The bivariate
Mean vs. Standard Deviation, Facies 4

Figure 5.14: Bivariate plot of mean grain size versus standard deviation for facies 4. Note that the aeolian sands tend to be finer and slightly better sorted than the fluvial sands.

The sandy, matrix-supported gravel sediments of facies 5 are located within the alluvial terraces at the bottom of the valley. The gravel units were deposited within a stream channel, most likely as a channel lag. Meandering and braided stream systems exhibit gravel lags at the base of the channel. The gravel lag would be buried by finer-grained sediments as the stream aggraded or migrated laterally. The gravel units are associated with both channel sands (facies 4) and more poorly sorted fluvial sediments (facies 2B).

Debris flows, an alternative mechanism of deposition, can be rejected with a reasonable degree of confidence. In describing debris flow deposits, Rust and Koster
(1984) state that a muddy sand matrix is expected. The gravel deposits within the study site clearly have a sandy matrix, containing little silt or clay (Table 5.1; Appendix E). Debris flow deposits would also be expected to contain large outsized clasts, (Lewis and McMonnie, 1994B) which were not evident within these deposits. An additional argument against a debris flow origin is interbedding of facies 5 and facies 4 sand units at the Newo Asiniak site (Chapter 6.2.3). It would be unlikely that a debris flow would produce an interbedded sedimentary sequence of this nature.
Chapter 6
Landform Stratigraphy Within Wanuskewin Heritage Park

Introduction

Each of the main physiographic elements found within the study site including: 1) the till plain; 2) alluvial terraces; 3) mass movement landforms and 4) the modern channel and floodplain will be discussed in this chapter. The nature of sediments and depositional environments will be described for each site investigated during this study. The integration of individual sites, as well as the consideration of possible driving forces responsible for shaping the landscape, will be discussed in Chapter 7. The elevations in this chapter are reported relative to a datum (0 m), located on an alluvial terrace central to the study site (Fig. 6.1, 6.2).

6.1 Till Plain

The till plain is the single largest landscape element within the study site (Fig. 6.1, Fig. 6.3). The bedrock surface (mapped as the Cretaceous Oldman Formation, consisting of interbedded sands, silts and clays, carbonaceous material and concretionary zones) is approximately 70 m below the ground surface (Christiansen, 1970). The bedrock surface is overlain by the Sutherland group then Saskatoon group deposits. The upper 8 m of sediment in the study site consists of Battleford formation till and proglacial sediments.
Figure 6.1: Location of landforms and individual sites within Wanuskewin Heritage Park. Source: Meewasin Valley Authority, photo number 87GDS 055 87-89.

A  Datum, Newo Asiniak site
B  Till plain
C  Till section that was sampled
D  Tipperary Creek site
E  Juniper Flats site
F  Thundercloud site
G  Dogchild site
H  Delta
I  Beaver Dam
J  Terrace Remnant NA
K  Terrace Remnant NO
L  Terrace Remnants SOU and SOL
M  Redtail site
N  Amisk site
O  Terracettes surveyed and sampled
P  Rotational slide
Q  Dry tributary channel north of Newo Asiniak site
R  Shallow broad gullies
S  Narrow gullies
T  Manmade dam
U  Dry tributary channel north of Dogchild site
Figure 6.3: Air photo of hummocky till plain adjacent to study site. Source: Meewasin Valley Authority, photo number 87GDS-85.
Surficial soil maps of the Saskatoon area indicate that the till plain is exposed at the surface east of the valley. West of the valley, the till is overlain by sandy glacio-lacustrine sediments deposited in Glacial Lake Saskatchewan (Ellis and Stonehouse, 1970). Boulders exposed at the surface exhibit evidence of glacial action including a dew drop shape, striations and chattermarks.

Samples obtained from a section facing the South Saskatchewan River, approximately 3 m below the surface of the till plain (Fig. 6.1, 6.2), were analysed for grain size and organic and inorganic carbon content. The grain size fractions reported by Christiansen (1970) for both the Battleford and Floral Formations do not include the gravel fraction. By re-calculating the grain size distribution without the gravel fraction, it can be determined that the samples fall close to the range published for the Battleford Formation (Table 6.1).

### 6.2 Alluvial Terraces

The alluvial terraces located near the valley bottom are the focus of this study (Fig. 6.4). Three archaeological sites were examined, including the Tipperary creek and Newo Asiniak sites which were re-excavated, and the Thundercloud site which was open (Fig. 6.1, 6.2). The paucity of boulders meant that locations not yet excavated

<table>
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<th>Percent Sand</th>
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<th>Percent Clay</th>
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<tr>
<td>Battleford Formation</td>
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<td>Floral Formation</td>
<td>19 - 32</td>
<td>36 - 46</td>
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Figure 6.4: Landforms within Wanuskewin Heritage Park. A) Tipperary Creek site alluvial terrace located near the mouth of the Opimihaw Creek. The South Saskatchewan river is pictured at the top right corner of the photograph. B) The modern stream channel and floodplain pictured in front of an alluvial terrace surface.
Figure 6.19: Landforms within Wanuskewin Heritage Park. A) Terrace remnant (NA) located north of the Newo Asiniak site. B) Colluvial slopes highlighting alternating gentle and steep, undercut slopes.
could be cored with relative ease. There is an extensive radiocarbon data base associated with the Tipperary Creek and Newo Asiniak sites constraining the deposition of the alluvium.

6.2.1 Tipperary Creek Site

The Tipperary Creek site is located on the terrace surface closest to the mouth of the valley (Fig. 6.1, 6.2). The elevation of this terrace is between 0.6 and 1.3 m below datum, sloping down towards the modern creek (Fig. 6.5). The surface of the excavation site is 0.75 m below datum. Meander scars were noted at the time of the original excavation, 1985 - 1987 (Ernest Walker, personal communication, 1995), but are no longer evident at the surface. A narrow terrace remnant, 3.0 to 3.4 m below datum is located adjacent to the modern stream where it joins the South Saskatchewan River.

A series of thirteen radiocarbon dates were obtained during the archaeological excavation of the site (Table 6.2). There are two date reversals within the data base (S-2806 and S-2816), possibly due to reworking of older material or contamination of the samples. The remaining dates are accepted as reasonable age estimates as they follow in chronological order. Individual cultural layers were easily identified as markers had been left in the wall of the pit by the original excavators.

A 1.77 m deep pit was dug, re-exposing the west wall of the original excavation. A 0.05 m diameter core was taken from the bottom of the pit to a depth of 2.57 m. The base of the section was composed of fine-grained, organic-poor sediments (facies 1B), succeeded by a 0.5 m unit of granular, sandy mud (facies 2B) (Fig. 6.6). A 0.22 m thick boulder layer in a matrix of facies 3 sediment extends from 1.87 m to 1.65 m below the
surface. Charcoal fragments were noted in this unit. Above the boulder layer is a thick unit of fine-grained sediments (facies 1B and 1A) to 0.19 m below the surface.

![Profile of Tipperary Creek Site](image)

Figure 6.5: Topographic profile of the Tipperary Creek site indicating the position of the excavation. The survey datum is at 0 m. The vertical exaggeration of the profile is 4X. Source: Topographic map, Meewasin Valley Authority; 1995 - 6 survey data.

Table 6.2: Cultural layer depths and uncorrected radiocarbon dates obtained from the Tipperary Creek site. Source: Morlan, 1992. The complete date list is located in Appendix 2.

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<td>790 ± 135</td>
<td>S-2810</td>
</tr>
<tr>
<td>7</td>
<td>0.54</td>
<td>855 ± 70</td>
<td>S-2811</td>
</tr>
<tr>
<td>8</td>
<td>0.63</td>
<td>889 ± 70</td>
<td>S-2812</td>
</tr>
<tr>
<td>9</td>
<td>0.65</td>
<td>945 ± 135</td>
<td>S-2813</td>
</tr>
<tr>
<td>10</td>
<td>0.72</td>
<td>1155 ± 75</td>
<td>S-2814</td>
</tr>
<tr>
<td>11</td>
<td>0.79</td>
<td>1235 ± 75</td>
<td>S-2815</td>
</tr>
<tr>
<td>12</td>
<td>0.91</td>
<td>1790 ± 75</td>
<td>S-2816</td>
</tr>
<tr>
<td>13</td>
<td>0.96</td>
<td>1535 ± 75</td>
<td>S-2885</td>
</tr>
</tbody>
</table>
Figure 6.6: Tipperary Creek excavation section diagram. Uncorrected radiocarbon dates are noted beside cultural layers (>). (C) indicates a coarse mean grain size; (F) indicates a fine mean grain size.
Sediments at the base of this unit are organic-poor, shifting to organic-rich at 0.97 m below the surface. Cultural layers 2 through 13 are found within the organic-rich sediments, associated with distinct laminations, generally less than 0.01 m thick. Laminations with wood material in the lower, organic-poor section are not associated with cultural layers and are not dated. A 0.01 m thick facies 2A layer represents a brief interruption of the fine-grained sediments. Cultural layer 1 is associated with a return to fine-grained sedimentation at 0.18 m. This continues to 0.07 m below the surface where sediments once again shift to facies 2A. The mean grain size of the units in this section display little variation with the exception of the facies 3 matrix supported boulder layer (Fig. 6.6). Thick sand units with cross-trough bedding were noted during the initial excavation (Ernest Walker, personnel communication, 1995), but could not be confirmed during the current study.

Sedimentation rates were calculated for the upper metre of the section using published radiocarbon dates and the depth of cultural layers measured during the 1995 excavations (Table 6.3). Layers associated with date reversals are excluded from the sedimentation rate calculations. There is an order of magnitude increase in sedimentation rates ca. 850 years BP (cultural layers 7 - 8) and ca. 150 yrs BP (cultural layers 1 - 3), the latter corresponding with deposition of the coarser facies 2A sediment noted above (Fig. 6.7).

The domination of facies 1 vertical accretion sediments within the section suggests deposition by a meandering stream system. The facies 1B vertical accretion
Table 6.3: Rates of sedimentation at the Tipperary Creek site calculated from cultural layer depths and published radiocarbon dates. The average sedimentation rate for the dated portion of the section is also noted.

<table>
<thead>
<tr>
<th>Layers Compared</th>
<th>Depth of Layers Compared (m)</th>
<th>Sedimentation Rate (mm a⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 - 3</td>
<td>0.19 - 0.38</td>
<td>1.90</td>
</tr>
<tr>
<td>3 - 4</td>
<td>0.38 - 0.40</td>
<td>0.22</td>
</tr>
<tr>
<td>4 - 5</td>
<td>0.40 - 0.45</td>
<td>0.23</td>
</tr>
<tr>
<td>5 - 6</td>
<td>0.45 - 0.50</td>
<td>0.18</td>
</tr>
<tr>
<td>6 - 7</td>
<td>0.50 - 0.54</td>
<td>0.62</td>
</tr>
<tr>
<td>7 - 8</td>
<td>0.54 - 0.63</td>
<td>3.60</td>
</tr>
<tr>
<td>8 - 9</td>
<td>0.63 - 0.65</td>
<td>0.31</td>
</tr>
<tr>
<td>9 - 10</td>
<td>0.65 - 0.72</td>
<td>0.33</td>
</tr>
<tr>
<td>10 - 11</td>
<td>0.72 - 0.79</td>
<td>0.87</td>
</tr>
<tr>
<td>11 - 13</td>
<td>0.79 - 0.96</td>
<td>0.57</td>
</tr>
<tr>
<td>Average</td>
<td>0.19 - 0.96</td>
<td>0.54</td>
</tr>
</tbody>
</table>

Figure 6.7: Graphical presentation of sedimentation rates in mm per year at Tipperary Creek site calculated from the radiocarbon date data base and measured distances between cultural layers.
sediments at the base of the section indicate a distal channel location. The migration of the stream towards the site is recorded in the transition to facies 2B. The matrix-supported boulder layer was likely derived from the slopes adjacent to the site. Mass movements would have introduced the boulders and coarse matrix into the stream channel. This colluvial material would have been washed by the stream removing the fine sediment, leaving the coarser fraction behind. Charcoal in the matrix suggests that fire may have removed the vegetation from the slopes, leading to instability. The boulder layer is undated although the application of the average sedimentation rates (0.54 mm a\(^{-1}\)) to the under-lying organic-poor unit suggests deposition may have been initiated ca. 2.8 ka BP. This would place deposition of the boulder layer prior to 2.8 ka BP. This estimate should not be considered definitive, however, as sedimentation rates may have varied over time.

The return to facies 1 sediments suggests the stream migrated away from the site. The thin unit of coarser-textured sediments (facies 2A) within the facies 1A unit indicates deposition due to a higher energy flood event capable of transporting and depositing coarser sediment. A pulse of slightly coarser material was noted in the field between cultural layer 7 and 8. The upper facies 2A unit likely records either larger flood events prior to the abandonment of the floodplain or the migration of the channel towards the site.

6.2.2 Juniper Flats Site

The Juniper Flats site is located on a large alluvial terrace, upstream from the Tipperary Creek site (Fig. 6.1, 6.2). The terrace surface is 0.79 to 1.13 m below datum,
sloping down to the modern stream (Fig. 6.8). Archaeological excavations have not been conducted at this site, resulting in the use of core data. Two cores (Core 3 and Core 4) were selected for detailed study. Core 3, located near the centre of the terrace, is 0.91 m below the datum. Core 4, located at the base of the colluvial slope, is 0.40 m above the datum.

Core 3

Core 3 records a simple history (Figure 6.9). With the exception of a thin unit of facies 2A sediment, 0.25 to 0.26 m below the surface, the entire 1.86 m core is classified as facies 1A vertical accretion deposits suggesting a history similar to the upper units at the Tipperary Creek site. The organic-rich laminations observed in the Tipperary Creek excavation were not evident within this core.

![Profile of Juniper Flats Site](image_url)

Figure 6.8: Topographic profile of Juniper Flats Site indicating the position of core 3 and core 4. The survey datum is at 0 m. The vertical exaggeration of the profile is 4X. Source: Topographic map, Meewasin Valley Authority; 1995 - 6 survey data.
Figure 6.9: Juniper Flats core 3 diagram. (C) indicates a coarse mean grain size; (F) indicates a fine mean grain size.

There are no radiocarbon dates at this site making it impossible to determine either when deposition occurred or sedimentation rates. Deposition at this site, and the Tipperary Creek site were likely contemporaneous so by applying the Tipperary Creek
average sedimentation rates to this site, it can be suggested that deposition would have
been initiated ca. 3.3 ka BP.

**Core 4**

Core 4, 1.99 m in length, is coarser in texture than either core 3 or the Tipperary
Creek excavation site. The base of the core consists of facies 2 sediments varying from
organic-poor facies 2B at the base, to organic-rich facies 2A (1.41-1.63 m below the
surface) then back to organic-poor facies 2B (Fig. 6.10). There is considerable variation
in mean grain size, particularly within the lower facies 2B unit, which corresponds with
changes in colour from light olive brown (coarser texture) to very dark greyish brown
and brown (finer texture).

There is a generally fining-upwards trend in mean grain size from 1.41 to 0.46 m
which includes units of both facies 2B and facies 3 sediments. At 0.46 m below the
surface, the mean grain size coarsens despite classification as facies 2B, due to a large
increase in sand. Above 0.38 m, the sediment fines upward sharply through alternating
units of facies 2A and facies 1A sediments.

The variations in texture and organic carbon, as well as the elevation and
proximity to the colluvial slope, suggest the inter-fingering of alluvial and colluvial
sediments, common near the interface of floodplains and valley slopes (Chapter 4.6). A
combination of slope-wash and mass movements would deposit colluvium on the
floodplain surface. These sediments would then be buried with alluvium during flood
events. The coarser sediment within the facies 2 units likely record pulses of colluvial
material. The colour variations reflect shifts from organic-rich to organic-poor
Figure 6.10: Juniper Flats core 4 diagram. (C) indicates a coarse mean grain size; (F) indicates a fine mean grain size.

sediments recording periods of exposure and soil formation and subsequent burial. Facies 1 vertical accretion deposits in the upper 0.2 m of the core confirm fluvial deposition did occur at this site.
6.2.3 Newo Asiniak Site

The Newo Asiniak site is centrally located within the study site, on the western side of the modern stream (Fig. 6.1, 6.2). A meander scar is located at the base of the valley wall. Units from both the main terrace surface and meander scar were excavated by Kelly (1986). The main terrace surface is between 0.25 m below and 0.5 m above the datum (Fig. 6.11). The terrace slopes gently down toward the front edge riser then down onto a second, smaller terrace before reaching the modern stream. A small remnant of the main terrace is evident on the lower terrace surface.

The west wall of a 2 m x 2 m block from the main terrace surface was re-excavated. It was not possible to re-excavate any of the meander scar units as the majority of this area now lies underneath a park path. Instead, a core (Core 7) was obtained from a location determined to have been left intact during the original investigation. The dig site is 0.44 m above the datum and Core 7 is 0.89 m below the datum.

Sediment descriptions and section diagrams produced during the current survey could not be correlated with Kelly’s original work. The position of the cultural layers defined by Kelly (1986) were determined with a reasonable degree of confidence, however, as the layers were described as being of similar depth below the surface throughout the site.

Eleven radiocarbon dates have been obtained from the Newo Asiniak site; nine during the original excavation and two during the current study (Table 6.4; Appendix D). Two of the radiocarbon dates are from the meander scar (S-2763, S-2528). The
Figure 6.11: Topographic profile of Newo Asiniak site indicating the position of the excavation, core 7 and the datum. The survey datum is at 0 m. The vertical exaggeration of the profile is 4X. Source: Topographic map, Meewasin Valley Authority; 1995 - 6 survey data.

Table 6.4: Cultural layer depths and uncorrected radiocarbon dates obtained from the Newo Asiniak site. Source: Kelly, 1986. The depth of each cultural layer was calculated from a stratigraphic diagram produced by Kelly. The last two dates in the table are from the meander scar. An asterisk (*) indicates dates obtained during the current study.

<table>
<thead>
<tr>
<th>Cultural Layer</th>
<th>Depth (m)</th>
<th>Date (yrs BP)</th>
<th>Sample Number</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.06</td>
<td>750 ± 70</td>
<td>S-2529</td>
</tr>
<tr>
<td>2</td>
<td>0.15</td>
<td>915 ± 70</td>
<td>S-2533</td>
</tr>
<tr>
<td>3</td>
<td>0.25</td>
<td>2235 ± 70</td>
<td>S-2530</td>
</tr>
<tr>
<td>4</td>
<td>0.36</td>
<td>3025 ± 250</td>
<td>S-2764</td>
</tr>
<tr>
<td>5</td>
<td>0.46</td>
<td>2525 ± 210</td>
<td>S-2765</td>
</tr>
<tr>
<td>6</td>
<td>0.55</td>
<td>3420 ± 85</td>
<td>S-2532</td>
</tr>
<tr>
<td>7</td>
<td>0.71</td>
<td>3455 ± 230</td>
<td>S-2766</td>
</tr>
<tr>
<td>*</td>
<td>1.47</td>
<td>4630 ± 60</td>
<td>TO-5763</td>
</tr>
<tr>
<td>*</td>
<td>2.05</td>
<td>4580 ± ?</td>
<td></td>
</tr>
<tr>
<td>meander scar</td>
<td>surface</td>
<td>185 ± 190</td>
<td>S-2763</td>
</tr>
<tr>
<td>meander scar</td>
<td>0.25</td>
<td>1540 ± 70</td>
<td>S-2528</td>
</tr>
</tbody>
</table>

original data base exhibits a date reversal (S-2764, S-2765). Four dates have standard deviations greater than ± 200 years (185 ± 190, S-2763; 3025 ± 250, S-2764; 2525 ±
and are therefore considered too unreliable for further consideration in this study. The upper meander scar date has a standard deviation larger than the actual date indicating a modern sample.

**Excavation**

The re-excavated section is the deepest sequence within the alluvial terraces (Fig. 6.12). The bottom units of the section to 2.16 m alternate between channel sand (facies 4) and finer-textured muddy sands (facies 2B). These sediments represent channel deposition, possibly on a point bar, associated with channel aggradation and migration.

Above 2.16 m the section is characterised by alternating channel sand and gravel units to 1.19 m below the surface. Of interest is the occurrence of thin units of mud-rich sediment directly overlying gravel units or forming the matrix of the upper 0.01 - 0.02 m of the gravel, characteristic of bar formation in braided stream channels (Collinson, 1986). Radiocarbon dates from this portion of the section are contemporaneous suggesting extremely rapid sedimentation as is often associated with braided channels. If a stream is close to a threshold in discharge, load or gradient, it may fluctuate between braided and meandering patterns (Schumm, 1981). It is not possible to state with certainty, however, that the gravel units were deposited within a braided system without observing the internal architecture of the deposit.
Figure 6.12: Newo Asiniak section diagram. Uncorrected radiocarbon dates are noted beside cultural layers 0. (C) indicates a coarse mean grain size; (F) indicates a fine mean grain size.
Above 1.19 m there is a decline in both gravel and sand. A thick unit of facies 2B channel sediments records the migration of the channel away from this location. A facies 3 unit, 0.75 - 0.45 m below the surface, suggests either increased discharge and a coarser sediment load or that more colluvial material is being deposited at the site. Deposition of this unit did not occur as a single event as indicated by evidence of human occupation at 0.71 m (layer 7) and 0.55 (layer 6). The migration of the stream away from the site is recorded in the upper 0.45 m of the section in a facies 2B unit followed by a facies 2A unit. Within these units there is decrease in mean grain size, increasing organic carbon accumulation and repeated human occupation. Facies 2A indicates a proximal channel location before final abandonment.

Rates of sedimentation have been calculated using radiocarbon dates cultural layers and new samples (Table 6.5). The rejection of three dates considered unreliable (S-2764 [4], S-2765 [5], S-2766 [7]) means that this comparison has a coarse resolution, so only general trends are apparent. Rapid sedimentation is evident in the lower units of this section, declining after 4.3 ka BP. Sedimentation rates increased ca. 0.8 - 0.9 ka BP, but not to former levels.

Table 6.5: Rates of sedimentation at the Newo Asiniak site calculated from cultural layer depths and reliable radiocarbon dates. The average rate of sedimentation for the dated portion of the section is also noted.

<table>
<thead>
<tr>
<th>Layers Compared</th>
<th>Depth of Layers Compared (m)</th>
<th>Sedimentation Rate (mm a⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 - 2</td>
<td>0.06 - 0.15</td>
<td>0.55</td>
</tr>
<tr>
<td>2 - 3</td>
<td>0.15 - 0.25</td>
<td>0.08</td>
</tr>
<tr>
<td>3 - 6</td>
<td>0.25 - 0.55</td>
<td>0.14</td>
</tr>
<tr>
<td>6 - non-cultural date</td>
<td>0.55 - 1.47</td>
<td>2.97</td>
</tr>
<tr>
<td>average</td>
<td>0.06 - 1.47</td>
<td>0.36</td>
</tr>
</tbody>
</table>
Core 7

Core 7, obtained from a meander scar at the back of the main terrace surface, is essentially a fining upwards sequence. The base of the core to 1.5 m is gravel (facies 5), indicating the bottom of the channel (Fig. 6.13). This is followed by sand (facies 4) to 1.4 m below the surface. Bedding structures cannot be seen in core material but the juxtaposition of sand on gravel is expected in fluvial deposits. As on the main terrace surface, there is an increase in gravel at facies 3, 1.4 m to 1.35 m below the surface. This unit is overlain by sand to 1.02 m below the surface.

Above 1.02 m the core continues to fine upwards and there is an increase in organic carbon with the shift to facies 2B then facies 2A. Above 0.33 m, the core is made up of organic-rich fine-grained facies 1A deposits. The upper 1.02 m of core likely represents a chute or channel which has been abandoned or cut-off allowing deposition of fine grained material.

6.2.4 Thundercloud Site

The Thundercloud site consists of an alluvial terrace, 3.3 m to 3.55 m above the datum, and a colluvial slope with terrace remnants 6.46 m and 10.95 m above the datum (Fig. 6.14). The Thundercloud site is the location of the current University of Saskatchewan archaeological field school and remained open throughout the course of this investigation. Two shallow pits, examined during the investigation of the colluvial slope, will also be considered in this section.
Excavation

The south face of the Thundercloud excavation was logged and sampled during the 1995 field season. This included an auger hole into the bottom of the excavation.

The 1996 archaeology field school extended the north face of the excavation below the
Figure 6.14: Topographic profile of Thundercloud Site indicating the position of the excavation, pit 1 and pit 2. The position of two terrace remnants, Thundercloud Upper (TCU) and Thundercloud Lower (TCL) are also indicated. The survey datum is at 0 m. The vertical exaggeration of the profile is 4X. Source: Topographic map, Meewasin Valley Authority; 1995 - 6 survey data.

Although the samples analysed were obtained from the south face, it is possible to correlate these units with units on the north face.

Only one radiocarbon date has been obtained from this site at present. There are, however, relative dates based on projectile points for many of the cultural layers (Table 6.6). There is a considerable range in age associated with the tool traditions observed at this site making the calculation of rates of sedimentation (beyond an average rate of 0.12 mm a⁻¹) impossible at this time.

The base of the section to 0.9 m is a sandy matrix-supported coarse gravel (facies 5; Fig. 6.15). The unit fines upward into a gravelly, muddy sand unit (facies 3) 0.8 m - 0.9 m below the surface. This is followed by a facies 2B unit extending from 0.8 m to between 0.47 m and 0.5 m. Cultural layer 7 is located within this unit at 0.75 m below
Table 6.6: Depth and relative age of cultural layers at Thundercloud. Source: Lis Mack (personal communication, 1996). An uncorrected radiocarbon date (S-3645) for cultural level 6 is also presented (Ernest Walker, personal communication 1997).

<table>
<thead>
<tr>
<th>Cultural Layer</th>
<th>Depth (m)</th>
<th>Source of Relative Age (points); Radiocarbon Age</th>
<th>Estimated Age (BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.05</td>
<td>metal points; Plains Triangular</td>
<td>280-100</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>400-100</td>
</tr>
<tr>
<td>2</td>
<td>0.1</td>
<td>Plains Triangular</td>
<td>400-100</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Plains Side-notched</td>
<td>550-250</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Prairie Side-notched</td>
<td>1200-800</td>
</tr>
<tr>
<td>3</td>
<td>U0.15+</td>
<td>Besant</td>
<td>2000-1200</td>
</tr>
<tr>
<td></td>
<td>0.19</td>
<td>Pelican Lake</td>
<td>3800-1850</td>
</tr>
<tr>
<td>4</td>
<td>U0.24+</td>
<td>no points</td>
<td>no date</td>
</tr>
<tr>
<td></td>
<td>0.27</td>
<td></td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>0.36</td>
<td>Duncan</td>
<td>4150-3100</td>
</tr>
<tr>
<td>6</td>
<td>0.46</td>
<td>McKean Lanceolate</td>
<td>4150-3100</td>
</tr>
<tr>
<td></td>
<td></td>
<td>radiocarbon date</td>
<td>4040 ± 90</td>
</tr>
<tr>
<td>7</td>
<td>0.75</td>
<td>no points</td>
<td>no date</td>
</tr>
</tbody>
</table>

The surface. There is some organic staining associated with the cultural layer. Facies 2A records a continued decrease in the sand fraction and an increase in organic carbon. Cultural layer 6 is located at the base of this unit. There is an undulating colour boundary between the facies 2B and facies 2A units, likely the result of pedogenesis. Root casts extend into the top of the underlying facies 2B unit.

Cultural layer 5, 0.34 m to 0.37 m below the surface, is in an overbank facies 1A unit. There is continued overbank sedimentation above this unit but the organic carbon content decreases resulting in a facies 1B classification. Cultural layer 4 is in a discontinuous, undulating facies 2A unit, ranging from 0.24 to 0.3 m below the surface.
Figure 6.15: Thundercloud section diagram. Uncorrected radiocarbon dates are noted beside cultural layers (). (C) indicates a coarse mean grain size; (F) indicates a fine mean grain size.

The undulating lower boundary observed in the colour of the sediment is likely due to pedogenesis. The discontinuous nature of the layer suggests that erosion removed material after deposition and pedogenesis had originally occurred. The section becomes
significantly coarser in mean grain size above the facies 2A unit, with the shift to facies 3. The upper boundary of this unit undulates between 0.15 m and 0.19 m below the surface. The top unit is organic-rich (facies 2A) and contains 3 cultural layers. Cultural layer 3 is situated at the bottom of the unit; cultural layer 2 is at 0.1 m and cultural layer 1 is at 0.05 m. Black rootcasts extend into the underlying very dark greyish brown sediment.

The clearly visible root casts which extend from each of the narrow, organic-rich, cultural layers suggests that the paleosols are most likely cumulic Regosols. These are recognised by the lack of B horizon development. Pedogenesis would have been halted by deposition of new sediment. Even within the top facies 2A unit, there are alternating bands of black and very dark brown sediment indicating a cumulic profile.

The Thundercloud excavation records a fluvial depositional history. The lower gravel unit is likely a channel lag deposited prior to ca. 4 ka BP (4040 ± 90 S-3645), the radiocarbon date on cultural layer 6. The channel then migrated away from the site, burying the channel lag with point bar then floodplain sediments. The fining-upward sequence, culminating in facies 1A vertical accretion deposits reflects continued migration of the channel away from the site.

Above the facies 1 units, the deposition of facies 2A proximal channel sediments suggests the stream migrated back towards the site. It may also mark a shift to generally higher discharges capable of transporting a higher sediment load. Cultural layer 4 is discontinuous in nature suggesting erosion of the site possibly during a larger flood event. Facies 3 indicates deposition of coarser material, possibly reflecting the introduction of
more sediment form the slopes. Repeated human occupation and organic accumulation indicates periodic subaerial exposure of the site.

**Pits 1 & 2**

Two shallow pits were dug near the base of the colluvial slope. Pit 1 is 3.9 m above the datum and pit 2 is 4.37 m above the datum (Fig. 6.14). Core samples were obtained extending pit 1 to 1.5 m in depth and pit 2 to 1.05 m in depth. The stratigraphy of each pit was similar (Fig. 6.16). The lowest unit of pit 1, extending to 0.78 m, is organic-poor, classified as facies 2B. The overlying facies 2A unit is slightly darker as expected with organic enrichment. Above facies 2A the sediments become coarser-grained resulting in classification as facies 3. At the top of the pit, 0.07 m to 0.26 m below the surface, gravel decreases and the sediment becomes organic-rich. The uppermost 0.07 m is the modern sod layer. Pit 2 exhibits a similar stratigraphy except that it does not contain an organic-rich facies 2A unit between the facies 2B and facies 3 units.

Based on their location above the main terrace surface, it is likely that the pits record a mainly colluvial depositional history. Facies 3 represents colluvial material deposited by more intensive mass movement or slope wash processes than the underlying facies 2A and 2B units. There may have been inputs of fluvial sediments, but this cannot be confirmed. The lack of radiocarbon dates makes it impossible to determine how far above the floodplain these sections would have been deposited. The lack of vertical accretion sediments supports deposition by slope processes above the main floodplain surface.
Figure 6.16: Thundercloud pits 1 and 2 diagrams. (C) indicates a coarse mean grain size; (F) indicates a fine mean grain size.
There was no evidence of the three distinct organic-enriched layers found in the terrace section suggesting slow continuous deposition of the uppermost unit allowing pedogenesis to occur. The lower facies 2A unit in pit 1 reflects stability of the surface. The lack of organic accumulation in pit 2 indicates that this site may have been subject to continuous sedimentation.

6.2.5 Dogchild Site

The Dogchild site is the narrowest and northernmost terrace investigated in this study (Fig. 6.1, 6.2). The terrace is 4 m - 5 m above the datum, sloping towards the modern stream (Fig. 6.17). No archaeological excavations have been conducted at this site so two auger holes (Core 9 and Core 10) were selected for analysis. Core 9 was taken from near the base of the colluvial slope, 6.25 m above the datum. Core 10 was taken from a meander scar on the south-east corner of the terrace.

![Profile of Dogchild Site](image)

Figure 6.17: Topographic profile of Dogchild site indicating the position of core 9 and core 10. The survey datum is at 0 m. The vertical exaggeration of the profile is 4X. Source: Topographic map, Meewasin Valley Authority; 1995 - 6 survey data.
Core 9

Core 9 is 1.0 m long (Fig. 6.18). The base of the core to 0.34 m is classified as facies 2B. There is an increase in gravel from 0.34 m to 0.27 m resulting in classification as facies 3. Above 0.27 m, the mean grain size fines then coarsens upwards to the top of the section. This upper unit has a significant colour change from olive brown at the base to very dark greyish brown to black at the top. It is organic-rich and therefore classified as facies 2A. There is also variation in mean grain size and inorganic carbon within this upper unit.

The elevation of the core above the main terrace surface suggests a dominantly colluvial origin (slope wash and/or mass wasting). The facies 3 unit indicates an intensification of these processes. There has been a considerable accumulation of organic material in the upper unit suggesting that the slopes have been stable for quite some time, thus allowing pedogenesis. The organic-rich unit may be cumulic which would suggest very slow sedimentation allowing organic material to accumulate.

Core 10

Core 10, 4.06 m above the datum, records a different depositional history. The base of the core, from 0.92 m to 0.81 m is a matrix-supported gravel classified as facies 5 (Fig. 6.18). This gravel unit marks the position of the stream channel. The texture, mean grain size and inorganic carbon content of the core varies considerably within facies 2B (0.81 m to 0.12 m). Pulses of coarser, sandier sediment alternate with finer, muddy sediments. The variations in texture may reflect the migration of the stream channel and associated point bar formation. Alternatively, the alternating sediments
Figure 6.18: Dogchild cores 9 and 10 diagrams. (C) indicates a coarse mean grain size; (F) indicates a fine mean grain size.

may be due to changes in sediment load or discharge. Changes in the sediment load may be due to changes in slope activity upstream of the site with the coarser sediments representing an influx of colluvial material. There is much less organic material
accumulation in core 10 than in core 9 suggesting that pedogenesis has been occurring over a shorter period of time.

This terrace is different from the other terraces in the study site. Ten cores were taken from various locations across the terrace surface and near the base of the slope. The majority of these bottomed out in boulders or large stones between 0.05 m and 0.4 m below the surface. The alluvial material overlying the colluvial boulders is not of a consistent thickness. The stream likely cut a the terrace into the colluvial slope similar to the terraces noted above the Thundercloud site and below the Opimihaw outlook (Chapter 6.4). This terrace may subsequently have been draped with alluvium, burying the boulders. The alluvium found in core 10 represents the position of the stream as the original terrace surface was incised. There are no radiocarbon dates from this site so it is impossible to confirm when deposition of the alluvium occurred.

6.3 The Modern Stream Channel and Floodplain

The stream incised to its present elevation some time within the past few hundred years (Fig. 6.4). The radiocarbon date on cultural layer 1 at the Tipperary Creek site (Table 6.2) and burial of metal points in cultural layer 1 at the Thundercloud site (Table 6.6) suggest this may have occurred as recently as this century. A radiocarbon date from the meander scar at the Newo Asiniak site also indicates recent abandonment.

Several small terrace remnants below the Juniper Flats and Tipperary Creek site, as well as the dissected nature of the lower Newo Asiniak terrace indicates that incision was not continuous. The stream is no longer actively incising and has begun to form a small floodplain through lateral erosion. Several reaches of the stream are currently
stagnant due to an active beaver population (Fig. 6.1). A small delta has formed at the mouth of the Opimihaw creek.

6.4 Terrace Remnants

Terrace remnants have been noted at several elevations within the study site (Fig. 6.19). Two remnants are located on the colluvial surface above the Thundercloud site, one on a slope located north of the Newo Asiniak site, one above the Amisk site, one north of the Opimihaw Outlook and two south of the Opimihaw outlook (Fig. 6.1, 6.2). The elevations near the front edges of each of these terrace remnants are recorded in Table 6.7

The surfaces of the remnants have more boulders than the surrounding colluvial slopes or till plain. Sand and mud would have been removed by fluvial action leaving a concentration of boulders at the surface. The high concentration of boulders precluded the investigation of the internal structure of the remnants with the auger.

As discussed in chapter 4, the correlation of terrace surfaces by altitude matching can be problematic. Unfortunately, so little has been preserved of the original surfaces that correlation by any other method is impossible. The highest, and probably earliest, remnants are found closest to the mouth of the valley. NO and SOU are approximately 16 m above the datum (Fig. 6.20). NA and SOL represent the next surface, located 12 m above the datum (Fig. 6.21). The remnants above the Thundercloud site are smaller in scale, TCU in particular. This remnant is represented by a break in the slope of the main colluvial slope (Fig. 6.14). This and the Amisk site remnant (AM) may represent
Table 6.7: Location of terrace remnants, including the names to be used in the text, and their elevation above the datum. The elevations were taken from the front of each remnant. In many cases it is difficult to find the exact break in slope, accounting for differences in elevations.

<table>
<thead>
<tr>
<th>Location of Remnant</th>
<th>Name of Remnant</th>
<th>Elevation Above A</th>
</tr>
</thead>
<tbody>
<tr>
<td>Thundercloud site (upper)</td>
<td>TCU</td>
<td>10.95</td>
</tr>
<tr>
<td>Thundercloud site (lower)</td>
<td>TCL</td>
<td>6.46</td>
</tr>
<tr>
<td>N of Newo Asiniak site</td>
<td>NA</td>
<td>12.24</td>
</tr>
<tr>
<td>Amisk site</td>
<td>AM</td>
<td>10</td>
</tr>
<tr>
<td>N of Opimihaw Outlook</td>
<td>NO</td>
<td>16.12</td>
</tr>
<tr>
<td>S of Opimihaw Outlook (upper)</td>
<td>SOU</td>
<td>16.63</td>
</tr>
<tr>
<td>S of Opimihaw Outlook (lower)</td>
<td>SOL</td>
<td>12.07</td>
</tr>
</tbody>
</table>

Figure 6.20: Topographic profile of terrace remnants south of Opimihaw Outlook indicating the position of the upper remnant (SOU) and the lower remnant (SOL). The survey datum is at 0 m. The vertical exaggeration of the profile is 4X. Source: Topographic map, Meewasin Valley Authority; 1995 - 6 survey data.
Figure 6.21: Topographic profile of Newo Asiniak Terrace Remnant (NA). The survey datum is at 0 m. The vertical exaggeration of the profile is 4X. Source: Topographic map, Meewasin Valley Authority; 1995 - 6 survey data.

The terrace remnants were most likely formed by the stream meandering as it incised into the till plain during the initial incision of the valley. In this scenario, no geomorphic thresholds are crossed; the terrace remnants are simply the natural result of slow incision. The lack of radiocarbon dates makes it impossible to determine when each surface was formed or rates valley incision.

6.5 Colluvial Surfaces

The colluvial slopes vary in gradient from long gentle slopes to steeper slopes formed as the stream undercut the valley walls (Fig. 6.19). Slope wash and mass movements would likely have contributed to deposition on the slopes throughout the
Figure 6.22: Topographic profile of Amisk site terrace remnant (AM) and the excavation site. The survey datum is at 0 m. The vertical exaggeration of the profile is 4X. Source: Topographic map, Meewasin Valley Authority; 1995-6 survey data.

valley’s history. Turf covered boulders indicate that creep has been an important processes.

6.5.1 Redtail Site

There are several small valleys incised into the till plain that drain directly into the South Saskatchewan River extending west from Opimihaw Creek valley. The Redtail site is located in one of these valleys, adjacent to a small dry channel (Fig. 6.1, 6.23). One tributary joins with the valley downslope from the excavation site.

Rudimentary particle size analysis was completed in conjunction with the archaeological excavation (Ramsay, 1993) but these are unsuitable for correlating the two studies, however, as the sampling techniques were flawed and the method used for statistical calculations was inappropriate for the coarse resolution of analysis. The percent of gravel, sand, silt and clay as well as textural groupings were not reported.
Figure 6.23: Topographic profile of Redtail site indicating the position of the excavation. The survey datum is at 0 m. The vertical exaggeration of the profile is 4X. Source: Topographic map, Meewasin Valley Authority; 1995 - 6 survey data.

An additional source of concern is the high numbers of Ah, Ae and varied B soil horizons identified by Ramsay (1993) as weakly developed chernozems.

There are eight radiocarbon dates associated with this site (Table 6.8). Dates on cultural levels 11, 12, 13 and 15 were obtained during the original excavation. Depths of these layers were estimated from descriptions and diagrams by Ramsay (1993). One additional date was obtained during the course of this study, possibly relating to cultural layer 9 (Appendix D). There are no date reversals and the reported standard deviations are small, indicating that this is a reliable data base. Sedimentation rates were calculated from the radiocarbon dates and the depths of cultural layers (Table 6.9).

A short section of the south wall of the original archaeological excavation was re-excavated. Compaction of the upper 0.6 m of the section occurred when the site was
Table 6.8: Depth and uncorrected radiocarbon dates for cultural layers at the Redtail site. Source: Ramsay, 1993. It should be noted that these depths were taken from the depth of layers described in a stratigraphic column, not the actual depth of the sample. This stratigraphic column does not describe the re-excavation section.

<table>
<thead>
<tr>
<th>Cultural Layer</th>
<th>Depth (m)</th>
<th>Date (yrs BP)</th>
<th>Sample Number</th>
</tr>
</thead>
<tbody>
<tr>
<td>This Study - possibly 9</td>
<td>0.89</td>
<td>3100 ± 80</td>
<td>TO-5764</td>
</tr>
<tr>
<td>11</td>
<td>1.1 - 1.12</td>
<td>3480 ± 80</td>
<td>S-3372</td>
</tr>
<tr>
<td>12(1)</td>
<td>1.14 - 1.16</td>
<td>3470 ± 80</td>
<td>S-3373</td>
</tr>
<tr>
<td>12(2)</td>
<td>1.14 - 1.16</td>
<td>3660 ± 75</td>
<td>S-3008</td>
</tr>
<tr>
<td>13(2)</td>
<td>1.18 - 1.23</td>
<td>3860 ± 70</td>
<td>S-3374</td>
</tr>
<tr>
<td>13(2)</td>
<td>1.18 - 1.23</td>
<td>3880 ± 70</td>
<td>S-3375</td>
</tr>
<tr>
<td>13(4)</td>
<td>1.27 - 1.34</td>
<td>4280 ± 80</td>
<td>S-3009</td>
</tr>
<tr>
<td>13(4)</td>
<td>1.27 - 1.34</td>
<td>4280 ± 70</td>
<td>S-3007</td>
</tr>
<tr>
<td>13(4)</td>
<td>1.27 - 1.34</td>
<td>4280 ± 70</td>
<td>S-3007</td>
</tr>
<tr>
<td>15</td>
<td>2.02 - 2.05</td>
<td>5010 ± 90</td>
<td>S-3007</td>
</tr>
</tbody>
</table>

Table 6.9: Sedimentation rates at Redtail site calculated from cultural layer depths and radiocarbon dates listed in table 6.8. The average rate of sedimentation for the dated portion of the section is also noted.

<table>
<thead>
<tr>
<th>Layers Compared</th>
<th>Depth of Layers Compared (m)</th>
<th>Sedimentation Rate (mm a⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>9(?) - 11</td>
<td>0.89 - 1.11</td>
<td>0.58</td>
</tr>
<tr>
<td>11 - 12</td>
<td>1.11 - 1.15</td>
<td>0.44</td>
</tr>
<tr>
<td>12 - 13(2)</td>
<td>1.15 - 1.20</td>
<td>0.17</td>
</tr>
<tr>
<td>13(2) - 13(4)</td>
<td>1.20 - 1.30</td>
<td>0.24</td>
</tr>
<tr>
<td>13(4) - 15</td>
<td>2.04 - 1.30</td>
<td>1.01</td>
</tr>
<tr>
<td>average</td>
<td>0.89 - 1.30</td>
<td>0.61</td>
</tr>
</tbody>
</table>

originally backfilled, rendering it useless. A short core, 0.5 m south of the pit was logged and analysed instead of the compacted sediments.

The Redtail site has a complex stratigraphy, particularly in the upper part of the section (Fig. 6.24). The base of the section to 2.82 m is classified as facies 2B, coarsening upwards to facies 3 sediment to 2.62 m below the surface then fines to 2.5 m below the surface (facies 2B). There is a thin sandy (facies 4) unit from 2.5 m - 2.42 m
Figure 6.24: Redtail section diagram. Uncorrected radiocarbon dates are noted beside cultural layers (). (C) indicates a coarse mean grain size; (F) indicates a fine mean grain size.
below the surface. Above the sand unit is a thick (1.95 m - 2.42 m) unit of coarsening upwards facies 2B sediment. A thin gravel unit (facies 5) is located above the facies 2B unit.

The mean and standard deviations of the facies 3 and facies 4 units were compared with the bivariate plots presented in Chapter 5. These plots suggest a fluvial origin for the sediments in the lower part of the section. The interbedded channel sand, gravel and more poorly sorted facies 2 sediment suggests fluctuating stream discharges and sediment loads. Based on the depths recorded by Ramsay (1993), cultural layer 15 should be located underneath the gravel unit indicating subaerial exposure of the site (Fig. 6.24, Table 6.8). No evidence of this cultural layer was observed during the re-excavation.

Above the gravel is a thick, coarsening upwards unit of facies 2B sediment, 1.86 m - 1.09 m below the surface. Thin laminations of clay-enriched sediment were observed near the base of the unit, while cobbles are scattered throughout the upper portion of this unit, above 1.45 m. Three dated cultural layers were reported by Ramsay (1993) at depths indicating they should be in the upper part of the unit but were not observed during the current study. This unit is topped by a thin (1.04 m - 1.09 m) unit classified as facies 3.

This portion of the section records the transition from a fluvial depositional environment to a colluvial depositional environment. As illustrated in chapter 5, the depositional environment of facies 2 cannot be interpreted on the basis of mean grain size and sorting characteristics. Additional information is necessary for interpretation. The
distinct clay laminations in the lower part of the unit indicate a stream environment with fluctuating discharges. The clay would have been draped onto the floodplain or point bars during waning flows. The abrupt shift from channel gravel to sediments containing clay laminations indicates that a change in stream discharge occurred.

The upper part of this unit contains cobbles but no clay laminations. The cobbles would have been introduced from the slopes suggesting that this unit is colluvial. It is not possible to determine if the stream would have reworked this sediment. Likewise, the overlying facies 3 unit lies close to the boundary between sediments that have been re-worked and those that have not on the bivariate plot in Chapter 5. The occurrence of several cultural layers in the upper unit indicates subaerial exposure. Organic enrichment was not observed suggesting that the site was probably experiencing continuous deposition.

Radiocarbon dates indicate that deposition would have changed to mass movement and slope wash processes by ca. 4.3 ka BP and would have been active to at least 3.5 ka BP. Sedimentation rates declined after the deposition of the gravel unit ca. 5 ka BP through the transition from fluvial to colluvial sediments to ca 3.6 ka BP (Table 6.9). After ca. 3.5 ka BP, this trend reversed to increasing rates of sedimentation.

Above facies 3, cobbles are no longer present in the section. A sharply fining upwards sequence of facies 2A (0.99 m to 1.04 m) to facies 1A (0.87 m - 0.99 m) sediment contains several layers of varying colour and organic matter. A weakly developed paleosol extends from 0.99 m to 0.95 m below the surface. All of the paleosols observed in the Redtail section were likely Regosols as they lack both a
Chernozemic A horizon and a B horizon. Above this, the organic carbon decreased slightly then increased at 0.9 m. Root casts extend from the 0.01 m thick paleosol into the layer below. These units record renewed fluvial deposition, but this time on the floodplain. A bone sample obtained from the upper paleosol indicates deposition of vertical accretion sediments by ca. 3.1 ka BP. This constrains the return to fluvial deposition to between 3.5 and 3.1 ka BP.

The domination of fluvial sedimentation continues to the top of the section. A second channel sand unit is located above the organic-rich facies 1A sediment. The sand is overlain by organic-rich facies 2A floodplain sediments which extend to 0.75 m below the surface. Root casts extend into the sand from the organic-rich layer.

The upper 0.25 m of the excavation and the short core are characterised by repeated cycles of coarse-grained facies 2B sediments overlain by finer-grained facies 2A sediment. The changes in mean grain size associated with the repeated cycles of organic-poor sediment likely record deposition by larger flood events. As each flood wanes there is a decrease in mean grain size allowing the accumulation of organic matter. Cultural layers 1 - 8 occur somewhere in these upper units, but were not observed during the current study. There are no radiocarbon dates associated with these cultural layers.

6.5.2 Amisk Site

The Amisk site is located on a colluvial slope, west of the modern stream (Fig. 6.1, 6.2). The western most wall of the original archaeological excavation, 0.93 m above the datum, was re-excavated during the current study (Fig. 6.22).
There is one date reversal in the radiocarbon data base at cultural layer 6 (S-2534) (Table 6.10). This date is rejected as being unreliable. The remaining dates, including the date generated by this study, are in chronological order. All radiocarbon dating at this site was carried out on bone material, except S-2537 which was carried out on charcoal. This resulted in an earlier date than the bone date from the same cultural horizon. The bone date will be used in this study as it has a lower standard deviation and is consistent with the material used to date other units. Large standard deviations were reported for several other dates (S-2536, S-2535, S-2534) calling into question the accuracy of the dates. However, the chronological order and reliance on bone material allows the data to be accepted with a reasonable level of confidence.

This site is unusual in that five of the seven cultural layers originally described could be identified, with reasonable confidence, on the basis of organic carbon and colour (Table 6.10). The location of the two remaining layers was approximated by comparing the original measurements of the depth of the layers with depths in the new excavation. The sediments originally described (Amundson, 1986) do not correspond with the stratigraphy of the new excavation reflecting the different methods used to determine texture.

The excavated section was 2.65 m in depth, bottoming out in a facies 2B matrix-supported boulder layer (Fig. 6.25). Similar high concentrations of boulders were noted on the terrace remnants. The Amisk site boulder layer may have formed during the initial incision of the valley and was subsequently buried.
Table 6.10: Cultural layer depth and uncorrected radiocarbon dates at the Amisk site. Source: Amundson, 1986; Morlan, 1992. The last date listed was obtained during the current investigation.

<table>
<thead>
<tr>
<th>Cultural Layer</th>
<th>Depth (m)</th>
<th>Date (yrs BP)</th>
<th>Sample Number</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.23</td>
<td>480 ± 65</td>
<td>S-2531</td>
</tr>
<tr>
<td>1</td>
<td>0.23</td>
<td>905 ± 155</td>
<td>S-2537</td>
</tr>
<tr>
<td>2</td>
<td>0.40</td>
<td>3055 ± 70</td>
<td>S-2769</td>
</tr>
<tr>
<td>3</td>
<td>0.72</td>
<td>3530 ± 110</td>
<td>S-2767</td>
</tr>
<tr>
<td>4</td>
<td>0.80</td>
<td>4015 ± 195</td>
<td>S-2536</td>
</tr>
<tr>
<td>5</td>
<td>0.95</td>
<td>4120 ± 190</td>
<td>S-2535</td>
</tr>
<tr>
<td>6</td>
<td>1.34</td>
<td>3895 ± 195</td>
<td>S-2534</td>
</tr>
<tr>
<td>7</td>
<td>1.90</td>
<td>5340 ± 120</td>
<td>S-2768</td>
</tr>
<tr>
<td>---</td>
<td>2.15-2.20</td>
<td>5600 ± 70</td>
<td>TO-5762</td>
</tr>
</tbody>
</table>

The lowest unit, classified as facies 2B, extends to between 1.05 and 1.10 m below the surface. The mean grain size of this unit is fairly consistent with a slight fining upward trend. Several cobbles were noted within this unit. Two radiocarbon dates, 5340 ± 120 (S-2768) and 5600 ± 70 (TO-5762), indicate deposition occurred by ca. 5.6 ka BP. The lack of organic carbon accumulation suggests continuous sedimentation throughout this unit. By comparing the depths of dated units at this and the Newo Asiniak site, it can be determined that the facies 2B unit was deposited above the main floodplain. This, and the occurrence of cobbles, indicate that the facies 2B unit would likely have been deposited by a combination of mass wasting and slope wash.

Organic carbon levels increase above this unit changing the facies classification to facies 2A. The undulating boundary between this and the underlying facies 2B unit reflects colour variations due to pedogenesis. The facies 2A unit, which extends to 0.80 m below the surface, includes two identified cultural layers; layer 4 at 0.80 m and
Figure 6.25: Amisk section diagram. Uncorrected radiocarbon dates are noted beside cultural layers (). (C) indicates a coarse mean grain size; (F) indicates a fine mean grain size.
layer 5 at 0.95 m. Each of these cultural layers are associated with increased organic matter indicating pedogenesis. The two layers are separated by a very dark greyish brown unit. The lower paleosol has a well defined Ah horizon, 5 - 10 cm thick, indicating long-term stability in the landscape. There is also some organic enrichment below the Ah horizon, indicating the presence of a Bm horizon. The thickness of the A horizon and presence of a B horizon suggests this may have been a chernozem. The upper paleosol is likely a cumulic Regosol, suggested by a thin Ah horizon.

Above cultural layer 4, there is an organic-poor unit (facies 2B), 0.80 - 0.74 m below the surface. Cultural layer 3 is located in the overlying facies 2A unit. This unit has an undulating upper boundary suggesting erosion. It is not possible to determine how thick the Ah horizon would have been. The radiocarbon dates and slight colour variations in the lower 0.05 m of the unit suggest gradual sedimentation. Low organic carbon and lighter brown colours accompanied by a decrease in mean grain size record the shift to facies 2B extending to 0.43 m below the surface. Sedimentation rates (Table 6.11) indicate this unit was deposited more rapidly than the underlying unit.

Table 6.11: Sedimentation rates at Amisk site calculated from cultural layer depths and radiocarbon dates listed in Table 6.10. The average rate of sedimentation for the dated portion of the section is also noted.

<table>
<thead>
<tr>
<th>Layers Compared</th>
<th>Depth of Layers Compared (m)</th>
<th>Sedimentation Rate (mm a⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 - 2</td>
<td>0.23 - 0.40</td>
<td>0.07</td>
</tr>
<tr>
<td>2 - 3</td>
<td>0.40 - 0.72</td>
<td>0.67</td>
</tr>
<tr>
<td>3 - 4</td>
<td>0.72 - 0.80</td>
<td>0.16</td>
</tr>
<tr>
<td>4 - 5</td>
<td>0.80 - 0.95</td>
<td>1.43</td>
</tr>
<tr>
<td>5 - 7</td>
<td>0.95 - 1.90</td>
<td>0.78</td>
</tr>
<tr>
<td>7 - non-cultural layer</td>
<td>1.90 - 2.15</td>
<td>0.96</td>
</tr>
<tr>
<td>average</td>
<td>0.23 - 2.15</td>
<td>0.38</td>
</tr>
</tbody>
</table>
The texture coarsens upward, changing to facies 3 from 0.43 m to 0.25 m below the surface. Cultural layer 2 is located at the bottom of this unit with root casts extending into the underlying unit. The history of this unit is complex as pedogenesis is associated with the bottom of the unit, not the top of the underlying facies 2B. This indicates that the deposition of the facies 3 sediment was intermittent with initial stability and soil formation ca 3.0 ka BP followed by deposition. There is some organic enrichment above the cultural layer but it is not certain if this developed during deposition or is a result of pedogenesis after the overlying unit had been deposited. The uppermost unit, classified as facies 2A, is a fining-upwards unit. Cultural layer 1 is located at the base of this unit.

The Amisk site was most likely a colluvial site throughout its history; sediment deposition was influenced by a combination of slope wash and mass wasting. While the elevation of the site on the colluvial slope was probably low enough that larger flood events could have reached it, it is not necessary to invoke this situation to account for the sedimentary sequence. There is no evidence in the form of vertical accretion deposits (facies 1), channel gravel (facies 5) or sand (facies 4) to indicate fluvial deposition. The cultural layers and evidence of pedogenesis in the form of organic-rich layers present throughout the section indicate periods of slope stability.

There are some significant discrepancies between the original archaeological and current studies. Cultural layer 7 was reported as being associated with a possible loess and/or coarse sand deposit ascribed to a glacio-lacustrine or glacio-fluvial origin (Amundson, 1986); however, laboratory grain size analysis conducted during the current study did not reveal the existence of either a loess or sand unit. Even allowing for
textural variations arising from different sample locations, the explanation originally presented is clearly impossible as it would require human occupation and deposition of cultural artifacts dated ca. 5.3 ka BP in an ice-proximal environment, 5 ka after deglaciation.

The original depositional explanation related to cultural layer 6, also located with the lower-most unit, is flawed. Large cobbles were noted in the original excavation and confirmed as present throughout the unit by the current study. These were explained as "manuports" (deposited by humans) as there "is no evidence of a natural transport mechanism with the energy to deposit these boulders" (Amundson, 1986, p 52). This is an unsatisfactory explanation as boulders are at present moving down slope via soil creep.

A 4 m long clay lens noted during the original excavation was interpreted as an overbank fluvial deposit. Unfortunately, the lens was not observed during the current study so this interpretation cannot be confirmed.

6.6 Terracettes

Terracettes are found throughout the study site on both east-facing and west-facing slopes. They are most clearly defined in upper slope positions. One set of well developed terracettes located near the mouth of the valley was selected for analysis (Fig. 6.1; Fig. 6.26). The upper 14 terracettes at this location were surveyed in order to determine their slope and configuration. These are normal terracettes with gently
Figure 6.26: Terracettes located on a steep, undercut colluvial slope.
sloping, vegetated treads and steep, vegetated risers (Fig. 6.27). They are roughly parallel, although in some cases, individual treads merge (Fig. 6.28).

A one meter deep trench was dug into terracette #8 (Fig. 6.29). The texture varies slightly with granular muddy sands (facies 2B) at the top and bottom of the section, separated by a thick sand (facies 4). A bivariate plot of facies 4 indicates that there is separation of fluvial and aeolian sands within the study site (Chapter 5). The difference in texture, as well as the position in the landscape, indicates that this sand is aeolian.

An undulating boundary of heavily stained sediments 0.05 to 0.10 m thick, 0.3 m below the surface separates the sand into a poorly sorted, olive brown, upper unit and

![Profile of Terracettes 0 - 13](image)

Figure 6.27: Topographic profile of terracettes. The survey datum is at 0 m. The vertical exaggeration of the profile is 3.4X. Source: 1995 survey data. The location of the profile is indicated on figures 6.1 and 6.28.
Figure 6.28: Plan view of terracettes. Source: 1995 survey data.

moderately well sorted, light yellowish brown, lower unit. Several granular laminations are located within the central sand facies and clay stringers with iron staining were found near the base of the sand at 0.85 m. The staining of the sediments indicates ground water flow.

There has been limited soil-profile development with only a Regosol present. Organic carbon content decreases from two percent at the surface to less than one
percent 0.06 m below the surface. This may be attributed to the instability of the terracettes possibly compounded by the sandy texture of the parent material.

An auger hole into terracette #12 located downslope from the terracette #8 revealed a similar sedimentary sequence (Fig. 6.27; Fig. 6.28). A fining upwards facies 2B unit overlies a pebbly, clay rich sediment, probably till. The sand in this section
contains more fine gravel and is not as well sorted as in terracette #8. The heavily stained unit was not noted in this core.

No dateable material was obtained from either terracette. It can be stated with confidence, however, that terracette formation post-dates the formation of other landforms as they are superimposed on top of cut banks formed during channel migration and small, shallow gullies.

Several mechanisms of terracette formation are outlined in chapter 4. Of these, soil creep is the most likely mechanism to have produced the terracettes occurring within the study site. Rates of soil creep range up to 15 mm a\(^{-1}\) in temperate continental climates and up to 10 mm a\(^{-1}\) in semiarid climates (Young and Saunders, 1986). Although creep is difficult to measure on a short-term basis, numerous turf-covered boulders within the study site suggest that this process is active today.

6.7 Rotational Slide

There is one well preserved rotational slide within the study site (Fig. 6.1; Fig. 6.30) identified by its distinctive slump block which is tilted down towards the headwall. The base of the slump block has been undercut by the modern stream. No dateable material has been obtained from the slide so it is not possible to determine when it occurred.

The lower-most units, classified as facies 2B and facies 3 likely represent colluvial material (Fig. 6.31). Above these units there is a repeated sequences of organic-poor facies 2B sediments overlain by an organic-rich facies 2A unit. The sediment is slightly darker at 0.6 m below the surface indicating pedogenesis. Root
Figure 6.30: Rotational slide. The manmade dam is just visible on the right (centre) edge of the photograph.

casts extend from the lower organic-rich unit into the underlying organic-poor unit indicating that this is a Regosol. The upper soil layer has both an A and weak B horizons suggesting it may be a rego-chernozem.

Several additional sections were cleaned towards the edge of the slump block and beyond. In these sections, the two organic-rich units were closer together. Beyond the slump, only one organic rich unit was observed. This suggests that the lower organic-rich layer may represent the ground surface before the slide occurred. A portion of the slump block would have buried this surface and a new soil subsequently formed at the surface.
Figure 6.31: Rotational slide section diagram. (C) indicates a coarse mean grain size; (F) indicates a fine mean grain size.
Chapter 7

Composite Model of Valley Development

Chapter 7 presents a composite model of valley development. The geomorphology of the valley is integrated with the sedimentology of the individual excavation units, pits and cores presented in Chapter 6. The various mechanisms which may have affected the development of the landscape are also considered.

An important consideration is the lack of chronological control during the early and mid-Holocene. Without adequate chronological control, landform development can only be placed in an event sequence. It is not possible to state when many of these events actually took place. It is also difficult to determine how the various driving mechanisms would have interacted with each other to shape the landscape at a particular site.

7.1 Driving Mechanisms

Changes in climate and base level would have been the primary driving mechanisms behind the formation of the modern landscape. Depending on the direction of the change in these variables, periods of channel incision or aggradation occurred.

Climate change would have influenced both fluvial and hillslope systems causing variations in both stream discharge and sediment load in the stream. The direct influence of base level would have been confined to the fluvial system which, in turn, would have
influenced hillslope systems. An example of this would be oversteepening of a slope through stream migration leading to slumping.

7.1.1 Stream Discharge and Sediment Load

Overland flow and stream discharge would have been high during the early post-glacial cool, moist period, particularly before an abundant vegetation cover was established. Discharge would have decreased as the region became increasingly arid during the Hypsithermal, with peak aridity ca. 6.0 ka BP (Vance et al., 1995). Increased moisture after ca. 6 ka BP would have increased stream discharge. An increase in the vegetation cover as the climate moistened would have tempered the increase in stream discharge. Vegetation reduces overland flow by improving the infiltration capacity of sediment. It also increases the friction on the slopes, which in turn reduces overland flow velocity. Discharge from the Dalmeny Aquifer remains an unknown quantity. If the hydraulic head of the aquifer dropped below the elevation of the valley, the aquifer would have ceased to supply moisture to the stream.

Changes in the vegetation cover would have also influenced the sediment load carried by the stream. Based on the relationship between precipitation and mean annual sediment yield (Langbein and Schumm, 1958), it is expected that sediment yields would have been highest during and immediately after the Hypsithermal (see Fig. 4.17). The post-Hypsithermal increase in vegetation cover would have reduced slope wash, resulting in declining quantities of sediment delivered to the system.
7.1.2 Base Level Fluctuations

The South Saskatchewan River is base level for the study site. Base level for the river would originally have been Glacial Lake Saskatchewan. After retreating northward, Glacial Lake Saskatchewan drained into Lake Agassiz forming a new base level at the Nipawin Delta. This delta formed ca. 10.2 - 9.5 ka BP during the Emerson phase of Lake Agassiz (Christiansen et al., 1995). Lake Agassiz gradually retreated northward with the Laurentide Ice Sheet, draining ca. 7.6 ka BP (Thorleifson, 1996). The South Saskatchewan River incised in response to this drop in base level. Radiocarbon dated paleosols on the Saskatoon terrace indicate this must have occurred by at least 8.1 ka BP (Chapter 4). There is 4.5 - 6 m of alluvium beneath the South Saskatchewan River (Cherry, 1962) indicating incision to 460 m a.s.l., 14 - 15 m below datum.

The river subsequently aggraded until ca. 0.1 - 0.2 ka BP, then incised to its modern level. The proximity of the Tipperary Creek and Juniper Flats alluvial terraces to the South Saskatchewan River indicates that the river must have aggraded to near these terrace elevations, averaging 1 m below the datum. It would not have been possible for the terraces to have built up above the current river level without the river also being higher.

Isostatic rebound would have been initiated with deglaciation and has continued into modern times, raising the portion of the continent formerly under the continental ice sheet. This would cause rivers to incise in an attempt to reach ultimate base level. The migration of channel incision upstream from the oceans would likely have contributed to the post-glacial incision of the South Saskatchewan River. This in turn would have
contributed to incision within the study site. To date, there have not been any estimates of isostatic rebound and its’ effects on landscape development published for central Saskatchewan.

Isostatic rebound would only cause base level to drop. For this reason it cannot be used to explain either the aggradation or modern incision of the South Saskatchewan River or Opimihaw Creek. It is possible that isostatic rebound would have reduced the amount of aggradation within the system.

7.2 Composite Model

A threshold diagram (Fig. 7.1) records changes in the altitude of a stream channel over time. Reversals in stream process from channel incision to aggradation, and vice versa, indicates that geomorphic thresholds were crossed three times. This effectively divides the evolution of the main valley into three phases; post-glacial incision (maximum ca. 10.5 ka BP extending to some time prior to ca. 4.6 ka BP), aggradation (prior to ca. 4.6 ka BP extending to ca. 100 - 200 yrs BP), and modern incision (100 - 200 yrs BP extending to the present time). The aggradation phase of the composite model is based on the changes evident at the fluvial sites (Newo Asiniak, Thundercloud and Tipperary Creek), as these are best represented within the data. Poor chronological control means that it is not possible to identify the reaction time which elapsed prior to the crossing of each threshold.
Figure 7.1: Threshold diagram for the Opimihaw Valley indicating periods of incision and aggradation separated by geomorphic thresholds. Threshold 1 marks the beginning of post-glacial incision. Threshold 2 indicates the shift to an aggrading system while threshold 3 records the beginning of modern incision. Note that the end of the initial incision and early aggradation of the valley are not dated.

7.2.1 Postglacial Incision Phase

The deglaciation of the Saskatoon area represents the maximum age for the incision of the valley (Chapter 4). Current estimates place this event ca. 10.5 ka BP (Thorleifson, 1996). Incision of Opimihaw Creek, driven primarily by base level change, would have post-dated the early stages of the channelisation of flow. Radiocarbon dates obtained from terraces formed during the incision of the South Saskatchewan River cannot be used to date incision within the study site as the features cannot be correlated. Incision of the valley probably occurred soon after deglaciation when the combined
influences of glacial meltwater, extra-glacial water, aquifer discharge and an early post-glacial cool, moist climate would have been strongest.

The depth at which incision ceased is presently unknown. The Opimihaw Creek would not have incised below the lowest level of the South Saskatchewan River, 460 m a.s.l. (14 m - 15 m below the datum), so this represents the maximum possible depth for the valley (Cherry, 1962). Providing chronological control for the end of incision is also problematic. Radiocarbon dates obtained from alluvial deposits at the Newo Asiniak site provide a minimum date of ca 4.6 ka BP. At least 0.85 m of sediment had been deposited by this time so this does not date the beginning of aggradation in the valley. A radiocarbon date obtained from the contact between the till and alluvium in the valley bottom would provide a better indication of the timing of events.

Without knowing when and where incision ceased, it is not possible to determine why it ceased. There are, however, two plausible alternatives. First, the Opimihaw Creek may have reached local base level. A core obtained from the Tipperary Creek site by the Saskatchewan Research Council (data pertaining to the core is no longer available) indicates incision to at least 7 m, possibly 11 m, below datum. Alternatively, incision may have ceased because stream discharge declined as the climate became increasingly arid.

It is most likely that a sinuous or meandering stream was responsible for channel incision. Braided channels are generally associated with channel aggradation as opposed to incision. Unpaired terrace remnants suggest gradual incision with a significant lateral component in response to a gradual drop in base level (cf. Rains and Welch, 1988). The limited extent of the terrace remnants indicates that there has been substantial
modification of the valley. Migration of the stream channel as it incised would have caused undercutting and slumping of the valley walls.

Undercutting would have likely occurred during the aggradation and modern incision phases as well. At the Newo Asiniak site, a meander scar at the base of the slope indicates channel migration and undercutting during the aggradation phase. Adjacent to the Thundercloud and Tipperary Creek sites, the modern channel is in a position for undercutting the banks. Sites like Juniper Flats, where there is no evidence of meander scars at the base of the slope, indicate that the undercutting predated the end of aggradation.

Shallow, broad gullies that extend back into the till plain are located on the west side of the valley on gentle slopes, primarily downstream from the dry tributary channel (Fig. 6.1). These gullies likely formed during the initial incision phase. Some of the gullies terminate part way down the slopes, indicating the elevation of the Opimihaw Creek when they were active. Other gullies, which terminate at the alluvial terrace surface, cannot be placed in this phase with as much confidence, but their morphology suggests contemporaneous formation. Above the Newo Asiniak site, the lower reaches of the broad gullies have been removed as the valley walls were undercut during stream migration. Chronologically, this places the gullies between incision and undercutting. As previously discussed, undercutting was likely active during both the incision and aggradation phases.

Throughout the study site there are narrow gullies which incise into the steep undercut slopes on both the east and west sides of the valley (Fig. 7.2). They do not
extend as far back into the till plain as the gullies discussed previously. These clearly post-date formation of the undercut slopes. Downstream of the man-made dam (Fig. 6.1), one gully has cut through the alluvial terraces to the modern stream, dating it to the modern incision phase. This indicates that gullying has been active throughout the history of the valley.

Figure 7.2: Narrow gullies incised into a steep, undercut colluvial slope.
There are two dry tributary networks on the west side of the valley; a small, second order one at the Dogchild site and a larger, third order one north of the Newo Asiniak site (Fig. 6.1). Both of these tributary networks are the site of park paths which precluded excavation. Parts of both networks have been eroded by undercutting during stream migration. They have also been superimposed with smaller runnels. The tributaries are not dated and could have formed during either the post-glacial incision of the valley or during the early part of the aggradation phase.

Aeolian re-working of the sandy, glacio-lacustrine deposits located on the till plain west of the valley is not constrained by radiocarbon dates. These sediments are found on a steep slope inside a large meander bend bounded on either side by terrace remnants (Fig. 6.1). Re-working would have occurred after the terrace remnants and meander loop were cut but prior to the formation of the terracettes as these features are composed of aeolian sand.

Several periods of aeolian activity have been noted at the site of the Saskatoon Terrace and at Beaver Creek, south of the study site (Turchenek et al., 1974). Burial of paleosols dated 9 940 ± 160 (S-442), 8 160 ± 150 (S-296), and 7 640 ± 150 (S-443) record the initiation of aeolian phases at these sites. There is no indication of the duration of aeolian deposition in the study area. While the Saskatoon terrace, Beaver Creek and Opimihaw Creek sites have not been correlated, it is plausible that the aeolian sand within the study site may also date from the early to mid-Holocene.

Small valleys, such as the one containing the Redtail site, extend west from Opimihaw Creek valley. The oldest radiocarbon date obtained from the Redtail site
indicates incision of the slopes prior to 5.0 ka BP. At this time, it is not possible to further constrain incision. There is no evidence of similar gullies extending eastwards.

7.2.2 Aggradation Phase

Aggradation of the channel would have begun prior to ca. 4.6 ka BP, the earliest date from the Newo Asiniak site. Fluvial sand and gravel deposited at the Redtail site indicates aggradation would have begun outside of the main valley prior to 5.0 ka BP. It is possible that aggradation would have been initiated at both locations at roughly the same time, but this cannot be confirmed without obtaining additional radiocarbon dates. Deposition would have continued to near modern times.

The change to aggradation in the valley generally corresponds with the passage of peak aridity towards the end of the Hypsithermal; a time of change from dry to increasingly moist conditions. Aggradation due to high sediment loads would be the expected result of an increase in precipitation in arid or semi-arid environments (Chapter 4). The vegetation cover would likely have been sparse at this time, offering little protection to the slopes. Runoff during storm events would cause slope wash, filling the stream channel with sediment. The timing of precipitation would have been as important as the absolute amount of precipitation. Concentration of precipitation either during the winter (contributing to high spring runoff) or to a few storm events would cause variable discharge and high sediment loads possibly leading to aggradation. Unfortunately, it is not yet possible to test this hypothesis. The rise in base level described in section 7.1.2 would have also led to aggradation due to a decrease in channel gradient, and therefore stream power.
The dominant landforms developed during this phase of landscape evolution are the alluvial terraces found along the length of the valley. Stream migration would have continued to undercut the valley walls throughout the aggradation phase, eroding terrace remnants, gullies and other landforms dating from the incision phase of development. The sparse vegetation on the slopes during the early aggradation (post peak aridity) phase would have allowed additional gully ing and possibly the formation of the dry tributary networks. There are numerous published accounts of rapid erosion and arroyo formation in the dry, sparsely vegetated American Southwest, leading to aggradation in the main channel downstream (e.g. Womack and Schumm, 1977). Sparse vegetation combined with increased discharge during the early post-Hypsithermal period may have led to similar conditions within the study site.

Contemporaneous dates obtained from interbedded fluvial sand and gravel units at the Newo Asiniak site indicate rapid aggradation ca. 4.6 ka BP. The facies 5 fluvial gravel unit at the Thundercloud site probably dates to this period. This estimate is based on both relative dates and a radiocarbon date of ca. 4 ka BP on the overlying cultural layer 6. These early aggradational sediments may represent deposition in a braided system which later shifted to a meandering system as sediment loads decreased. The Holocene transition from braided to meandering channels has been documented at sites around the world (Schumm and Brakenridge, 1987), but cannot be proved at this site.

After ca. 4.6 ka BP, there is a change in the nature of sedimentation at the Newo Asiniak site to facies 2 lateral accretion and proximal channel deposits indicating that the stream is both aggrading and meandering. Sedimentation rates decreased at this time
The fining-upwards sedimentary sequence at the Thundercloud site also indicates migration of the stream away from the site.

Alternating facies 4 fluvial sand and facies 2 proximal channel units were deposited at the Redtail site indicating channel migration prior to subaerial exposure and human occupation ca. 5.0 ka BP. Fluvial deposition continued until some time prior to 4.3 ka BP. Sedimentation rates declined over the period of fluvial deposition (Fig. 7.3B).

Sedimentation rates were also high on the lower slopes during the early aggradation phase. The lack of organic carbon enrichment in the lower portion of the Amisk site unit dated ca. 5.6 to 4.1 ka BP suggests continuous deposition. The upper date is from a paleosol which indicates stability at this site ca. 4.1 ka BP.

There is a general change in the sedimentary record ca. 4.0 ka BP. At the Newo Asiniak site, sedimentation rates declined (Fig. 7.3A). The wide range in the relative dates at the Thundercloud site preclude the placement of specific events into a chronological framework. Cultural layers 5 and 6 (radiocarbon dated ca. 4.0 ka BP), have contemporaneous relative dates suggesting slow rates of sedimentation. There was also a trend from continuous deposition toward frequent instability recorded in several closely spaced paleosols at the Amisk site ca. 4.1 to 3.5 ka BP. A slight increase in sedimentation rates ca. 4.0 ka BP was followed by a progressive decline (Fig. 7.3C).

These changes suggest that the watershed was becoming more vegetated, possibly induced by a gradually moistening climate. The Waldsea Lake record indicates that conditions were moister than present ca. 4 to 3 ka BP. A thicker vegetation cover would have provided increased slope stability such as recorded at the Amisk site.
that conditions were moister than present ca. 4 to 3 ka BP. A thicker vegetation cover would have provided increased slope stability such as recorded at the Amisk site.

A

Sedimentation Rates at Newo Asiniak Site

B

Sedimentation Rates at Redtail Site
Figure 7.3: Sedimentation rates at the A) Newo Asiniak, B) Redtail, C) Amisk, and D) Tipperary Creek sites.
Increased discharge and a reduced sediment load would result in lower sedimentation rates as the stream would be able to transport its load more easily. There may also have been a change to more stable stream discharge as the climate became more humid. This would be expected to reduce aggradation.

Erosion of a cultural layer dated ca. 3.5 ka BP, followed by deposition of an organic-poor unit, indicates a return to higher sedimentation rates at the colluvial Amisk site. Charcoal fragments in the boulder layer at the Tipperary Creek site indicate that fires occurred within the study site. The temporary removal of the vegetation cover at the Amisk site may have initiated a period of increased slope wash ca. 3.5 ka BP. The lack of paleosols indicates fairly continuous sedimentation prior to a return to formerly stable conditions (Fig. 6.23). A cultural layer located at the bottom of the overlying unit, dated ca. 3.0 ka BP, constrains the end of the period of slope instability. Pedogenesis at the Amisk site records very slow sedimentation and periodic stability after 3.0 ka BP.

Prior to ca. 2.2 ka BP, deposition of point bar and proximal channel sediments at the Newo Asiniak indicates migration of the channel. The upper most unit, containing several dated cultural layers is organic-rich indicating slow sedimentation prior to abandonment of the site. Three cultural layers located within the organic-rich upper unit at the Thundercloud site suggests slow sedimentation at this site as well.

The Saskatchewan Research Council core at the Tipperary Creek site suggests deposition of fine-grained facies 1 vertical accretion deposits at this site may have been initiated fairly early in the aggradation phase. A boulder layer indicates that mass movements occurred in the vicinity of this site roughly estimated ca. 3.0 ka BP (based on
the application of sedimentation rates to undated units). The high resolution dated vertical accretion unit at Tipperary Creek was deposited starting ca. 1.5 ka BP. There is some variation in sedimentation rates with a generally decreasing trend punctuated by a brief period of higher sedimentation rates ca. 0.85 ka BP and ca. 0.175 ka BP (Fig. 7.3D). The increased sedimentation rates correspond with deposition of coarser floodplain sediments reflecting short term larger flood events rather than a general change in the environment.

The average sedimentation rates are higher at the Tipperary Creek site than at other fluvial sites (Table 7.1). The years of record are not directly comparable, however, as only the upper section of the Tipperary Creek site has been dated. The dated section indicates deposition of nearly 1 m of sediment in the last 1.5 ka BP. In comparison, 0.1 - 0.25 m of sediment was deposited at the Thundercloud site, 0.25 m of sediment was deposited in the meander scar and less than 0.25 m on the main terrace surface at the Newo Asiniak site. This difference in sedimentation between the valley mouth and upstream reaches is evident in the elevations recorded on the threshold diagram (Fig. 7.1). More rapid sedimentation at the mouth of the valley than in upstream reaches is expected where a rise in base level has led to aggradation.

The Redtail site does not fit into the composite model developed for the Opimihaw Creek valley during the aggradation phase. The periods of slope stability recorded by pedogenesis are out-of-phase at the two colluvial sites (Redtail and Amisk). At about the same time that the Amisk site is stabilising and undergoing pedogenesis (4.1 to 3.5 ka BP), the Redtail site shifted from a fluvial to colluvial depositional
Table 7.1: Comparison of average sedimentation rates at three fluvial sites within the Opimihaw valley, calculated from uncorrected radiocarbon dates and depths of cultural layers. The upper Thundercloud date is a relative date based on tool traditions. Sources: Kelly, 1986; Lis Mack, personal communication 1996; Morlan, 1992.

<table>
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<tr>
<th>Site</th>
<th>Layers Compared</th>
<th>Depths of Layers Compared (m)</th>
<th>Years Compared</th>
<th>Average Sedimentation Rates (mm a⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tipperary Creek</td>
<td>1 - 13</td>
<td>0.19 - 0.96</td>
<td>&lt; 100 - 1535</td>
<td>0.54</td>
</tr>
<tr>
<td>Newo Asiniak</td>
<td>1 - non-cultural</td>
<td>0.06 - 1.47</td>
<td>750 - 4630</td>
<td>0.36</td>
</tr>
<tr>
<td>Thundercloud</td>
<td>1 - 6</td>
<td>0.05 - 0.46</td>
<td>100* - 4040</td>
<td>0.12</td>
</tr>
</tbody>
</table>

environment (ca. 4.3 ka BP). No evidence of slope stability, in the form of pedogenesis, was noted at the Redtail site that corresponds with the paleosols at the Amisk site at this time. Renewed sedimentation and a lack of pedogenesis at the Amisk site corresponds with the shift back to fluvial deposition and the start of pedogenesis at the Redtail site. The core data obtained from the slopes above the Dogchild and Thundercloud sites do not help to resolve this issue as neither core is chronologically constrained.

Numerous cultural layers and organically-enriched units dating to the last 3.0 ka BP indicate periodic flooding followed by pedogenesis and re-occupation of the Redtail site. This type of sedimentation is similar to that noted at the Tipperary Creek site, but the sediments themselves have little in common indicating deposition in a different environment. The Redtail section indicates deposition within and proximal to a channel throughout most of this period while the Tipperary Creek site indicates a distal channel location. There are no noted similarities between the Redtail and Newo Asiniak or Thundercloud sites during this period.
These inconsistencies illustrate the need for both larger sections and an improved radiocarbon data base within the study site. Trenches extending across individual terrace surfaces and into the colluvial slopes would allow a better interpretation of the interaction between hillslope and fluvial sites. This might allow the discrepancies between the main valley and the Redtail site to be resolved.

7.2.3 Modern Incision

At some point during the last 100 - 200 yrs BP, a threshold was crossed and the Opimihaw Creek began to incise its floodplain. The radiocarbon date from cultural level 1 at Tipperary Creek, found 0.19 m below the surface, indicates incision within the past 100 years BP. Relative dates from cultural level 1 at the Thundercloud site also indicates incision may have occurred as recently as 100 years BP. A modern radiocarbon date (185 ± 190, S-2763) from the uppermost sediments deposited in the meander loop at the Newo Asiniak site likewise indicates recent abandonment. Small terrace surfaces (1 - 3.5 m below datum) were formed at the Newo Asiniak site and around the Tipperary Creek and Juniper Flats sites during the modern incision phase.

The modern floodplain is at a steeper gradient than that of the alluvial terraces. The creek is approximately 2 m below the Dogchild terrace, 2.5 m below the Newo Asiniak terrace and 4.75 to 5 m below the main terrace surfaces by the time it has reached the Juniper Flats and Tipperary Creek sites. This variation in channel gradient is evident in the topographic profiles presented throughout chapter 6. The change in elevation indicates that the modern stream has not reached base level along its length and theoretically is still incising. Currently, there is very little flow in the creek, particularly in
the upstream reaches, limiting the amount of incision that can actually occur. Aerial photographs taken in 1987 indicate that flow actually ceased upstream from the study site. Numerous beaver dams along the length of the creek have caused ponding, further limiting channel erosion and incision.

Sediment carried by Opimihaw Creek was deposited in the South Saskatchewan river forming a small delta. This has been exposed and dissected by the creek during low flow periods. The delta was inundated by the river during the high discharge floods across the prairies during the summer of 1995. There is a similar deposit at the mouth of the valley containing the Redtail site, although this delta in not dissected by a perennial stream.

A drop in base level would have caused this most recent period of incision. There are no records of a change in the elevation of the South Saskatchewan River channel, however, the fact that the river is 6 m below the Tipperary Creek site terrace indicates that it did incise. As discussed previously, it would not have been possible for the alluvial terraces within the study site to have formed with the South Saskatchewan river at its current elevation. Incision would have occurred during the same time period as incision within the Opimihaw Creek valley. There are no records of environmental change in central Saskatchewan corresponding with this most recent period of incision. Placing the incision within the context of what is known about global changes in the last 100 - 200 years would require a better understanding of when incision was initiated.

There are two remaining landscape elements which have not been constrained chronologically. There is a large, well-preserved rotational slide located upstream from
the man-made dam (Fig. 6.1; 6.30). There has been little modification to this block, suggesting that the slide is fairly recent, but a lack of comparable slides precludes an age estimate. The base of the slide has been undercut on the downstream side where the modern stream flows through a culvert in the dam. The slide was likely caused by the stream undercutting the base of the original valley wall. This may have happened during stream migration during the later stages of the aggradation phase. Alternatively, incision of the stream may have caused the bank to be oversteepened, resulting in the slide. Ponding of the modern stream behind the dam has raised the water level so that it is not possible to determine where the modern channel was located or how much incision occurred. Based on the trend of decreasing incision towards the head of the stream, approximately 1 m of incision would have occurred. Knowing where the modern channel is located may have allowed the age of the slide to be estimated. If the modern channel was at the base of the slide, the slide would probably have dated from the incision phase. If the modern channel was separated from the base of the slide by an alluvial terrace, the slide would have occurred before incision, during the aggradation phase.

There are several well developed sets of terracettes on both the east and west valley walls, located on the steep undercut slopes (Fig. 6.1, 6.26). No dateable material was recovered from the excavated terracettes, but they appear to be fairly recent landforms. The terracettes extend across the gullies located within the meander bends, post-dating both undercutting and gully formation. The terracettes may date from either the aggradation or modern incision phases of the valley.
Chapter 8

Summary and Directions for Future Research

8.1 Summary

The story of the Opimihaw creek began ca. 10.5 ka BP when the Saskatoon area was deglaciated and Glacial Lake Saskatchewan drained. Channelisation of flow into the South Saskatchewan River (local base level) allowed small tributary streams, such as the Opimihaw Creek, to form. As the South Saskatchewan River incised, likely in response to the gradual draining of Glacial Lake Aggasiz, the meandering Opimihaw Creek incised through both surficial sediments and till. A combination of an early post-glacial cool, moist climate, glacial and extra-glacial water and aquifer discharge would have removed sediment from the study site.

The migrating stream undercut the valley walls forming large meander loops, eroding away floodplains dating to earlier stages of post-glacial incision. Only small, isolated terrace remnants perched on the colluvial slopes have been preserved. It is likely that this migration and undercutting have also occurred during the aggradation and modern incision phases of landscape formation.

There are several features within the study site that likely date to the incision phase of valley formation but which cannot be constrained with greater precision. Two small tributary networks and numerous shallow broad gullies were incised into the
colluvial slopes. Some of the gullies have been partially removed by stream migration, while others are still discernible along the full length of the slopes. Aeolian re-working of glacio-fluvial sediments deposited on the till plain west of the valley led to the deposition of well sorted sands on the upper colluvial slopes.

Prior to ca. 4.6 ka BP, the crossing of a geomorphic threshold caused the stream to shift from an incising to an aggrading system. This was likely due to a combination of a rise in base level and climate change. Although peak aridity of the Hypsithermal would have passed, the climate would still have been dry. Slope wash, resulting from a sparse vegetation cover, would have overloaded the stream with sediment whenever precipitation occurred. This, combined with the fluctuating discharge characteristic of drier environments, may have led to the rapid aggradation recorded at the Newo Asiniak site.

As the environment became increasingly moist, a denser vegetation cover would have become established reducing slope wash. This change is recorded in a series of paleosols at the Amisk site. A reduced sediment load combined with increasingly regular discharge led to a gradual decline in sedimentation rates at the Newo Asiniak site.

Deposition of an organic-poor unit at the Amisk site indicates that the general trend toward reduced sedimentation was interrupted ca 3.5 - 3 ka BP. Charcoal fragments in the Tipperary Creek boulder layer matrix suggests fire may have occurred. This would remove the vegetation cover leading to a temporary increase in slope wash. The colluvial phase at the Redtail site is out of phase with the Amisk site. Periods of
reduced sedimentation and pedogenesis do not occur at the same time at both sites. A satisfactory explanation for this cannot be forwarded at this time.

As the stream aggraded, it migrated laterally across its floodplain. This migration is recorded in the interbedded channel lag, channel sands and proximal channel sediments at both the Newo Asiniak and Thundercloud sites. The nature of sedimentation was not uniform across the floodplain. Downstream, at the Juniper Flats and Tipperary Creek sites, repeated deposition of fine-grained vertical accretion sediments was followed by pedogenesis. The downstream reaches were also characterised by more rapid sedimentation rates as expected where there has been a rise in base level. A similar series of repeated flooding followed by pedogenesis is recorded in the upper fluvial phase at the Redtail site.

Within the last ca. 0.1 to 0.2 ka BP, a geomorphic threshold was crossed and the stream began to incise into its floodplain. The proximity of the South Saskatchewan river to the Juniper Flats and Tipperary Creek terraces indicates that the Opimihaw Creek was likely responding to a change in base level. The modern floodplain is at a steeper gradient than the floodplain recorded in the valley bottom alluvial terraces. This suggests that the stream has not reached base level. Currently, there is very little flow in the creek, due in part to several beaver dams along its length.

Normal terracettes are found along the length of the valley on the steeper colluvial slopes. They most likely formed by soil creep, a process that has remained active through to modern times. The terracettes cannot be chronologically constrained, but likely formed during the aggradation or modern incision phases of valley formation. Likewise,
a single well-preserved rotational slide is not chronologically constrained. This feature may also date from either the aggradation or modern incision phases.

Repeated occupation of the Opimihaw valley is recorded in the numerous cultural layers at both alluvial and colluvial sites. The earliest dated layer is at the Amisk site, recording occupation ca. 5.3 ka BP. The Tipperary Creek site has the best record of occupation with 13 radiocarbon dated cultural layers since ca. 1.5 ka BP.

8.2 Directions for Future Research

There are several avenues for future research that could be explored within the study site, that would improve and extend the work conducted during this study. Little or no disturbance of the study site would be required in order to pursue these lines of research.

It has long been recognised that sedimentary facies which are found stacked upon each other in the sedimentary record (in continuous sections) are those which are found next to each other at the present. This rule of succession is commonly known as Walther’s law (Miall, 1990). Similar facies sequences will be repeated in a particular section if a depositional environment shifts back and forth over time. An example of this would be the migration of a stream away from a site recorded in a fining-upwards sequence of channel lag, point bar and finally vertical accretion sediments. If the stream were to migrate back towards the site, a second cycle of sediments would be deposited.

The cyclical nature of sedimentary deposits may be investigated statistically through Markov chain analysis (Miall, 1990). Markov chain analysis is used to predict which facies is most likely to be found next in a section, based on the what the preceding
facies is, allowing composite models of facies successions to be developed. A composite model of expected facies transitions within the study site would aid in the interpretation of future sections. It should be noted, however, that this would just be a general guideline and that individual sections should not be expected to conform exactly to the model.

One of the more significant limitations during the course of this study was the inability to excavate in locations likely to be of great geomorphic and sedimentological significance. Excavation was confined to small pits, exposing less than one horizontal metre of sediment. This meant that there was no opportunity to observe either the internal architecture or three dimensional changes across the sites. Several of the alluvial terraces have not had archaeological excavations, limiting the current study to coring, providing an even more limited picture of the nature of the sediments. The use of ground penetrating radar may provide additional information on the large scale changes in the sediment record preserved in the alluvial terraces. This technique is non-invasive causing minimal disturbance.

Pollen and plant macrofossil studies would be a valuable addition to knowledge of the site. A better understanding of vegetation changes through time would increase the reliability of estimates of slope stability and sediment delivery to the stream. This type of study would also improve our understanding of environmental changes since deglaciation. Radiocarbon dates obtained from individual macrofossils would improve the chronology of the site. It is likely, however, that pollen and macrofossil preservation would be limited to sites with anaerobic conditions.
There are several lines of research that could be pursued outside of the study site. A better understanding of fluctuations in the South Saskatchewan River would improve interpretations of possible driving mechanisms behind the changes at the site. This may include dating the postglacial incision, aggradation and modern incision of the river, describing the nature and thickness of alluvium beneath the river in the vicinity of the study site, and determining what caused the changes in the river. A system of this size would be expected to be responding to a more complex interaction of variables than the Opimihaw Creek valley so establishing an accurate picture of change would be a large undertaking.

On a smaller scale, there are several valleys similar to the Opimihaw Creek which drain into the South Saskatchewan River. Studies similar to the current one could be done at these sites allowing a broader picture of post-glacial landscape change to be developed. A better understanding of the regional response to climate and base level change may help to answer the questions raised by the Redtail site. In time, it may be possible to separate out site specific and broader response signals.

Landscape evolution studies in central Saskatchewan will continue to be highly speculative until there is a better understanding of Holocene climate change. There is plenty of room for the investigation of the fossil record. This might include plant macrofossils, pollen, molluscs, ostracodes and insects. A high-resolution, interdisciplinary approach to investigating climate change, such as the Palliser Triangle Project, would be highly beneficial.
Bibliography


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APPENDIX A

Modified Loss on Ignition Procedure and Calculations
Modified Particle Size Analysis Procedure and Calculations
Modified Loss on Ignition Method

The following loss on ignition method for the determination of percent organic carbon and percent inorganic carbon in a sediment sample is taken from Dean (1974) and Scott (1991).

Procedure
Notes: Crucibles must be engraved as pencil markings are removed in the 1 000°C burn. Tongs must be used to lift crucibles as oil from hands will affect results.
1. A powdered sample weighing approximately 10 g is dried in an oven at 90° - 100°C for a minimum of one hour and then cooled to room temperature in a dissector.
2. Approximately 5 g of this dried sample is placed in a crucible that has been pre-weighed to 0.0001 g. The sample and crucible are then weighed to 0.0001 g in order to determine the dry weight of the sample to be analysed.
3. The samples and crucibles are placed in a 550°C muffle furnace for one hour then cooled to room temperature in a dissector and weighed to 0.0001 g.
4. The samples and crucibles are then placed in a 1 000°C muffle furnace for one hour, cooled to room temperature in a muffle furnace then weighed to 0.0001 g.
5. Each sediment sample is analysed at least twice to provide a check.

Calculations
The percent of organic carbon in the sample is calculated as follows.

\[
\% \text{ organic carbon} = \frac{\text{mass of oven-dried soil before ignition}}{\text{mass of soil after 550°C burn}} \times \frac{\text{mass of soil after 550°C burn}}{\text{mass of oven-dried soil before ignition}} \times 100
\]

(Appendix A.1)

The percent of inorganic carbonate in the sample is calculated as follows.

\[
\% \text{ loss} = \frac{\text{mass of soil after 550°C burn}}{\text{mass of soil after 1 000°C burn}} \times \frac{\text{mass of soil after 1 000°C burn}}{\text{mass of soil after 550°C burn}} \times 100
\]

(Appendix A.2)

\[
\% \text{ carbonate content} = \frac{\% \text{ loss}}{0.44}
\]

(Appendix A.3)
Modified Particle Size Analysis Method

The following particle size analysis method is taken from Day (1965), Isphording (1972), Kunze (1965), Lewis and McConchie (1994) and Scott (1991).

Procedure

Sample Preparation
1. Place 30 - 50g of air dry sample in labelled 400 ml beaker. Should use more if gravely, less if fine.
2. Remove root hairs, bone, wood fragments etc... rinsing with distilled water to ensure that no sediment particles are on the removed material.
3. Cover the sample with 3% hydrogen peroxide to a depth of 2 - 4 cm. Place on hot plate ensuring that temperature does not exceed 90°C. Add more peroxide as needed till all organic material has been digested. Stir regularly.
4. Record weight of a clean dry 2 mm sieve and labelled 600 ml beaker.
5. Wet sieve sample through 2 mm sieve using distilled water.
6. Dry screen and sample in oven and record weight. The > 2 mm fraction (gravel) may be stored for later analysis if required.
7. Dry beaker and sample and record weight of < 2 mm fraction (sand, silt and clay).
8. Add exactly 50 ml of 10% calgon solution to sample in beaker. Wash into baffle cup with distilled water and stir for 15 min.
9. Record weight of clean, dry 63μ screen.
10. Wet sieve the sample into labelled 1 000 ml cylinder using distilled water. Fill cylinder to 1 000 ml with distilled water. Cover with watch glass and allow to continue defloculating for at least 24 hours. before proceeding with pipette analysis (silt and clay fraction).

Dry Sieving
1. Record mass of clean dry sieves. Half phi intervals should be used (1.4 mm, 1 mm, 710 μm, 500 μm, 325 μm, 250 μm, 180 μm, 125 μm, 90 μm, 63 μm).
2. Place sand sample in top of first nest of sieves (1.4 mm, 1 mm, 710 μm, 500 μm, 325 μm and pan).
3. Place in Ro-Tap for 15 min.
4. Record weights of sieves and sample.
5. Place sample from pan into the top of the next nest (250 μm, 180 μm, 125 μm, 90 μm, 63 μm and pan). Record the weight of the pan for the second sieving.
6. Record weights of sieves and sample.
7. The sample can now be discarded.
**Pipette Analysis**
1. Fill fish tank with distilled water and let sit for at least two days to allow temperature to equilibrate with room temperature.
2. Assemble 5 clean small beakers for each sample to be processed. Label then weigh to 0.0001 g.
3. Place 1 000 ml cylinders in the fish tank.
4. Record temperature in fish tank and check extraction times at this temperature (Table 43-1 Day, p 548, 1965)
5. Record all extraction times (time zero, 4 min. - 20µm, 1 hour - 5µm, 7 hours - 2µm and 29 hours - 1µm) for each sample to be processed.
7. Immediately upon completion of stirring insert 20 ml pipette to a depth of 10 cm and extract the first sample from the suspension. This extraction (time zero) is critical. Expel the sample into labelled beaker.
8. Fill pipette with distilled water and discharge into the beaker. Rinse the pipette into the beaker a second time.
9. Place beaker in oven to dry. Record weight to 0.0001 g.
10. Repeat steps 7 - 9 at the specified times. Do not re-stir sample between extractions.

**Preparation of Calgon Solution**
1. Dissolve 50g of calgon in 1000 ml of distilled water.
2. Weigh a clean dry beaker to 0.0001 g.
3. Place exactly 50 ml of calgon solution into beaker and dry.
4. Record weight to 0.0001 g.

**Calculations**

*Initial Sample Preparation*

Mass of gravel.

\[
\text{mass of gravel} = \text{mass 2 mm sieve} - \text{mass 2 mm sieve and sample} \\
\text{ (Appendix A.4)}
\]

Mass of sand, silt and clay.

\[
\text{mass sand, silt and clay} = \text{mass 600 ml beaker} - \text{mass 600 ml beaker and sample} \\
\text{ (Appendix A.5)}
\]

Mass of sand.

\[
\text{mass sand} = \text{mass 63 µm sieve and sample} - \text{mass 63 µm sieve} \\
- \text{mass < 63 µm after dry sieving} \\
\text{ (Appendix A.6)}
\]
Mass of silt and clay.

\[ \text{mass silt and clay} = \text{mass sand, silt} - \text{mass sand} + \text{mass < 63 \(\mu\)m and clay} \]

After dry sieving

(Appendix A.7)

Mass of sample fraction on sieve after dry sieving.

\[ \text{mass of sample on} \ n \ \text{sieve} = \text{mass of soil and on} \ n \ \text{sieve} - \text{mass of} \ n \ \text{sieve after sieving} \]

(Appendix A.8)

Total mass of original sample.

\[ \text{total mass} = \text{mass gravel} + \text{mass sand, silt and clay} \]

(Appendix A.9)

Calculation of percent of total mass that each size fraction represents.

\[ \text{percent of total mass} = \frac{\text{mass of sample fraction}}{\text{total mass}} \times 100 \]

(Appendix A.10)

**Pipette Analysis**

Pipette analysis calculations are to be completed in conjunction with table.

<table>
<thead>
<tr>
<th>extraction time</th>
<th>mass sample and calgon</th>
<th>mass sample</th>
<th>mass</th>
<th>percent still in suspension</th>
<th>percent change</th>
<th>percent of original sample</th>
</tr>
</thead>
<tbody>
<tr>
<td>zero time</td>
<td></td>
<td></td>
<td></td>
<td>100</td>
<td></td>
<td></td>
</tr>
<tr>
<td>4 min.</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1 hour</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>7 hours</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>29 hours</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>colloidal clay</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Mass of sample and calgon.

\[ \text{mass sample} = \text{mass beaker and and calgon} - \text{mass beaker and calgon extraction} \]

(Appendix A.11)
Mass of sample alone.

\[
\text{mass sample} = \text{mass sample} - \text{mass calgon and calgon}
\]

(Appendix A.12)

Mass of sample multiplied by 50.

\[
\text{mass} = \text{mass sample} \times 50
\]

(Appendix A.13)

Percent of silt and clay still in suspension at \( n \) time.

\[
\% \text{ in suspension at } n \text{ time} = \frac{\text{mass at } n \text{ time}}{\text{mass at zero time}} \times 100
\]

(Appendix A.14)

Percent change from previous reading.

\[
\% \text{ change} = \frac{\text{previous } \% \text{ in suspension} - \% \text{ in suspension at } n \text{ time reading}}{\% \text{ in suspension at } n \text{ time reading}}
\]

(Appendix A.15)

Percent of original sediment sample at this size.

\[
\% \text{ original sample} = \% \text{ change} \times \% \text{silt and clay}
\]

\[
\frac{100}{100}
\]

(Appendix A.16)

Percent colloidal clay

\[
\% \text{ colloidal clay} = \% \text{silt and clay} - (\% 4 \text{ min} + \% 1 \text{ hour} + \% 7 \text{ hours} + \% 29 \text{ hours})
\]

(Appendix A.17)

Percent silt

\[
\% \text{silt} = \% 4 \text{ min} + \% 1 \text{ hour}
\]

(Appendix A.18)

Percent clay

\[
\% \text{ clay} = \% 7 \text{ hours} + \% 29 \text{ hours} + \% \text{ colloidal clay}
\]
Calgon Solution

Mass of calgon in 1000 ml cylinder.

\[
\text{mass calgon in cylinder} = \text{mass of beaker} - \text{mass of beaker and calgon}
\]

Mass of calgon in 20 ml pipette sample.

\[
\text{mass calgon} = \frac{\text{mass calgon in cylinder}}{1000} \times 20
\]
APPENDIX B

Formulas and Verbal Scales for Graphic Size Parameters
Formulas and Verbal Scales for Graphic Size Parameters

The phi (ϕ) units corresponding to each of the percentiles indicated in the formulas are determined from a cumulative frequency percent graph plotted on probability paper. The formulas are directly from Folk and Ward (1957). These equations consider the region between the 5th and 95th percentile. Earlier equations generally only considered the central portions of the curves, not taking into account the tails present within most distributions.

**Graphic Mean**

\[ M_z = \frac{\phi_{16} + \phi_{50} + \phi_{84}}{3} \]

(Appendix B.1)

**Inclusive Graphic Standard Deviation**

\[ \sigma_1 = \frac{\phi_{84} - \phi_{16} + \phi_{95} - \phi_{5}}{4.6} \]

(Appendix B.2)

- < 0.35ϕ  very well sorted
- 0.35 to 0.50ϕ well sorted
- 0.50 to 0.71ϕ moderately well sorted
- 0.71 to 1.0ϕ moderately sorted
- 1.0 to 2.0ϕ poorly sorted
- 2.0 to 4.0ϕ very poorly sorted
- > 4.0ϕ extremely poorly sorted
APPENDIX C

Radiocarbon Date List
Years BP refers to uncorrected radiocarbon date published in referenced source.

Calibrated dates are given in years BP unless otherwise stated.
<table>
<thead>
<tr>
<th>Years BP</th>
<th>Calibrated Date (BP)</th>
<th>Sample Number</th>
<th>Material</th>
<th>Location</th>
<th>Comments</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>10230 ± 140</td>
<td>ASA-D95-1</td>
<td>Wood</td>
<td>Andrews Site, SK 50°20'N 105°52'W</td>
<td>500 cm below surface</td>
<td>Yansa, 1995</td>
<td></td>
</tr>
<tr>
<td>10200 ± 140</td>
<td>ASA-D95-2</td>
<td>Wood</td>
<td>Andrews Site, SK 50°20'N 105°52'W</td>
<td>455 cm below surface</td>
<td>Yansa, 1995</td>
<td></td>
</tr>
<tr>
<td>2340 ± 70</td>
<td>Beta-6507</td>
<td>plant fiber matrix</td>
<td>Waldsea Lake, SK</td>
<td>Lake core 27, 190-199 cm</td>
<td>Last and Schweyen, 1985</td>
<td></td>
</tr>
<tr>
<td>2920 ± 70</td>
<td>Beta-6508</td>
<td>Disseminated organics in clay matrix</td>
<td>Waldsea Lake, SK</td>
<td>Lake core 27, 283.5-290 cm</td>
<td>Last and Schweyen, 1985</td>
<td></td>
</tr>
<tr>
<td>1230 ± 50</td>
<td>Beta-6891</td>
<td>Disseminated organics in clay matrix</td>
<td>Waldsea Lake, SK</td>
<td>Lake core 27, 86-90 cm</td>
<td>Last and Schweyen, 1985</td>
<td></td>
</tr>
<tr>
<td>3970 ± 90</td>
<td>Beta-6892</td>
<td>plant fiber matrix</td>
<td>Waldsea Lake, SK</td>
<td>Lake core 27, 379-384 cm</td>
<td>Last and Schweyen, 1985</td>
<td></td>
</tr>
<tr>
<td>12 100 ± 160</td>
<td>GSC-1319</td>
<td>Peat</td>
<td>49°47'N 98°35'W Rossendale, MB</td>
<td>Phase six date. Depth 4.26 m in alluvium in channel</td>
<td>Klassen, 1972</td>
<td></td>
</tr>
<tr>
<td>Age (±)</td>
<td>Sample ID</td>
<td>Type</td>
<td>Location</td>
<td>Depth (cm)</td>
<td>Notes</td>
<td></td>
</tr>
<tr>
<td>--------</td>
<td>-----------</td>
<td>------</td>
<td>----------</td>
<td>------------</td>
<td>-------</td>
<td></td>
</tr>
<tr>
<td>6000 ± 170</td>
<td>GSC-1335</td>
<td>Organic Sediment</td>
<td>Cycloid Lake, SK</td>
<td>257-262 cm</td>
<td>Mott, 1967; 1972</td>
<td></td>
</tr>
<tr>
<td>14300 ± 320</td>
<td>GSC-1369</td>
<td>Organic detritus</td>
<td>49°14'N 98°14'W, Thornhill, MB</td>
<td>Phase four date. Depth 3.65 m near base of alluvium in gully</td>
<td>Teller, 1976</td>
<td></td>
</tr>
<tr>
<td>7590 ± 220</td>
<td>GSC-1506&lt;sup&gt;1&lt;/sup&gt;</td>
<td>Organic sediment</td>
<td>50°52'N 107°56'W, Clearwater Lake, SK</td>
<td>1250-1260 cm. Uncorrected age.</td>
<td>Mott, 1973</td>
<td></td>
</tr>
<tr>
<td>9310 ± 150</td>
<td>GSC-1506&lt;sup&gt;2&lt;/sup&gt;</td>
<td>Carbonate sediment</td>
<td>50°52'N 107°56'W, Clearwater Lake, SK</td>
<td>1250-1260 cm. Uncorrected age.</td>
<td>Mott, 1973</td>
<td></td>
</tr>
<tr>
<td>7580 ± 220</td>
<td>GSC-1506&lt;sup&gt;1&lt;/sup&gt; &amp; GSC-1506&lt;sup&gt;2&lt;/sup&gt;</td>
<td></td>
<td>50°52'N 107°56'W, Clearwater Lake, SK</td>
<td>8&lt;sup&gt;13&lt;/sup&gt;C corrected age.</td>
<td>Mott, 1973</td>
<td></td>
</tr>
<tr>
<td>1120 ± 190</td>
<td>GSC-1563&lt;sup&gt;1&lt;/sup&gt;</td>
<td>Organic sediment</td>
<td>50°52'N 107°56'W, Clearwater Lake, SK</td>
<td>60-70 cm. Uncorrected age.</td>
<td>Mott, 1973</td>
<td></td>
</tr>
<tr>
<td>1260 ± 190</td>
<td>GSC-1563&lt;sup&gt;2&lt;/sup&gt;</td>
<td>Carbonate sediment</td>
<td>50°52'N 107°56'W, Clearwater Lake, SK</td>
<td>60-70 cm. Uncorrected age.</td>
<td>Mott, 1973</td>
<td></td>
</tr>
<tr>
<td>1170 ± 190</td>
<td>GSC-1563&lt;sup&gt;1&lt;/sup&gt; &amp; GSC-1563&lt;sup&gt;2&lt;/sup&gt;</td>
<td></td>
<td>50°52'N 107°56'W, Clearwater Lake, SK</td>
<td>8&lt;sup&gt;13&lt;/sup&gt;C corrected age.</td>
<td>Mott, 1973</td>
<td></td>
</tr>
<tr>
<td>Age (±)</td>
<td>Sample Code</td>
<td>Type</td>
<td>Location</td>
<td>Age Date</td>
<td>Notes</td>
<td></td>
</tr>
<tr>
<td>--------------</td>
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<td>-----------------</td>
<td>---------------------------------</td>
<td>-------------------</td>
<td>------------------------------------------------</td>
<td></td>
</tr>
<tr>
<td>11 090±160</td>
<td>GSC-642</td>
<td>Organic lake</td>
<td>West of North Battleford, SK</td>
<td>Site 4, basal lake sediments</td>
<td>Mott, 1967</td>
<td></td>
</tr>
<tr>
<td>8 520±170</td>
<td>GSC-643</td>
<td>Organic lake</td>
<td>55°16'N 105°16'W Cycloid Lake, SK</td>
<td>basal lake sediments - 323-328 cm</td>
<td>Mott, 1967</td>
<td></td>
</tr>
<tr>
<td>10 260±170</td>
<td>GSC-647</td>
<td>Organic lake</td>
<td>53°48'N 106°04'W Waskesiu, SK</td>
<td>basal lake sediments - 496-504 cm</td>
<td>Mott, 1967</td>
<td></td>
</tr>
<tr>
<td>11 560±640</td>
<td>GSC-648</td>
<td>Organic sand</td>
<td>53°14'N 105°43'W Prince Albert, SK</td>
<td>Phase seven date. Base of pond deposit, 5.9-6.2 m</td>
<td>Lowden et al., 1967; Mott, 1967</td>
<td></td>
</tr>
<tr>
<td>10 690±190</td>
<td>GSC-677</td>
<td>Wood</td>
<td>50°58'N 101°24'W Shellmouth, MB</td>
<td>From alluvium below slump deposits 3 m above modern Assiniboine River floodplain</td>
<td>Klassen, 1989</td>
<td></td>
</tr>
<tr>
<td>10200 ± 80</td>
<td>GSC-5822</td>
<td></td>
<td>Andrews Site, SK 50°20'N 105°52'W</td>
<td>490 cm below surface</td>
<td>Yansa, 1995</td>
<td></td>
</tr>
<tr>
<td>11 610±450</td>
<td>GX-2254</td>
<td>Organic silt</td>
<td>53°46'N 106°54'W Ladder Valley, SK</td>
<td>Phase seven date. 10 m below surface under till</td>
<td>Christiansen, 1979</td>
<td></td>
</tr>
<tr>
<td>13 900±400</td>
<td>I-1268</td>
<td>Spruce wood</td>
<td>Bemis Moraine, central Iowa</td>
<td>Base of till overlying silt and sand.</td>
<td>Teller et al., 1980</td>
<td></td>
</tr>
<tr>
<td>Date ± Range</td>
<td>Site Code</td>
<td>Material</td>
<td>Location</td>
<td>Context Description</td>
<td>Source</td>
<td></td>
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<td>----------</td>
<td>---------------------</td>
<td>--------</td>
<td></td>
</tr>
<tr>
<td>14 200±500</td>
<td>I-1402</td>
<td>Spruce wood</td>
<td>Bemis Moraine, central Iowa</td>
<td>In upper 0.3 m of loess overlain by till.</td>
<td>Teller et al., 1980</td>
<td></td>
</tr>
<tr>
<td>13 900±240</td>
<td>I-3476</td>
<td></td>
<td>Sewell Lake, MB</td>
<td>Phase four date. Upper Assiniboine Delta deposits</td>
<td>Ritchie, 1976</td>
<td></td>
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<td>Christiansen, 1968</td>
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<td>15 850 ± 225</td>
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<td>Teller et al., 1980</td>
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<td>Spruce wood</td>
<td>Bemis Moraine, central Iowa</td>
<td>Within loess beneath till</td>
<td>Teller et al., 1980</td>
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APPENDIX D

Radiocarbon Dates Obtained During Study
Newo Asiniak Site Date

Sample Number: NA-0719-B-32
Depth of Sample: 1.47 m below surface
Description of Sample: Bison hyoid fragment
Description of Sediments: Facies 5 - organic-poor muddy sandy gravel.
Munsell Colour: 2.5Y5/5 moist (matrix).
Weight Used (mg): 2300
Isotrace Lab Number: TO-5763
Age (years BP): 4630 ± 60

**Calibration Report**

Calibrated ages calculated by calibration program C14CAL using dendro calibration data.

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<th>95.5 % confidence interval (1σ)</th>
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<td>100%</td>
<td>3365 cal. BC</td>
<td>3380 BC - 3345 BC</td>
<td>3525 BC - 3300 BC</td>
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Redtail Site Date

Sample Number: RT-0707-B-18R

Depth of Sample: 0.90 m below surface

Description of Sample: Bison humerus shaft fragment.

Description of Sediments: Facies 1A - organic-rich sandy mud. Munsell Colour: 2.5Y2/2 moist.

Weight Used (mg): 3300

Isotrace Lab Number: TO-5764

Age (years BP): 3100 ± 80

Calibration Report

Calibrated ages calculated by calibration program C14CAL using dendro calibration data.

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<th>68.3 % confidence interval (1σ)</th>
<th>95.5 % confidence interval (1σ)</th>
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231
Amisk Site Date

Sample Number  A-0731-B-05
Depth of Sample  2.15 - 2.20 m below surface
Description of Sample  Bison proximal metatarsal
Description of Sediments  Facies 2B - organic-poor slightly gravelly muddy sand. Munsell Colour: 10YR8/1 moist, 2.5Y6/2 moist
Weight Used (mg)  3300
Isotrace Lab Number  TO-5762
Age (years BP)  5600 ± 70

Calibration Report

Calibrated ages calculated by calibration program C14CAL using dendro calibration data.

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232
APPENDIX E

Detailed Sample Descriptions:
Particle Size, Organic and Inorganic Carbon Content, Colour,
Statistical Parameters and Facies Classification
**Munsell Colour**
D - dry colour
M - moist colour
U - undulating boundary between units

**Organic and Inorganic Carbon**
Determined by modified loss-on-ignition procedure outlined in Appendix A.

**Percent Gravel, Sand, Silt and Clay**
Determined by sieving and pipette procedures outlined in Appendix A.

**Class**
Based on Folk and Ward triangular classification diagrams (Fig E.1).

G - gravel
mG - muddy gravel
msG - muddy sandy gravel
sG - sandy gravel
gM - gravelly mud
gmS - gravelly muddy sand
gS - gravelly sand
(g)M - slightly gravelly mud
(g)sM - slightly gravelly sand mud
(g)mS - slightly gravelly muddy sand
(g)S - slightly gravelly sand
S - sand
cS - clayey sand
mS - muddy sand
zS - silty sand
sC - sandy clay
sM - sandy mud
sZ - sandy silt
C - clay
M - mud
Z - silt

**Mean and Standard Deviation**
Calculated according to procedure outlined in Appendix B
Reported in phi (φ) units.
Figure E.1: Triangular diagram used to determine the textural class of a sediment. A) Used for samples containing gravel. Based on the percentage of gravel and the ratio of sand to mud. B) Used for samples that do not contain gravel. Based on the percentage of sand and the ratio of silt to clay. Adapted from Folk, 1968.
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<td>1b</td>
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Newo Asniak Site

NA-0719-GS-02

NA-0719-GS-03
NA-0719-GS-04

UO.09+0.11-

10YR2/1 to

UO.29+0.31

10YR2.5/2

UO.32+0.34-

5.46

1.38

0.5

60.34

20.07

19.09

(g

10YR3/2.5

4.38

3.23

2.11

53.54

21.45

22.9

(g

10YR5/3

4.13

14.28

10YR4.5/3

2.33

14.44

6.17

69.57

12.67

11.59

g

5.08

71.17

12.42

11.33

g

13.47

13.3

(g

;

NA-0719-GS-06

0.37

I

NA-0719-GS-07

0.37-

UO.47+0.48
NA-0719-GS-08

UO.47+0.480.54

NA-0719-GS-09

0.54-0.59

IOYR5/3

NA-0719-GS-12

0.64-

IOYR3/3

1.79

11.71

10YR2.5/3

2.07

11.4

2.5Y4.5/2

1.74

9.95

3.02

70.21

UO.68+0.71
-

NA-0719-GS-13

UO.68+0.71UO.72+0.75

NA-0719-GS-15

UO.80+0.85-

N
W

0.93

'"

NA-0719-GS-17

1.00-1.09

2.5Y5/4

1.06

7.77

2.37

78.22

9.73

9.69

(g

NA-0719-GS-IS

1.09-1.15

2.5Y5/3

2.1

16.41

0.44

52.14

24.53

22.89

(g

NA-0719-GS-20

1.19-1.24

matrix

32.57

51.01

10.03

6.38

m

42.69

52.32

3.55

1.43

m

2.5Y5/4

NA-0719-GS-21

1.25-1.37

matrix

0.52

31.58

0.71

7.06

0.27

88.36

4.44

27.6

0.12

37.03

35.32

27.53

14.3

65.7

10.27

9.73

g

36.81

52.83

6.32

4.04

m

0.92

94.8

2.5Y5/5

NA-0719-GS-24

1.47-1.61

2.5Y5.5/4

NA-0719-GS-25

1.61-1.625

2.5Y6/2

NA-0719-GS-26

1.625-1.66

2.5Y5/4

NA-0719-GS-29

1.83-1.92

matrix

6.94
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2.13

(g)

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(g

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-_

-

10YR5/3
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NA-0719-GS-30

1.92-2.09

2.5Y6/4

0.38

10.45

NA-0720-GS-40
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2_.Q9-�._!i__ �Y5.5/2

0.95
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2.03
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2.25
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21.14
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