THE UNIVERSITY OF ALBERTA

A COMPARISON OF TWO SOILS FROM THE MILK RIVER RIDGE, SOUTHWEST ALBERTA

BY

J. ANTHONY BRIERLEY

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE OF MASTER OF SCIENCE IN PEDOLOGY

DEPARTMENT OF SOIL SCIENCE

EDMONTON, ALBERTA

SPRING, 1988

ISBN 0-315-42800-7

THE UNIVERSITY OF ALBERTA

RELEASE FORM

NAME OF AUTHOR: J. ANTHONY BRIERLEY

TITLE OF THESIS: A COMPARISON OF TWO SOILS FROM THE MILK RIVER RIDGE, SOUTHWEST ALBERTA

DEGREE: MASTER OF SCIENCE

YEAR THIS DEGREE GRANTED: SPRING 1988

J Brierley

12603 49 Ave

Edmonton Alberta

Date: April 28/88

THE UNIVERSITY OF ALBERTA

FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled A COMPARISON OF TWO SOILS FROM THE MILK RIVER RIDGE, SOUTHWEST ALBERTA submitted by J. ANTHONY BRIERLEY in partial fulfillment of requirements for the degree of MASTER OF SCIENCE

M. J. Dudas
Supervisor

C. L. Kershaw

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Date: ~~9 A~~ Apr/28/88

ABSTRACT

The Del Bonita Plateau is an unglaciated area on the Milk River Ridge in southwest Alberta. There are many unique periglacial features associated with this unglaciated plateau including ice wedge casts, involutions and asymmetric valleys. The soils in this area were mapped in a recent soil survey of the M.D. of Cardston. Orthic Black Chernozemic soils from the Plateau, the Del Bonita series, and from the adjacent area developed on Laurentide till, the Beazer series, were sampled and analyzed. These soil series have similar gross profile morphologies and occur in the Fescue Grass Ecoregion.

There have been various suggestions about the origin of the Del Bonita soil series parent material ranging from fluvial to eolian. By comparing the results of physical, chemical, mineralogical and micromorphological analyses a basis of hypothesizing the origin of the Del Bonita parent material was obtained. Texture, pH and organic matter values were similar for the two soils. $CaCO_3$ equivalent values of the Del Bonita soil were consistently greater than the Beazer soil. Similarly the calcite and dolomite X-ray diffraction peaks of the two soils differed indicating different weathering regimes or geomorphic processes between the two parent materials. The suite of clay minerals (smectite, mica and kaolinite) were

identical for both soils indicating that the Del Bonita material is likely derived from Laurentide drift. Heavy and light mineral fractionation indicated preferential selection of quartz minerals within the Del Bonita material. Because the heavy mineral content and proportion of feldspars are consistently less within the Plateau materials, they are probably of eolian origin. Quartz sand grain surface features were analyzed using the scanning electron microscope. Del Bonita sand grains surface features suggested that eolian and fluvial processes were important in their evolution. The Beazer grains displayed features characteristic of morainal environments. The Del Bonita soil thin sections exhibited unique micromorphological features. Conglomeric and fuscitic fabric, products of periglacial processes, were present in the Del Bonita Ck but absent in the Beazer samples.

The origin of the Del Bonita series parent material appears to be loess, which originated from a Laurentide drift source area. This eolian material was subsequently altered by periglacial activity. These processes altered some diagnostic characteristic of loess, such as grain size distribution, complicating the identification of the Del Bonita parent material.

ACKNOWLEDGEMENTS

The author is indebted to Agriculture Canada, Land Resource Research Centre, by whom he was employed during the course of the research project; to Dr. M. Dudas, thesis supervisor, for his guidance and direction during the project from its embryonic stages to the final manuscript, and for the financial support of the laboratory analysis at the university; and to Dr. W. Pettapiece for initiating the research project, and his unfailing encouragement, and suggestions throughout this endeavour.

Numerous people assisted in the field work of the M.D. of Cardston soil survey. A. Rodvang, A. Kjearsgaard, J. Tajek as well as W. Pettapiece assisted in developing the conceptual map units for describing the distribution of soils on the Plateau and the neighbouring areas. Discussions and the occasional field trips with Dr.s A. Stalker and R. Barendregt as well as the exchange of ideas in the office with Dr. I. Shetsen, were all helpful in piecing together the geological history of the surficial materials in the study area.

The guidance and assistance of Dr. S. Pawluk in reviewing the micromorphological descriptions and photographing the soil thin sections is acknowledged.

The technical support of W. McKean, A. Schwartzer and J. Beres of the Agriculture Canada - Alberta Research Council soils lab for performing routine soils analysis on the samples; P. Yee for instructing me in the operation of laboratory equipment in the Soils Department; and G. Bayrock for the direction in operating the scanning electron microscope are acknowledged. In particular, I extend my thanks to M. Abley for his contribution in preparing the thin sections as well as developing and modifying computer programs for the production of X-ray diffraction patterns.

Discussions regarding different aspects of the project with numerous personnel of the Alberta Soil Survey community and fellow graduate students contributed to the success of this thesis.

Last, but not least, sincere appreciation is extended to Diana, my wife, for her moral support, continued encouragement, forbearance and patience throughout the production of this thesis.

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CHAPTER I. INTRODUCTION

The Del Bonita Plateau is a unique area within the Milk River Ridge Upland in the south west corner of Alberta. Surficial geologists Calhoun (1906), Stalker (1962), Westgate (1968), and Shetsen (1980) have recognized the Plateau as being unglaciated during the Wisconsin Ice Age. As a nunatak, the Del Bonita Plateau has many similarities with a more fully researched unglaciated area, the Cypress Hills, which lie in the south east corner of Alberta.

In 1982 Agriculture Canada soil survey personnel initiated a soil survey of the M.D. of Cardston which included the Del Bonita Plateau. The Plateau was distinctive from the surrounding area on the Milk River Ridge in that it was level. Ck horizons on the Plateau contained noticeably more carbonates than neighboring soils on typically hummocky morainal landscapes but otherwise the soil morphologies were very similar. Both soils are classified as Orthic Black Chernozemics.

The loam textured parent material of the Del Bonita soil has been described by Stalker (1962) and the U.S.D.A. (1980) as alluvium. Whether the origin of the alluvium was from the Rocky Mountains or from Laurentide till remained unanswered. Also, the similarities of this area with the Cypress Hills suggested that the material could be loess (Westgate 1968; Catto 1981).

During the routine soil survey procedures when the distribution of different soils in the area was mapped, these queries regarding the origin of the parent materials of the Del Bonita Plateau remained unanswered. Therefore in 1983, this project was initiated in an attempt to provide some answers regarding the origin of the Del Bonita surficial materials. Samples from the soils and parent materials on the level landscapes positions of the Plateau were compared with soils developed on Laurentide till in the neighboring area of the Milk River Ridge. The samples were analyzed using detailed mineralogical and physical analytical procedures and their similarities and differences compared in order to determine the origin of the Del Bonita surface materials.

CHAPTER II. DESCRIPTION OF STUDY AREA IN THE SOUTHWESTERN ALBERTA SETTING

A. Geographic Setting

1. Location

The Del Bonita Plateau and the associated area studied within this project are situated within the Municipal District of Cardston in the southwest corner of the province of Alberta (Figure 1). The plateau is approximately 20,000 hectares in size with the majority occurring on the Canadian side of the International Border. The remainder of the plateau extends south into the state of Montana. The area lies between $48^{o}90'$ and $49^{o}10'$ north latitude and $112^{o}20'$ and $113^{o}00'$ west longitude. The study area, on the Canadian side of the border, is in Township 1, Ranges 21 and 22, west of the 4th meridian.

The plateau (where the hamlet of Del Bonita resides) is approximately 80 km south of Lethbridge, 45 km south of Magrath and 40 km southeast from the town of Cardston. The Cypress Hills, an area which will be referred to frequently, is approximately 200 km east-northeast of Del Bonita on the Alberta-Saskatchewan border. The Rocky Mountains lie 75 km to the west of the study area.

The Plateau is delineated by prominent landscape features. The northern and western edges are

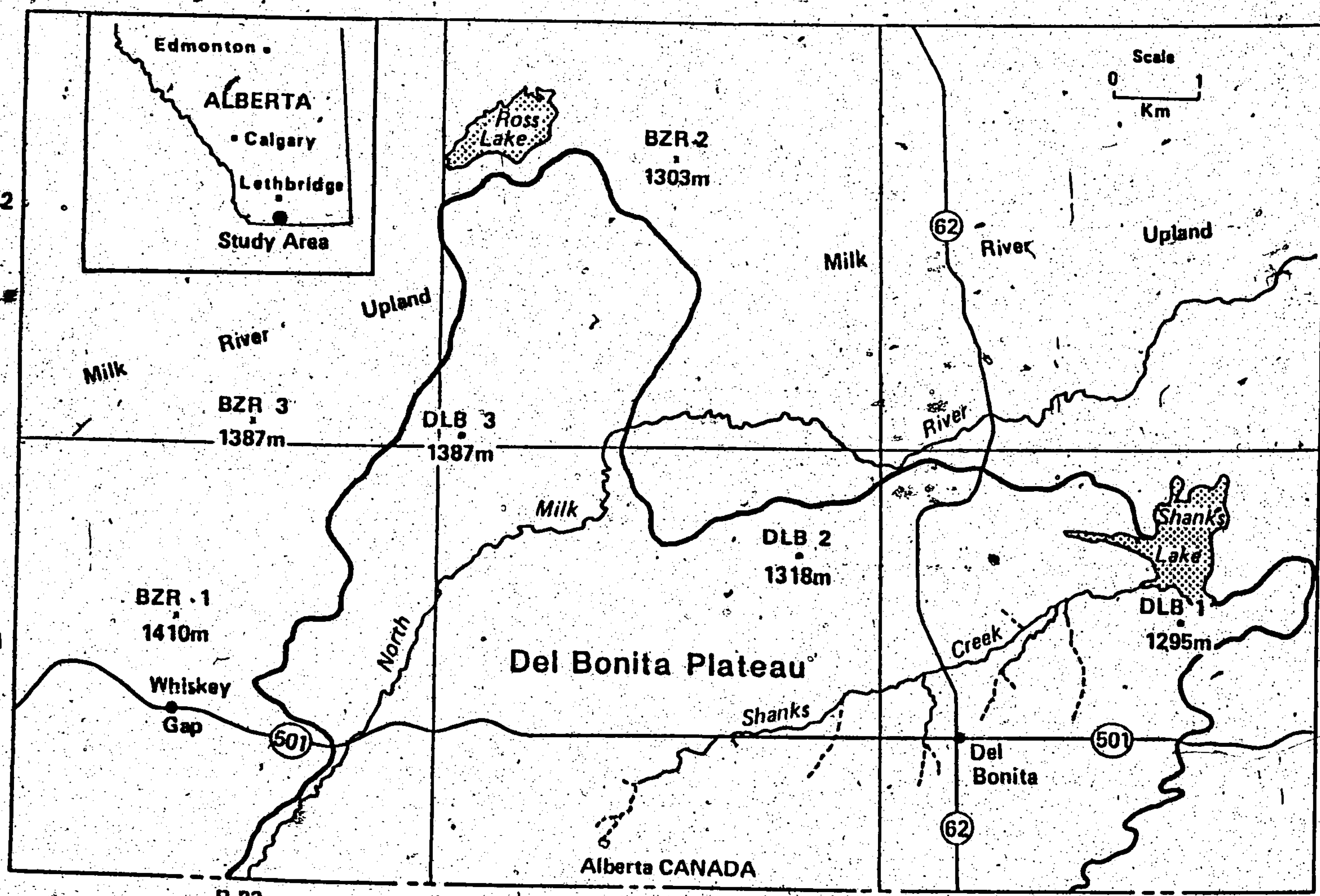

Figure 1 Map of Study Area with Site Locations and Physiographic Areas

approximately outlined by the north fork of the Milk River. The eastern boundary is defined by steep slopes associated with a large preglacial erosional valley. This valley joins with an intricate erosional channel network to the south, which dissects a series of unglaciated plateaus in Montana.

The study area also includes the adjacent area roughly within a 2 km radius north, and northwest of the Del Bonita margin. This is part of the Milk River Ridge where the elevation is greater than 1270 m above sea level (asl). Most of the Plateau is at 1300 masl, with some outlying unglaciated areas at 1390 masl.

The Milk River Ridge forms part of the Continental Divide separating south and east flowing rivers. The two forks of the Milk River drain into the Missouri-Mississippi river system which ultimately flows into the Gulf of Mexico. Just to the west of the study area, the rivers of the Rocky Mountain foothills, principally the St. Mary, Belly and Waterton flow northeastward and join the Hudson's Bay drainage network.

2. Vegetation, Climate, Soils and Land Use

The study area occurs in the Fescue Grass Ecoregion which is characterized by a modified cool continental climate (Strong and Leggat 1981) and occurs within the

Black soil zone (Brierley et al. 1986). The native vegetation of the Fescue Grass Ecoregion, synonymous with the Festuca scabella Association recognized by Moss and Campbell (1947), is dominated by rough fescue *(Festuca scabella)* in the well drained landscape positions. Under ideal conditions this lush grass forms large bunches or tussocks from 25-50 cm in diameter and 30-90 cm in height (Moss and Campbell 1947). Parry oatgrass *(Danthonia parryi)* is found in secondary amounts within this ecosystem.

As environmental conditions change other grass and forb species appear. On moister north aspects, shrubs such as buckbrush *(Symphoricarpos occidentalis)* and wolf willow *(Elaeagnus commutata)* occur (Strong and Leggat 1981). Conversely on drier south aspect slopes or coarser textured parent materials fescue loses dominance to Parry oatgrass and speargrass *(Stipa* spp.) (Moss and Campbell 1947). When overgrazed rough fescue is replaced with increaser species such as lupine *(Lupinus* argenteus), sage *(Artmesia* spp.), and sedges *(Carex* spp.) in depressional areas.

The climate is strongly influenced by the Rocky Mountains to the west. Chinook winds moderate the winter temperatures such that the winters are significantly milder than a true continental climate. The December to February temperatures for this ecoregion are the warmest in the province because of these chinook winds (Strong and Leggat 1981). In fact, the wind blows incessantly in this area with an average of only fifteen "calm" non-windy days per year recorded at the Lethbridge airport weather station (Environment Canada 1982).

The study area is within the 2A(H) zone on the Agroclimatic Map of Alberta (Bowser 1967). This zone is characterized by precipitation limiting crop growth roughly 50% of the time and frost damage for wheat being a hazard (Bowser 1967). This broad definition is supported by the climatic data available for the area. Annual precipitation is approximately 400 mm on average and the frost-free period is around 90-100 days (U.S.D.A. 1980) (Table I). The moisture deficit balance, as defined by the Alberta Agrometeorology Advisory Committee (1986) for this portion of the province is approximately -350 mm. (This value is calculated using Baier's methodology on the available climatic data - precipitation minus potential evapotranspiration)

Table I. Climatic Data for the Del Bonita study area and vicinity

	Total precip (mm)	Frost-free period (days)	Degree days (>5°C)	Mean daily temp (°C)	Agro-climatic zone
Babb, Montana[1]	490	60	-	-	-
Cardston, Alberta[2]	550	111	1543	4.8	2H
Cutbank, Montana[1]	292	108	-	-	-
Del Bonita, Alberta[2]	397	-	1390	4.3	2A(H)
Lethbridge, Alberta[2] (airport)	423	124	1775	5.3	2A
Milk River, Alberta[2]	316	124	1788	5.2	2A
Pincher Ck, Alberta[2] (town)	589	106	1395	4.1	2H

data from [1]U.S.D.A. (1980) and [2]Brierley et al. (1986)

The Del Bonita Plateau is in the Black Soil zone, with Orthic Black Chernozemic being the typical soil. Dark Brown soils are dominant immediately to the east and to the north of the study area. The surface Ah layer is generally 10 cm thick under native vegetation and the depth to lime is between 30-40 cm. The depth of Ah, organic matter content, and depth to lime increases as one proceeds west of the Plateau concomitant with increased precipitation. By contrast, Dark Brown soils are found in the Lethbridge area despite receiving more precipitation than Del Bonita (Table I). The controlling climatic factor is the mean annual temperature. At Del Bonita, the mean daily temperature is $1^{o}C$ less than to the north. Therefore, because the area is cooler, there is less evaporation and a smaller moisture deficit, and Black soils occur on the western portion of the Milk River Ridge.

The majority of the Del Bonita Plateau is or has been cultivated. Wheat and barley are the dominant grain crops, while other areas are used for forage production and grazing of sheep and cattle on improved pastures. Areas adjacent to the Plateau that are hummocky and steep are dominated by native range and are extensively grazed by cattle.

B. Physiography of the Study Area

1. Physiographic significance, surface characteristics, and areal extent

The Del Bonita Plateau and adjacent area on the Milk River Ridge is a unique and intriguing region of southern Alberta. Unlike the majority of the land surface in southern Alberta, the Plateau was not covered by ice sheets during the Wisconsin Ice Age. The fact that the Plateau like the better known example of the Cypress Hills was a nunatak during the last glacial period has been recognized by numerous American and Canadian surficial geologists (Calhoun 1906; Alden and Stebinger 1913; Alden 1932; Horberg 1954; Stalker 1962; Westgate 1968; Shetsen 1980; Prest 1984).

The principal evidence indicating that the area escaped glaciation is based upon the composition of the surface material coarse fragment content and the overall surface form. The surficial materials of the Plateau lack granitic stones or fragments transported from the Canadian Shield by Laurentide ice sheets. The presence of these rock fragments is a key characteristic used to trace the extent of continental ice sheets (Dawson 1895; Wagner 1966; Beaty 1975; Stalker 1977). The surficial material on the Milk River Ridge directly to the north of the Plateau, within the study area, has been identified as till of Laurentide origin, based on the presence of such granitic fragments (Alden 1913 and 1932; Stalker 1962).

The other striking feature of the Del Bonita Plateau is the surface form. The area appears level (flat as a table top) to the naked eye. The overall slope of the plateau is approximately 0.4%. It slopes down from the southwest to the northeast - away from the Rocky Mountains. Also present on the plateau are a series of distinct asymmetrical valleys which run in approximately the same direction as the overall plateau slope. These series of valleys all empty into Shanks Creek which cuts the Plateau in half on the Canadian side by flowing in a more easterly direction before entering Shanks Lake.

In comparison, the surface form of the region, to the north, east and west of the Plateau are typical morainal landscapes. Hummocky landforms with the slopes ranging from 5-20% characterize this area.

South of the International Border, the Del Bonita Plateau extends for approximately 3 km before a series of preglacial erosional valleys terminate this remnant pediment surface (Calhoun 1906; Alden 1913). Within Montana the unglaciated region extends east to Cutbank, Longitude 112°10' to 113°20' on the west, and south to latitude 48°20'. Till of Cordilleran origin surrounds this recognized unglaciated area to the west and south of the International boundary. East of longitude 112°10' on both sides of the border, and west and north of the Del Bonita Plateau on the Canadian side of the 49th parallel, the till is of Laurentide origin (Calhoun 1906; Alden and Stebinger 1913) (Figure 2).

2. Bedrock Geology and Surficial Materials

The study area is underlain with bedrock of three Cretaceous formations. The Bearpaw formation, the oldest of the three is composed almost exclusively of dark grey to brownish grey marine shales, which weathers to small

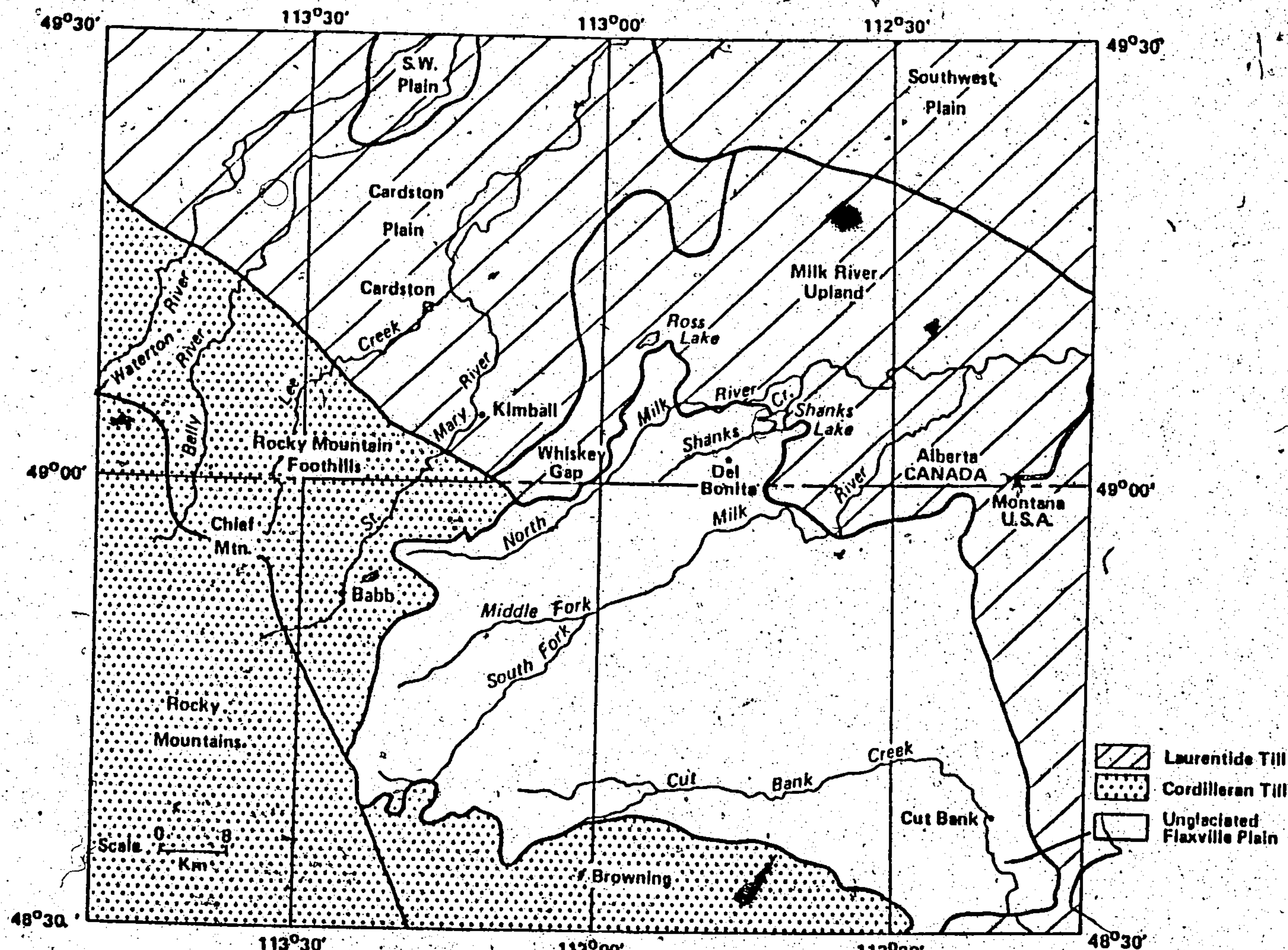

Figure 2 Map of Physiography of Southwestern Alberta and Distribution of Pleistocene Deposit in SW Alberta and NW Montana.

Physiographic lines from Pettapiece 1986, and distribution of Pleistocene Deposits based on Calhoun's 1906, Alden and Stebinger's 1913 map with alterations on the Canadian side based on the M.D. of Cardston Soil Survey.

angular fragments. This formation is restricted to the eastern portion of the study area.

Directly to the west, adjacent to and overlying the Bearpaw formation is the slightly younger aged Blood Reserve formation. Green (1972) recognized a thin, north-south orientated band of this ,dominantly sandstone composed formation in the study area. The sandstone is medium grained, light grey to buff grey in color of marine and non-marine origin.

The youngest and most extensive of the three bedrock types is the St. Mary formation, occurring on the western part of the study area. It is distinguished by alternating layers of sandstone and shales (Russel and Landes 1940; Irish 1968; Green 1972). The sandstone layers are prominent as they are harder and more resistant to weathering then thenon-marine greenish grey shales which are poorly bedded and friable.

The Laurentide till north of the Del Bonita Plateau is of early Wisconsin age (Alden 1913; Horberg 1954; Westgate 1968; Stalker 1962 and 1977; Shetsen 1980). This material is of variable thickness usually greater than 5 m where the surface form is undulating and hummocky, but thinner on prominent bedrock ridges. In certain locations gravels of unknown origin are exposed beneath the till overlying the bedrock. These gravels are referred to as Saskatchewan gravels (Stalker 1962).

A different sequence of surficial materials is found on the Del Bonita Plateau. Directly above the bedrock is approximately 6 m of pink quartzitic gravels, but the thickness of these gravels varies from 1.5-15 m (Shetsen 1980). These gravels are of late Tertiary to early Pleistocene [illegible] and are referred to as Flaxville gravels (Calhoun 1906; Alden 1913; Horberg 1954; Stalker 1962). They are definitely of pre-Wisconsin origin (Shetsen 1980). The source of these gravels has been traced to quartzitic bedrock outcrops in the Rocky Mountains, in the state of Montana (Collier and Thom 1917; Dr A. Stalker pers. comm. 1982)[1].

Overlying these Flaxville gravels is a deposit of material of unknown origin. The average depth of this material is 2 m, but it ranges from 0-6 m across the Plateau (Shetsen 1980; Brierley et al. 1986). Surficial geologists have differed in opinion as to its origin. An alluvial origin has been suggested by several (Wyatt et al. 1939; Stalker 1962; U.S.D.A. 1980). Since other portions of the Flaxville gravel plains have been overridden by glaciers a morainal origin has also been suggested (Alden 1913; Horberg 1954). More recently the

[1] Dr. A. Stalker, Geological Survey of Canada, Ottawa, Ontario.

material has been referred to as cryoturbated fluvial-eolian (Dr. I. Shetsen pers. comm. 1982)[2].

This material overlying the Flaxville gravels on the Del Bonita Plateau has received little attention from surficial geologists (Dr. A. Stalker pers. comm. 1983)[3]. The majority of the literature is concerned with the occurrence of remnant erosional plains and the origin of the Flaxville gravels. However, the surficial material capping the gravels on the Cypress Hills has been extensively studied by Westgate (1968) and Catto (1981 and 1983). From their research they have postulated that the material on the Cypress Hills is loess. Since there are landform and surficial material similarities between the Del Bonita Plateau and the Cypress Hills, it would appear that the material overlying the Flaxville gravels on the Plateau is loess, as well.

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[2] Dr. I. Shetsen, Alberta Research Council, Edmonton, Alberta.

[3] Dr. A. Stalker, Geological Survey of Canada, Ottawa, Ontario.

CHAPTER III. LITERATURE REVIEW

A. Geomorphic History

1. Introduction

The Del Bonita Plateau is a remnant erosional surface which has been preserved since the pre-Wisconsin period because of several coincidental factors. The principal factor is its geographic location. The study area straddles the east-west Continental Divide of North America. This has preserved the whole of the Milk River Ridge from extensive fluvial erosion. Alden and Stebinger (1913) also noted that because the Milk River drainage system drains the Milk River Ridge and associated Del Bonita Plateau, this upland area has remained relatively unaffected by fluvial erosive processes since deglaciation.

2. Blackfoot Peneplain and associated benches

The sequence of events leading to the present form of the Del Bonita Plateau begins with the formation of the Blackfoot peneplain, originally described by Willis (1902) reported in Alden and Stebinger (1913). Willis' original concept of the Blackfoot peneplain was a plane surface formed of flat-lying Cretaceous and Tertiary beds of rock in the zone of folding and faulting bordering the mountain front in Montana. This plane surface was thought to have formed in the early Tertiary so it was perhaps an antecedent to the Lewis overthrust.

Alden and Stebinger (1913) considered the Blackfoot peneplain surface almost non-existent in southern Alberta. They confined the use of the term "Blackfoot peneplain" to the highest benches existing in Glacier Park in Montana.

Because there were difficulties in applying the Blackfoot peneplain term to such elevated plateaus at lower levels, Alden (1932) developed a widely accepted chronological sequence of benches or plateaus based upon Collier and Thom (1917) findings. These are still widely accepted. Horberg (1954) summarized and correlated these associated benches to geographic locations in southern Alberta. Four benches and their approximate ages are recognized and collectively referred to as the Flaxville surfaces (Collier and Thom 1917).

No. 3 bench	Late Pleistocene
No. 2 bench	Early Pleistocene
No. 1 bench (Flaxville)	Pliocene
No. 0 bench (Cypress Plain)	Oligocene

The No. 0 bench which is also referred to as the Cypress Plain is of Oligocene age (Alden 1932; Westgate 1968). This age was determined from fossils found by Cope (1891), reported in Collier and Thom (1917) and Westgate (1968). This plain represents the initial level at which rivers flowed from the Rocky Mountains out onto the plains. Examples of this oldest bench are the Cypress Hills of southeastern Alberta and Flattop Mountain near the St. Mary Lakes in Montana. The elevation of the Cypress Hills is between 1441 m (4800 ft) and 1111 m (3700 ft) above sea level sloping from west to east. Flattop Mountain which is closer to the Rocky Mountains is at an elevation of 2492 m (8300 ft) above sea level (Horberg 1954).

The next oldest plain is the Flaxville Plain or No. 1 bench, believed to be of Miocene or early Pliocene age. The age is based upon vertebrate remains recovered from the gravels (Collier and Thom 1917; Alden 1932). Westgate (1968) places the Flaxville Plain 210-300 m (700-1000 ft) below the Cypress Plain in northeastern Montana, but up to 450 m (1500 ft) lower near the Rocky Mountains. Therefore in southwestern Alberta this second bench is thought to be represented by the flat surface on Makowan Butte which is at an elevation of 1711-1801 m (5700-6000 ft) above sea level (Horberg 1954). In northwestern Montana remnants of the Flaxville plain are on Lee Ridge and other areas with corresponding elevations. The flattish summit areas within the Milk River Ridge at elevations above 1381 m (4600 ft) are believed to be a portion of the Flaxville Plain (Horberg 1954). However, this area has been intermittently covered with morainal drift material which complicates the correlation (Brierley et al. 1986).

The No. 2 bench of early Pleistocene age is extensively preserved. It occurs at elevations between 1652-1261 m (5500-4200 ft) and lies approximately 150 m (500 ft) above the present drainage systems (Horberg 1954). Near the Rocky Mountains the No. 2 bench is roughly 90 m (300 ft) below the Flaxville plain but to the east the difference between the two peneplains increases to between 150-180 m (500-600 ft) (Westgate 1968).

In southwestern Alberta the No. 2 bench may be represented by old valley floor remnants at 1500 m (5000 ft) on the west side of Waterton Lake, and on Cloudy Ridge at an elevation of between 1562-1652 m (5200-5500 ft) above sea level. These areas are covered with drift which makes determination of bedrock elevation uncertain (Horberg 1954).

On the Milk River Ridge the unglaciated area between the North and South forks of the Milk River, called the Del Bonita Plateau, is equivalent to the No. 2 bench (Alden 1932). Further to the east corresponding portions of this remnant surface are overlain with drift, again masking the recognizable features, principally the level terrace-like landform. Around the Cypress Hills, the upland at elevations of 961-1021 m (3200-3400 ft) above sea level occurring southeast of Lake Pakowki is thought to be a remnant of the No. 2 bench (Westgate 1968).

The last set of benches or terraces represent "another stage of downcutting by the streams followed by lateral planation and deposition of river gravel" (Alden 1932:vii). They are referred to as the No. 3 bench and are of Middle to Late Pleistocene age. This bench is about 30-150 m (100-500 ft) lower than the No. 2 bench and it represents broad valley bottoms of streams which dissected the No. 2 bench (Westgate 1968). The deposition of quartzitic gravels on this bench ceased when the rivers were blocked by advancing Laurentide ice sheets (Westgate 1968). Horberg (1954) recognized the same discrete No. 3 bench remnant as Alden (1932). This bench occurs along the North and Middle (now called the South) forks of the Milk River, south of the International border in Montana. North of the border, in southwestern Alberta the only remains of the No. 3 bench are beneath till. These areas are distinguished by the presence of quartzitic gravels underlying till. The gravels in such situations are referred to as Saskatchewan gravels (Horberg 1952; Stalker 1968; Westgate 1968).

The term "Saskatchewan sands and gravels" has been used to describe the deposits which were laid down by the last series of preglacial rivers, prior to the Quaternary glaciation. The gravels deposited on the upper, older benches (ie. No. 0,1,2) may not be referred to as Saskatchewan gravels (Stalker 1968). However confusion in applying the term does exist. Generally it is applied to the majority of gravels which underlie Laurentide morainal material and overlie bedrock, irrespective of elevation and correlation to the Blackfoot peneplain sequence of benches.

3. Glacial History of the Study Area

The landscapes of southwestern Alberta and northwestern Montana have been modified by a sequence of glacial advances during the Pleistocene epoch. These glaciers were of Cordilleran and Laurentide origin. Four major advances with associated warmer interglacial periods during the Pleistocene have been recognized for North America. The terminology and approximate ages of these events are given in Table II.

Table II. Pleistocene Glaciations in North America

Glacial	Interglacial	Age (thousands of yrs)
	Recent	7 - present
Late Wisconsin		22 - 7
Middle Wisconsin		54 - 22
Early Wisconsin		128 - 54
	Sangamon	235 - 128
Illinoian		375 - 235
	Yarmouthian	? - 375
Kansan		?
	Aftonian	?
Nebraskan		?

(from Harrison 1976:14)

Laurentide or Cordilleran glaciers did not override the Del Bonita Plateau prior to the Wisconsin ice age. Pre-Wisconsin aged Cordilleran ice sheets have been traced on to the plains up to 80 km (50 miles) east of the Rocky Mountains, on the Canadian side of the border (Stalker 1962; Harrison 1976). This oldest Cordilleran till deposit found in southern Alberta is referred to as "Albertan till" and was originally described by Dawson (1895). Exposures of this till are evident where burial or erosion by successive mountain ice sheets was not too extensive; the marginal areas of later glaciations (Stalker 1963). On the plains the Albertan till is overlain by a Laurentide till which is of Kansan (Harrison 1976) or Illinoian (Rutter 1984) age. The continental ice sheet responsible for this Labuma till advanced into the Rocky Mountain foothills to elevations of 1500-1652 m (5000-5500 ft) (Harrison 1976). On the Milk River Ridge along the south fork of the Milk River, (Twp 1, Rge 19), outcrops of Labuma till have been recognized beneath more recent geologic deposits (Dr. A. Stalker and Dr. R. Barendregt pers. comm. 1983)[4].

Alden and Stebinger (1913) traced pre-Wisconsin aged extensions of mountain glaciers as far east as the St. Mary Ridge in Montana. The till here described by Alden

[4] Dr. A. Stalker, Geological Survey of Canada, Ottawa, Ontario. Dr. R. Barendregt, Dept. of Geography, University of Lethbridge, Lethbridge, Alberta.

(1932) corresponds in age to the Kansan or Illinoian-aged deposits described by Horberg (1954) and Dr. E. Karlstrom (pers. comm. 1983)[5] on Makowan Butte in Alberta. Thus from the descriptions in the literature no till, especially of pre-Wisconsin age, has been described within the study area.

The position of the maximum Laurentide ice margin during the Wisconsin glaciation is well agreed upon by the many surficial geologists who have worked in the area. Calhoun (1906) depicted the maximum ice sheet position. (Figure 2). Alden's (1932), Stalker's (1962) or Prest's (1984) demarcation does not differ significantly from Calhoun's original map. During the course of the Agriculture Canada soil survey in the M.D. of Cardston, numerous inspections of materials were recorded in the occurrence of soils developed on Laurentide till closely correspond with Calhoun's 1906 representation (Brierley et al. 1986). The Del Bonita Plateau stood out as a nunatak while the surrounding area (except directly to the north) was beneath the Wisconsin aged Laurentide ice sheet (Westgate 1968; Shetsen 1980).

The time within the Wisconsin glaciation when the Laurentide ice sheet attained this southern maximum is disputed. During the late or classical Wisconsin, for

[5] Dr. E. Karlstrom, Dept. of Geography, N. Arizona University, Arizona.

example, Stalker (1977) and Reeves (1973) place the southern limit 40 km north, parallel to Etzikom Coulee near Lethbridge. However other researchers, Westgate (1968), Shetsen (1980), and Mickelson et al (1983), oppose placing Classical Wisconsin ice margin so far north, since there is no firm chronologic or stratigraphic evidence for Stalker's ice margin position. Also there are no morphological differences in the appearance of hummocky moraines or river valleys south of Stalker's ice margin to indicate a difference in landscape maturity (Shetsen 1980; and concurred by G. Richmond pers comm reported by Rutter 1984). Catto's (1981) research around the Cypress Hills, dates the southern limit of glaciation as "pre-late" Wisconsin or more than 30,000 years B.P. Discrepancies in surficial geologist's opinions exist with respect to what, how, and when glaciers moved in southern Alberta, especially relative to the vicinity of the Milk River Ridge and the Del Bonita Plateau.

For the purpose of this project the actual position of the Classical Wisconsin ice sheet is not crucial. The southern limit of the Laurentide glacier relative to the Del Bonita Plateau is of more importance. The fact that the Del Bonita Plateau was not glaciated is agreed upon, and this is of paramount importance. There is extensive agreement amongst the numerous surficial geologists who have worked in southern Alberta and neighboring areas, regarding this conclusion. So whether glaciers retreated in stages with no real major readvance or retreated completely then readvanced to form the Lethbridge moraine (Stalker 1977) and leave an ice free corridor parallel to the Rocky Mountains extending to the Yukon (Rutter 1984), is not relevant to this discussion.

B. Loess

Loess may be defined as "wind blown silt deposits", however no single definition has been universally accepted (Pye 1984). Loess deposits have been recognized throughout the world, with large areal extents in China, Central Asia, Northern Europe, and North America. In North America the largest loessial plains are associated with the Missouri River drainage system in the south central States (Chorley et al. 1984). In Canada loess deposits are currently being mapped by Dr. I. Smalley (pers. comm. 1984)[6]. Presently, recognized loess deposits in southern Alberta are confined to the Cypress Hills area (Westgate 1968; Jungerius 1969; Catto 1981). However, the similarities between the Cypress Hills and this study area, some of which have been described previously, suggest that the material of disputed origin on the Del Bonita Plateau is loess. Therefore the theories of

[6] Dr. I. Smalley, Dept. of Earth Sciences, University of Waterloo, Ontario.

origin, characteristics, and identification techniques for loess material will be considered.

1. Theories of Loess Origin

The definition of loess often includes its origin. For example, the Glossary of Geology (1972) describes loess as "windblown dust of Pleistocene age, carried from desert surfaces, alluvial valleys and outwash plains lying south of the limits of the ice sheets, or from unconsolidated glacial or glaciofluvial deposits uncovered by successive glacial recessions but prior to invasion by a vegetation mat" (Gary et al. 1972:416).

The association of loess deposits with the presence of Pleistocene ice sheets has been long recognized. Hardcastle (1890), as reported by Smalley (1983), states "no other agent than ice could have produced so great a quantity of such fine material", (Smalley 1983:482). The fine material referred to is glacially produced rock meal present within morainal deposits. This material is further concentrated by fluvial action associated with floodplains. Upon desiccation, this dust-like material is easily carried by wind currents (Smalley 1983). Similarly Chamberlain (1897), quoted by Smith (1942), expands upon Hardcastle's original idea by describing "the development of extensive flats over which glacial silts were spread during the great periodic extensions of glacial waters caused by periods of warm weather in the melting season..." (Smith 1942:141). He continues to state that after waters had retreated the extensive silt-covered plates would become exposed to the sweeping influence of the wind, and when they had dried the silt would be borne in great quantities over the adjoining uplands (Smith 1942). The derivation of loess deposits involving katabatic winds from retreating ice sheets blowing over extensive glaciofluvial outwash plains during the Pleistocene, is a widely accepted scenario (Bryan 1943; Hobbs 1947; Wascher et al. 1947; Franzmeier 1970; Chorley et al. 1984).

In North America examination of loess deposits support this hypothesis. The majority of the loess deposits in the eastern part of the Great Plains and Mississippi Valley are between 25,000-13,000 years old, as determined by C^{14} radiocarbon dating. This age corresponds with the retreat of the Wisconsin ice sheets (Pye 1984). Also there is a high correlation between loess deposits and rivers which carried glacial meltwater. For example, the Peorian loess deposits of the eastern Great Plains and Mississippi Valley originated from the reworking of coarse textured river floodplain sediments by northwesterly winds (Wascher et al. 1947; Pye 1984). Similarly, the loess deposits on the Cypress Hills have been correlated with the retreat of the Wisconsin ice sheet from south eastern Alberta (Westgate 1968; Jungerius 1969; Catto 1981). There is some dispute that glacial grinding may not be the sole contributor to producing silt sized quartz grains, an important component of loess. Pye (1984) reports that processes such as wetting and drying, salt weathering and eolian abrasion all produce silt sized grains. However, since the majority of the world's loess deposits are associated with the margins of ice sheets and flood plains of glacial meltwater channels, glacial grinding is widely accepted as the chief agent producing silt (Smalley 1966; Chorley et al. 1984).

Therefore an origin for loess that involves glaciers, glaciofluvial activity and wind is the widely accepted theory for explaining the occurrence of the world's loess deposits (Smith 1942; Smalley 1966; Pye 1984).

2. Characteristics of Loess

Loess is typically a well sorted, non stratified, buff colored, non indurated material (Lewis et al. 1975; Pye 1984). The word loess comes from the German words löss or lösch meaning loose (Gary et al. 1972). The material is stable (able to stand as vertical sections) when dry, but quickly loses its internal shear strength when moistened. Classically, loess consists of 50-80% silt sized particles and contains less than 10% clay sized particles. The median grain size is in the range of 20-40 μm (or 5.75-4.60 phi) and the distribution is positively skewed (Chorley et al. 1984; Pye 1984). Typical loess has a bulk density of 1.5-2.0 g cm^{-3} (Pye 1984).

Other types of loess have been characterized on the basis of texture and deposition. If there is more than 20% sand sized grains then loess is defined as "sandy loess". Conversely loess with more than 20% clay in the material is described as "clayey loess". The term "reworked loess" is applied to loess that has been eroded and redeposited through fluvial and colluvial processes. When lacustrine and fluvial material have been mixed with loessial material, sediments may be referred to as "loessoid deposits" (Pye 1984). Deposits of loess altered by weathering or pedogenic processes are called "weathered loess". In these deposits, the homogeneity of the loess material may be masked by the presence of clay accumulations (Rutledge et al. 1975a; Pye 1984).

Another feature of loess deposits noted by numerous researchers is the change of particle size and thickness with increasing distance from the source area (Smith 1942; Swineford et al. 1943; Frazee et al. 1970; Rutledge et al. 1975a; Lewis et al. 1975; Souster et al. 1977; Catto 1981). The mean particle size of loess material decreases with the log of the distance from the source. This decrease in particle size is due to sorting within the silt and sand sized fractions. Fine and medium silt fractions consistently increase, while the very fine sand fraction decreases farther from the source area (Frazee et al 1970; Rutledge et al. 1975a; Lewis et al. 1975; Souster et al. 1977). The coarse silt sized fraction generally decreases with distance in accordance with very fine sand (Frazee et al. 1970; Rutledge et al. 1975a; Lewis et al. 1975) however, Souster et al. (1977) found that the amount of coarse silt did not alter significantly with distance from the source.

The thickness of the loessial deposits decreases according to the log of the increasing distance from the source (Smith 1942; Frazee et al. 1970; Souster et al. 1977). Therefore the rate of decrease is not constant. Near the source area the loess deposit decreases in thickness at a rapid rate until a steady state thinning trend is established with increasing distance (Frazee et al. 1970; Souster et al. 1977; Catto 1983). Research in southern Saskatchewan indicated that the loess deposit decreased in thickness from 80 to 45 cm over a distance of 50 km (These values are based upon the regression equation $Y=80.3-20.2\log x$ with an r value equal to -0.85) (Souster et al, 1977).

The interrelationship between thickness and sorting of loessial deposits is a function of wind velocity. As wind velocity dissipates with distance, coarse particles drop out resulting in a decrease in median grain size but an increase in the degree of sorting within the deposit. Also, as the distance from the source area increases the "purer" fractionated loess deposits become thinner (Frazee et al. 1970; Catt 1983).

3. Identification Techniques

Identification of loess is based upon the assessment of physical and mineralogical properties of the material, in addition to recorded field characteristics (Price et al. 1975). Particle size analysis is the most important and widely accepted procedure (Smith 1942; Lewis et al. 1984). Other diagnostic procedures involve determination of the mineralogical composition of the sand, silt, and clay fractions by heavy liquid separations and X-ray diffraction (Westgate 1968; Rutledge et al. 1975b). The scanning electron microscope has also been employed to characterize the surface morphology of sand and silt sized particles (Krinsley and Doornkamp 1973). The results from these procedures are compared with similar values obtained for other parent materials of different depositional modes. Therefore, loess is identified by means of deduction/induction, piecing together the evidence and eliminating the alternatives.

The results of particle size analysis may be expressed in various ways. Graphs, cumulative frequency curves, and histograms are more commonly used to indicate the percent distribution of sand, silt, and clay-sized particles. From these, median grain sizes, skewness, and sorting indices for the deposit are determined (Smith 1942; Folk and Ward 1957; Catto 1983; Pye 1984). Different transformation techniques are also employed in order to remove pedogenic influences (Rutledge et al. 1974a). For example, in situations where the formation of a Bt horizon has masked the homogeneity of the material, particle size distributions are represented on a clay free basis (sand and silt added together equals 100%). This transformation allows for testing the uniformity of parent materials when pedogenesis has complicated the situation (Barnhisel et al. 1971; Rutledge et al. 1975a).

The identification of components within certain fractions is another tool used for deducing the origin of a deposit. Westgate (1968) used the presence /absence relationship of specific minerals within the heavy mineral fraction to conclusively state that the material on the Cypress Hills is loess. Similarly the ratio of quartz to feldspar within the light mineral fraction is a useful parameter for distinguishing parent materials (Price et al. 1975). X-ray diffraction analysis of the clay fraction may be used in a similar fashion (Rutledge et al. 1975b) although weathering due to pedogenic processes may complicate the interpretation.

The description of quartz sand and silt sized grain surfaces using the scanning electron microscope (SEM) has proven useful in determining the mode of depositional history. Numerous researchers (Smalley and Cabrera 1970; Brown 1973; Krinsley and Doornkamp 1973; Whalley and Krinsley 1974; Bull 1981; Hill and Nadeau 1984) have applied this procedure. The description of surface textures is based upon the fact that transportation and deposition processes result in characteristic surface textures (Brown 1973). Since no single feature may be diagnostic of a depositional environment, a series of features must be used to develop a classification (Krinsley and Doornkamp 1973; Bull 1981; Hill and Nadeau 1984). For example, surface sand grain features such as conchoidal fractures, arc shaped steps, and high relief are considered glacial in origin. On the other hand, aeolian grains have upturned plates, dish shaped concavities, low relief, and are generally more rounded in appearance (Krinsley and Doornkamp 1973; Bull 1981). By recording the presence or absence of characteristic surface features for a number of grains per sample, different parent materials may be recognized.

This SEM method of analysis is subjective since visual interpretation and individual bias is involved. However, when Culver et al. (1983:136) compared eight samples analyzed by five scanning electron microscopists, they found that, "SEM analysis of quartz grain surface textures appears to be a reliable and statistically valid means of discriminating between samples from different environments".

f. The identification of loessial material from other parent material using present procedures is not a clean cut process. The evidence gathered from field observations must be combined with laboratory analysis in order to build a tight case when an unknown material may be loess (Price et al. 1975). No one mode of identification solves the mystery when such a puzzle exists.

C. Periglacial Features

1. Introduction

Péwé (1969) stated that a precise definition of the term "periglacial" has yet to be universally accepted. Originally, Lozinski, 1909 (as reported by Washburn 1973) introduced the term to designate "the climate and the climatically controlled features adjacent to the Pleistocene ice sheets" (Washburn 1973:1). This definition is too confining since periglacial environments exist today, irrespective of age or proximity to glaciers (Washburn 1973). In 1942, Sharpe (as reported by Schaefer 1949) characterized a periglacial environment as one with "low temperatures, strong winds and many fluctuations across the freezing point at certain seasons" (Schaefer 1949:155). Nowadays, the term "periglacial" is applied to environments where frost action is the dominant process (Gary et al. 1972). At first this appears ambiguous until the definition of frost action is considered.

Frost action is "the mechanical weathering process caused by alternate or repeated freezing and thawing of water in pores, cracks and other openings usually at the surface" (Gary et al. 1972:280). Frost action encompasses frost heaving, frost wedging and frost sorting which are the result of freezing and thawing cycles. The intensity of frost action processes are a function of the amplitude of the temperature range and duration (Berg 1969; Washburn 1973; French 1976). Frost action, a significant factor within periglacial environments, is a reflection of climate which in turn may be controlled by numerous factors such as latitude, altitude and global Ice Ages.

The Del Bonita Plateau was a nunatak during the Wisconsin Ice Age (Westgate 1968), therefore the area must have been subjected to periglacial environmental conditions. Relict features within the surficial materials on the Plateau are present. They are a product of periglacial activity and the subsequent amelioration of the climate. Of principal interest are the ice wedge casts, involutions and associated calcium carbonate concentrations, the distribution of gravel sized quartzitic coarse fragments throughout the upper surface material as well as the asymmetric valleys. The theories of their formation are considered in this section

2. Ice Wedge Casts

Ice wedge casts or fossil ice wedges are reliable indicators of relict periglacial environments in present-day temperate latitudes (Péwé 1973). In North America many fossil wedges have been described in the area bordering the southern limit of Wisconsin ice sheets. Schaefer (1949) described periglacial features in central Montana. Mears (1981) found fossil wedges 650 km south of the Laurentide ice sheet margins in the Rocky Mountains of Wyoming. Ice wedge casts have also been discovered in glaciated areas. Westgate and Bayrock (1964) and Berg (1969) described fossil wedges found within the Saskatchewan gravel underneath Laurentide till of

Wisconsin age near Edmonton, Alberta. Numerous ice wedge casts have been described by geologists all over the northern hemisphere: Péwé (1969) in Alaska; Johnsson (1959) and French (1976) in northern Europe; and Kostyaev (1969) for example.

Ice wedge casts are "filled in" ice wedges. To understand their formation the development of ice wedges must be understood as well. The mechanism and significance of the subsequent fossilization of the ice wedge then becomes apparent.

i. Origin of Ice Wedges

Ice wedges are wedge-shaped accumulations of ground ice which are vertically orientated, the apices downward. Active ice wedges are present-day features restricted to areas of continuous permafrost, where the mean annual air temperatures are between -5 to -8 ^{o}C or colder (Berg 1969; Péwé 1973).

The most widely accepted mode of ice wedge formation is based upon the "contraction theory". This hypothesis was first proposed by Leffingwell in 1915 and later substantiated by Lachenbruch (1962). Leffingwell (1915) initially noted the appearance of cracks in the snow cover in the Arctic during the winter months. The following spring, he discovered that these cracks were not confined to the snow layer, but extended into the surface mineral layer. The cracks were usually several millimeters wide and up to a meter in depth.

The "contraction theory" is based upon the linear coefficient of thermal expansion which varies for different materials. Ice has a value of $50 \times 10^{-6}/^{o}C$ while quartz has a value of roughly $10 \times 10^{-6}/^{o}C$ (Lachenbruch 1960; MacKay 1972; Black 1983). The significance of this is best shown with an example. With a drop in temperature of 10 ^{o}C, a 10 m long block of ice would shrink in length by 5 mm. Under the same conditions a similar sized rock would contract by 1 mm. In a soil material, the amount of contraction and depth of the crack is therefore a function of soil moisture which in turn is dependent upon the texture of the soil material. To initiate a crack in ice cemented frozen gravel requires a $10^{o}C$ drop in temperature over the period of a day (Black 1983). Alternatively, in saturated fine grained sediments, a temperature change of 4-8 ^{o}C can induce the formation of a crack (Black 1976). Péwé (1965) stated that a mean annual temperature of $-5^{o}C$ was required for surface crack initiation. For a contraction crack to develop into an active ice wedge, the following sequence of events was proposed by Lachenbruch (1960 and 1962).

1. Initially a crack develops during the winter.

2. The following spring, runoff from the melting snow drains into the crack. When the permafrost layer is encountered this meltwater is refrozen. A vertical ice vein within the permafrost develops.

3. The subsequent winter, thermal tension reopens the crack. The ice vein is a zone of weakness since the linear coefficient of thermal expansion of the ice in the vein is different than the frozen material around it.

4. The next spring more runoff water freezes within the crack, so the ice vein or wedge grows annually or seasonally.

5. In each freeze-thaw cycle an ice wedge grows in width by 5-20 mm, at the top of the feature. The rate of width growth decreases with depth (Washburn 1973).

Mackay (1984) has shown that ice-wedge cracks originate near the top of permafrost and then propagate both upward to ground surface and also downward into the subjacent ice wedge ice. Therefore, ice wedges may grow in two directions from the permafrost layer.

ii. Origin and Characteristics of Ice Wedge Casts

The replacement of the ice within active wedges with another material occurs during climatic warming. This replacement process preserves the shape of the previous active ice wedge. The resulting relict periglacial feature is called an ice wedge cast or fossil ice wedge.

Ice wedge casts are divided into two principal groups, based on the nature of the infilling material. Group 1 includes fossil wedges where the material infilling the wedge is primary, while Group 2 has secondary infill material. These two broad groups are associated with different paleoenvironments (Jahn 1975).

Group 1, primary wedge casts, are best exemplified by sand wedges. These features are characteristic of arid, windy, periglacial environments. Frost cracks are not initially filled by runoff water due to its absence, but by shifting sand or silt. In this desert like environment, the combination of desiccating winds and little snowfall keep the ground surface snow free. However, due to the excessively low temperature contraction cracks still form. The internal fabric of these wedge casts are typically characterized by a vertical foliation pattern and containing very well sorted material with a grain size less than 2 mm. This shows that these features grow in a pattern analogous to true frost wedges (Washburn 1973; Mears 1981; Carter 1983).

Secondary ice wedges, Group 2, are ubiquitous periglacial features indicative of moist periglacial environments (Black 1976). The formation of these wedge casts is a two step process involving the initial development of a true ice wedge and the subsequent replacement with another material upon the disintegration of periglacial conditions.

As the mean annual temperature increases, the permafrost layer in the periglacial area melts from the surface down. Ice within the frost wedge will melt faster than the surrounding material due to the purity of the ice. Preferential melting of ice occurs along the ice-wedge ground material interface, and voids appear in this zone. (This process is analogous to the melting of an ice cube in its container - initial melting occurs at the edges) These voids are subsequently filled in with sediments from runoff water derived from melting snow and surface permafrost deterioration. This preferential melting along the interface extends to substantial depths within the intact underlying permafrost. Ultimately as the thawing process continues, ice of the original frost wedge is completely replaced by sediments derived from runoff and "slumping" processes (Péwé 1969; Washburn 1973; Jahn 1975; Walters 1978). During the later stages of cast formation, the surface mineral material becomes supersaturated since there is more ice than water pore space within the underlying material. Therefore a perched water table is developed. The "soupy" surface material then flows or slumps into the ice wedge depression (Black 1976).

Due to their origin and development, secondary filled wedge casts are larger and more variable in appearance than sand wedges. The fabric of these casts are less distinctly vertical and there is a wider range of particle size distributions.

There are two distinct zones within ice wedge casts; the lining and the middle. The "lining" portion of the cast is recognized along the edges of the wedge cast where the initial melting and replacement of ice occurred. This lining is characterized by orientated fabric parallel to the edge of the fossil wedge (Jahn 1975). The middle portion of the cast represents the "slumped in" secondary material which occurred during permafrost degradation. The fabric of this portion of the wedge is often concave in orientation, "resembling stacked dishes" (Mears 1981). This internal fabric is often altered by freeze-thaw processes after initial deposition (Walter 1978).

Features similar to "true" ice wedge casts may be the result of glacial thrusting, solidification, disturbances through landslides and soil tonguing due to desiccation (Yehle 1954; Johnsson 1959). To eliminate possible confusion Johnsson (1959) listed five criteria for true wedge cast identification:

1. The filling material must seem to have come from above.

2. The lines of the sides of true ice wedges usually dip downwards.

3. Stones should be vertically aligned in the wedge, while in the adjacent side layers they are mainly horizontal.

4. The top of the wedge must be widest, with the apex pointing downward.

5. The ice wedge cast should be more or less vertical.

Berg (1969) noted a size criteria as an additional diagnostic feature of true fossil ice wedges. The ratio of width to height must be less than 1:3.

If these criteria are satisfied then a feature may be reliably recognized as an ice wedge cast. Subsequently paleoclimatic characteristics of the area can be postulated (Péwé 1963; Péwé 1973; Carter 1983; Chorley et al. 1983).

Patterned ground is generally associated with active and fossil ice wedges. Symmetrical or regular patterns are formed by the intersection of ice wedges. The polygonal areas bounded by wedges are generally in the 30 m diameter range (Péwé 1973) but some up to 100 m in diameter have been described by Washburn (1956). Due to this relationship, patterned ground can sometimes be found in conjunction with fossil ice wedges, as a relict periglacial feature (MacKay 1972; Washburn 1973).

3. Involutions

Involutions are another periglacial feature which may be associated with ice wedge casts. Sharpe (1942) as reported by Washburn (1973) described involutions as "aimless deformation, distribution and interpenetration of beds produced by frost action (Washburn 1973:147). French (1976) states that these structures may be the result of pressures induced by freezing in the active layer of a permafrost environment. Also Schaefer (1949) felt that cryostatic pressure exerted on surficial materials during ice wedge and patterned ground formation at depth could form involutions.

Involution-like structures may be produced by processes other than frost action. Some of these include ice shoving or glacial thrusting, volume change due to the presence of bentonite, differential loading, mass movement, and tectonic deformation (Schaefer 1949). However, when these structures are found in association with other periglacial features these structures are termed "periglacial involutions" (French 1976). In this review since intensive frost action is assumed to be the dominant process for the production of involutions in this study area, only periglacial frost action forces will be discussed.

The most widely accepted theory for the origin of periglacial involutions is differential freezing (French 1976). During cooling in the autumn, or due to a change in climatic regime, the freezing front descends unevenly, squeezing unfrozen saturated sediments. The cryostatic pressure is created due to the presence of the perennially frozen zone beneath the unfrozen materials (Schaefer 1949; Washburn 1956; French 1976). The freezing front descends unevenly since the amount of interstitial water present depends on the material texture, which may be variable. This results in contortions of bedding plains or pocket formation. Differential vegetative cover also affects the rate of freezing front penetration.

The presence of the permafrost layer is crucial to inhibit drainage of the meltwater from the active layer. A high degree of saturation favors expansion as ice forms during freeze-up. These are the circumstances needed for frost heaving of cryoturbation processes. Such frost action processes have been shown to duplicate involution deformations, under laboratory conditions (French 1976). Since involution formation is controlled by moisture, density of materials and frost heaving, involutions are not necessarily solely a product of periglacial conditions. Therefore as Washburn (1973) suggests, the identification of past periglacial environments based on the presence of involutions must be used with extreme caution.

4. Sorting due to Frost Action

The movement of stones and the sorting of different particle sizes due to freeze-thaw cycles has been noted by numerous researchers (as reported by Washburn 1973) These processes have been related to the existence of freeze-thaw cycles since they've been experimentally reproduced in the laboratory (Corte 1963; Manikowska 1982; Reiger 1983).

Two methods have been proposed to explain the upward movement of stones in a medium. They are the frost pull and frost push theories.

The frost push theory is based on the premise that stones have a greater thermal diffusity than the surrounding medium. As a freezing front approaches the stone, ice forms around it and at its base. Pore water is drawn towards the cold zone and segregation ice or an ice lense develops at the base of the stone sooner than in the surrounding material. The expanding ice then heaves the stone upwards. Upon melting, fine-sized particles seep in from the sides, thus preventing the stone returning to its initial position (Washburn, 1973). The controlling factors of this process are the initial soil moisture and the hydraulic conductivity of the medium (Taber 1943; French 1976).

The frost pull hypothesis depicts stones being moved upwards during the period of frost penetration from the surface. As the freezing front descends, a void above stones forms due to frost heaving. As the front continues to penetrate and the material at the sides of the stone becomes frozen, while the soil beneath the stone is unfrozen. Thus the stone becomes detached from the subsoil as heaving occurs and it is lifted towards the surface. Since thawing progresses from the surface downward, the cavity at the base of the stone is filled with slumped-in material, from the sides. Viscosity and other tensile parameters of the unfrozen material are some controlling parameters of this method in addition to the previous factors mentioned for the previous method (Kaplar 1965; Washburn 1973; French 1976). Kaplar (1970 as reported by Washburn 1973) used the following relationship to approximate the rate of upward movement in one freeze-thaw cycle by the frost pull method.

$$D = \frac{H_R L}{R_f}$$

D = vertical distance of stone
H_R = heave [illegible] of the medium(mm/day)
L = vertical height of the stone below its greatest horizontal distance
R_f = rate of frost penetration(mm/day)

Manikowska (1982:112) indicates that "the maximum heave during one freeze-thaw cycle is proportional to the effective height of the object". This is similar to the relationship depicted by Kaplar's equation. However Manikowska continues to say that since stones rotate so the long axis is vertically aligned (this fact was substantiated by her research in Poland) the heaving values increase with successive freeze-thaw cycles.

Of the two hypotheses, the frost pull theory is most widely accepted (Kaplar 1965; Washburn 1973; Manikowska 1982). In addition to stone up-freezing materials subjected to frost action show unique particle size distribution variations and fabric arrangements. The degree of particle size sorting and soil micro fabric arrangements depends upon soil moisture, soil temperature, rate of freezing and orientation of freezing plane, just as does stone movement (Corte 1963). These factors all affect the degree of cryoturbation (Reiger 1983). Corte (1963) from laboratory experiments has reported that fine particles (< 74 μm) in laboratory experiments migrate downwards before an advancing frost plane while coarser particles will move upwards in the same way as stones.

Therefore, vertical and lateral sorting are dependent upon the orientation of the freezing front relative to the particles of the unfrozen medium.

With evidence that stones and soil particles move due to freeze-thaw cycles the micro fabric of the soil must also undergo rearrangement. Koniscev et al. (1973) subjected loam textured soils to numerous freeze-thaw cycles and then observed orientated thin sections with a polarizing microscope. Fabric arrangements in the form of irregular, oval or elongated shapes developed in the soils. They termed this frost action process "micropolygonization" (Koniscev et al. 1973). Other researchers, Bunting and Fedoroff (1973), Fox and Protz (1981), Pawluk (1983), and Mellor (1986) to name a few, have noted similar rounded aggregations of fabric in present day "cold" soils.

The process of describing the upward movement of stones and particle and fabric redistribution is still subject to controversy. The frost pull theory is most widely accepted for stone up-freezing (Washburn 1973; Manikowska 1982), but wetting and drying cycles have been noted to cause stone movement as well. Similarly, wetting and drying can not be eliminated as a possible agent of micropolygonization of soil fabric (Pawluk 1983; Jim 1986).

5. Asymmetric Valleys

An asymmetric valley is "a valley with one side steeper than the other" (Gary et al. 1972:44). These valleys usually occur in parallel sets with regular spacing between them (French 1972; Washburn 1973). Due to their origin these valleys are better developed in regions formerly subject to periglacial environments at lower latitudes than in the polar regions (Washburn 1973). Often associated with asymmetric valleys are features called asymmetric dells. These are smaller valleys which don't contain a present day stream, but where the diagnostic shape is maintained (Washburn 1973). Within a fossil drainage pattern system, asymmetric dells and valleys often exist in an intimate relationship (Grimbérieux 1982).

In describing asymmetric valleys, orientation and degree of asymmetry are the key components used. The alignment of the valley with flow direction and the aspect of the steeper slope are factors considered in great detail An asymmetry index is calculated using the following:

$AI = 1 - g/s$ g = angle of gentle slope in degrees

s = angle of steep slope in degrees

This index provides a basis for comparing different valley systems in terms of asymmetry and orientation. Symmetrical valleys have an AI value of 0, because both slopes are equal. As the discrepancy in slope increases the AI value increases. In addition, direction of the steep slope may be indicated by positive and negative signs. East facing steep slopes which are referred to as having "inverted asymmetry" are indicated by negative AI values (Grimbérieux 1982).

The generally accepted scenario for the origin of asymmetric valleys is related to differences in insolation, which is a function of aspect. These variables cause differential thawing which creates further consequences in a permafrost environment. Valley slopes facing south, southeast and southwest receive more solar radiation than slopes with northerly aspect, in the Northern hemisphere. Assuming that initially the area is underlain with permafrost, maximum thawing and subsequent gelifluction occurs on these south-facing slopes. The asymmetry of the valley is further developed as the soliflucted sediment from the southerly slope forces the meltwater stream up against the opposing slope. The colder slope is undercut and the valley is further incised. Since in polar latitudes the temperature difference between exposures is less, the asymmetric valleys are not a well defined as in the previous periglacial zones of the lower latitudes (French 1972; Washburn 1973).

Recognition of periglacial environments by the presence of asymmetric valleys needs to be judiciously assessed because non periglacial processes can cause asymmetry. Micro climate variation, stream evolution as well as bedrock structural control may be responsible for the occurrence of asymmetric valleys irrespective of periglacial activity. Therefore the existence of these valleys provides evidence that should be used in conjunction with other features to delimit former periglacial environments (Washburn 1973; French 1976).

CHAPTER IV. METHODS

In this chapter the procedures used for the field collection of the samples and their subsequent analyses are briefly described. In addition, the approach and rationale for some specific analyses are justified.

A. Field Procedures

The soils of the study area were mapped during the course of the soil survey of the M.D. of Cardston (an Agriculture Canada, Land Resource Research Centre initiated soil survey project). The survey met Survey Intensity Level (SIL3) specifications as defined by the Mapping Systems Working Group (1981). The distribution of soils for the area is represented on 1:50,000 scale photo mosaic base maps with soil and map unit descriptions in the accompanying report. The soil map units are recognized on the basis of the dominant and significant soil names present in the delineated polygons. Topographic classes, as defined in the Canadian System of Soil Classification (CSSC 1978a), are modifiers added to the soil map unit.

Sites within the study area for this project were selected on the basis of the soil survey. The Beazer soil was selected as the reference soil because of its many profile similarities with the Del Bonita soils. Also the parent material of this soil name was known to be

Laurentide till. Soil map units where Del Bonita (DLB) and Beazer (BZR) soil series were the respective dominant soils were the areas selected. Criteria for site selection within these potential areas included native vegetative cover (rough fescue) and representative soil profile characteristics for these Orthic Black Chernozemic soils. The sites were distributed across and around the Plateau, within the designated study area (Figure 1).

A total of six profiles (3 DLB and 3 BZR profiles) were classified according to the Canadian System of Soil Classification (CSSC 1978a) and described using the categories outlined in the CanSIS manual for describing soils in the field (ECSS 1983). Bulk samples were taken for each field-recognized horizon. Orientated samples for micromorphological descriptions were taken for BZR soil profile #1 and DLB soil profiles #1 and #3. From the gravel pits associated with DLB sites #1 and #3, additional samples were taken from depths below the control section (i.e. 120 cm). Surface samples, including the top 5 cm of soil and vegetation, were collected for phytolith extraction from all six locations.

B. Laboratory Analysis

The bulk samples were air dried at room temperature, ground and passed through a 2 mm sieve. The coarse fragment content (fragments greater than 2 mm in size) was estimated in the field. The subsequent analyses were performed on the fraction less than 2 mm.

The routine chemical and physical analyses were initially conducted by the Alberta Research Council - Agriculture Canada soils lab personnel. These analyses characterized the general similarities and differences of the two soil series. These routine procedures are documented in the Manual of Soil Sampling and Methods of Analysis (CSSC 1978b), except where otherwise indicated.

1. Soil reaction was determined using both 0.01 M $CaCl_2$ and water in a 2:1 ratio of solution to soil [3.11 and 3.13 respectively][1]

2. Calcium carbonate equivalent was determined by the manometric method (Bascombe 1961)

3. Organic carbon was determined by dry combustion using an induction furnace with gasometric detection of evolved CO_2 (Leco Model 577-100) [3.611]

4. Exchange capacity was calculated by displacement of ammonium with sodium chloride [3.321]

5. Exchangeable cations were measured following displacement with ammonium acetate at pH7. The displaced K, Na, Mg and Ca were determined by inductively coupled plasma spectroscopy [3.321]

6. Particle size analysis was determined using the hydrometer method with no pretreatments, to remove organic matter from Ah horizons or $CaCO_3$ from C horizons [2.121]

[1] Numbers in square brackets indicate method in CSSC (1978b)

C. Specific Analytical Procedures

1. Particle Size Separation

A different particle size analysis methodology was necessary for the nineteen samples selected for detailed mineralogical analysis. Samples from the A, B and lower C (after 80 cm) horizons for each site were used for analysis. The analyzed samples thus all come from roughly the same representative depths within the profiles. The fine earth fraction obtained by sieving the Flaxville gravels was the additional sample analyzed. Organic matter and calcium carbonate were removed from the appropriate samples using 30% H_2O_2 and concentrated HCl respectively. All samples were then dispersed using an ultrasonic unit. Sand, silt and clay fractions were separated by wet sieving and gravity separation based on Stokes Law (Jackson 1975). The separated clay was flocculated in another container with 1 N $CaCl_2$. The silt and sand were separated by wet sieving using a 270 mesh sieve.

The sand and silt fractions were further subdivided. Sands were dry sieved using a nest of sieves (sieve numbers 270, 140, 60, 35, 18) on a sonic sifter for seven minutes. Silts were subdivided into fine, medium and coarse fractions by gravity sedimentation. Percentages of the individual sand and silt fractions with the total sand and silt were determined on a weight basis.

2. Preparation of Cumulative Curves and Analysis

Cumulative particle size distribution curves were drawn for each sample using the five sand fractions, three silt fractions and the clay. Particle size diameter is plotted from coarse to fine on the X-axis of the graph, while cumulative percent from 0-100 is plotted on the Y-axis. The resulting graphical representations of particle size were then analyzed by statistical equations to compare, sorting, skewness and kurtosis (Folk and Ward 1957). Values for the equations were obtained from standardized curves (or graphs) by computer (GEM software package). Since it is difficult to obtain values from hand drawn representations of S curves between data points because of inconsistency or draftsmen "artistry" (Mason and Folk 1958; Folk 1966), these standardized point curves were used.

The particle size diameter values were expressed in a transformed scale ϕ. This scale is a log transformation (Folk and Ward 1957) which simplifies the computation of these statistical parameters and may be justifiably used for convenience (Swan et al. 1978).

$\phi = -\log_2$ (diameter in mm)

$\Longrightarrow$ ϕ = (log of the diameter) / -log2

The lower diameter for each class was used for the calculation of each representation ϕ value since clay by definition is everything less than 2 μm. The same arbitrary breaks were used for other fractions. For example:

(log 2 μ or .002)/(-log2) = 9 ϕ fine silt value

(log 5 μ or .005)/(-log2) = 7.6 ϕ med. silt value

On the phi scale, all particles were assumed to be less than the value 14. Therefore 14 ϕ on the X-axis corresponded to 100% on the Y-axis. The necessary values for the statistical parameter equations were read from the extrapolated curve from 9 ϕ to 14 ϕ (Folk and Ward 1957).

Particular values were calculated from these cumulative curves and were then used to compare materials. Such calculated values were:

i) Median - the phi value when Y-axis is 50

ii) Mz = mean size = $\dfrac{\phi16 + \phi50 + \phi84}{3}$

iii) Sorting - obtained from equation called Inclusive Graphic Standard Deviation

$$\Phi = \frac{\phi 84 - \phi 16}{4} + \frac{\phi 95 - \phi 5}{6.6}$$

($\phi 84$ means the diameter value at 84%)

iv) Skewness - this value refers to the symmetry of the distribution

$$SK_1 = \frac{\phi 16 + \phi 84 - 2\phi 50}{2(\phi 84 - \phi 16)} + \frac{\phi 5 + \phi 95 - 2(\phi 50)}{2(\phi 95 - \phi 5)}$$

v) Kurtosis - "measures the normality of a distribution by comparing the sorting in the central part of the curve with the sorting in the tails" (Mason and Folk 1958:218)

$$K_G = \frac{\phi 95 - \phi 5}{2.44(\phi 75 - \phi 25)}$$

Folk and Ward (1957) listed categories of values with different descriptors for sorting, skewness and kurtosis as a classification system. This system allowed for the descriptive comparison of different cumulative curves.

3. Heavy and Light Mineral Separation and Dissolution

The comparison of the mineralogy of the fine earth fraction of the Del Bonita and Beazer soils was based upon the fine sand fraction. A 1 gram subsample of the fine sand from each sample was analyzed. Each fine sand sample was separated into heavy (specific gravity > 2.95) and light fractions (< 2.95) using S-tetrabromoethane liquid.

The quartz and feldspar percentages of the light mineral fraction were determined using the HF-HCl dissolution technique (Pawluk 1967). Concentrations of Na, K and Ca were determined using atomic adsorption spectrometry. These values were used to calculate the content of Na-, K- and Ca- feldspar. The proportion of quartz was assumed to be the difference between the sum of the feldspars and the light mineral value.

4. X-ray Diffraction Analysis

The whole clay sample collected in part (a) of Particle Size Analysis was analyzed using X-ray diffraction. Slides were made from the Ca and K saturated clays using the paste method of Theisen and Harward (1962). The remainder of the unused Ca saturated clay was freeze-dried for preservation. The slides were subjected to the following treatments before being placed in the X.R.D.:

1. K-saturated clay, heated to 105°C and run 0% relative humidity
2. K-saturated clay, heated to 105°C, equilibrated to 54% relative humidity, and run at 54% relative humidity

3. K-saturated clay, heated to 300°C and run at 0% relative humidity
4. K-saturated clay, heated to 550°C and run at 0% relative humidity
5. Ca-saturated clay, equilibrated to 54% relative humidity and run at 54% relative humidity
6. Ca-saturated clay, equilibrated with ethylene glycol and run at ambient conditions
7. Ca-saturated clay, equilibrated with glycerol and run at ambient conditions

X-ray diffraction analysis was performed using a Phillips PW1730 X-ray generator, a PW1710 diffractometer, and using Co^{L} radiation. The step scanning speed was 2 sec for each step of 0.05 2θ.

For determination of calcite and dolomite values, the whole untreated fine earth samples from representative C horizons were analyzed with the X.R.D. Samples were ground to a very fine powder and packed in an aluminum specimen holder. The step scanning speed was 10 sec for each step of 0.02 2θ, for a region of 30-40°. The relative proportions of calcite and dolomite were approximated from the peak intensities of the carbonate minerals.

5. Scanning Electron Microscopy

Surface micromorphological features present on quartz sand grains were examined with a Cambridge Stereoscan scanning electron microscope. The fine sand (100-275 μm) size fraction was used for this analysis, as recommended by Krinsley and Doornkamp (1973) and Darmody (1985). A subsample was taken from the fine sand fraction and sprinkled on to stubs coated with adhesive (E-kote No. 3040). These stubs were then spatter coated with gold to prevent charging of the sample. The quartz grains on each stub were chosen at random. Each grain was identified as being quartz by means of a Kevex 7000, Energy Dispersive X-ray analyzer. Up to ten grains per each stub (no more to avoid biasing (Darmody 1985)) were described in terms of presence or absence of 13 surface morphological features. Features used for characterization were ones described by Krinsley and Doornkamp (1973), Whalley and Krinsley (1974) and Bull (1981). For each horizon per soil series between 25-30 fine sand sized quartz grains were described, thus attaining the minimum requirements recommended by Baker (1976) for sample analysis.

Phytoliths were extracted from the surface soil samples following the modified procedures as outlined by Jones and Beavers (1964) and Twiss et al. (1969). The phytoliths were extracted from the silt fraction by centrifuging a suspension of the silt and a mixture of bromoform and tetrabromoethane adjusted to a specific gravity of [illegible]. The samples were centrifuged three times and the phytoliths were cleaned and dried in acetone. Phytoliths were extracted from the plant material by means of dissolving the organic material with H_2O_2. The remaining residue was analyzed and the phytoliths described.

The phytoliths from these two sources were mounted on SEM stubs. The black TV tube Koat glue was found to be a suitable adhesive, and the phytoliths were blown on to the stubs using a small beaker and compressed air. Fifteen random phytoliths per stub were categorized using Twiss et al. (1969) four main shape criteria. A total of forty five phytoliths per soil series were characterized in this fashion. Phytoliths from the plant materials obtained at one DLB and one BZR location were similarly mounted on stubs.

6. Micromorphology

The orientated samples for micromorphological descriptions were prepared by impregnating the soil with Scotchcast epoxy resin #3 under vacuum. The sample blocks were then cut and mounted on slides so a vertical orientation was represented. Grinding and polishing of the thin sections to a 30 μm thickness was accomplished using a semi-automated procedure involving the Logitech thin section precision polishing system (M. Abley pers. comm. 1988)[7]

The micromorphology was described using the terminology of Brewer (1976), Brewer et al. (1983), Brewer and Pawluk (1975), as well as, Fox and Protz (1981).

[7] M. Abley, Deptartment of Soil Science, University of Alberta.

CHAPTER V. RESULTS AND DISCUSSION

This chapter is divided into two sections. The first consists of analysis of the field observations and field data. Several aspects warrant consideration, including the soil morphology and associated landscapes. Then the varied relict periglacial features which are unique to the Del Bonita Plateau are elaborated upon.

In the second section of the chapter, data obtained from physical, mineralogical, chemical and morphological analyses are presented. The results are examined for trends and comparisons within this data set as well as with other published data. Procedural differences are also discussed. The section is subdivided on the basis of different analyses.

A. Field Observations and Data

The Del Bonita (DLB) and Beazer (BZR) soils are differentiated in the field principally on the basis of landscape form and topography, coarse fragment lithology and carbonate content. Photo interpretation proved a reliable tool for differentiating DLB and BZR areas, during the Cardston soil survey, once the relationship of soil types to landscape was established for this portion of the Milk River Upland.

The soils of this study area are all classified as Orthic Black Chernozemics according to the Canadian System of Soil Classification (CSSC 1978a). In the adjacent area within Montana, the equivalent to the DLB soil name is the Michelson series. These are Argic Cryoborolls developed on alluvium. The BZR soil name equivalent is the Leavitt series, also an Argic Cryoboroll but developed on till (U.S.D.A. 1980).

A representative profile of these soils has a surface layer which is black, loam textured and 10 cm or more in thickness. The B horizon is dark brown, generally clay loam with visible clay skins on some of the subangular structured ped surfaces, and is 20-40 cm thick. The calcareous C horizon varies from loam to clay loam and occurs at depths between 30-60 cm below the surface. The coarse fragment content by volume is between 5-10% throughout the profile. More detailed site and soil descriptions as well as physical and chemical analyses for the DLB and BZR soil profiles used in this project are documented in Appendix 1.

Some of the soils in this study area are "borderline" Chernozemic soils. The depth of the Ah horizon is variable and often these horizons barely satisfy the 10 cm depth requirement of the Chernozemic order. However, due to the presence of "Ah tonguing" the depth requirement is attained so these thin black soils are still classified as Chernozems.

As previously mentioned, the lithology of the coarse fragments was a diagnostic tool for differentiating DLB and BZR soils. The DLB soils contain less than 10% coarse fragments, and all of the gravel-sized fragments are quartzites. These rocks are the same lithology as the Flaxville gra[illegible] underlie the control section at these DLB site[illegible]ZR soils contain up to 10% coarse fragments of [illegible]re sizes and lithology. They are generally gravel-sized but the occasional stone and boulder is present. The lithology is varied with granitic fragments occurring as well as some quartzites. The granitic rocks originated from the Canadian Shield, therefore the parent material of the BZR soil is Laurentide till (Calhoun 1906; Stalker 1963 and 1977; Shetsen 1980). The quartzitic rocks within the till were derived from areas north of the Plateau where the Flaxville plain once extended - as evidenced by the existence of Saskatchewan gravels (Stalker 1968).

Relict periglacial features were observed from the gravel pits on the Del Bonita Plateau. Ice wedge casts and involutions indicating previous interaction between the overlying materials and the Flaxville gravels are common Figure 3). These relict features of a past

Table IV. Particle Size Fractionation Data (% of each fraction)

Field ID	Lab No.	vcs	cs	ms	fs	vfs	Sand	csi	msi	fsi	Silt	Clay
BZR 102	T1	2.0	3.5	7.7	14.9	10.6	38.7	11.8	13.2	10.0	35.0	26.3
202	T2	3.1	4.8	9.2	18.6	12.0	47.7	12.0	6.8	5.2	24.0	28.3
302	T3	1.0	2.2	5.4	11.1	8.6	28.3	12.5	13.4	11.5	37.4	35.3
DLB 102	T4	0.7	1.6	3.8	6.6	8.5	21.1	21.4	18.9	10.0	50.3	28.6
202	T5	1.2	1.5	4.3	10.8	8.9	26.7	21.8	14.4	9.0	45.2	28.1
302	T6	0.7	1.1	4.4	10.8	8.0	25.0	15.9	19.4	10.5	45.8	29.2
BZR 101	T9	1.0	2.6	7.5	14.6	9.2	34.9	13.4	8.4	4.5	26.3	38.8
201	T10	3.7	4.2	7.3	15.5	9.1	39.8	11.6	8.4	4.2	24.2	36.0
301	T11	0.6	1.7	3.8	9.1	7.1	22.3	15.6	9.2	5.0	29.8	47.9
DLB 101	T12	1.0	2.0	6.6	10.1	7.2	26.9	22.5	9.3	5.3	37.1	36.0
201	T13	1.3	2.2	4.6	11.8	9.8	29.7	17.2	10.5	4.8	32.5	37.8
301	T14	1.7	1.9	3.7	9.1	7.2	23.6	17.8	9.3	4.8	30.9	44.5
BZR 104	T15	1.9	1.8	3.2	6.9	6.8	20.6	15.9	19.5	8.9	44.3	35.1
204	T16	2.2	5.1	8.8	14.3	8.5	38.9	12.3	10.8	7.1	30.2	30.6
305	T17	1.3	2.0	4.9	11.9	9.0	29.1	12.0	13.3	9.6	34.9	36.0
DLB 104	T18	1.1	1.4	5.2	17.1	9.1	33.9	9.1	[illegible]	4.3	20.9	45.2
204	T19	1.2	1.8	5.9	15.5	9.8	34.2	14.2	[illegible]	[illegible]	30.0	35.8
304	T20	0.5	1.6	9.1	30.2	10.4	51.8	5.1	[illegible]	2.3	10.8	37.4

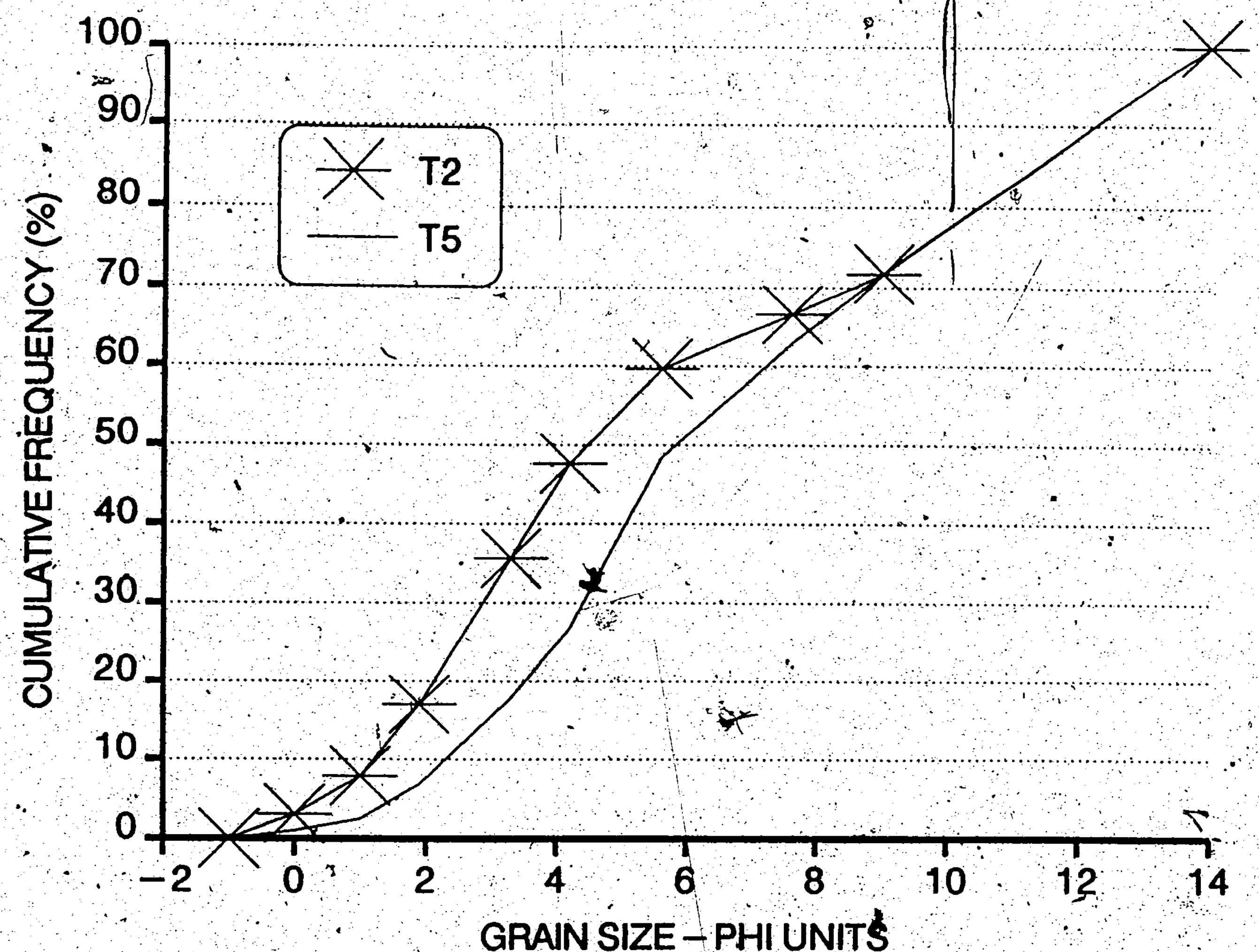

Figure 8. Example cumulative frequency curves for Bm horizons from the Beazer (T2) and Del Bonita (T5) soils.

the very poorly sorted category. For the other horizons no distinct pattern exists, however the range of the sorting values for the DLB series is less than the BZR series for each horizon grouping. Generally, differences of the sorting index values of the two soil series is negligible.

Catto (1981 and 1983), using the same Folk and Ward's sorting index reported that the material atop the Cypress Hills is well to very well sorted. He states that "deposits on the plateau are composed largely of coarse silt with the concentration of fine sand approximating 20%" (Catto 1981:183). From analyzing the sensitivity of the Folk and Ward sorting index it was determined that 90% of a sample needs to be in the same size fraction (i.e. coarse silt or fine sand) before the very well sorted category is attained. Using Westgate's (1968:57) cumulative curves of his loess samples from the Cypress Hills, the calculated sorting index values from these distributions are similar to those listed in Table V. Therefore, it is unclear how Catto (1981) obtained his well sorted index values for Cypress Hills loess.

The skewness values indicate the degree of sorting in the tails of the distribution. If a distribution is positively skewed then the finer particle sizes are concentrated in the tail. If the distribution is

Table V. Cumulative Curve Calculations

Lab No.	Median*	Mean*	Sorting*	Skewness	Kurtosis
T1	5.6	6.2	4.0	+0.229	0.80
T2	4.5	5.9	4.2	+0.405	0.75
T3	7.0	7.2	4.0	+0.057	0.74
T4	6.3	7.1	3.6	+0.253	0.92
T5	5.9	6.7	3.7	+0.276	0.87
T6	6.5	7.0	3.8	+0.176	0.84
T9	5.9	6.7	4.2	[illegible]244	0.65
T10	5.4	6.4	4.4	+0.238	0.69
T11	8.2	8.0	3.9	-0.084	0.70
T12	5.7	6.8	4.1	+0.311	0.73
T13	6.1	7.0	4.1	+0.244	0.71
T14	7.3	7.6	4.1	+0.060	0.72
T15	7.0	7.5	3.9	+0.143	0.85
T16	5.4	6.3	4.3	+0.261	0.71
T17	6.9	7.1	4.0	+0.079	0.73
T18	7.5	7.4	4.2	-0.006	0.63
T19	5.8	6.7	4.1	+0.275	0.68
T20	4.1	6.0	4.3	+0.573	0.61

* mean, median and sorting values are in ϕ units

negatively skewed, coarse particles are in the tail. For a normal distribution where the two tails are equal, the skewness value is 0 (Folk and Ward 1957). The values of skewness for the study area samples are variable within horizons and soil series. The greatest variability, representing the range of skewness values occurs within the DLB, Ck horizons. These Ck samples have values ranging from -0.006 to +0.573. Again, no distinct pattern is distinguishable for the samples from these soils.

Pye (1984) reported that loess typically has a positive skewness in the range of 0.3-0.7. This means that the fines are dominant, since the fine fraction is concentrated within the tail of the distribution. The coarse grains were left behind at the source area of the loess (Refer to Figure 9 - a histogram of T20 with a positive skewness value of +0.5).

Kurtosis is another tool for measuring normality of the distribution. This value compares the sorting in the tail to the central node as a ratio. A normal curve has a value of 1.0. All of the curves for the study area samples are very platykurtic and platykurtic as defined by Folk and Ward (1957). This means that the sorting in the central part of the distribution is worse than the tails. The Ah and Bm horizon samples from the DLB series have consistently higher kurtosis values than the corresponding samples from the BZR series. Therefore these samples from

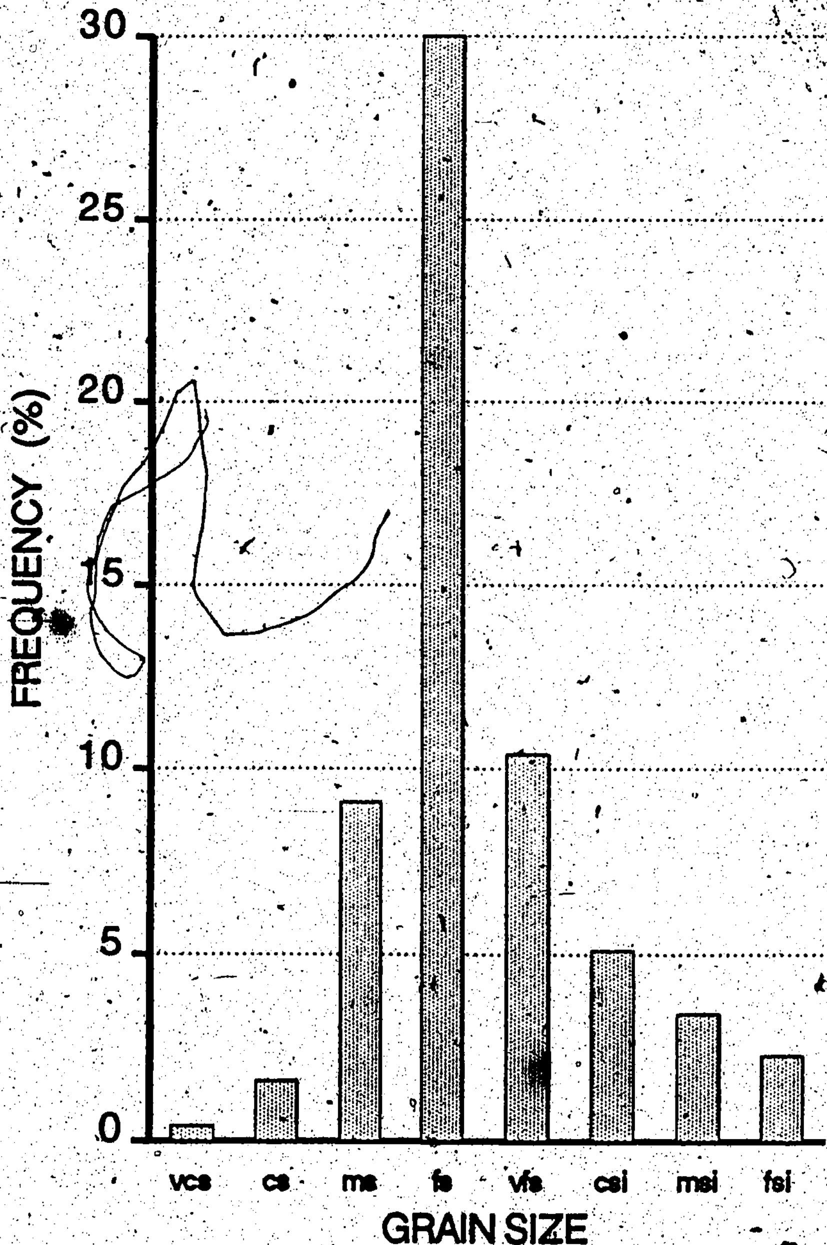

Figure 9. Example histogram of grain size distribution for the Del Bonita Ck2 horizon. (sample T20)

the DLB soil are a closer representation of a normal distribution. The converse occurs for the Ck horizon samples, with the kurtosis values of the Ck horizons from the BZR series being consistently greater than the values of the DLB soil. The variability of sample distributions appears to be less significant when expressed as kurtosis. The differences between the representative samples are similar in terms of kurtosis. Therefore, using this "more" sensitive test of normality, distinct differences between the particle size distributions of these soils are not apparent.

The variability and heterogeneity of the Del Bonita material as expressed by sorting indices as well as skewness and kurtosis values may be the product of periglacial processes after deposition of the material. Environmental conditions on the Plateau should have been similar to those on the Cypress Hills assuming that both areas were nunataks during the late Wisconsin glaciation. Westgate (1968), Jungerius (1969) and Catto (1981) all accounted for the occurrence of quartzitic stones and coarse sand within the upper surface material on the Cypress Hills as being a result of frost activity. They explained that these coarse components from the underlying Cypress Hills geologic formation were uplifted and incorporated within the loess during the period of periglacial conditions. Similar observations have been recorded in other areas of the world which have experienced periglacial activity (Korstaj et al. 1982). This scenario can explain the textural variability of the surface material on the Del Bonita Plateau. In this situation, the underlying Flaxville gravels provided the coarse material, which alters the particle size distribution frequency curves of the Del Bonita material. Laurentide till is naturally poorly sorted due to it's genesis.

3. Heavy and Light Mineral Fractionation

The percentage of heavy and light minerals within the fine sand fraction of the BZR and DLB soils are listed in Table VI. The quartz and feldspar composition of the light mineral fraction is also summarized. Some trends distinguishing the two soils are apparent from this analysis.

The DLB soils contain less heavy minerals than the BZR soils. The heavy mineral percentage values for the DLB soils varies from 0.5-2.0. The BZR samples range from 1.8 - 3.0%

In comparison to the heavy mineral percentages documented in the literature, these results are possibly inflated. Smith (1979) reported heavy mineral percentages of less than one percent within the fine sand fraction of Cordilleran till samples, which he reports as being

Table VI. Heavy and light Mineral Fraction, and Composition of the Light Mineral Fraction.

Field No.	Lab No.	Sp. gravity > 2.95 Heavy Mineral %	Sp. gravity < 2.95 Light Mineral %	Composition of Light Mineral Fraction				
				% Quartz	% Anorthite	% Albite	% Orthoclase	% Feldspar
BZR 102	T1	2.5	97.5	70	3	16	11	30
202	T2	2.6	97.4	71	3	16	10	29
302	T3	2.4	97.6	72	3	15	10	28
DLB 102	T4	2.0	98.0	75	2	12	11	25
202	T5	1.5	98.5	74	2	12	12	26
302	T6	0.7	99.3	77	2	12	9	23
BZR 101	T9	2.6	97.4	73	3	14	10	27
201	T10	3.0	97.0	73	4	14	9	27
301	T11	2.4	97.6	70	3	17	10	30
DLB 101	T12	1.6	98.4	75	2	11	12	25
201	T13	1.3	98.7	73	3	14	10	27
301	T14	0.7	99.3	76	2	12	10	24
BZR 104	T15	1.8	98.2	72	1	17	10	28
204	T16	2.0	98.0	76	1	15	8	24
304	T17	1.8	98.2	75	1	15	9	25
DLB 104	T18	1.0	99.0	80	1	10	9	20
204	T19	1.6	98.4	81	1	10	8	19
304	T20	0.5	99.5	82	1	9	8	18
Sieved gravels	T21b	0.7	99.3	77	3	13	7	23

comparable to other researcher's values for similar materials in the Rocky Mountains. Twardy (1969), from his analysis of Laurentide till materials around Edmonton, Alberta, found the heavy mineral fraction to range from 0.7-1.5%. On the other hand, Catto (1981) found the till and loess materials in the vicinity of the Cypress Hills to contain more heavy minerals than the materials in this study area.

The larger light mineral fraction of the DLB soils contains up to 12% more quartz than the fine sand fraction from the BZR series. Conversely the light mineral fraction of the BZR soil contains up to 12% more feldspars, primarily albite and orthoclase, than samples from the DLB series. The low amounts of anorthite, Ca feldspar, in both soils reflects the susceptibility of this mineral to weathering. Anorthite is the least stable feldspar whereas albite and orthoclase are the most stable feldspars based on Goldrich's (1938) stability series (Huang 1977).

The difference in quartz-feldspar ratios of the DLB and BZR soils is consistent and markedly striking. Catto (1981 and 1983) noticed a similar relationship between the loess on the Cypress Hills and associated tills in the area. The proportion of feldspar was consistently less within the loessial material than the till. He accounted for these mineralogical differences on the basis of two factors - weathering and differential sorting by wind. Feldspars are less resistant to the physical and chemical weathering processes than quartz. Therefore, at the surface of exposed morainal or glaciofluvial deposits upon deglaciation, the feldspar grains are more weathered than quartz. These weathered feldspars, pulverized by glacial action (Smalley 1966) and altered by chemical reactions which are favored by the ameliorating climate as well as the presence of water, resemble platy clay minerals (Huang 1977). Angular and spherical grains are more easily incorporated in the suspension load of wind currents. Therefore, due to the difference in grain shape of quartz and weathered feldspar, quartz is possibly differentially selected by eolian processes. The result is that the loessial deposits on the Cypress Hills contain more quartz than feldspar (Catto 1983). Similar quartz to feldspar relationships are observed between the different parent materials of this study area on the Milk River Ridge.

4. X-Ray Diffraction Analysis

i. Clay Mineralogy

The purpose of this analysis was to investigate a possible difference in clay mineralogy between the BZR and DLB soil parent materials. Initial qualitative - semi-quantitative analysis of clay minerals is possible by X-ray diffractometry (Alexiades and Jackson 1966). X-ray diffraction analysis was performed on the representative A, B, and C horizons from these two soils. Also the clay fraction from the sieved Flaxville gravel sample at the DLB site no. 1 was analyzed for comparative purposes. Selected X-ray diffraction patterns of the clay fraction from the DLB and BZR soils, as well as sieved gravels are shown in Figures 10 to 14.

The suite of clay minerals represented in these diffraction patterns are essentially identical, throughout all profile samples and even the sieved gravels. Smectite, mica and kaolinite are consistently present as the dominant clay minerals, with trace amounts of chlorite and quartz.

The smectite component of the clay fraction consists primarily of montmorillonite. This is indicated by the existence of a relatively sharp peak at 16 A for the Ca glycerol treatment. The slight broadening of this peak with increasing 2 θ angle in some samples (eg. #4 and #21b) indicates that some beidellite is present (Borchardt 1977).

Mica in the form of muscovite occurs in all samples. This clay mineral is identified by the consistent presence of the 10 A peak. The existence of the 5 A peak for the Ca 54% RH treatment and K 0% RH treatments shows that the

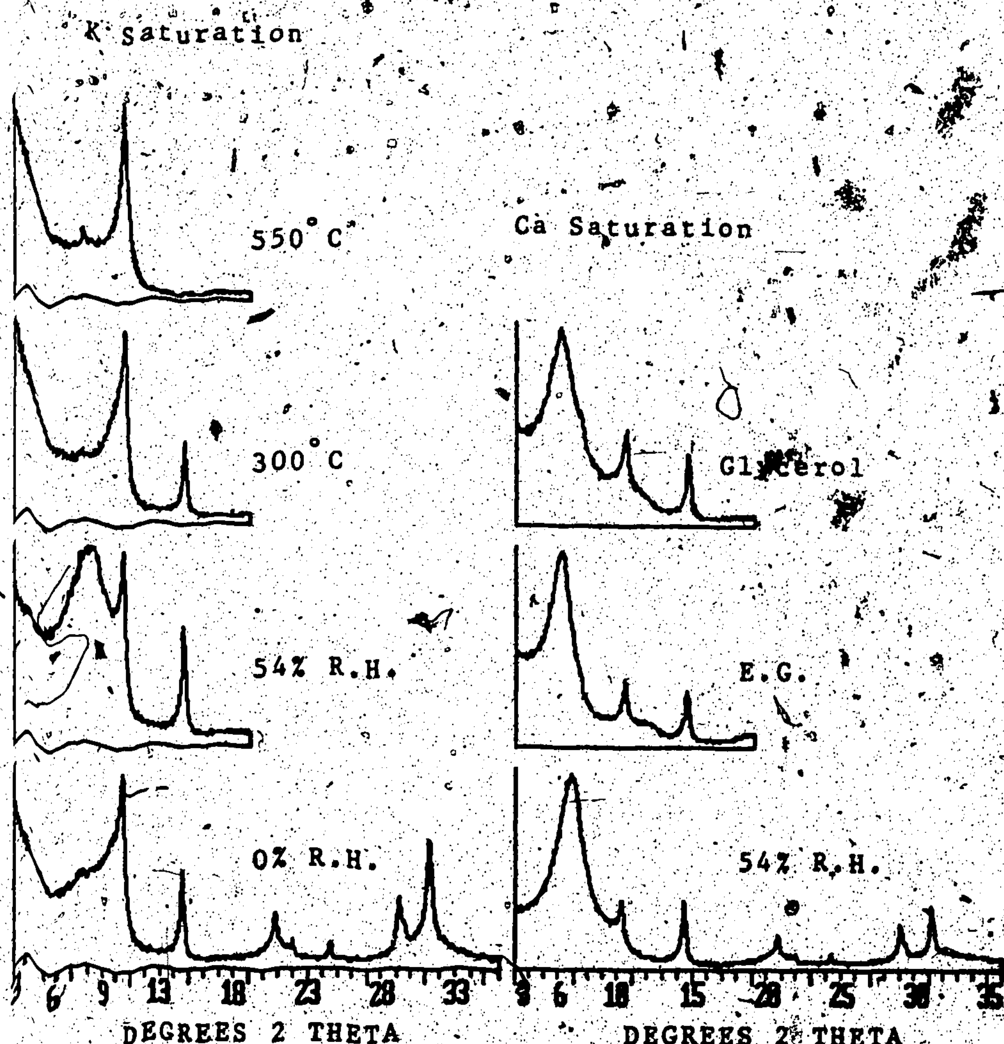

Figure 10. X-ray diffractogram of the total clay fraction from the Bm horizon, Beazer soil. (sample T2)

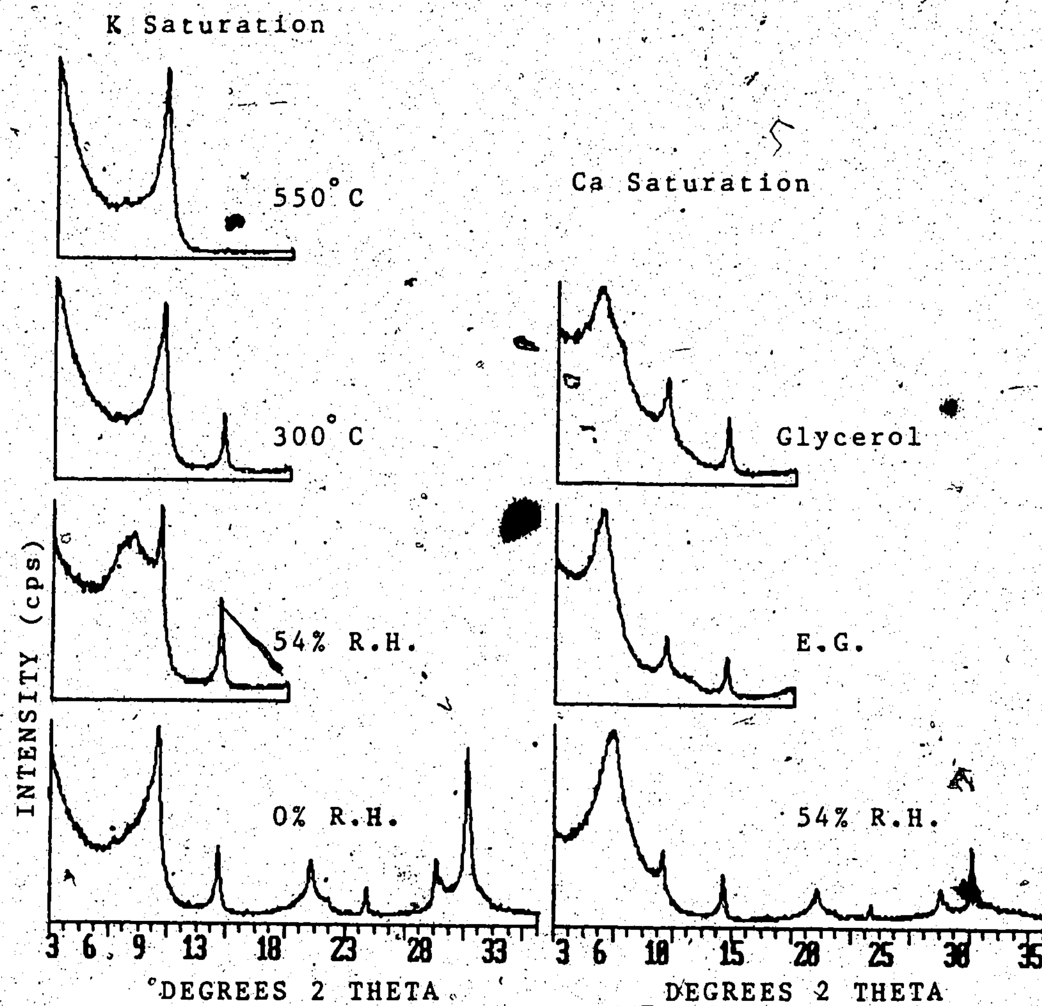

Figure 11. X-ray diffractogram of the total clay fraction from the Bm horizon, Del Bonita soil. (sample T4)

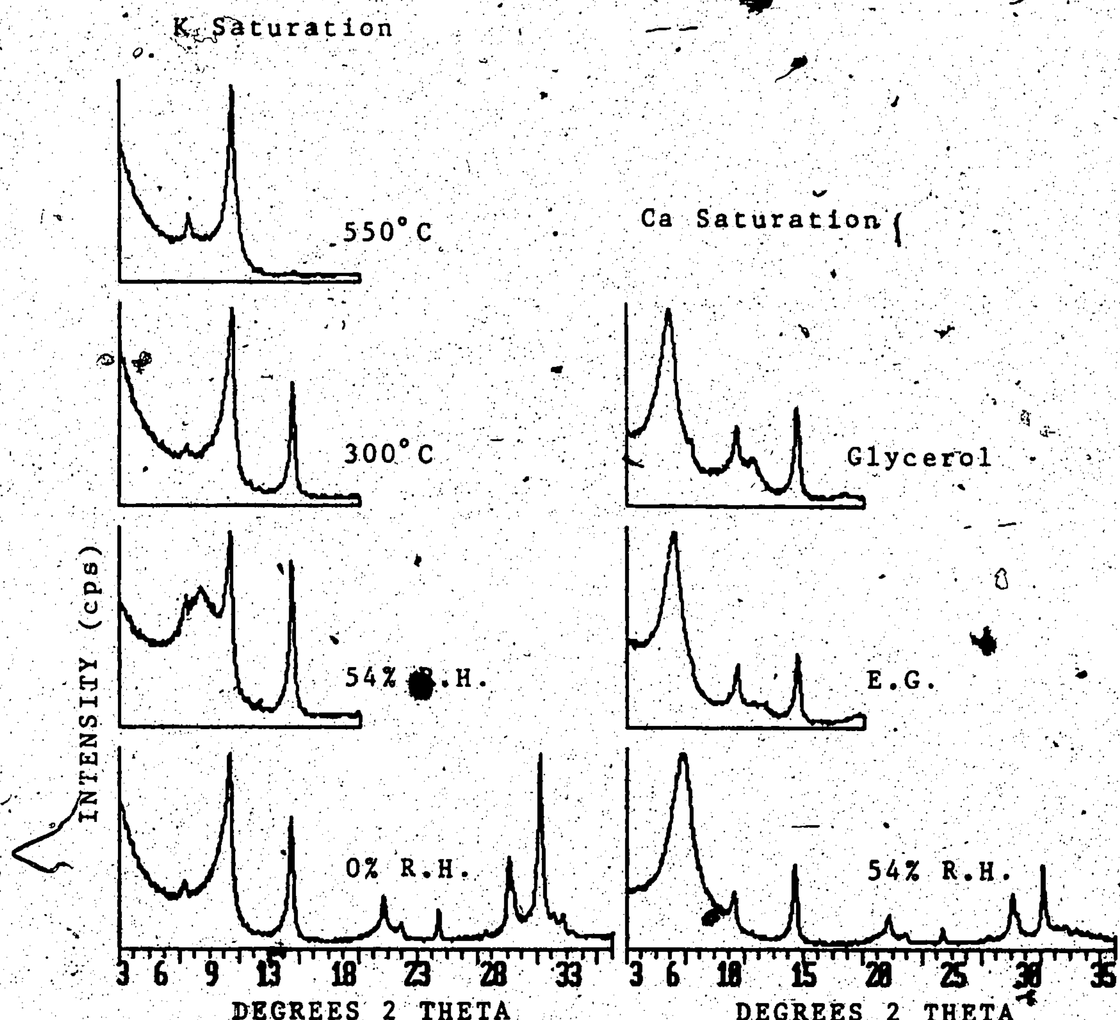

Figure 12. X-ray diffractogram of the total clay fraction from the Ck2 horizon, Beazer soil. (sample T16)

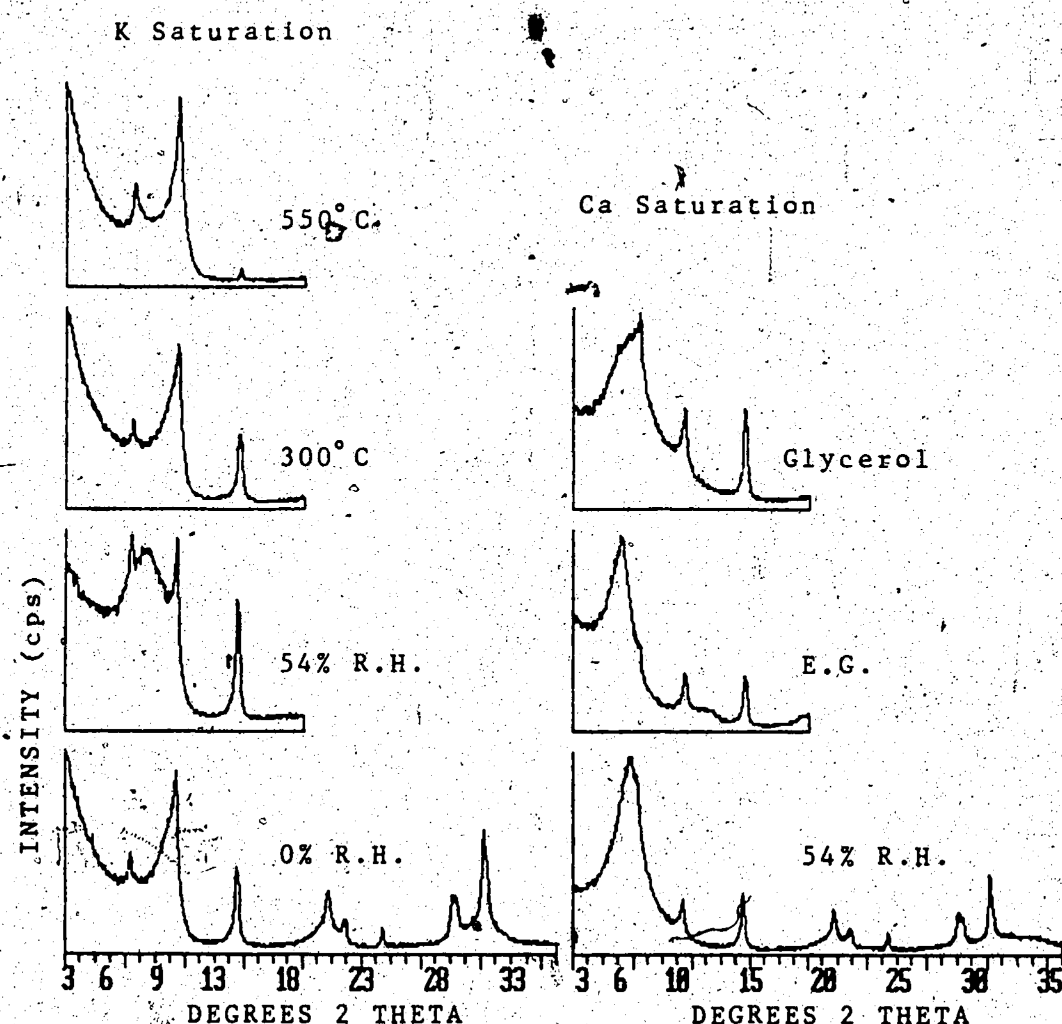

Figure 13. X-ray diffractogram of the total clay fraction from the Ck2 horizon, Del Bonita soil. (sample T19)

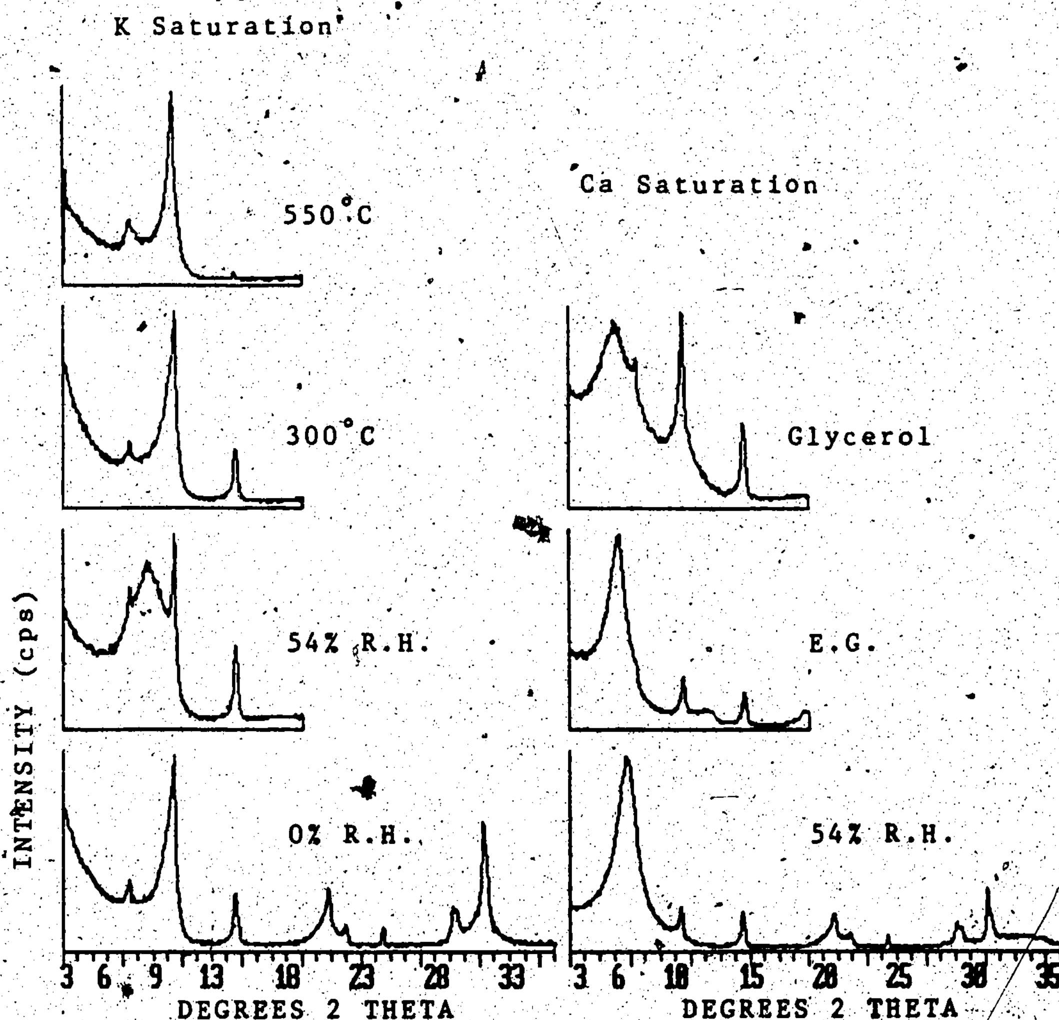

Figure 14. X-ray diffractogram of the total clay fraction from the sieved gravels. (sample T21b)

mica is the diotahedral form, muscovite (Fanning and Keranidas 1977).

Kaolinite is identified by the presence of the 7 A peak which disappears with the K 550°C treatment. The low amount of chlorite within these samples eliminated the possible confusion with second order chlorite peaks (Pettapiece 1970). Therefore kaolinite identification was straight forward.

Chlorite exists in very minor amounts as indicated by the low intensity peaks at 14 A. The peak at this angle for the K 550°C treated clay samples is the diagnostic peak for the mineral. The chlorite peak is more apparent with the collapse of smectite and associated intergrades (Barnhisel 1977). The intensity of chlorite peaks fluctuates slightly between samples, with some appearing quite prominent especially in the C horizon samples. As the second order peaks are insignificant the amount of chlorite is considered minimal. Similarly, the chloritic intergrades are assumed to be of minor significance because the diffraction peaks occurring as shoulders between 10-14 A for the K 550°C treated clays are non-existent.

As observed from Figures 10 to 14, the X-ray diffraction patterns and associated suite of clay minerals are consistent for all samples. Because of this fact it was deemed unproductive to analyze the clay mineralogy in further detail.

The suite of clay minerals present in the BZR soil horizons is consistent with documented clay mineralogy of Laurentide till samples. Kodama (1979), in his summary of clay mineralogical data of Canada, states that smectite and mica characterize the soils found on the Interior Plains. Dudas and Pawluk (1982) report that the clay mineralogy of continental till from the Calgary vicinity and southeastern Alberta consists of smectite, mica, kaolinite and chlorite in order of decreasing abundance. The smectite component was dominantly montmorillonite with some beidellite, and muscovite was the dominant mica constituent. Twardy (1969) from his analysis of till samples around Edmonton, Alberta reports that montmorillonite, illite and intergrades of these clay minerals are dominant, while chlorite and kaolinite are present in minor amounts. Spiers (1982) similarly found that smectite and mica were the dominant clay minerals of Laurentide tills around Fort McMurray, Alberta. However, he also reported abnormally high levels of chlorite and kaolinite than previously described for similar till materials in Alberta. These higher levels of chlorite and kaolinite were attributed to the incorporation of local bedrock lithology within the glacial till in the Fort McMurray area (Spiers 1982). The same hypothesis may be used to explain the presence of kaolinite as a dominant clay mineral in the BZR samples. The shale bedrock of the Bearpaw formation which is areally extensive in southern Alberta (Green 1972) provides a source of kaolinite for the Laurentide till on the Milk River Ridge.

For the purpose of comparison, Cordilleran tills contain a different suite of clay minerals. Tills originating from the Rocky Mountains characteristically lack smectite. Mica, chlorite, vermiculite and quartz are the principal clay minerals with some kaolinite present in minor amounts (Smith 1979; Pettapiece 1970).

The BZR soil parent material is therefore definitely Laurentide till due to the presence of smectite and absence of chlorite and vermiculite. Since the clay mineralogies of the DLB and BZR samples are identical (based on X.R.D.) the DLB parent material must be derived from Laurentide till. Because we know that the DLB parent material is not till due to the absence of granitic coarse fragments, the material on the Del Bonita Plateau must have been deposited by fluvial or eolian processes.

The coincidence that the sieved gravel clay fraction mineralogy is identical to the BZR and DLB soils is because of leaching and cryoturbation processes. The intermixing of the upper and lower deposits by these processes has occurred since deposition of the upper surficial material on top of the Flaxville gravels. The result is that the clay materials extracted from the gravels reflects the mineralogy of the overlying material. The suite of clay minerals of Cordilleran origin is not evident, as originally expected.

ii. Calcite - Dolomite Analysis

The approximate proportions of calcite and dolomite minerals present within the Ck horizons from the DLB and BZR soils was determined from analysis of the X-ray diffraction patterns. Prior to this analysis it was known that the calcium carbonate equivalent values for the DLB soils were greater than the BZR soil values. This further analysis sought to determine whether carbonate mineralogical differences correspond with the different $CaCO_3$ equivalent values.

In carbonate studies using X-ray diffraction patterns, the important d-spacings are 3.036 A for calcite and 2.89 A for dolomite (Goldsmith et al. 1955; Doner and Lynn 1977). The presence of broader asymmetrically shaped calcite peaks, at larger angles of 2 θ are attributed to the presence of secondary magnesium calcites (St. Arnaud and Herbillion 1973). X-ray methods of analysis are adequate in identifying the presence and proportion of calcite, dolomite and magnesium calcite intergrades in soil samples (St. Arnaud and Herbillion 1973). Mg-calcite reacts most like calcite, but due to the presence of magnesium within the calcite structure, these secondary intergrades have different properties. As the proportion of Mg increases, the d-spacing decreases and the solubility increases (Goldsmith et al. 1955; St. Arnaud and Herbillion 1973).

The proportion of calcite to dolomite minerals as determined from analysis of the X-ray patterns are summarized as follows:

BZR	Ck_1	calcite	>=	dolomite
	Ck_2	dolomite	>	calcite
DLB	Ck_1	calcite	>>	dolomite
	Ck_2	calcite	>	dolomite

Figures 15 to 18 show the diagnostic peak intensity differences between the carbonate samples obtained from the BZR and DLB soils.

The relative proportion of calcite to dolomite in the BZR Ck_2 horizon indicates that this material, at the solum depth of 80 cm, is possibly less modified by weathering processes than the other Ck horizon samples. The presence of secondary magnesium calcite is indicated by the broad asymmetric peak at 3.037 Å (Figure 16). Secondary Mg-calcites are more soluble that other secondary carbonates which in turn are more soluble that calcite and dolomite (Doner and Lynn 1977; St. Arnaud and Herbillion 1973).

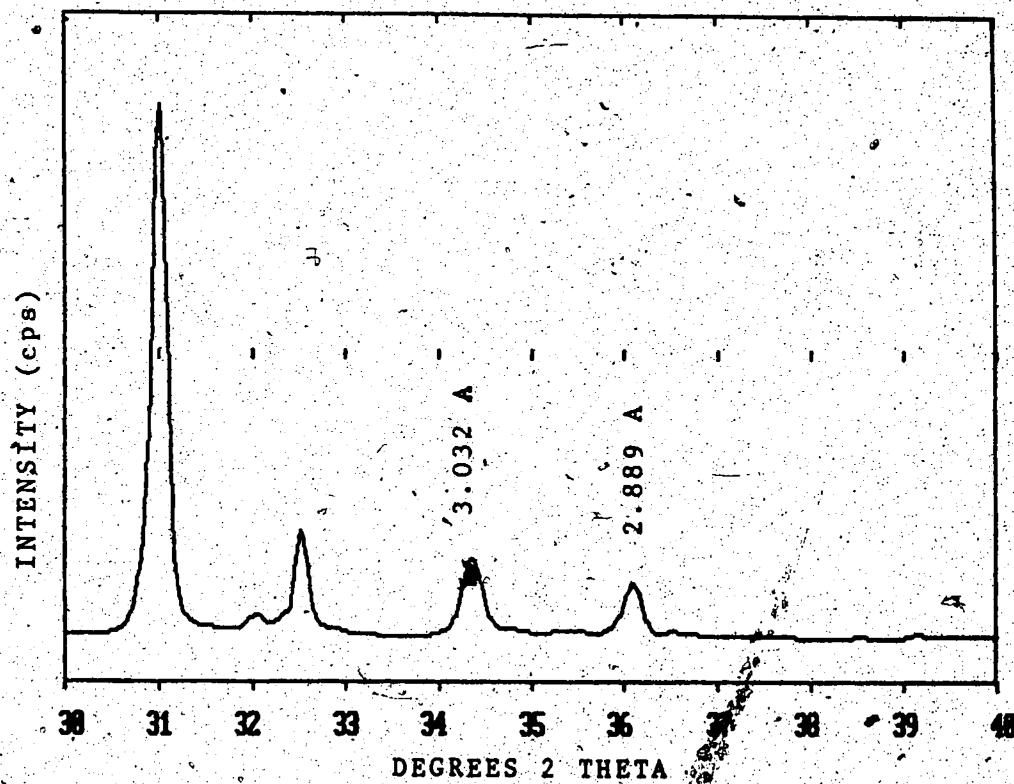

Figure 15. X-ray diffractogram of calcite and dolomite minerals from the Ck1 horizon, Beazer soil. (sample TBZCA203)

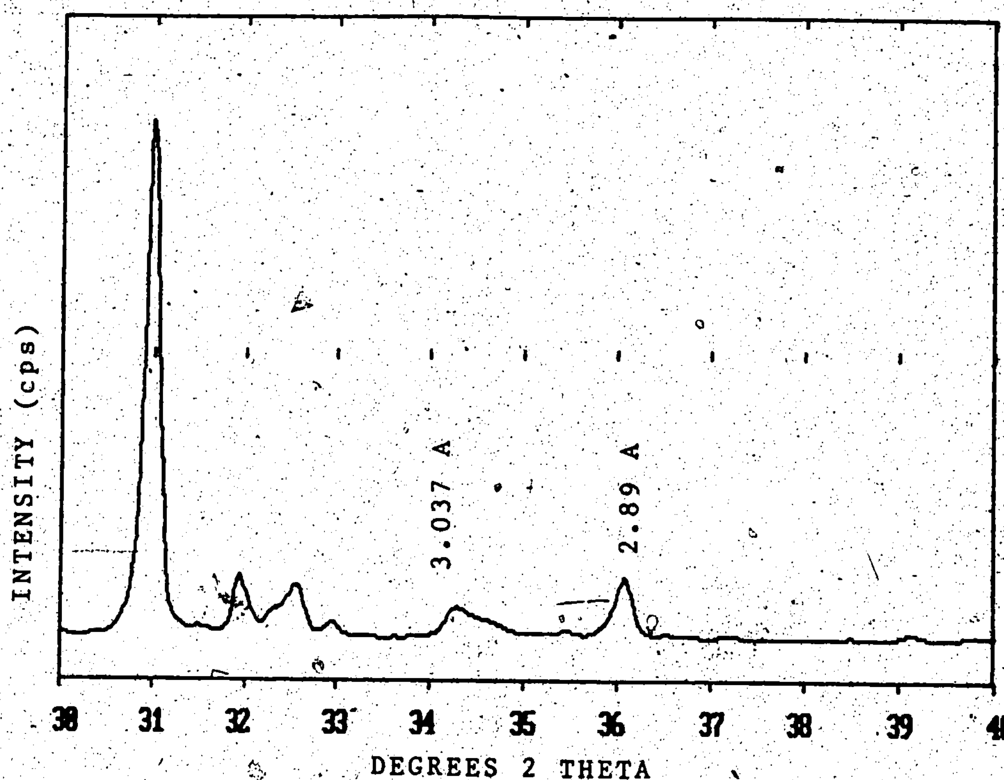

Figure 16. X-ray diffractogram of calcite and dolomite minerals from the Ck2 horizon, Beazer soil. (sample TBZCK204)

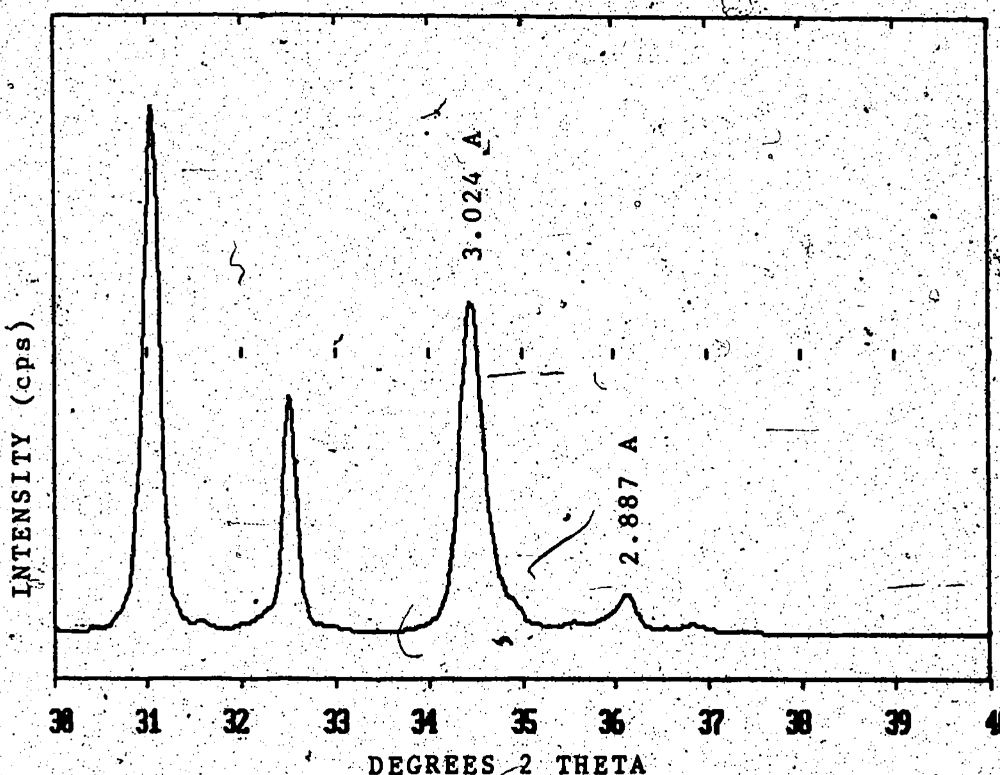

Figure 17. X-ray diffractogram of calcite and dolomite minerals from the Ck1 horizon, Del Bonita soil. (sample TDLCA203)

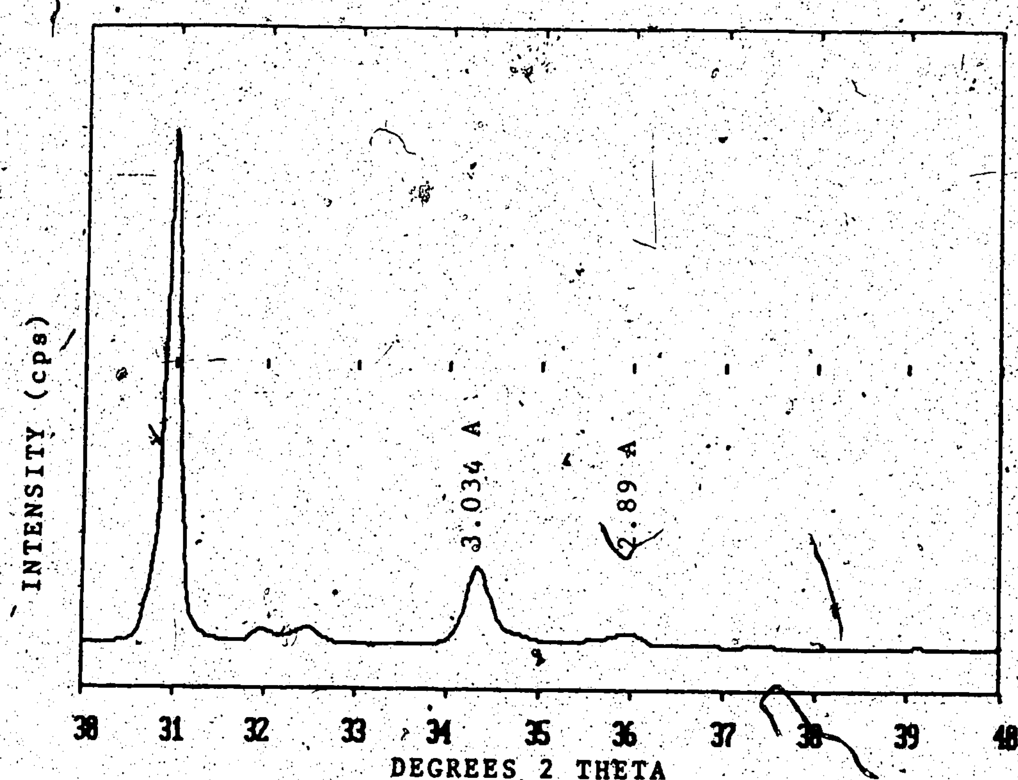

Figure 18. X-ray diffractogram of calcite and dolomite minerals from the Ck2 horizon, Del Bonita soil. (sample TDLCK204)

The presence of more dolomite than calcite in conjunction with the soluble secondary Mg-calcite in this BZR Ck_2 horizon is easily distinguishable in this sample. Because the calcite peak of this Ck_2 horizon is so small, the proportion of Mg-calcite skews the 3.037 Å peak. In the other samples, where the proportion of calcite is much larger, the presence of Mg-calcite is harder to detect.

Within the BZR Ck_1 horizons (Figure 15) the proportion of calcite exceeds dolomite therefore pedogenic processes are assumed active. The calcite X-ray diffraction peak at 3.032 Å is symmetrical so the calcite is assumed to be relatively pure (little Mg substitution). This accumulation of calcite in the Ck_1 horizon corresponds with the relative increase in $CaCO_3$ equivalent in comparison to the Ck_2 horizon (15% in the Ck_1, 12% in the Ck_2).

The X-ray diffraction patterns of the Ck horizons from the DLB soils consistently indicate that more calcite than dolomite is present (Figures 17 and 18). The corresponding $CaCO_3$ equivalent values of 31% and 17% for these DLB Ck_1 and Ck_2 horizons correlate well to the visible differences of the calcite peak intensities. The proportions of primary and secondary calcite within these peaks are unknown. There may have been more calcite originally present in the DLB soil material than in the

BZR parent material. This possibility warrants consideration because the intensity of the calcite peak for the DLB Ck_2 (Figure 18) horizon is equal to or greater than the BZR Ck_1 horizon peak (Figure 15). This latter peak has been secondarily enriched by pedogenic processes as previously described. Therefore, the presence of more calcite in the DLB Ck_2 horizon must be due to previous evolutionary stages or a longer weathering cycle of this Del Bonita Plateau material.

5. Scanning Electron Microscopy

i. Sand Grain Analysis

The surface characteristics observed on quartz sand grains by means of the scanning electron microscope are summarized in Table VII. A total of 106 grains from the B and C horizons of the DLB and BZR soils were described.

The quartz sand grains from the B and C horizons of the BZR soil exhibit characteristics of glacial materials. Conchoidal fractures, break blocks, arc shaped steps, parallel steps, subangular and dominantly medium to high relief are the crucial characteristics (Bull 1981; Krinsley and Doornkamp 1973; Whalley and Krinsley 1974). Since the parent material of the BZR soil is till these findings are not surprising.

Table VII. Surface Characteristics of Quartz Sand Grains

Surface Feature Characteristic	BZR B horizon total:26 grains	BZR C horizon 25 grains	DLB B horizon total:30 grains	DLB C horizon 25 grains
conchoidal fracture	92%	64%	60%	12%
breakage blocks	65%	64%	50%	24%
arc shaped steps	58%	36%	37%	28%
scratches and grooves	23%	12%	30%	4%
V-shaped marks	50%	40%	57%	48%
parallel steps	42%	56%	33%	32%
dish shaped concavities	4%	12%	17%	52%
upturned plates	85%	92%	97%	80%
roundness: rounded	4%	12%	27%	64%
subrounded	42%	52%	50%	28%
subangular	50%	16%	20%	8%
angular	4%	20%	3%	0%
carapace	31%	64%	60%	68%
silica ppt	50%	72%	60%	92%
solution pits	77%	56%	77%	80%
relief: low	8%	20%	30%	52%
medium	69%	64%	67%	48%
high	23%	16%	3%	0%

The high occurrence of silica precipitation and solution pits on the sand grains from the BZR soil samples, is also not believed to be unusual for basal deposits within moraines. Whalley and Krinsley (1974) noted appreciable amounts of silica precipitation as well as carapace on sample grains collected from a lateral moraine. They account for the precipitation on the grains surfaces being due to the high stress of interparticle contacts. A high pH solution is obtained if mafic mineral particles are ground in the presence of water. If the pH of the solution is greater than 9 then amorphous silica goes into solution. Even at pHs less than 9 dissolution will occur but at a much slower rate (Whalley and Krinsley 1974; Manker and Ponder 1978).

The occurrence of carapace or comminution debris on glacial sand grains is also a common occurrence (Smalley and Cabrera 1970; Krinsley and Doornkamp 1973; Whalley and Krinsley 1974). Whalley and Krinsley (1974) document this feature as being more frequent on subglacial grains. Other researchers have simply depicted comminution debris as being a product of glacial grinding, and this feature may persist on other deposits such as loess (Smalley and Cabrera 1970). Pye (1983) feels that this debris is held by electrostatic charge.

The sand grains from the Del Bonita Plateau, associated with the DLB soil developed on unknown parent material exhibit some trends which are different than the BZR examples. The DLB sand grains exhibit significant differences in relief and roundness (Table VIII) (Silk 1979; Steele and Torrie 1980). Also, significant differences occur with respect to dish-shaped concavities, conchoidal fractures, breakage blocks and carapace surface features. The remaining features such as V-shaped marks, upturned plates, silica precipitation, arc shaped steps and parallel features show no significant difference in their presence on the grains of these soil samples. Generally the quartz sand grains from the DLB soil are more rounded and have lower relief relative to the grains from the BZR soil.

The importance of these observations on the sand grains from the DLB soil remains ambiguous. Eolian or loess particles are classically characterized by the presence of dish shaped depressions, low rolling topography, meandering ridges, upturned plates and rounded corners (Krinsley and Doornkamp 1973). These features are believed to be a result of particle interaction during saltation. The DLB grains indicate some tendency of being loess due to the presence of dish shaped depressions, in conjunction with low relief androundedness (Hill and Nadeau 1984). Another feature indicative of loess is the

Table VIII. Statistical Parameters for Evaluation of the Chi-squared test by Surface Characteristics and Horizons

Surface feature characteristic	horizon	chi squared critical value	chi squared calculated value	decision
conchoidal fracture	Bm	3.86	7.76	reject H0, accept H1
	Ck	3.86	14.36	reject H0, accept H1
breakage blocks	Bm	3.86	1.35	accept H0
	Ck	3.86	8.12	reject H0, accept H1
arc shaped steps	Bm	3.86	1.49	accept H0
	Ck	3.86	0.34	accept H0
scratches and grooves	Bm	3.86	0.45	accept H0
	Ck	3.86	1.08	accept H0
V-shaped marks	Bm	3.86	0.25	accept H0
	Ck	3.86	0.32	accept H0
parallel steps	Bm	3.86	0.33	accept H0
	Ck	3.86	2.92	accept H0
dish-shaped concavities	Bm	3.86	3.88	reject H0, accept H1
	Ck	3.86	10.63	reject H0, accept H1
upturned plates	Bm	3.86	2.41	accept H0
	Ck	3.86	1.48	accept H0
roundness	Bm	7.81	8.4	reject H0, accept H1
	Ck	7.81	17.49	reject H0, accept H1
carapace	Bm	3.86	4.78	reject H0, accept H1
	Ck	3.86	0.27	accept H0
silica precipitate	Bm	3.86	0.56	accept H0
	Ck	3.86	3.38	accept H0
solution pits	Bm	3.86	3.32	accept H0
	Ck	3.86	0	accept H0
relief	Bm	7.81	7.90	reject H0, accept H1
	Ck	7.81	8.24	reject H0, accept H1

H0: There is no statistically significant difference in the distribution of proportions of a surface feature characteristic between DLB and BZR soil horizons.

H1: There is a statistically significant difference in the distribution of proportions of a surface feature characteristic between DLB and BZR soil horizons.

Level of significance = 5%

presence of a "dirty" surface (Pye 1983). Grains from both the DLB and BZR soil samples exhibit "dirty" rough surfaces, where silica precipitation was the dominant agent. The proportion of silica precipitation described on the grains from the C horizons is consistently greater because some $CaCO_3$ encrustations were probably included within these surface feature descriptions.

Pye (1983) has shown that some carapace may be removed by sonic vibration. Since these samples were initially exposed to sonification as well as a milk shake mixer during initial fractionation of the samples the proportion of communition debris may be underestimated however this is consistent between samples.

The presence of V-shaped markings are features of subaqueous or fluvial origin (Krinsley and Donahue 1968; Krinsley and Doornkamp 1973; Manker and Ponder 1978; Hill and Nadeau 1984). Krinsley and Doornkamp (1973) also state that upturned plates occur in conjunction with V markings however this combination has not been extensively noted by other researchers. Both these features are due to abrasion. These features are generally believed to be due to mechanical action when unorientated, and of a chemical origin when orientated in linear patterns (Bull 1981). The number of grains from the DLB and BZR soil samples exhibiting V markings are primarily the same.

Even though these markings were present they rarely were found at densities as high as 2 μm^2 which is a criteria for high energy beach environments (Hill and Nadeau 1984). Manker and Ponder (1978) question whether impact pits and meandering ridges are truly eolian features since they recognized similar features on fluvial deposit sand grains.

The presence of upturned plates and V markings on the sand grains from the DLB soil is camouflaged by silica precipitation. The amount of precipitation dominantly consisting of alumino silicates, with some $CaCO_3$ crustrations as determined by using the Kevex analyzer, may camouflage the existence of V markings and upturned plates. Therefore the presence of these features on the sand grains from the DLB soil may have been underestimated in this analysis.

The following conclusions may be drawn from the SEM analysis. The sand grains from the BZR soil exhibited the recognized surface texture features characteristic of glacially derived materials (Figure 19). In comparison, the DLB soil sand grains occasionally exhibited the typical rounded shape, low relief surface and abundant silica precipitation associated with grain surfaces of eolian origin (Figure 20). These "classic type" sand grains are the exception with the majority of the sand grains from both soil's exhibiting intermediate surface

Figure 19. Classic glacial quartz sand grain from the Beazer soil. Note the conchoidal fractures, arc shaped steps, angular shape and medium to high surface relief.

Figure 20. Classic eolian quartz sand grains from the Del Bonita soil. Note the spherical shape and smooth surface covered with silica precipitate.

Figure 21. Typical "intermediate" quartz sand grain from the Beazer soil. Note the conchoidal fracture, bottom right, but the associated medium relief and subangular to subrounded shape.

Figure 22. Typical "intermediate" quartz sand grain from the Del Bonita soil. Note the dish shaped concavity, upper right, conchoidal fracture, lower left, pitted surface, medium relief and subangular to subrounded shape.

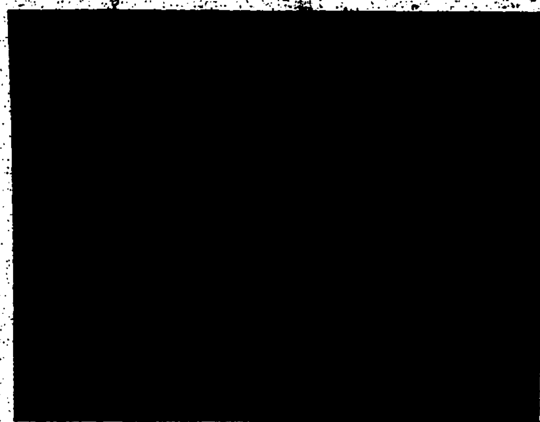

Figure 19

Figure 20

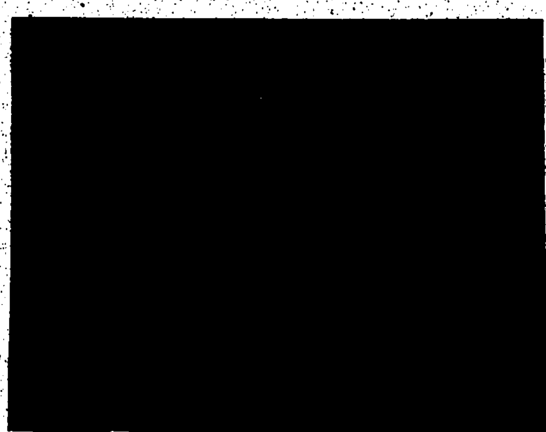

Figure 21

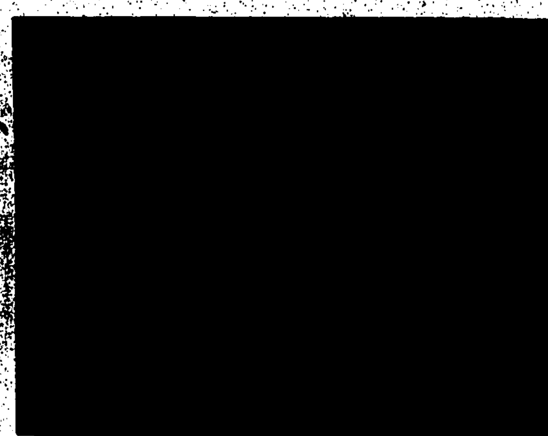

Figure 22

features (Figures 21 and 22). These latter grains have surface features which are a product of glacial, fluvial and eolian processes, andthere is an overlap of features between the two soil materials. The BZR soil sand grains are generally more "glacial" in origin, evidenced by the greater percentage of conchoidal fractures and subangular shaped grains. The sand grains from the DLB soil are generally more subrounded and have a lower overall surface relief. On the basis of these observations the Del Bonita material may have originated from glacial deposits but it has been reworked by fluvial and/or eolian processes. More conclusive statements concerning the origin of the sand grains from the DLB and BZR soils, on the basis of this analysis, are not possible.

ii. Plant Phytolith Assemblages

The purpose of investigating phytolith assemblages of surface soil samples is to determine whether different paleoecological regimes existed in a particular area. The premise of this procedure is that plant phytoliths provide a fossil record of past ecological systems that may have existed at a particular site (Lutwick 1969). Surface soil samples were collected from the Del Bonita Plateau (the DLB soil sample ssites) and the surrounding area on the Milk River Ridge (the BZR soil sample sites) to see if the phytolith assemblages do vary between areas. With the DLB parent material theoretically being older, a unique qualitative and/or quantitative assemblage was anticipated.

Plant opal is hydrated amorphous silica of biogenic origin (Wilding et al. 1977). Plants incorporate silica into the epidermal cells of their leaves resulting in the development of opaline silica bodies, referred to a phytoliths (Jones and Handreck 1967; Twiss et al. 1969). The rate of silica uptake varies between plant species. For example, grasses contain 10-20 time the concentration of silica found in legumes and other dicotyledons (Russel 1961:536 as reported by Jones and Handreck 1967). Therefore, since plant phytoliths "assume the shape of the cells in which they are deposited", some plant opals are diagnostic of certain plant species (Lutwick 1969:77).

The classification of phytoliths is based on morphology. The distinction between forest opal and grass opal is well established. Wilding et al. (1977:487) describe deciduous forest opal as consisting of "incrustations of cellular components with numerous thin sheet structures while grass opal consists mostly of solid polyhedral structures resulting from silicification of the entire cell". Norgren (1973) extracted plant opal directly from conifer needles. He noted that there was some similarity of features with phytoliths from grass species. However he felt that he was able to differentiate coniferous forest opals from grass opal.

Twiss et al. (1969) subdivided the Graminaceae family into four groups based on shape. Norgren (1973) expressed two reservations in using this classification. First, was that the three subfamilies Festucoid, Panicoid and Chloridoid contain many grass species. Second point was that the last group, the Elongate class, was too all encompassing since it is characterized by rod shaped phytoliths, and this phytolith is the most common. However, more recently it has been shown that these four groups are taxonomically significant and environmentally sensitive (Johnson el al. 1984). Therefore, this classification scheme was employed in this study, and forest opaline bodies were to be observed as a separate class, although none were present.

The results of the phytolith assemblage analysis are summarized in Table IX. Quite simply there is no difference between the DLB and BZR sample sites in terms of phytolith assemblages. The dominant proportion of phytoliths were classified as Elongate with Festucoid class phytoliths consistently present. The phytoliths extracted from the plants presently growing on these sites indicate similar phytolith assemblages. Thus the present day indigenous grass species, *Festuca scabella*, is the

Table IX. Summary of Phytolith Classification

	Classes of Phytoliths *					
Sample	Chloridoid	Festucoid	Panicoid	Elongate	Unclassified	Total
DLB	1 (2.4%)	11 (24.4%)	5 (11.1%)	25 (55.6%)	3 (6.7%)	45
BZR	1 (2.2%)	12 (26.7%)	2 (4.4%)	29 (64.5%)	1 (2.2%)	45
Plant extracted						
DLB	-	5 (33.3%)	-	10 (66.7%)	-	15
BZR	-	4 (26.7%)	3 (20.2%)	8 (53.3%)	-	15

* classes of phytoliths based upon Twiss et al. (1969) subdivision

principal supplier of phytoliths described in surface soil samples from these two areas.

On the basis of this analysis - the conclusion attained is that the Del Bonita Plateau and the surrounding area on the Milk River Ridge have experienced the same ecological regime.

In hindsight it appears that in order to detect a difference in phytolith assemblages a more detailed sampling regime needs to be implemented. Samples should have been collected at regular intervals to a depth of 2 m or to the contact between the surface material and the Flaxville gravels on the Del Bonita Plateau. Sampling at this increased density and depth would possibly enable a researcher to detect the presence of a paleosol and associated vegetation community. Previous paleoenvironments of the Del Bonita Plateau and neighboring Milk River Ridge area may have then been distinguishable. Upon visual inspection of the gravel pits on the Plateau no paleosols were noticed so this sampling procedure was not persued. Cryoturbation disturbances may account for the visual absence of relict horizons from paleosols.

6. Micromorphology

The micromorphological analysis of the thin sections obtained from the three profiles (1 BZR and 2 DLB soil profiles) revealed some distinct and unique features.

This analysis indicates that the parent materials of the DLB and BZR soils have experienced different climatic and geomorphic regimes. The DLB soils have fabric arrangements characteristic of cryogenic processes (Koniscev et al. 1973; Bunting and Fedoroff 1974; Fox and Protz 1981) while these are absent in the BZR soils. The micromorphological descriptions are summarized in Table X.

The Ah and Bm horizons of the BZR and DLB soils are similar in terms of f fabric and plasma fabric descriptions. The fabric of the Ahs vary from ortho granic to iunctic porphyric. These fabrics are similar with grain content of the plasma and accommodation of f members being the only difference (Brewer et al. 1983). The plasma of the Ah horizons is described as skel-masepic. The Ah of the DLB soil #1 profile has more silt and the plasma lacks domains of clay orientation. The plasma fabric is described as silasepic-inundulic. The other Ah horizons have less silt-sized grains, and the zones of orientated clay are associated with skeleton grains, voids and as islands within the matrix. Likewise, the Bm horizons are basically similar. The f fabrics vary from granoidic to porphyric and the plasma fabric is identical. The fabric descriptions of these Ah and Bm horizons are comparable to micromorphological descriptions

Table X. Micromorphology Descriptions of the BZR and DLB soils

BZR Samples - Site #1. NE21 - 1 - 23 - W4th

Sample	Horizon	f fabric	Plasma fabric	Additional features
BZR - B101	Ah	orthogranic	skel-masepic	Fecal pellets, plant roots
BZR - B102	Bm	matri grainoidic-vughy poyphyric	skel-masepic	Vughs, interconnected vughs
BZR - B103	Ck_1 @50 cm	fragmoidic - porphyric	masepic (zones of crystic)	Vughs, skew planes, $CaCO_3$ nodules present in patches
BZR - B104	Ck_2 @85 cm	fragmoidic - porphyric	siliasepic - crystic	Vughs, vesicles, craze and skew planes, crystallia chambers

DLB Samples - Site #1. WC21 - 1 - 21 - W4th

Sample	Horizon	f fabric	Plasma fabric	Additional features
DLB - D101	Ah	iunctic porphyric	siliasepic - inundulic	Parallel joint planes, vughs, fecal pellets
DLB - D102	Bm	granoidic-iunctic porphyric (zones of vughy porphyric)	skel-vo-masepic	Vughs, sesquioxide glaebules

DLB - D103	Ck_1 @35 cm	granoidic meta fragmoidic// granic crystic	weak vo-masepic to insepic -	Calcitans and matrans on skeleton grains. Concentric glaebules
DLB - D104	Ck_2 @80 cm	vughy porphyric//granoidic fragmoidic (zones of matri matri plectic porphyric)	crystic to weak skel-masepic	Vughs, concentric glaebules - different stages of development
DLB - D105	Ck_2 @100 cm	matri plectic // matri plectic porphyric // vughy	crystic to weak skelsepic porphyric	Vughs, concentric glaebules - different stages of development
DLB - D106	Ck_2 @150 cm	matri chlamydic//plectic porphyric//vughy porphyric	crystic to weak skel-masepic	Vughs, interconnected vughs, concentric glaebules disappeared-

DLB Samples - Site #3 - SC6 - 2 - 22 - W4th

Sample	Horizon	f fabric	Plasma fabric	Additional features
DLB - D201	Ah	orthogranic // granoidic	skel-masepic	Fecal pellets, craze planes, channels
DLB - D202	Bm	granoidic porphyric//vughy porphyric	skel-vosepic	Vughs, channels. Some sesquioxide glaebules
DLB - D203	Ck_1 @30 cm	meta fragmoidic - granoidic	weak masepic - siliasepic (with silt size clay domains); zones of crystic	Concentric glaebules calcitans on skeleton grains

DLB - D204	Ck_1 @80 cm	granoidic//granoidic vughy porphyric	mosepic and crystic	Concentric glaebules-less developed than D203. Matrans and calcitans on skeleton grains
DLB - D205	Ck_1 @100 cm	plectic porphyric // vughy porphyric	crystic banded with ma-skelsepic	Concentric glaebules different stages of development. Matrans and calcitans on skeleton grains
DLB - D206	Ck_1 @150 cm	vughy porphyric//granoidic (zones of plectic porphyric	ma-skelsepic and crystic	Concentric glaebules - different stages of development. Matrans and calcitans on skeleton grains

of other Black Chernozemic soils in Alberta (Sanborn and Pawluk 1983).

The micromorphological differences between the DLB and BZR soils are visually evident within the Ck_1 and Ck_2 horizons. Using the Brewer (1976), Brewer and Pawluk (1975), and Brewer et al. (1983) terminology the actual difference is not apparent. The description of the BZR soil Ck f fabric is fragmoidic + porphyric and this is not immensely different from the DLB soil Ck horizons (granoidic, fragmoidic and vughy porphyric). However this latter fabric is not as dense or massive as the f fabric of the BZR soil Ck horizon (Figures 23 and 24). From the micromorphological descriptions the discrepancy between the CK horizons of these two soils is only apparent in the additional features column. Here the feature termed "concentric glaebules" occurs with regularity for both DLB soil descriptions but not in the BZR soil description. These circular shaped f members are unique and warrant further discussion.

The f fabric of the DLB soil Ck horizons have zones containing "bubble like" matrans, and exhibit "micropolygonization", a term used by Koniscev et al. (1973) and Fox and Protz (1981). The fabric is characterized by the presence of circular features which maybe are suspended or incorporated within the matrix (Figure 24). These circular shaped features (concentric

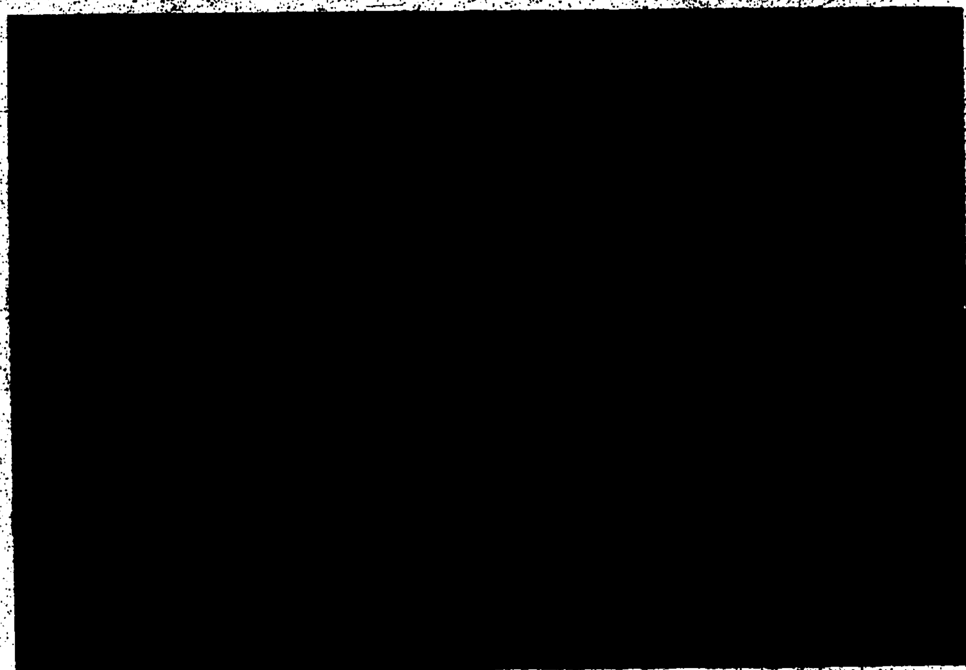

Figure 23. Fragmoidic-porphyric fabric of the Beazer Ck1 horizon. Vughs and skew planes are characteristic features. (plain light, mag. 10X)

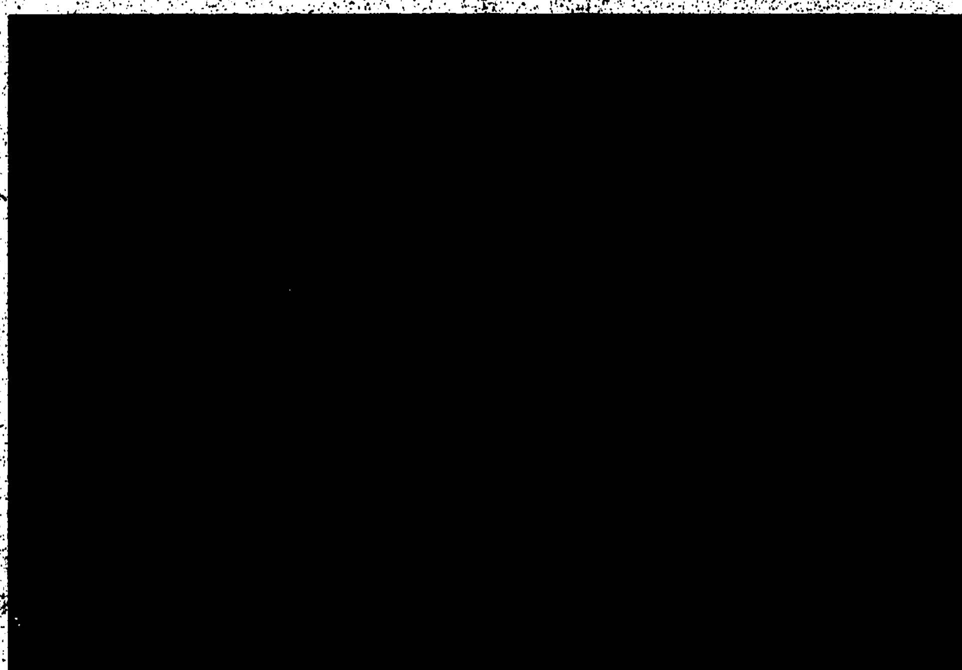

Figure 24. Matri conglomeric-granoidic fabric of the Del Bonita Ck1 horizon.(slide D103) Concentric glaebules and matrans on skeleton grains are characteristic features. (plain light, mag. 10X)

glaebules) consist of matrix material which may or may not be centered around a skeleton grain. In either case they exhibit concentric development. Their internal fabric composition varies considerably from silt to sand sized grains sometimes intermixed with orientated clay. In the best examples the clay as well as sand grains are orientated in a concentric pattern (Figure 25). These features are well developed in the Ck_1 horizons from both DLB soil profiles (#1 and #3) at depths of 30-35 cm below the surface. In the lower horizons these features are less common, but they are still present. With depth they appear less circular and less isolated from the surrounding matrix material. It is as if they are still in the embryonic stages with increasing depths. At a depth of 150 cm, the occurrence of these features decreases to rare.

Numerous researchers (Koniscev et al. 1973; Bunting and Fedoroff 1974; Fox and Protz 1981; Mellor 1986) have described similar features in soils from present day permafrost-affected soils in the Northern hemisphere. Koniscev et al. (1973:215) noted "aggregations of mineral components of an irregular, oval or elongated shape; micropolygons in the form of rings constituted by fragments of the primary minerals" of the sand size fraction. These concentric fabric aggregations were noted to disappear with depth, however the best examples

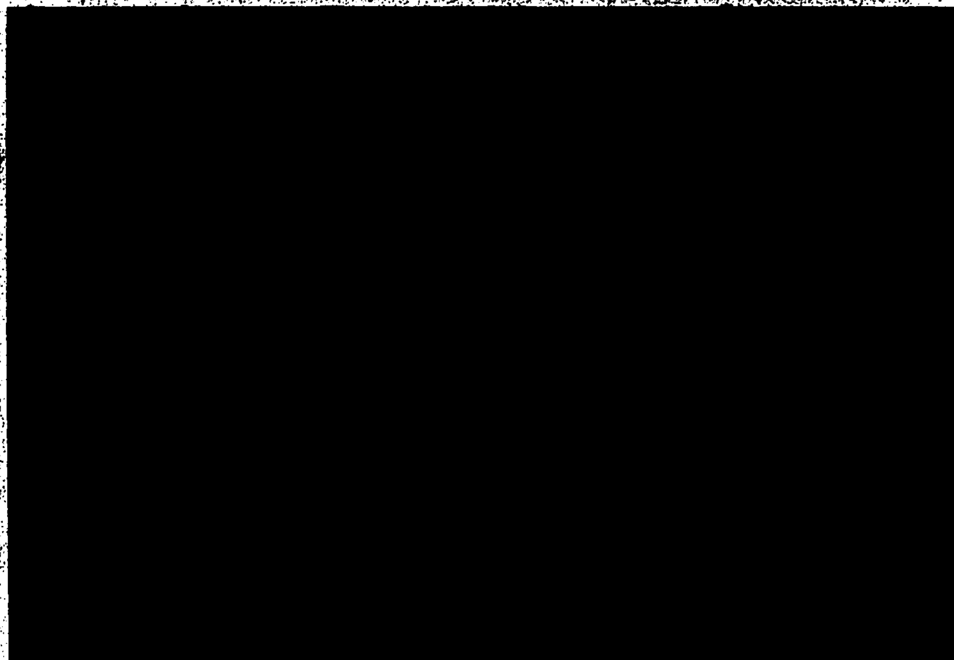

Figure 25. A well developed concentric glaebule within the Del Bonita Ck1 horizon.(slide D103) Note the circular distribution of clay and sand grains throughout the feature. (X-polarized light, mag. 10X)

Figure 26. Suscitic fabric of the Del Bonita Ck2 horizon. (slide D105) The majority of the sand grains and the few skeleton grains are vertically oriented and coated with matrans. (plain light, mag. 10X)

occurred at depths of 1.2-1.7 m (Koniscev et al. 1973). Similar features and their corresponding depth relationship have also been reported by Bunting and Fedoroff (1973) and Fox and Protz (1981) in their studies of soils from the Canadian Arctic. In these latter examples, the best developed features occur within 1 meter of the surface, close to the present day frost table (Fox 1983).

Fox and Protz (1981) in their analysis of Cryosolic soils from the MacKenzie Valley in Northern Canada showed that f members rearrange themselves within the plasma and f matrix. Because of this rearrangement of micromorphological components within the soil they feel that Brewer's (1964) and Brewer and Pawluk's (1975) terminology requires modification. They state that "the distinct fabric patterns cannot be characterized with Brewer's (1964) terminology for related distributions, since it is based on the concept that skeleton grains remain relatively immobile" (Fox and Protz 1981:32). They also indicate that Brewer and Pawluk's (1975) terminology does not account for "the distinct rearrangement of f members within the f matrix" (Fox and Protz 1981:32). Therefore due to these deficiencies of previous systems in describing the unique fabric and features found in present and relict permafrost soils, Fox and Protz (1981) proposed the creation of three new fabric types:

1. orbiculic - Fabric where skeleton grains or f members are distributed in a circular or ellipsoidal/ arrangement.

2. suscitic - Skeleton grains are orientated vertically or nearly vertically. Occasionally skeleton grains or f members have accumulation of fine material at their base.

3. conglomeric- A compound fabric arrangement in which the primary units are discrete f members (coarser units, usually derived from granic fabric distributions) that are randomly set into a ground mass of finer f matrix material (usually but not necessarily less than 10 om size) about which the associated voids delineate a secondary fabric of elliptical to rounded framework members such as fragmic or granoidic.

(Fox and Protz 1981:32)

On the thin sections of the DLB soils, examples of orbiculic fabric arrangements are not present. However examples of suscitic and conglomeric fabrics are present on thin sections from the DLB CK horizons. Using this terminology to modify the original descriptions, the f fabric of the DLB soil Ck_1 thin section (D103) could be described as orthosuscitic // matri-conglomeric // granoidic metafragmoidic. The suscitic component is present in minor amounts, however this term does account for the vertically orientated skeleton grains with accumulations of matrix material plastered on them in different places (Figures 26 and 27). Where the concentric aggregates of fabric are clustered together, conglomeric fabric is best exemplified. With depth, as the concentric features become enclosed within the matrix material, the terms conglomeric porphyric and conglomeric porphyskelic are applicable (Fox and Protz 1981; Fox 1983).

The genesis of conglomeric fabric has been theorized. Koniscev et al. (1973) showed that micropolygonization of fabric may be duplicated by repeated freeze-thaw cycles. Therefore these features are a product of frost action. Within permafrost soils the development of lenticular features is a product of differential sorting. Corte (1963) has shown that particles less than 74 microns move downwards from a freezing front, while coarser particles move upwards. It has also been proposed that these sorted features are the result of partial infilling of melting ice lenses (Harris and Ellis 1980). These lenses of sorted materials have been reported by numerous researchers in cold environment soils (Reiger 1983). Mellor (1986) theorized that these lenticular features may be partially disintegrated and rounded by subsequent cryoturbation processes, producing micropolygonized fabric. This hypothesis is similar to the "snow ball

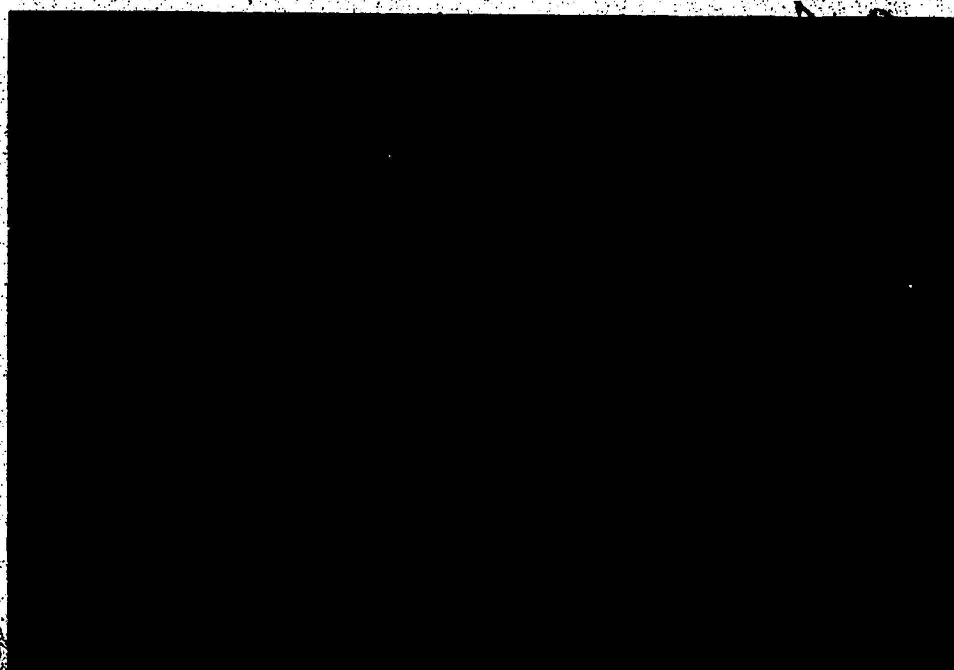

Figure 27. Well developed calcitan at the base of a skeleton grain, from the Del Bonita Ck1 horizon.(slide D103) Note the "layered" appearance of this feature.
(plain light, mag. 10X)

theory" which has been mentioned as a possible way that these features and associated fabric may develop (Dr. C.A. Fox pers. comm. 1987)[8]. The premise of this theory is that during "freeze-up" the advancing freezing front from the surface downward and upward from frost table, does not happen in a regular fashion - parallel to the surfaces. Therefore pockets of unfrozen soil may be moved or squeezed because of heaving as the surrounding material freezes. Turbulence or circular movement within these unfrozen pockets may result in the formation of these concentric glaebules. Credence to this hypothesis is evident from present day examples because these features are present in association with dynamic frost tables, i.e. the active layer of permafrost environments (Fox and Protz 1981). Also this corresponds with Koniscev et al.'s (1973) previously mentioned laboratory findings and the development of cryoturbated papules in the paleoargillic zone at a depth of 50-90 cm in soils in England (Bullock and Murphy 1979).

Features associated with suscitic fabric have been widely documented (Washburn 1973; French 1976; Harris 1981; Reiger 1983). The genesis of vertically orientated coarse

fragments within the soil matrix has been described in detail by Corte (1963), and this theory is widely accepted

[8] Dr. C.A. Fox, Land Resource Research Centre, Agriculture Canada, Ottawa, Ontario.

(Reiger 1983). The development of fine matrix cappings and pendants on coarse fragments is more nebulous.

Pendant formation on the sides and bottom of coarse fragments appears analogous to "small scale" ice wedge formation. Within the matran, distinct zones are recognizable (Figure 27). Each new "lining" or concentration of material represents a melting sequence. Development of segregation ice under and around coarse fragments in fine textured soils of cold environments has been noticed by numerous researchers, as reported by Harris and Ellis (1980). Upon melting of this ice lense, finer matrix material seeps in around the bottom of the stone. This process appears to be repeatable as visible by the different appearing linings of material within the pendant. This cryogenic based theory is identical to the frost push hypothesis for stone movement proposed by Corte (1963).

The occurrence and development of fine grained cappings on coarse fragments is disputed in the literature. Although cappings are common features of cold environment soils, as noted by Mellor (1986), similar entities have been discovered in soils of Australia, so it appears agents other than frost action may be involved (Brewer and Pawluk 1975). The most accepted method of cap development consists of infilling of fine material upon disintegration of the ice sheath around coarse fragments (Harris and Ellis 1980; Mellor 1986). Reiger (1983) proposes that as stones are uplifted within the soil, fine particles of the matrix are "plastered" on to the upper surface of the stone, thus creating surface cap. There is also a possibility that wetting and drying processes may result in these features, however this procedure appears to be less significant, using the recorded occurrences as a criteria. Whichever process is involved, surface cap features on coarse fragments are primarily confined to relict and present day cold environment, fine textured soils where segregation ice development is favoured (Harris 1981).

From the analysis of the micromorphological descriptions, the following conclusions may be obtained. First, the soils on the Del Bonita Plateau exhibit features within the CK horizons which are a product of periglacial activity. The BZR soils do not contain similar features. Second, the frost table on the Plateau varied in depth most frequently between 30-150 cm.

CHAPTER VI. CONCLUSIONS

This study compared an Orthic Black Chernozemic soil developed on parent material of unknown origin, the Del Bonita (DLB) soil, with a similar soil developed on Laurentide till, the Beazer (BZR) soil. The characteristics of these soils were compared and assessed to deduce the origin of the DLB parent material.

The Del Bonita soil parent material is loess, although it must be recognized that this material has been substantially affected by cryogenic or periglacial processes. The crucial facts in coming to this conclusion are:

1. The clay mineralogy of both soil materials is identical, as indicated by the X-ray diffraction patterns. This finding eliminated the possibility that the surficial material on the Plateau originated from the Rocky Mountains. The possibility that the material was fluvially derived from the Laurentide drift was eliminated since the overall Plateau slope reflects the mountain origin of the underlying Flaxville quartzitic gravels (0.4% slope down to the North east).

2. The $CaCO_3$ equivalent values of the DLB soils were consistently greater than the BZR soils. The X-ray diffraction patterns indicated that the proportion of

calcite to dolomite for the DLB soils were distinctly different from the BZR soils. Dolomite exceeded calcite in the BZR CK_2 horizons. Because the calcite peaks of the DLB CK_2 horizons were consistently greater than dolomite, it indicated that the loess source material had been altered, probably by fluvial action. Since the DLB surface material contains large amounts of $CaCO_3$, the loess must have originated from a carbonate rich area.

A modern day analogy explains the existence of such a carbonate rich source area. In the Crowsnest Pass area soils developed on the relict floodplains of the Oldman and Crowsnest rivers are extremely calcareous ($CaCO_3$ equivalent values greater than 30%). Alternatively the soils developed in till on the valley side slope positions have considerably lower $CaCO_3$ equivalent values.

This analogy indicates that a carbonate rich area may have been present due to glaciofluvial activity. A fluvial source area associated with a Laurentide ice sheet would be a rich source of calcuim carbonate minerals. Therefore, the carbonate concentrations, especially calcite, may be greater within the Del Bonita surficial material than the Laurentide till, if the loess originated from a glaciofluvial source area.

3. The light mineral fraction of the DLB soils was consistently greater than the BZR soils. Also the

proportion of quartz to feldspar within the DLB soil light mineral fraction was greater than the equivalent BZR soil. These light mineral characteristics are consistent with the findings for other loess deposits (Price et al. 1975; Catto 1981).

4. The fact that the Plateau experienced periglacial conditions as evidenced by the presence of ice wedge casts etc. was confirmed from the micromorphology of the CK horizons of the DLB soils. The BZR and DLB soils are similar with respect to profile characteristics such as depth of solum and texture. However, the existence of conglomeric and suscitic fabric only within the DLB soil CK horizons means that the Plateau has experienced periglacial conditions. The effect of cryogenic processes has not been altered by pedogenesis. This confirms the fact already deduced from the field observations, that the loess overlying the Flaxville gravels was present at some time when the Plateau was subjected to periglacial conditions.

The other analyses did not greatly contribute to this deduction process for a variety of reasons:

The particle size analyses were not conclusive because the cumulative curves of the DLB samples reflected the effects of cryoturbation. Typically loess deposits are well sorted. Due to cryogenic processes sands and gravels from the Flaxville gravels were incorporated within the overlying surface material. Poorly sorted loess deposits are the product of this process. Therefore, the parent materials of the DLB and BZR soils are both poorly sorted according to the sorting index classification proposed by Folk and Ward (1957).

The scanning electron microscopic surface morphology of quartz grains was not diagnostic of loess deposits because the material on the Plateau was relatively close to the source area. Typical loess quartz grains are the result of saltation over thousands of kilometers in desert situations (Krinsley and Doornkamp 1973). The loess material of the DLB soil is derived from Laurentide drift which may be as close a 10 km away. Because of the source material proximity, typical surface morphological features associated with loess quartz grains are not present in the DLB material in significant amounts. Only broad general differences between the sand grains from the BZR and DLB material were identifiable and these were not conclusive.

The following scenario is a simplified model accounting for the deposition of the loess on the Del Bonita Plateau. This is based on Hardcastle's (1890) hypothesis regarding glaciers and loess as described by Smalley (1983).

During the Wisconsin glacial period there were numerous glacial advances and retreats of continental ice sheets across southwestern Alberta. After each advance, as the climate ameliorated, glaciers retreated exposing freshly ground glacial detritus. With the improving climate, glacial meltwater containing dissolved minerals would have been abundant. This runoff water would flow away from the glacier on to the exposed moraine creating a floodplain of sludge containing "rock meal" (Hardcastle 1890 as reported by Smalley 1983) and precipitated carbonates. The materials on the floodplain dried and were exposed to katabatic winds blowing from the nearby glaciers. These northerly winds selectively picked up materials of particular grain sizes and carried it to the south. At breaks in elevation, changes in surface roughness and as wind speed decreased, the material was deposited as loess. After many hundreds of years the resulting loess deposits were significant - up to 2 m thick on the Del Bonita Plateau.

With a readvancing ice sheet, the Plateau was subjected to an environment dominated by cryogenic processes. The examples of the periglacial features such as ice wedge casts, up-freezing of stones and involutions are proof that the loess existed above the Flaxville gravels when the area was exposed to a permafrost environment. Ground ice must have been abundant in this moist periglacial environment. Upon the onset of another minor altithermal period, the Plateau surficial materials were modified by processes associated with the melting of the permafrost. Solifluction and altiplanation are two processes that resulted in the present day distribution of loess on the Plateau. For this reason the term cryoturbated - fluvial - eolian, proposed by Dr. I. Shetsen (pers. comm. 1982)[9] does not appear incorrect in describing the upper surficial material of the Del Bonita Plateau.

[9] Dr. I. Shetsen, Alberta Research Council, Edmonton, Ablerta.

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Appendix

Detailed Soil Descriptions and Results of Routine Analysis

Appendix 1

Series: Beazer (BZR)
Site No.: 1
Location: NE 21 Twp 1 Rge 23 W4th
Slope and Aspect: 12%, northeast
Sample site postion: upper slope
Parent material: Laurentide till
Vegetation: Rough fescue
Elevation: 1410 metres

Sample No.	Horizon	Depth	Description
BZR 101	Ah	0-17 cm	Very dark grayish brown (10YR 3/2 d);loam; moderated to strong, medium to coarse, prismatic breaking to moderate to strong, medium angular blocky; gradual irregular boundary; 16 - 22 cm thick.
BZR 102	Bm	17-36	Dark brown (10YR 4/3 d); clay loam; moderate to strong, coarse, prismatic, breaking to moderate, medium subangular blocky; clear wavy boundary; 14 - 20 cm thick.
BZR 103	Ck_1	36-74	Light brownish gray (10YR 6/2 d) loam; moderate,medium to coarse, prismatic breaking to weak, fine to medium, subangular blocky; diffuse wavy boundary; 38 - 44 cm thick.
BZR 104	Ck_2	74-100	Light brownish gray (2.5Y 6/2 d) clay loam; massive.

Results of Routine Physical and Chemical Analysis
BZR Site Number 1

Sample No.	% sand	% silt	% clay	textural class*	% coarse fragment (field estimate)
BZR 101	36	38	26	loam	5
BZR 102	39	32	29	clay loam	5
BZR 103	37	36	27	loam	10
BZR 104	22	49	29	clay loam	10

Sample No.	pH H2O	pH CaCl2	CaCO3 equiv. %	%C	Exchangeable cations cmol(+)/kg Na	K	Ca	Mg	TEC
BZR 101	7.3	7.1		4.20	0.01	1.1	24.0	4.7	33.4
BZR 102	7.5	7.0		1.66	0.01	0.8	15.6	5.7	25.5
BZR 103	7.9	7.5	14.42						
BZR 104	8.1	7.7	15.43						

* textural class based on hydrometer method with no pretreatments

Series: Beazer (BZR)
Site No. [illegible]
Location: [illegible] 28 Twp 2 Rge 22 W4th
Slope and aspect: 6%, southeast
Sample site position: upper slope
Parent material: Laurentide till
Vegetation: Rough fescue
Elevation: 1030 meters

Sample No.	Horizon	Depth	Description
BZR 201	Ah	0-8 cm	Very dark brown (10YR 2/2 d); Loam; moderate, medium, subangular blocky breaking to moderate, medium, granular; clear irregular boundary; 4 - 15 cm thick.
BZR 202	Bm	8-36	Dark grayish brown (10YR 4/2 d); loam; moderate, medium, prismatic breaking to moderate fine to medium subangular blocky; clear, wavy boundary; 19 - 28 cm thick.
BZR 203	Ck_1	36-80	Light brownish gray (2.5YR 6/2 d); loam; massive; diffuse, wavy boundary; 44 - 60 cm thick.
BZR 204	Ck_2	80-100	Brown (10YR 5/3 d); clay loam; massive.

Results of Routine Physical and Chemical Analysis
BZR Site Number 2

Sample No.	% sand	% silt	% clay	textural class*	% coarse fragment (field estimate)
BZR 201	44	32	24	loam	5
BZR 202	50	28	22	loam	8
BZR 203	43	30	27	loam	10
BZR 204	43	30	27	loam	10

	pH	pH	CaCO3		Exchangeable cations cmol(+)/kg				
Sample No.	H2O	CaCl2	equiv. %	%C	Na	K	Ca	Mg	TEC
BZR 201	6.9	6.5		6.71	0.01	1.9	21.2	6.0	35.1
BZR 202	6.5	6.0		1.32	0.02	0.5	11.9	4.6	20.8
BZR 203	7.8	7.4	15.50						
BZR 204	8.2	7.9	11.75						

* textural class based on hydrometer method with no pretreatments

Series: Beazer (BZR)
Site No.: 3
Location: SE 10 Twp 2 Rge 23 W4th
Slope and Aspect: 13, northeast
Sample site position: middle
Parent material: Laurentide till
Vegetation: Rough fescue
Elevation: 1387 meters

Sample No.	Horizon	Depth	Description
BZR 301	Ah	0-12 cm	Black (10YR 2/1 d); clay loam; moderate, medium, subangular blocky breaking to weak, fine granular; clear, irregular boundary; 9 - 20 cm thick.
BZR 302	Bm_1	12-40	Dark brown (10YR 3/3 d); clay loam; strong, coarse prismatic breaking to strong, fine to medium subangular blocky, gradual wavy boundary; 20 - 28 cm thick.
BZR 303	Bm_2	40-57	Dark brown to brown (10YR 4/3 d) clay loam; moderate to strong, medium, prismatic, breaking to strong, fine to medium, subangular blocky; clear wavy boundary; 15 - 20 cm thick.
BZR 304	Ck	57-100	Grayish brown (2.5YR 5/2 d); clay loam; massive.

Results of Routine Physical and Chemical Analysis
BZR Site Number 3

Sample No.	% sand	% silt	% clay	textural class*	% coarse fragment (field estimate)
BZR 301	28	38	34	clay loam	5
BZR 302	31	32	37	clay loam	5
BZR 303	33	31	36	clay loam	5
BZR 304	31	33	36	clay loam	10
BZR 305	30	35	35	clay loam	10

Sample No.	pH H2O	pH CaCl2	CaCO3 equiv. %	%C	Exchangeable cations cmol(+)/kg Na	K	Ca	Mg	TEC
BZR 301	6.6	6.1		9.15	0.07	1.6	26.9	7.0	47.2
BZR 302	6.1	5.8		2.00	0.03	0.0	19.0	7.7	34.9
BZR 303	6.7	6.3		1.15					
BZR 304	7.8	7.4	11.41						
BZR 305	8.1	7.7	12.						

* textural class based on hydrometer method with no pretreatments

Series: Del Bonita (DLB)
Site No.: 1
Location: WC 21 Twp 1 Rge 21 W4th
(Shanks Lake gravel pit)
Slope and Aspect: 0%, N/A
Sample site position: N/A
Parent material: eolian
Vegetation: Rough fescue
Elevation: 1295 meters

Sample No.	Horizon	Depth	Description
DLB 101	Ah	0-9 cm	Very dark brown (10YR 2/2 d); clay loam; weak fine subangular blocky breaking to structureless, single grain; gradual wavy boundary; 7 - 15 cm thick.
DLB 102	[illegible]	9-35	Dark yellowish brown (10YR 4/4 d); clay loam; moderate, medium, prismatic breaking to moderate to strong, medium, subangular blocky clear wavy boundary; 20 - 26 cm thick.
DLB 103	Ck_1	35-80	Very pale brown (10YR 7/3 d); clay loam; massive; diffuse wavy boundary; 35 - 50 cm thick.
DLB 104 @80cm DLB 105 @100cm	Ck_2	80-100	Very pale brown (10YR 7/4 d) loam; massive.