QUATERNARY GEOLOGY AND STRATIGRAPHY

OF THE

ASSINIBOINE FAN DELTA AREA, SOUTHWESTERN MANITOBA

A THESIS

PRESENTED TO

THE FACULTY OF GRADUATE STUDIES

UNIVERSITY OF MANITOBA

IN PARTIAL FULFILMENT

OF THE REQUIREMENTS FOR THE DEGREE

OF MASTER OF SCIENCE

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April 1993

ISBN 0-315-81783-6

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BY

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A Thesis submitted to the Faculty of Graduate Studies of the University of Manitoba in partial fulfillment of the requirements for the degree of

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ABSTRACT

The Assiniboine fan delta covers an area of 6400 km^2, and consists of a lower region separated from a higher region to the west by an escarpment known as the Campbell beach escarpment. The topographically higher part of the fan delta is the target of this thesis.

The higher region consists of 5 major depositional settings: a braided river plain leading to the fan delta, a Gilbert-type delta, underflow fan deposits, subaerial debris flow deposits, and pro-delta deposits. The braided river plain deposits are composed of tabular cross bedded and trough cross bedded gravelly sand. The Gilbert delta, which is separated from the braided river plain to the west by an escarpment, is dominated by gravel foreset beds. The underflow fan deposits, which lie east of these foresets, are dominated by massive sand and rippled sand. The pro-delta deposits consist of silty clay and clayey silt. The subaerial debris flow deposits occur in the eastern braided river plain and in the proximal fan delta near the modern Assiniboine valley.

The Assiniboine River eroded a broad and shallow channel across the northern fan delta, and a broad and deeper channel across its southern side. Coarse channel fill sediments in a river terrace north of Glenboro were dated at 10,600 $\pm$150 BP (GSC 383).

During late Wisconsinan time, the Red River Lobe advanced southwestward across the Assiniboine valley to the Darlingford and Alexander moraines. The Alexander moraine, which is the northwestern extension of the Darlingford moraine, became an ice barrier to water flowing from the west and formed Lake Hind. Subsequently Lake Brandon formed to the east of the ice barrier when the Red River Lobe retreated northeastward. At first, Lake Hind drained to Lake Agassiz through the Pembina Spillway along a southern route, bypassing the Assiniboine fan delta. When the ice barrier failed along the Assiniboine bedrock valley, water from Lake Hind and the Assiniboine River flooded into Lake Agassiz near Brandon, starting deposition of the Assiniboine fan delta.

The earliest episode of flooding was catastrophic because sheet-like poorly sorted gravels and boulders were deposited over an area of 350 km^2 at the base of the proximal fan delta sequence. After the water level fell to an elevation about 380 m (the middle Herman beach level), a Gilbert-type delta was deposited by a series of floods. Each of the floods probably deposited multiple foreset beds. Subsequent falling of lake level caused the excavation of the broad and shallow channel (called the Chater-Douglas Station channel) across the northern fan delta. Meltwater flowed through this channel to the middle fan delta area, where it became a density underflow into Lake Agassiz, depositing sediments on to the lake floor to form the underflow fan delta.

The Assiniboine River shifted to the southern fan delta before 10,600 BP, and before the deposition of massive sandy gravels in the proximal fan delta. As the level of Lake Agassiz rose during the Emerson phase (9900-9500 BP), fluvial and fluvio-lacustrine sediments were deposited in the Assiniboine River valley. These sediments were later incised as the lake fell during the Nipigon phase (9500-8500 BP), leaving many paired and unpaired terraces at an elevation between 330 and 350 m.

ACKNOWLEDGMENTS

I would like to express my sincere thanks to Mr. B. Wolfe, MSc student, Department of Geological Sciences, University of Manitoba, for his help in the field and for his advice in the preparation of this thesis. Special thanks are directed to my advisor, Dr. J.T. Teller, Department of Geological Sciences, University of Manitoba, for his financial support through his N.S.E.R.C. Operating Grant, and for advice and help in the field and guidance in the preparation of this document. I am grateful to Dr. Susan Pullan of Geological Survey Canada for running several kilometers of seismic reflection survey in my study area. Mr. Harry Jol and Dr. Derald G. Smith (Dept. of Geography, University of Calgary) kindly provided equipment and expertise for a ground penetrating radar survey in 1990. The Freshwater Institute (Winnipeg) provide a canoe for access to cuts along the Assiniboine River. Samples and logs collected by P.F.R.A. from the delta area were made available for this study.

Chapter 1

INTRODUCTION

Location of the Study Area

The Assiniboine fan delta is located in southern Manitoba between Brandon and Portage La Prairie (Fig. 1-1). It is confined to the north by Riding Mountain and to the south by Tiger Hills upland. The thesis study area is mostly concentrated in the upper and middle part of the fan delta between Brandon and the Campbell Beach (Fig. 1-1). In addition, the Assiniboine Valley has been studied from Virden to Brandon to help understand the nature of the flow to the fan delta. The total study area covers about 4400 km^2 and is bounded by latitudes 49° 35' and 50° 09, and longitude 98° 45' and 100° 30'. The area is covered by 6 1:50,000 topographic map sheets, Brandon (62 G/13), Wawanesa (62 G/12), Carberry (62 G/14), Glenboro (62 G/11), Treherne (62 G/10), and Alexander (62 F/16). The study area is included on the Surficial Geology Maps of Manitoba(1:250,000), Brandon sheet (62 G) and the eastern part of Virden sheet(62 F).

Purpose Of the Study

The purpose of this thesis is: 1. to study the facies and depositional environments of facies in the Assiniboine fan delta. 2. to reconstruct the paleo-hydrological conditions necessary to account for both the erosional and depositional features. 3. to integrate the Assiniboine fan delta history in to the history of Lake Agassiz.

Previous Studies

The Assiniboine delta was first described by Upham (1895) and later was mapped by Johnston (1934) and Elson (1956, 1960, 1967). Elson(1956) was perhaps

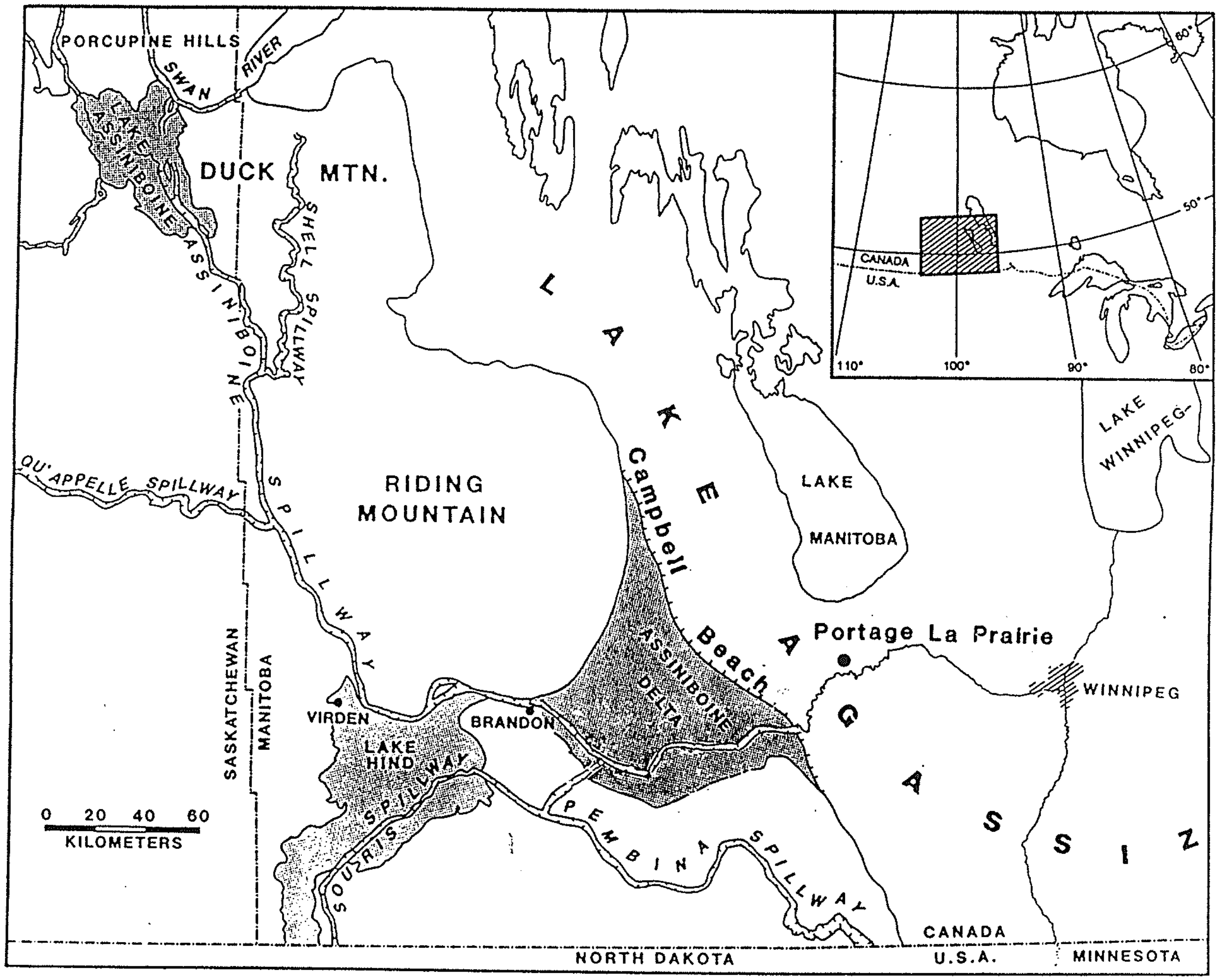


Fig. 1-1 Location of the Assiniboine fan delta, showing spillways and lakes that might have contributed to the construction of the Assiniboine fan delta.

the first to try to associate the development of the Assiniboine valley to the history of Lake Agassiz and related some of the Assiniboine River terraces to beaches of Lake Agassiz. The Manitoba Water Resource Branch and P.F.R.A. (1986) have drilled hundreds of wells in the Assiniboine delta area, their stratigraphic records and lab analyses are on file with the Manitoba Water Resources Branch. There are dozens of reports and maps on water availability in the Assiniboine delta, such as the Ground Water Availability Study in Brandon Map Area (Halstead, 1959), Ground Water Availability Study in the South Cypress Area (Manitoba Water Resources Branch, 1977), and Ground Water Resources in the Cypress Planning District (Manitoba Water Resources Division, 1978). Underwood McLellan Associate Limited (UMA Group, 1977) prepared a report of Sand and Gravel Resources of Assiniboine delta. Klassen(1972, 1975, 1983) studied the Assiniboine River fluvial deposits from its headwaters at glacial Lake Assiniboine to Portage La Prairie (Fig.1-1). He speculated about the age of the fan delta and the correlation between the delta and the fluvial fill of the Assiniboine channel. The importance of the Assiniboine fan delta study has been highlighted by Teller and others (1990), who stated that the Assiniboine fan delta might have been formed by more than one catastrophic flood during the late Wisconsinan. These floods, in turn, might have played a role in catastrophic flooding from Lake Agassiz to the Great Lakes through the eastern outlets (Teller and others 1990).

Methods of Study

Field work

The Assiniboine fan delta study is based on examination of field exposures, new sediment cores, well log studies, and geophysical data studies. Most of the field work was conducted in the upper and middle part of the fan delta during the summers of 1989 and 1990. 32 gravel pits on the upper fan delta area were examined and samples were collected for sieve analyses. In addition, 31 river cuts were studied along the Assiniboine River from east of Brandon to the Spruce Wood Provincial Park in the summer of 1989.

In order to obtain continuous profiles of major sedimentary units within the fan delta, shallow seismic surveys were carried out by the Geological Survey of Canada in the summer of 1989. Equipment used in the shallow seismic survey included 24 geophones and a computer control and storage system with in-hole shotgun as the explosive sources (Fig. 1-2). During the survey, 3 m offsets between the source and the first receiver and 3 m offsets between receivers were used. 11 lines which had deep well control were shot. However, low ground water levels and thick dry eolian sand near the surface did not result in good resolution of the seismic data, and only major lithological contacts were identified, not specific sedimentary structures.

A ground penetrating radar survey was conducted in 1990 to further define sedimentary units and internal bedding of the fan delta. Equipment used in the ground penetrating radar survey included a transmiting unit, a receiving unit, a control unit, and a display and storage unit (Fig. 1-3). The principle of the ground penetration radar technique is similar to that of the shallow seismic survey, which is that the lithological (grain size) interfaces within the delta have pronounced changes in the acoustic properties. The radar produces a short pulse of high frequency electromagnetic energy, which is transmitted to the subsurface by the transmitting unit. Changes in acoustical properties near bedding interfaces will cause a portion of the energy to be reflected back to the receiving unit (Davis and Annan, 1989). This signal is then amplified and digitized before being stored in the computer. The higher the clay and silt content near the surface, the shallower the radar signal can

Fig. 1-2 In-hole shot gun and geophones used in the seismic survey. Left: shot gun; right: geophone

Fig. 1-3 Equipment used in the radar survey. Upper: Computer control system and storage system; Lower: two 200 megaHz Antennas.

reach. The best areas to use ground penetration radar is a dry surface with a low water table and low silt and clay content. Three kinds of antennae with frequencies of 200 mHz, 100 mHz, and 50 mHz were used in the survey. By using the 200 mHz antennae, detailed sedimentary textures at shallow depth were displayed on output profiles, but deeper reflections were lost. The 50 mHz antennae was used to penetrate deeper strata and identify major sedimentary interfaces at the expense of losing detail. The 100 mHz antennae commonly provided sufficient detail and depth in the Assiniboine fan delta area.

During the survey, the antennae were spaced one meter apart and pulses of electromagnetic energy were shot at 0.5 m or 1 m intervals. Sediments in the Assiniboine fan delta are mostly moderately to poorly sorted and commonly contain some silt and clay, so attenuation was a problem. Therefore the best reflections on the radar profiles were in the depth range of 2 m to 6 m and occasionally to a depth of 10 m.

Two wells were drilled in the Assiniboine fan delta (90AD-1 and 90AD-2) using the rotasonic drilling method to examine vertical sequences within the fan delta. This drilling system used high frequency resonance, produced by two synchronized rollers, imparted down the drill stem to the rotating cutting bit to facilitate rapid penetration (Fig. 1-4). Cores were taken continuously during drilling in the hollow stem core barrel which lay inside the hollow stem of the drill rod. The advantage of this drilling system is that it can drill fast and deep, can penetrate hard material (eg. boulders and bedrock), and can produce a continuous suite of core samples. The main disadvantage is that many sedimentary structures are destroyed by the vibration technique, especially in sand and coarse silt sized units. Two wells were drilled: one 41 m deep (core 90AD-1) located northeast of Shilo (NE$\frac{1}{4}$ of SE$\frac{1}{4}$, Sec.16, Tp10

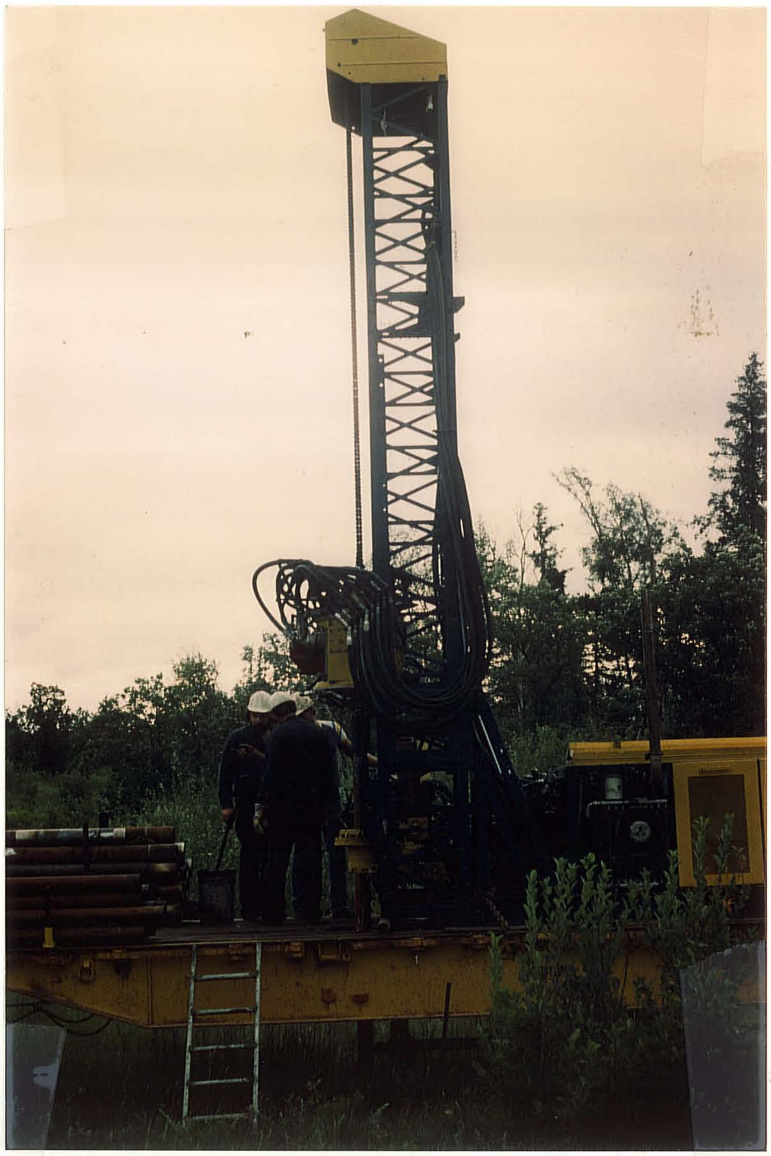


Fig. 1-4 Sonic drilling rig.

R16W), and another 71 m deep (core 90AD-2) located south of Carberry at NE$\frac{1}{4}$ of SE$\frac{1}{4}$, Sec.10, Tp10, R14W.

Lab analyses

Subsamples from rotasonic cores 90AD-2 and 90AD-1 were taken at intervals of 0.5 m for grain size analyses. Some samples from selected field exposures also were analyzed to aid the interpretation of depositional environments. These samples were dried in an oven at 200°F (93°C) and were then disaggregated by hand. Only samples containing at least 75 percent sand were sieved. Dry samples were sieved into 0.5 phi intervals using a RoTap shaker. The weight percentages of samples in each 0.5 phi interval have been used to construct frequency curves and cumulative curves. The grain size diameters (phi) of the sample at the weight percentages of 5, 16, 25, 50, 75, 84 and 95 from the cumulative curve were used to calculate the mean size(Mz), standard deviation(σ_I), skewness(Sk_I), and kurtosis(K_G) by using Folk's(1966) equations:

$$Mz = \frac{\phi16+\phi50+\phi84}{3}$$

$$\sigma_I = \frac{\phi84-\phi16}{4} + \frac{\phi95-\phi5}{6.6}$$

$$Sk_I = \frac{\phi16+\phi84-2\phi50}{2(\phi84-\phi16)} + \frac{\phi5+\phi95-2\phi50}{2(\phi95-\phi5)}$$

$$K_G = \frac{\phi95-\phi5}{2.44(\phi75-\phi25)}$$

These parameters have been used to plot graphs showing the changes of mean size, standard deviation, and skewness with depth. These in turn, have been used to help interpret the depositional environments based on the studies of Friedman (1961) and Moss (1962) (Table 1).

Table 1. **Grain Size parameters and Environments**

Sorting Interval	Degree of Sorting	Depositional Environment
When mean is	between1.0 – 2.0ϕ	
$\sigma \leq 0.35$	Very well sorted	Most barrier bars and lake dune sands, many beach sands, many marine sands above wave base.
0.35 – 0.5	Well sorted	Most beach sands($Sk \leq 0$) , most marine sands above wave base, many lagoonal sands, many inland dune sands, Some river sands
0.5 – 0.8	Moderately well sorted	Beach sands($Sk \leq 0$), River or delta sands, many lagoonal sands, most offshore sands below wave base, most inland dune sands.
0.8 – 1.4	Moderatly sorted	Many river sands($Sk_{max} = 1.4$), many restricted lagoonal sands, some offshore sands below wave base and many glaciofluvial sands.
1.4 – 2.0	Poorly sorted	Many glaciofluvial sands.
2.0 – 2.6	Very poorly	Many glaciofluvial sands.
> 2.6	Extremely poor	Some glacio-fluvial sands, most tills.

When mean is	less than 1.0ϕ	
0.5 – 0.8	Moderately well	Many beach sands.
0.8 – 1.4	Moderate	Most beach sands, some river or delta sands.
1.4 – 2.0	Poor	Some river sands, many glaciofluvial sands.
2.0 – 2.6	Very poor	many glaciofluvial sands.
> 2.6	Extremely poor	Some glaciofluvial sands, Many tills

After Friedman(1961)

Geological Setting

Bedrock geology

Bedrock topography is an important factor that contributes to the thickness and distribution of till and fan delta deposits. The composition of bedrock commonly affects the composition and texture of till and outwash deposits.

The bedrock geology in southern Manitoba has been summarized by Klassen (1970), Teller (1976), and Teller and Bluemle (1983). The bedrock can be subdivided into three major groups: 1. Precambrian, which occupies a wide zone from northwestern Manitoba to southeastern Manitoba; 2. Paleozoic rocks, which are exposed in a NW-SE belt through the Lake Winnipeg and Lake Manitoba basins; 3. Mesozoic rocks, which outcrop west of Lake Manitoba westward into Saskatchewan (Fig. 1-5). The subsurfaces of Mesozoic and Paleozoic rocks dip to the southwest and wedge out to the northeast (Teller and Bluemle, 1983,) (Fig. 1-5).

Precambrian Rock: Precambrian rocks in Manitoba have been divided into two geological provinces: The Churchill province of northwestern Manitoba and the Superior province of eastern Manitoba (Fig. 1-5). The Archean rocks of the Superior province are composed of granite, greenstone, schist, and gneiss rocks. The Churchill province includes two types of rocks: the older Archean and Proterozoic metavolcanic granulites and granitoids and the younger Proterozoic metasedimentary supracrustal rocks (Manitoba Mineral Resources Division, 1979).

Paleozoic Rock Paleozoic rock consists of carbonate, mainly dolomite, dolomitic limestone intercalated with some shale, and occasionally sandstone, conglomerate, and anhydrite (McCabe, 1971). Cambrian rock is absent from this region, so Ordovician rocks unconformably overlie Precambrian rocks. The lithology of the lower

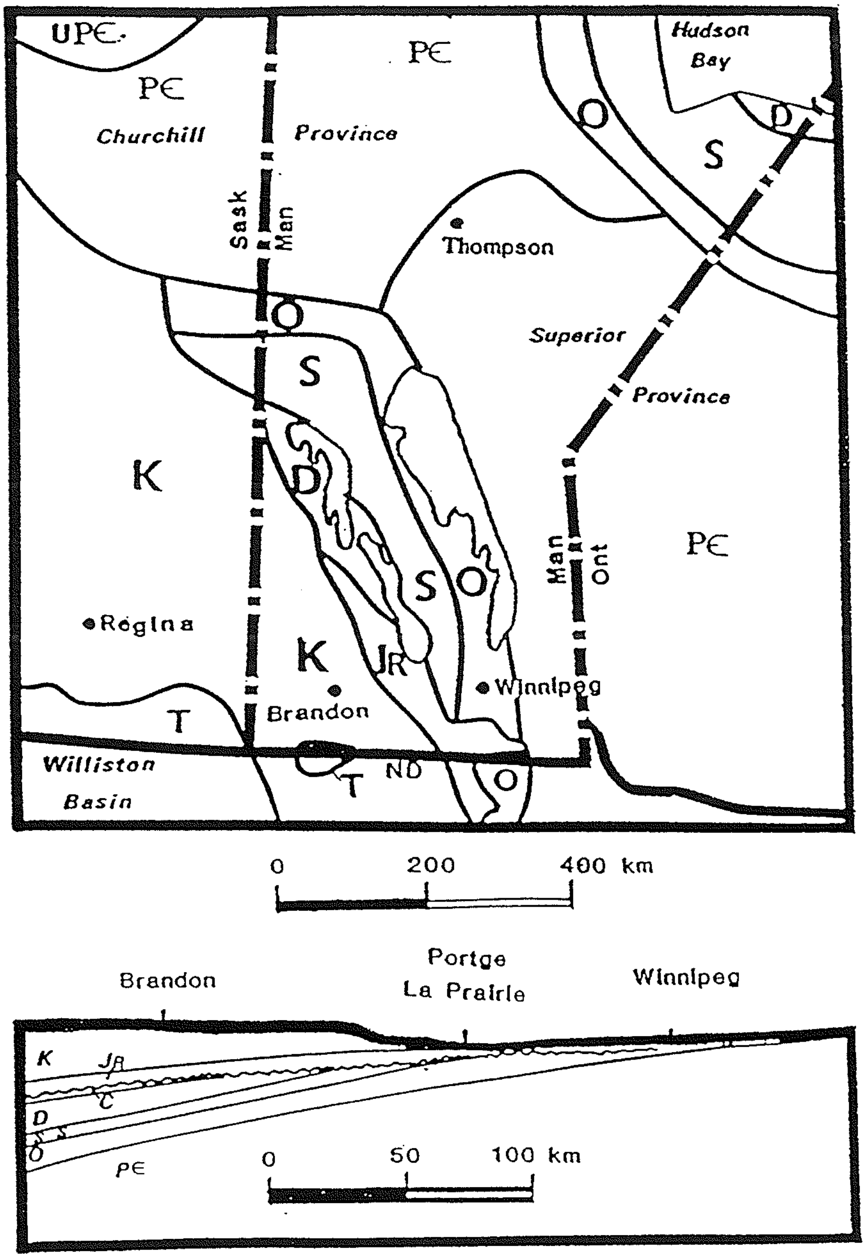


Fig. 1-5 Bedrock geology of Manitoba with W-E cross section through Brandon and Winnipeg, Manitoba (after Teller and Bluemle, 1983).

Ordovician is composed of sandstones, while the upper Ordovician and lower Devonian are dominated by dolomite (Teller and Bluemle, 1983; McCabe, 1971). The Siluriun, upper Devonian, and the Mississippian rocks are marked by the co-existance of anhydrite and limestone-dolomite.

Mesozoic Rock Mesozoic rocks are mainly composed of shales with some sandstone and thin limestones, and are overlain directly by Quaternary sediments in southwestern Manitoba and Saskatchewan. These rocks have been discussed by Wickenden (1945), Halstead (1959) and Teller and Bluemle (1983). The oldest Mesozoic rocks in this area are Jurassic in age, which consist of red shale, sandstone, and gypsum (Halstead, 1959). Cretaceous rocks in southern Manitoba are composed of shale, calcareous shale, sandstone, and lower rank coal.

Bedrock topography in the study area

The bedrock topography in southern Manitoba has been mapped by Teller (1976). On his map, the elevation of the bedrock surface is shown to be lower under the Assiniboine fan delta and to rise to the west toward the Souris Basin, northwest toward the Assiniboine River Plain, and south to the Tiger Hills Upland (Fig. 1-6). The bedrock surface beneath the Assiniboine fan delta clearly shows a bedrock channel from Brandon extending eastward to Carberry and Austin where it joins a northwest-southeast tending bedrock valley (Fig. 1-6). A smaller bedrock valley lies along the present Souris River channel, extending from the Pembina Spillway northward to north of Wawanesa (Fig. 1-6).

Physiographic division

Before discussing the Quaternary geology, it is necessary to introduce the physiographic divisions in southwestern Manitoba. In southwestern Manitoba, the Camp-

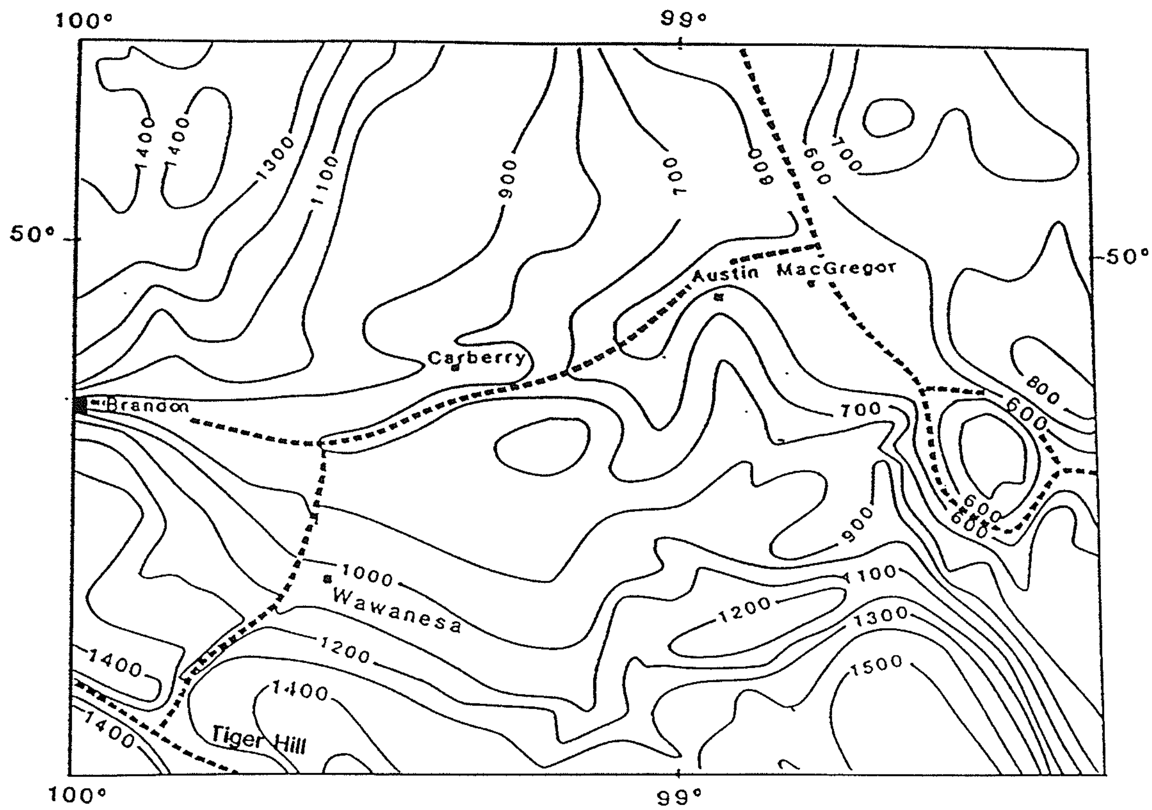


Fig. 1-6 Bedrock topography of the study area. Dashed lines show the bedrock valley from Brandon to Austin and the from Tiger Hills northward to Wawanesa (after Teller, 1976). Elevation in feet.

bell beach forms the boundary between two major physiographic divisions, the Manitoba Plain and the Saskatchewan Plain (Klassen,1979; Conley 1986). Most of the Assiniboine fan delta is in the Assiniboine Delta Plain physiographic sub-province (which is part of the Saskatchewan Plain) although part of the distal fan delta is located in the Manitoba Plain. These and other physiographic divisions are shown in Figure 1-7.

Quaternary deposits

Quaternary deposits are the surface materials that blanket or veneer the bedrock, and form the local landforms. Bedrock controls the broad landforms (Klassen, 1989). Most of the Quaternary deposits in southwestern Manitoba, including till, glacial lacustrine deposits, and glacial fluvial deposits, were deposited duing Pleistocene glaciations. Only a small portion of the Quaternary deposits were deposited during the Holocene.

Till is a term "applied to a sediment that has been transported by glacier ice and subsequently deposited by or from it with little or no sorting by water (Dreimanis and Lindquivist, 1984; Klassen, 1989). In southwestern Manitoba, till is about 10 to 100 m thick (Klassen, 1989, Figure 2-17), and occurs to the surface in the Riding Mountain upland, Assiniboine River plain, Souris River plain, Tiger Hills upland, and Boissevain plain. Elsewhere till is overlain by fluvial, aeolian, and lacustrine sediments.

Glacial lake deposits occur in former glacial lakes, such as Lake Agassiz and Lake Hind. They consist of clay and silt, as well as sand. Glacial lacustrine sediments in Lake Agassiz are over 30 m thick (Teller, 1976), and consist dominantly of clay and silt. In the Lake Hind area, glacial lacustrine sediments are 10 to 90 m thick, consists of 0 to 20 m sand overlying 10 to 70 m silt and clay.

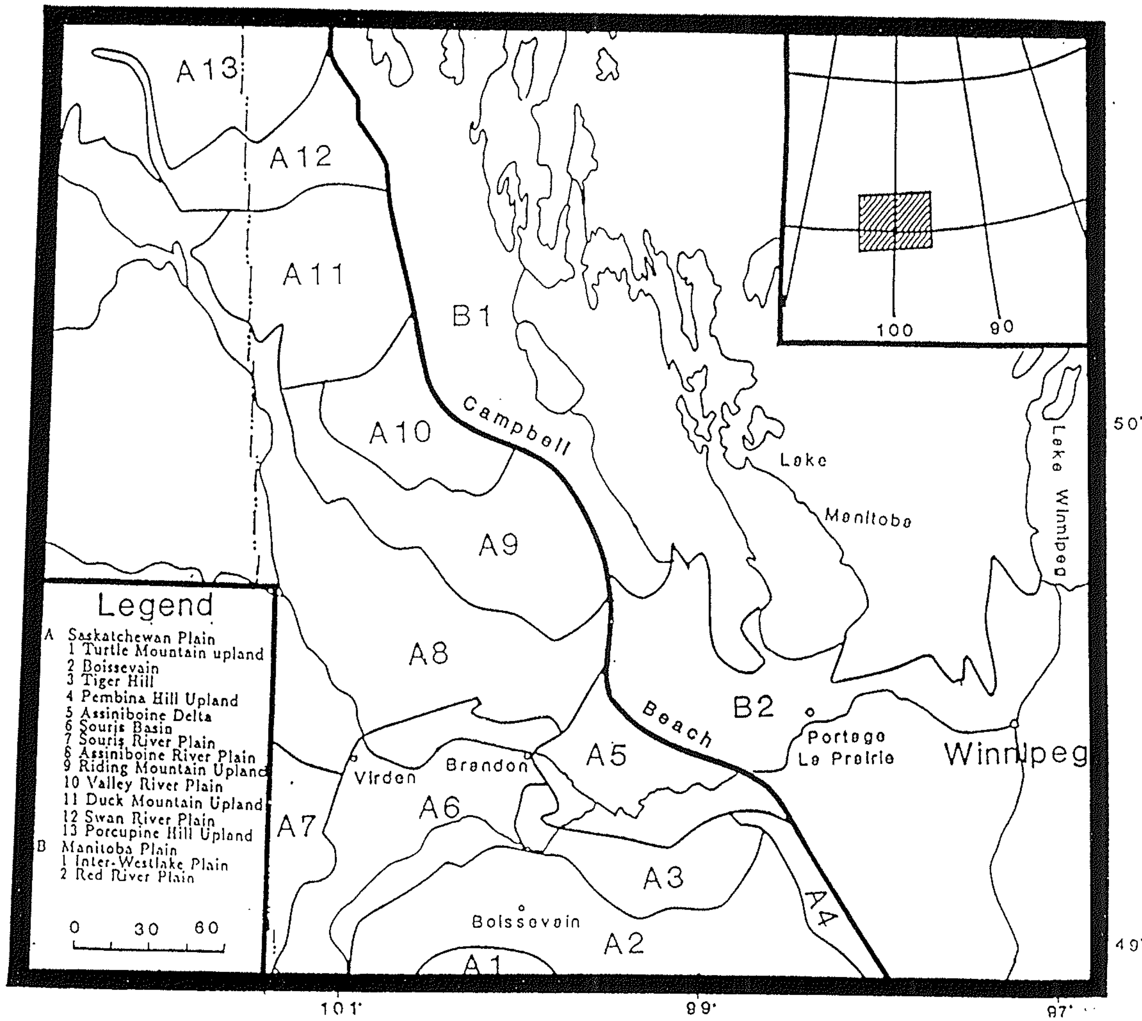


Fig. 1-7. Physiographic divisions of southwestern Manitoba (after Klassen, 1979).

Glacial fluvial deposits consist of sand and gravel that are mostly stratified. They occur as kames, eskers, deltas, and fans. They are commonly associated with glacial drainage systems. The largest landforms of glacial fluvial deposits are deltas in the western Lake Agassiz basin, such as the Assiniboine fan delta that was deposited by meltwater from the Assiniboine River, and the Pembina delta that was deposited by meltwater from the Pembina Spillway.

Drainage system

The Assiniboine River is the major river system in the study area (Fig. 1-8). It flows south from glacial lake Assiniboine near Kamsack, Saskatchewan, to Virden, Manitoba, where it turns and flows east toward Brandon (Fig. 1-1). The river flows in a shallow and relatively straight channel from Brandon to the junction of Souris River, and then flows in a meandering deep channel to the Campbell beach and beyond (Fig. 1-8). The modern Epinette Creek heads at Sewell Lake in an old distributary channel of the Assiniboine River, and drains southeast across the middle fan delta, joining the Assiniboine River north of Glenboro. After flowing north from North Dokota and turning east into the Pembina Valley near the town of Souris, the Souris River flows north from the Pembina Spillway through the town of Wawanesa and joins the Assiniboine River 12 km northeast of Wawanesa (Fig. 1-8). Other small rivers that drain to the Assiniboine fan delta include the Cypress River, Little Souris River, and Oak Creek (Fig. 1-8).

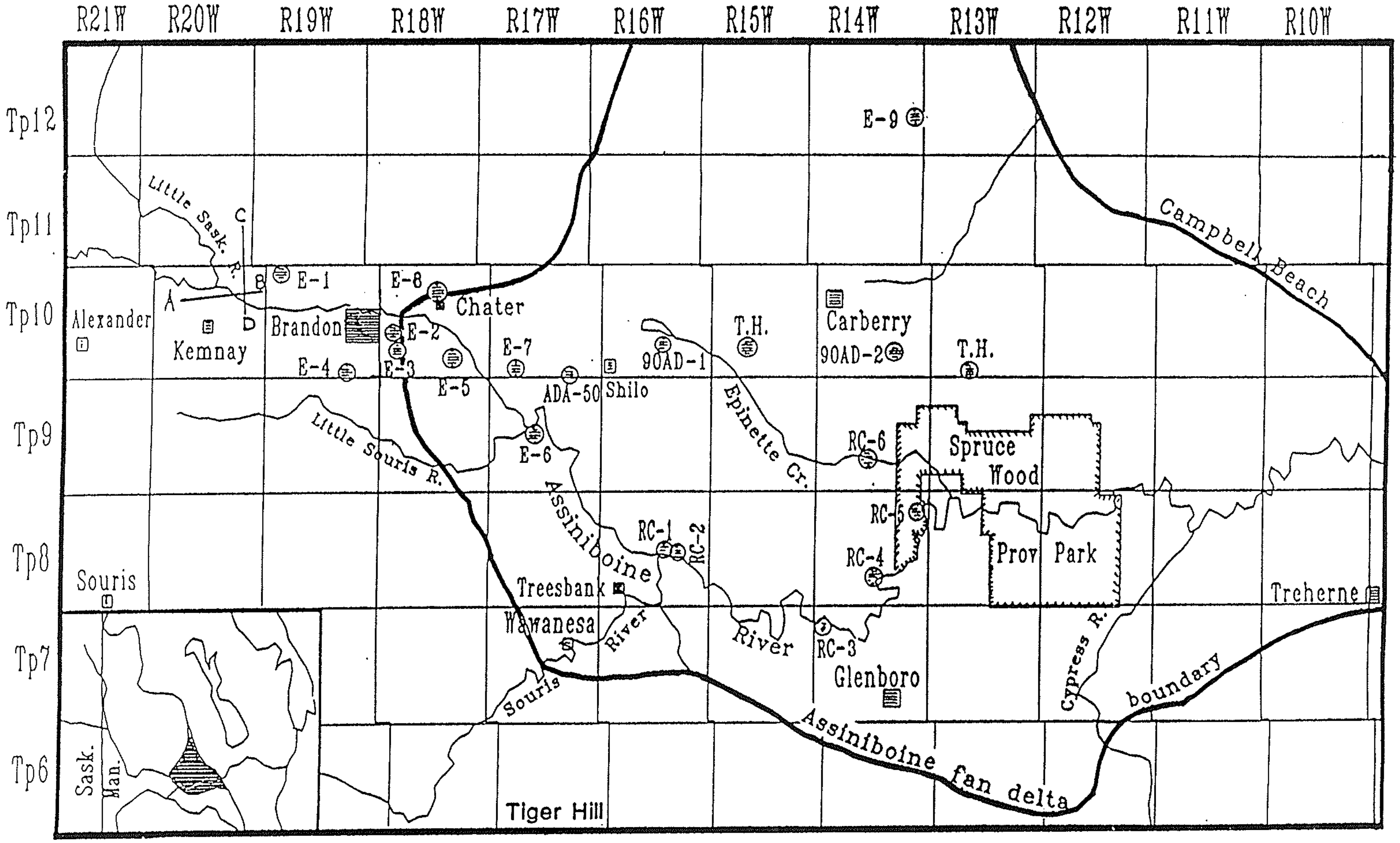


Fig. 1-8 Drainage systems, and location of exposures, river cuts, and wells in the Assiniboine fan delta area. E = exposures, RC = river cuts, A-B and C-D are cross sections, ADA-50, 90AD-1, 90AD-2, and T.H. are cores.

Chapter 2

TILL

Introduction

Till stratigraphic studies have been done by several workers in this area. Welsted and Young (1980) studied the till and superglacial fluvial deposits in the Brandon Hills area. Till studies in the Tiger Hills Upland and Boissevain Plain have been carried out by Elson(1956, 1967) and Conley(1986). Important detailed till stratigraphic studies have been done by Klassen (1979) in the Riding Mountain Upland, Duck Mountain Upland, and Assiniboine River Plain area.

Till Stratigraphy

Based mainly on the carbonate content and on stratigraphic position, 7 till units in the Riding Mountain, Duck Mountain, and Assiniboine River Plain have been identified by Klassen(1979), and 7 equivalent till units in the Killarney and Holmfield area have been reported by Conley(1986). These tills are, in ascending order: the Largs Formation, Tee Lake Formation, Shell Formation, Minnedosa Formation, Lennard Formation, Zelena Formation, and Arran Formation (Table 2). Only the Lennard Formation, Zelena Formation and Arran Formation have been interpreted to be late Wisconsinan in age and, therefore, only these may have any relationship to the history of the Assiniboine fan delta.

Lennard formation

The Lennard Fm is overlain by the Zelena Fm and Arran Fm on the Tiger Hills upland north of the Pembina Spillway, and is exposed at the surface in the Boissevain Plain and south of Pembina Spillway (Conley, 1986). It is the youngest

Table 2. Till formation southwest of Manitoba After 1) Klassen (1979); 2) Conley (1986)

Formation	Age	Direction of Ice Flow 1	Direction of Ice Flow 2	Carbonate Content 1	Carbonate Content 2
Arran	late Wisconsinan	NE-SW	NE-SW	34 – 65	27 – 44
Zelena	late Wisconsinan	NE-SW	NE-SW	26 – 36	18 – 25
Lennard	late Wisconsinan	NW-SE	NW-SE	14 – 19	13 – 16
Minnedosa	early Wisconsinan	NE-SW	NW-SE	14 – 17	14 – 17
Shell	pre Wisconsinan	E-W	NW-SE	24 – 36	12 – 16
Tee Lake	pre Wisconcene	NW-SE	NW-SE	20 – 26	23
Largs	early Pleistocene	NE-SW	NE-SW	9 – 21	8 – 10

till exposed on the Assiniboine River Plain and possibly the youngest till under the Lake Hind lacustrine sediments south of Virden (Klassen, 1979). The carbonate content of the Lennard Fm ranges from 14 to 19 percent, lower than that of the Zelena Fm and Arran Fm (Table. 2). Till fabric studies by Klassen (1979) and Conley(1986) indicates that the Lennard Fm was deposited by glacial ice flowing southeastward during the late Wisconsinan.

Zelena formation

The Zelena Fm is the youngest till on the Riding Mountain Upland (Klassen, 1979). According to the studies of others, it underlies the Arran Formation on the Valley River Plain, Swan River Plain, West Lake Plain (Klassen, 1979), and on the Tiger Hills Upland (Conley, 1986). The carbonate content of the Zelena Fm is 26 to 36 percent according to Klassen (1979), and 18 to 25 percent according to Conley's (1986) work on the Tiger Hills Upland (Table 2). The analysis of three till samples underneath the Assiniboine fan delta yield carbonate contents of 26, 28, and 30 percent (Fig. 2-1, Appendix I). Another two till samples from outcrops south of the Assiniboine fan delta, near the town of Wawanesa, show a carbonate content of 20 percent (Fig. 2-1).

The Zelena Fm was deposited by ice flow from the northeast during the late Wisconsinan (Klassen, 1979). Klassen(1979) indicated that the Zelena and Lennard formations are time equivalent, however Conley (1986) stated that the Zelena Fm overlies the Lennard Fm north of the Pembina Spillway.

Arran formation

The Arran Formation is the youngest till found in the Valley River Plain, Swan River Plain, and Interlake-Westlake Plain (Klassen, 1979) and is the youngest till on the Tiger Hills Upland (Conley, 1986). The carbonate content of the Arran

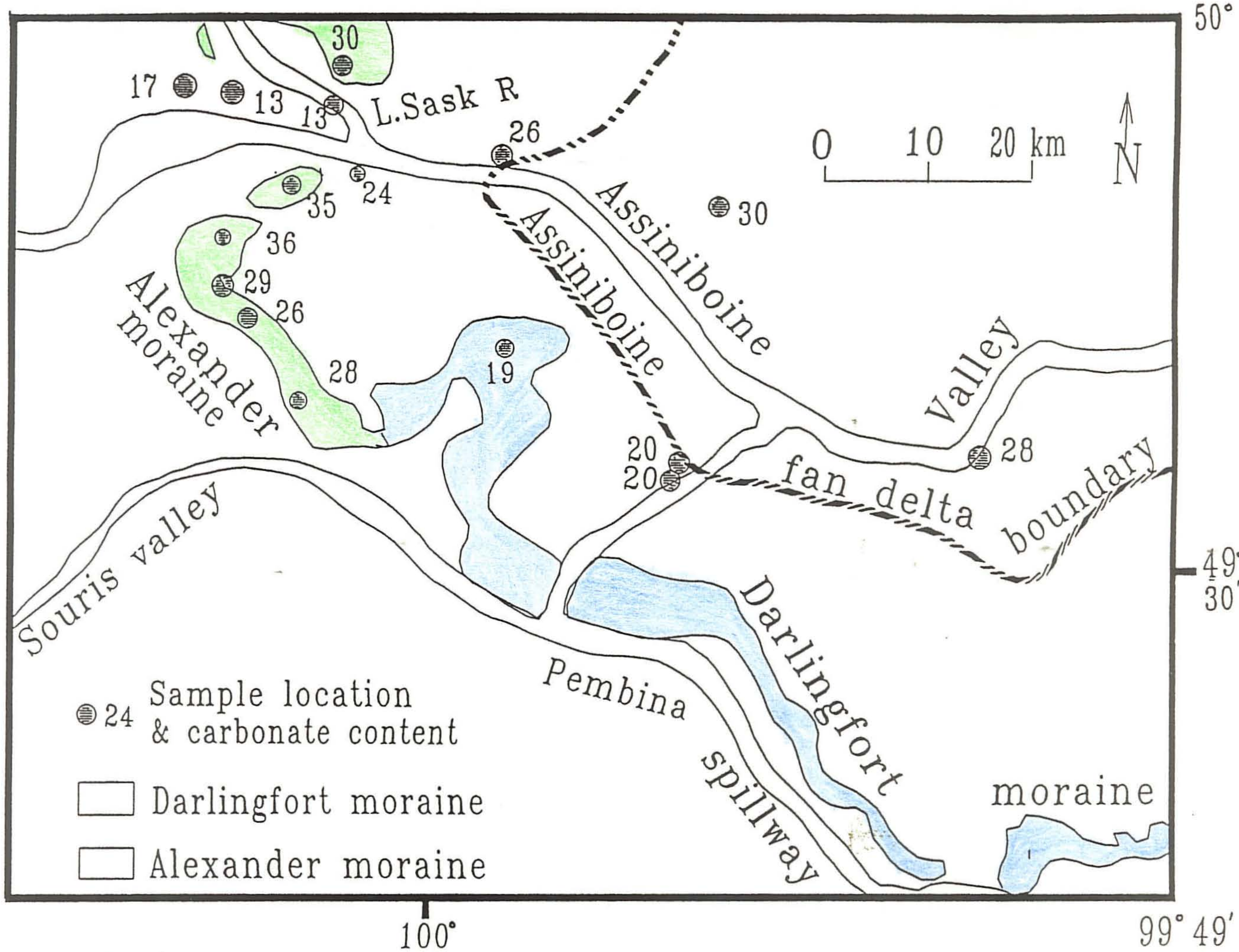


Fig. 2-1 The Darlingford Moraine and its northwestern extension (Alexander moraine). Numbers represent the percentage of carbonate in the clay-sand size fraction determined in this study.

Formation is reported as 34 to 65 percent by Klassen (1979) and 27 to 44 percent by Conley (1986) who studied the southernmost region of its extent (Table 2). The Arran Formation was deposited by southwestly ice flow during the last glaciation (Klassen, 1979; Conley, 1986).

The Elevated Belt of Till

An elevated belt of till lies across the topographically low area between the Assiniboine River and the Souris River and extends from north of Alexander south to the town of Souris. This hummocky ridge then turns eastward to Brandon Hills and the Tiger Hills Upland (Fig. 2-1). This till belt is about 3 to 8 km wide (Fig. 2-1). The elevation of the till belt between Alexander and Souris ranges from 435 to 450 m, which is lower than the Tiger Hills Upland to the southeast and the Assiniboine River Plain to the north, but higher than the elevation of lacustrine sediments in Lake Hind to the west and Lake Brandon to the east. The surface of this till belt contains many closely spaced shallow potholes with low relief (Fig. 2-2). In contrast, lacustrine silt and clay sediment adjacent to the till belt contains only a few pot-holes (Fig. 2-2), and the number of potholes decreases to the west and the east. The division between till and lacustrine clay approximately coincides with the 435m (1427 ft.) contour line between the towns of Alexander and Souris. In the northern end of the till belt, the number of potholes decreases sharply to the higher Assiniboine River Plain, while eastward from the town of Souris, pothole topography merges with the hummocky topography on the Darlingford Moraine along the southern Tiger Hills upland.

Based on 7 samples, I have determined that the till in this belt is composed of 5 to 10 percent gravel, 30 percent sand and 60 to 65 percent of silt and clay. The carbonate content of till ranges from 24 to 36 along the till belt (Fig. 2-1,

Fig. 2-2 Airphoto of the pot-holed till. Notice the abundant pot-holes (in till) below the red line and fewer pot holes (in silty clay) in the upper right (above the red line) of the photo.

Appendix I), which is generally within the range determined for the Zelena till by Klassen (1979). In a local area north of the Assiniboine River and west of the elevated till belt, where there are no potholes on the surface, the carbonate content of till is much lower, ranging from 13 to 17 percent only (Fig. 2-1). For comparison, a till sample from the Virden area was analyzed, which yielded 19 percent carbonate (Appendix I). Analyses of till by Klassen (1979) from the Assiniboine River Plain northwest of the till belt show the carbonate content there also to be lower than 20 percent. Therefore it is interpreted to be part of the Lennard Formation.

Interpretation

The carbonate content of the Zelena Fm, which was deposited by ice flowing from the northeast across more carbonate-rich bedrock, ranges from 26 to 36 percent (Klassen, 1979) (18 to 25 percent according to Conley, 1986). The carbonate content of the Arran Fm, which also was deposited by northeast-southwest flowing ice, ranges from 34 to 65 percent (Klassen, 1979), or 27 to 44 percent according to Conley (1986) (Table 2). Thirteen of my till samples (three under the Assiniboine fan delta, two samples near the town of Wawanesa, one on the Brandon Hills, and seven samples on the till belt) have their carbonate content ranging from 19 to 36 percent with an average of 27 percent. These data suggest that these till samples belong to either the Zelena or Arran Fm. They are the Zelena Fm if Klassen's (1979) standard is used. However, it can be either the Zelena or Arran Formation according to Conley's study (1986). I prefer to interpret them as the Zelena till, partially because they merge into the Zelena till in the Riding Mountain upland, and into the Darlingford moraine, which consists of mainly Zelena till (Conley, 1986) in the Tiger Hills upland.

The pot-holed nature of the till belt from the town of Alexander to Souris

(Fig. 2-1) indicates that these tills were deposited by stagnant ice, and the elevated surface of the till belt (at least 15 m higher than the area immediately to the west or east of the belt) suggests that this till belt may be part of an end moraine. In fact, the southern end of this elevated till belt merges with the Darlingford Marine (Fig. 2-1), which marks the maximum extent of deposition of the Zelena and Arran formations (Conley, 1986). Therefore it is highly possible that the elevated and hummocky till belt is an extension of the Darlingford Moraine, being deposited at the western end of the Zelena ice, which flowed from the Interlake area. For convenience, this till belt is named the Alexander end moraine.

Silt and clay sediments west of the till belt were deposited in Lake Hind while silt and clay sediments east of the till belt were deposited from Lake Brandon. There is no trace of lacustrine or fluvial sediments over the summit of the till belt, suggesting that Lake Hind and Lake Brandon water levels were lower than the ice barrier, and that these lakes were largely drained by the time the ice barrier decayed and the hummocky moraine belt was finally deposited.

Chapter 3

GENERAL CHARACTERISTICS OF THE ASSINIBOINE FAN DELTA

Introduction

In this chapter I describe the general geomorphology of the fan delta, the thickness of the fan delta sediments and the aerial distribution of particle size, and the thickness of the lacustrine silt and clay. The detailed sediment description and stratigraphy of the fan delta will be presented in the next chapter.

The Assiniboine fan delta covers an area of about 6400 km^2 between Brandon and Portage La Prairie (Klassen, 1983). The main part of the fan delta lies between Brandon and the Campbell beach (Fig. 1-1), and is the target of this study.

East of the Campbell beach escarpment the surface elevation of the distal fan deposits ranges from 275 to 305 m with a uniform gradient of 1.9m/km towards the east. West of the beach, the fan delta surface is higher in the northeast area and lower to the south and southwest. West of Brandon, the fan delta joins with the river plain leading to the delta, and the rivr plain is at an elevation between 380 m and 410 m.

Geomorphology of the Fan Delta

Based on the degree of scouring and the nature of the surficial sediments, the Assiniboine fan delta is arbitrarily divided into 3 areas (Fig. 3-1): (1) Brandon and west of Brandon region (the river plain leading to the fan delta, zone 1 in Fig. 3-1); (2) Brandon to Douglas Station region (the proximal fan delta, zone 2 in Fig. 3-1); and (3) Douglas station to the Campbell beach escarpment (the middle to distal fan delta, zone 3 in Fig. 3-1).

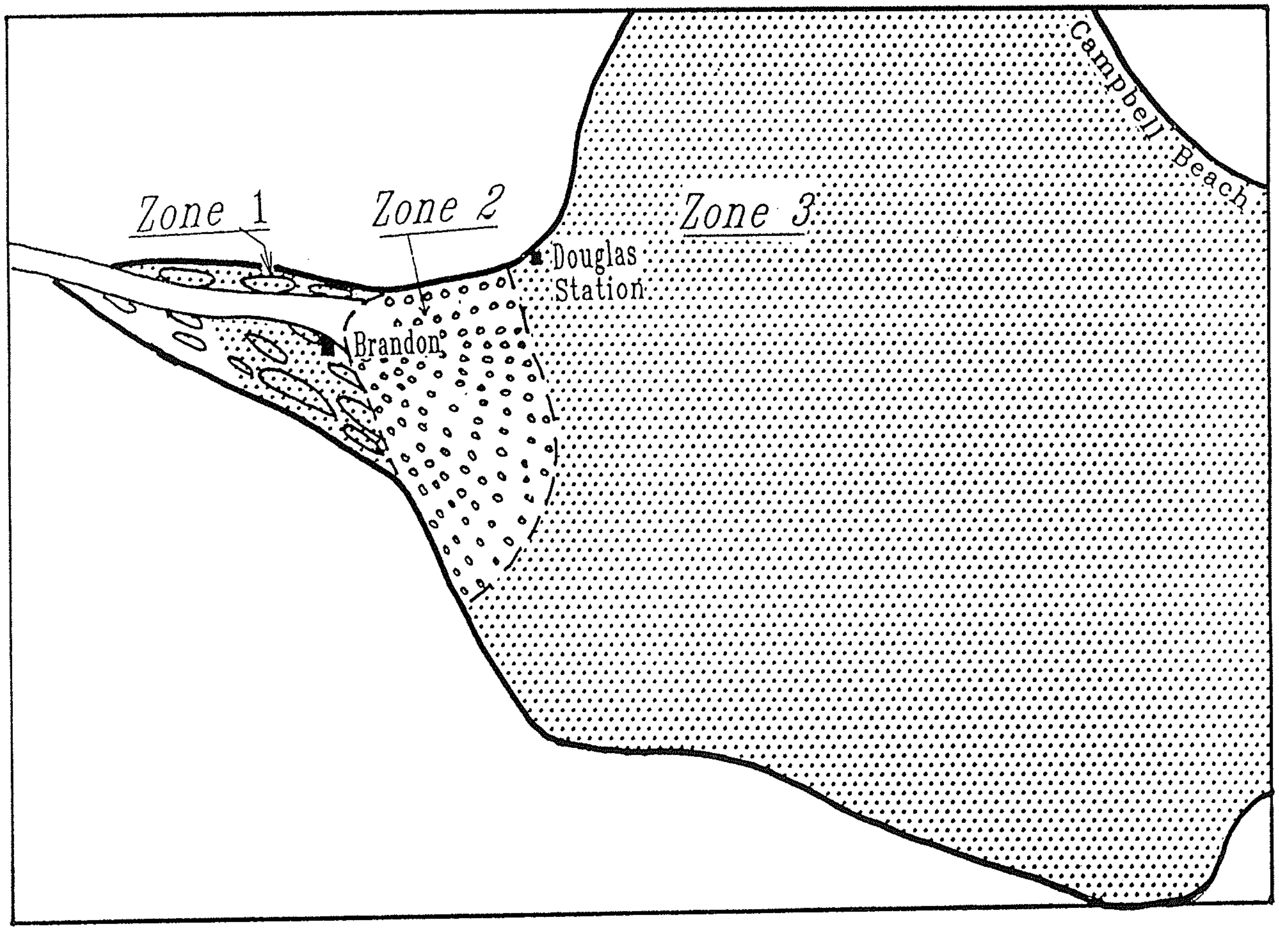


Fig. 3-1 Diagram showing 3 zones of the Assiniboine fan delta. Zone 1 is the braided river plain leading to the fan delta, which is separated from zone 2 by an escarpment. Zone 2 is dominated by gravel deposits. Zone 3 is dominated by sand deposits.

Brandon and west of Brandon region, zone 1

This region is a wide scoured region bounded to the north and south by lacustrine silty clay at higher elevations. The Assiniboine Valley divides this region (Fig. 3-2). North of the Assiniboine River, fluvial deposits occur at an elevation of about 403 to 417 m, with a sharp lower contact to the underlying till or lacustrine silty clay, and a possible erosional lateral contact to the lacustrine silty clay to the north and south (Fig. 3-3).

Sediments on the south side of the river are at an elevation between 380 m to 410 m. The landscape consists of shallow scoured channels (Braid channels) and low residual hills that are elongated parallel with the modern Assiniboine River. In the western area, till is exposed both on the streamlined hills and in the scoured channel floors, and patches of cross bedded gravels and boulders (50 percent shale fragments) occur in scoured channels and in downstream side of residual hills. In the eastern area, surficial sediments are dominated by sand and sandy gravel, which do not have a high percentage of shale.

Brandon to Douglas Station region, zone 2

The Assiniboine fan delta in this region is bounded by unscoured glacial sediments to the north and south that lie at least 5 m above the level of the fan delta. The fan delta is severely scoured in the southern area in the vicinity of the Assiniboine valley (from 3 km east of Brandon to the Little Souris River junction) where most of the fan delta sediments has been stripped away, leaving only channels and residual hills with their summit at or lower than 380 m. The No.1 hill and the No.2 hill are the major residual hills (Fig. 3-2).

The No.1 streamlined hill is about 3 km long, 300 to 900 m wide, and 8 m high, with its top at an elevation of 374 m (Fig. 3-4). The upstream end of the hill is

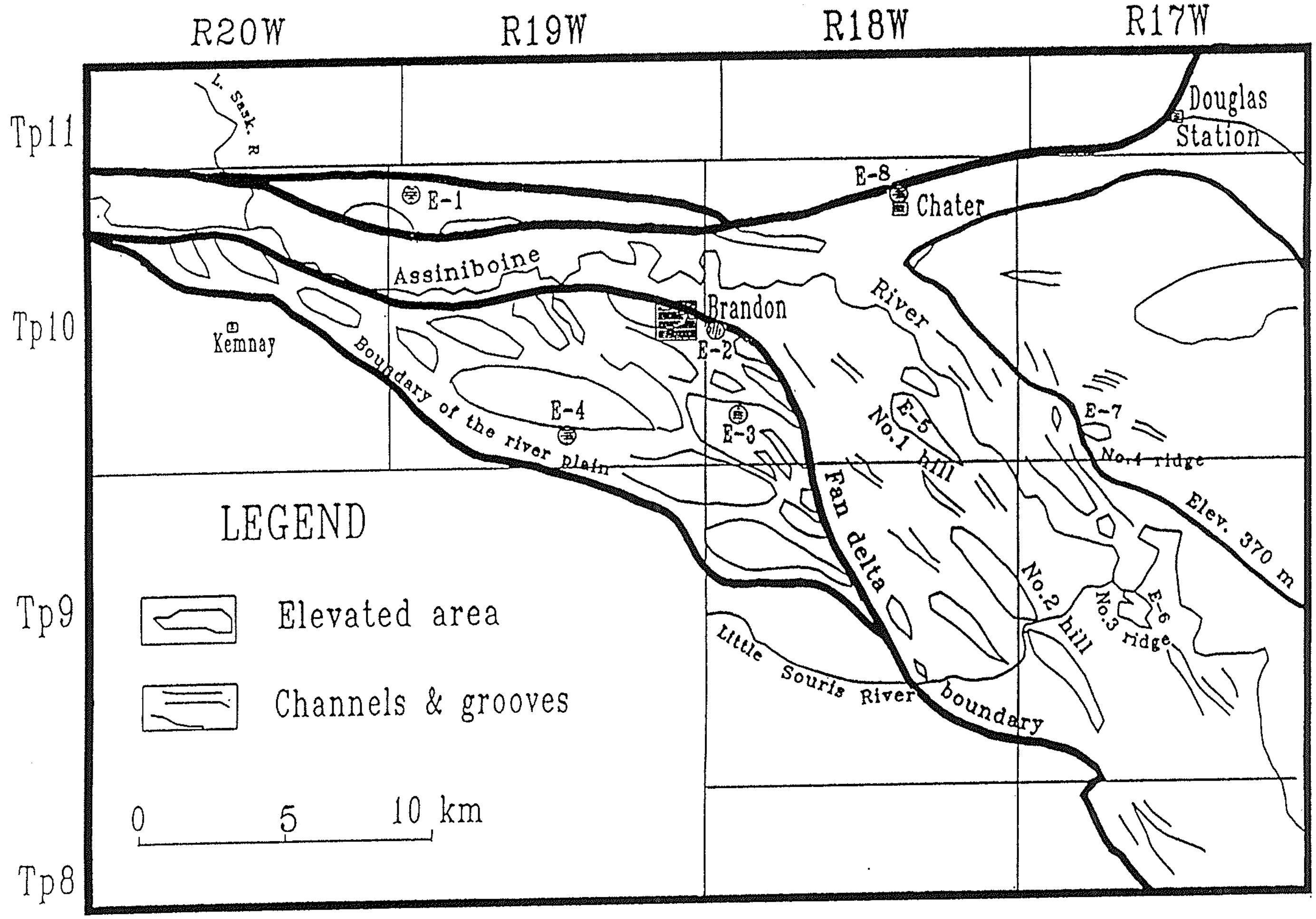


Fig. 3-2 Geomorphology of the western Assiniboine fan delta. This figure is mapped based mainly on interpretation of airphotos (scale 1:20,000).

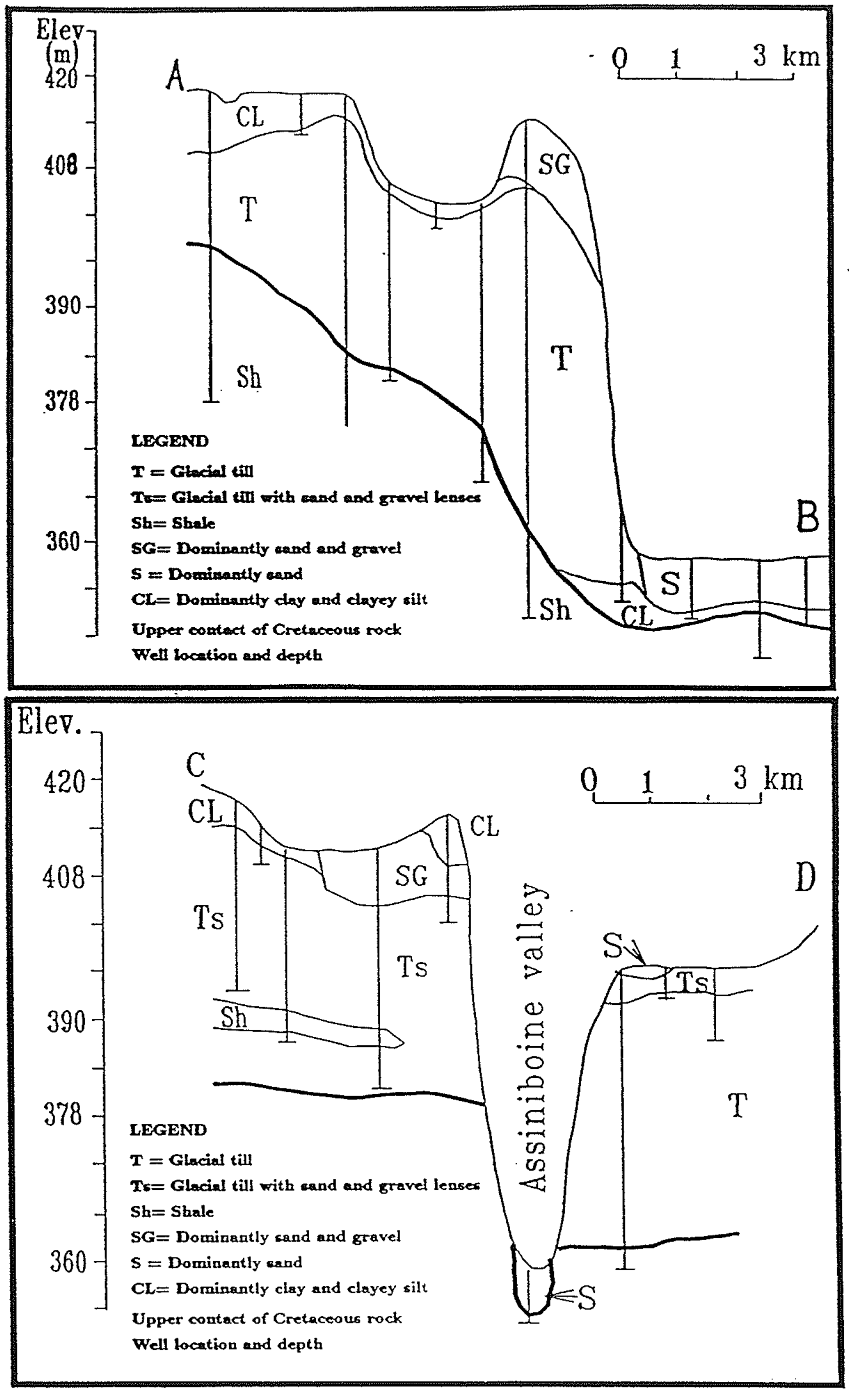


Fig. 3-3. Cross sections west of Brandon (A-B', and C-D' in Fig. 1-8). Notice that sand and gravel occur in low elevations, whereas silty clay commonly is present at higher elevations.

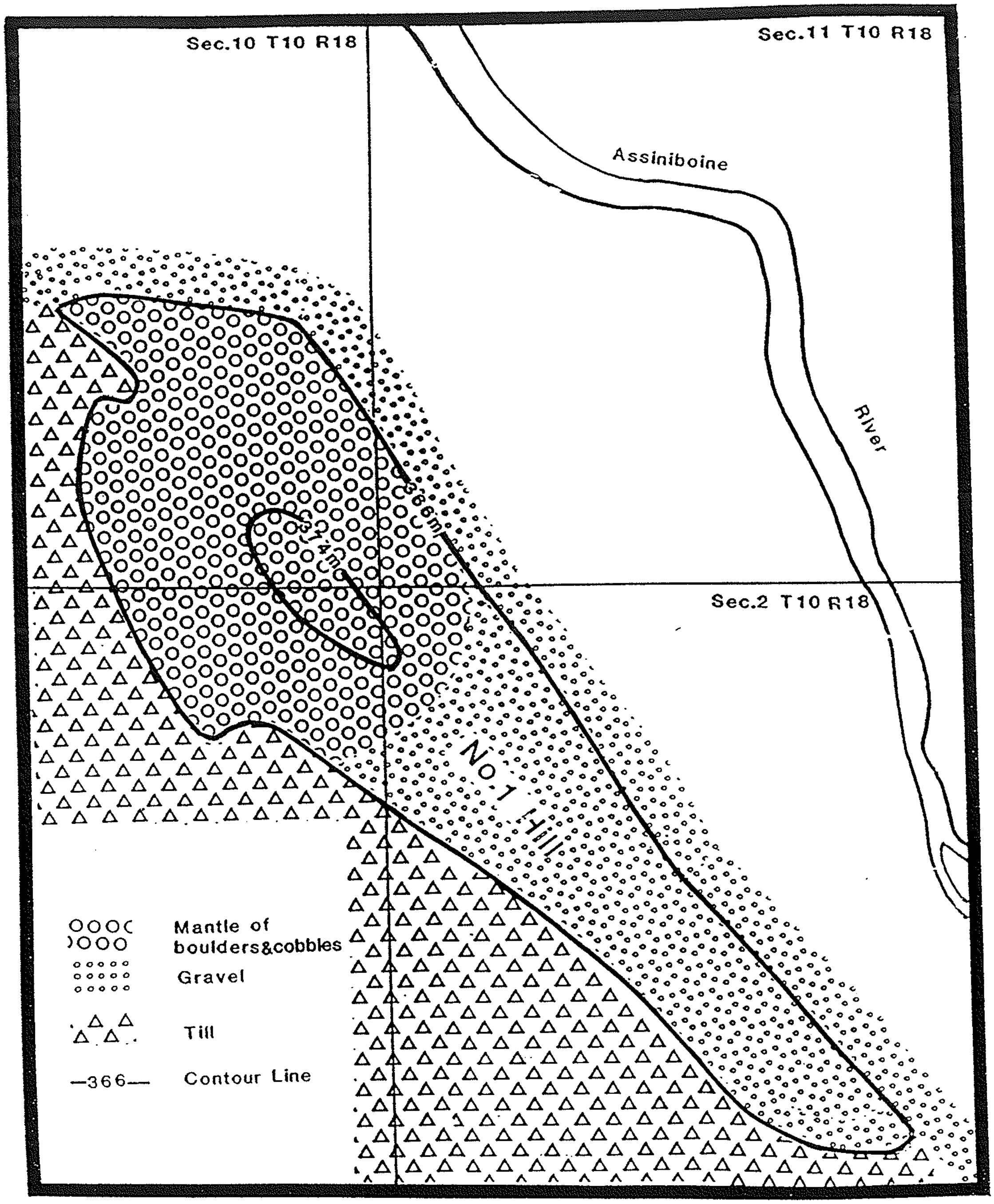


Fig. 3-4 No.1 streamlined hill at 5.5 km east and 1 km south of Brandon (E-5 in Fig.1-8, and No.1 hill in Fig. 3-7).

an irregular shape, and is covered with boulders and cobbles. The downstream end of the hill is narrower and near-surface sediments are fine gravels and sands. The grain size and stratigraphy in the interior of the hill will be discussed in the next chapter. Scoured channels lay both to the west and east of the hill. The floor of the scoured channel on the west side of the hill is composed of till, while the channel floor on the east side of the hill is composed of gravels.

The No.2 streamlined hill is about 8 km long, 1 km wide, and 10 to 12 m high (Fig. 3-5). It is outlined by the 370 m contour line with its summit at an elevation of 380 m (Fig. 3-5). The upstream portion of the hill is armored with cobbles and boulders up to 40 cm in diameter. The summit of the hill is mostly composed of silty sand overlying sandy gravels, and the downstream end is composed of mainly fine to medium grained sand. The grain size in the interior of the hill is unknown. The lowland area surrounding the hill is composed of gravelly clay till with a lag of boulders on the surface. The Little Souris River, instead of flowing around the hill cuts across the hill in the middle (Fig. 3-5).

The northern half of the proximal fan delta seems relatively well preserved because there are only a few scoured grooves and ridges (Fig. 3-2).

Douglas Station to the Campbell beach escarpment, zone 3

The highest point of elevation in this area, 390 m, is in the northeastern fan delta (Fig. 3-1). A relatively level plain stands at an elevation of about 388 m north of Carberry and at 383 m south of Carberry. Much of the remainder of the area is covered by inactive sand dunes (Manitoba Water Resources Department, 1970). Generally the land slopes gently southward from the plain to the Assiniboine River and east to the Campbell beach escarpment.

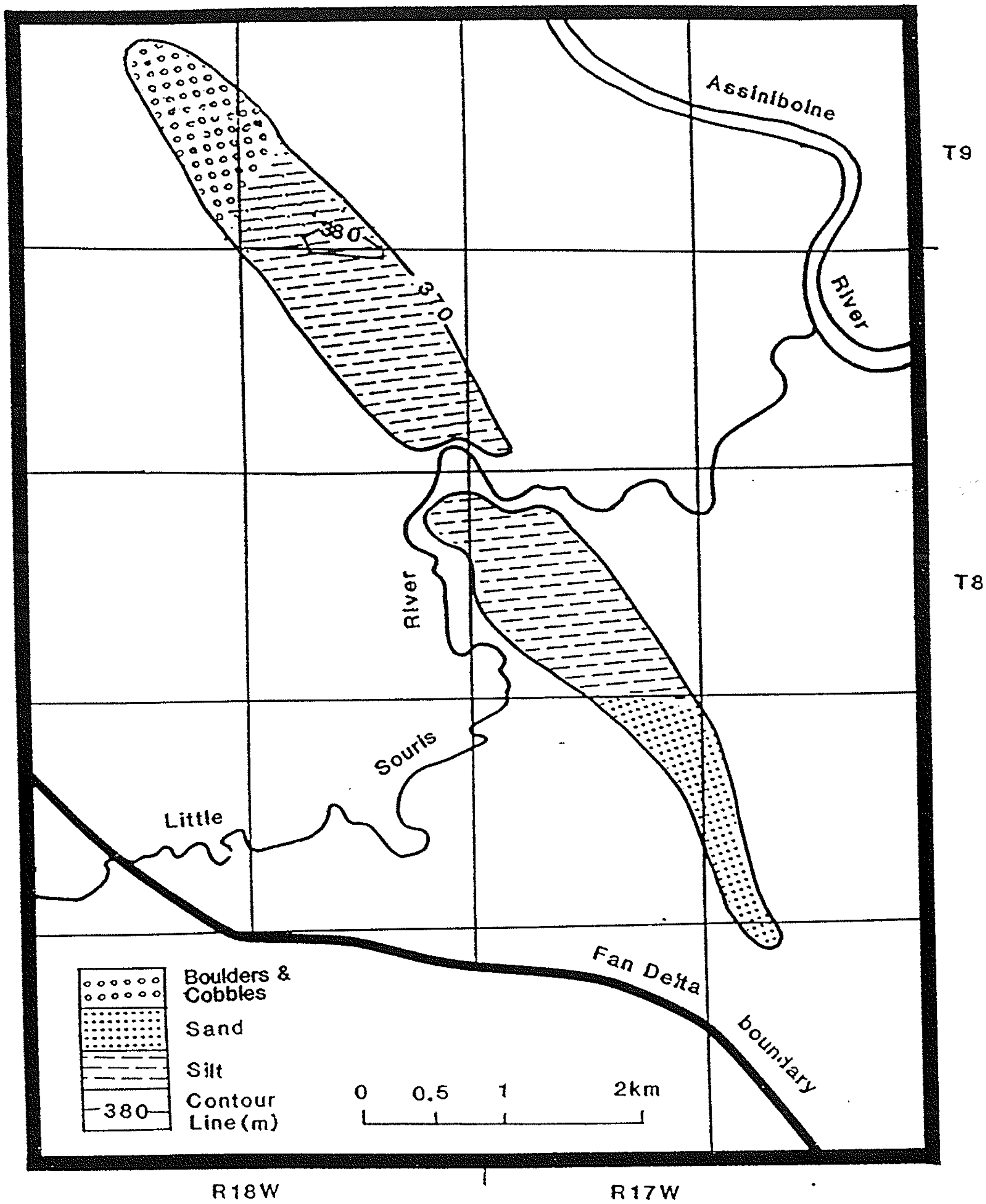


Fig. 3-5 No.2 streamlined hill at 3 km west of the Little Souris River junction (No.2 hill in Fig. 3-2).

Thickness of the fan delta sediments

The fan delta sediments west of the Campbell beach escarpment are thickest in the northeastern region, and thiner toward south and southwest (Fig. 3-6). The thickest part of the fan delta is located 10 km north and 10 km east of Carberry, while the thinest part is located near Brandon, where either till is exposed on the scoured channel floor or is overlain by less than 10 m of fan delta sediments. Although eolian dunes are locally thick, Eolian sand does not affect the general trend of the thickness of the fan delta very much. As Fenton et al. (1983) note, there is a general relationship between the eaolian dunes in the Lake Agassiz basin and fine to medium grained delta sediments. In the northeastern part of the fan delta, where fan delta sediment is the thickest, no sand dunes are marked on the Quaternary Geology Map (Manitoba Mineral Resources Division, 1980). In the southern part of the fan delta, where it is relatively thin, sand dunes are well developed.

Grain size distribution of the fan delta

About 400 well logs were used to study the aerial distribution of grain size near the fan delta surface. In each well the average particle size of the upper 3 m of the core were recorded. Based on these data the general distribution of particle size in the fan delta was mapped (Fig. 3-7). This figure is characterized by gravels near the fan delta apex (about 200 km^2), fining eastward to fine sand and silty clay 10-20 km west of the Campbell beach escarpment, and then becoming coarser along the Campbell beach escarpment. For comparation purpose the particle size near the bottom (0 to 3 m above till) of the proximal fan delta area was also recorded. Gravels dominate a 350 km^2 area extending from Brandon to 20 km east of Brandon.

Lacustrine Silt and Clay

A lacustrine clayey silt and clay unit underlies the fine to coarse grained sandy

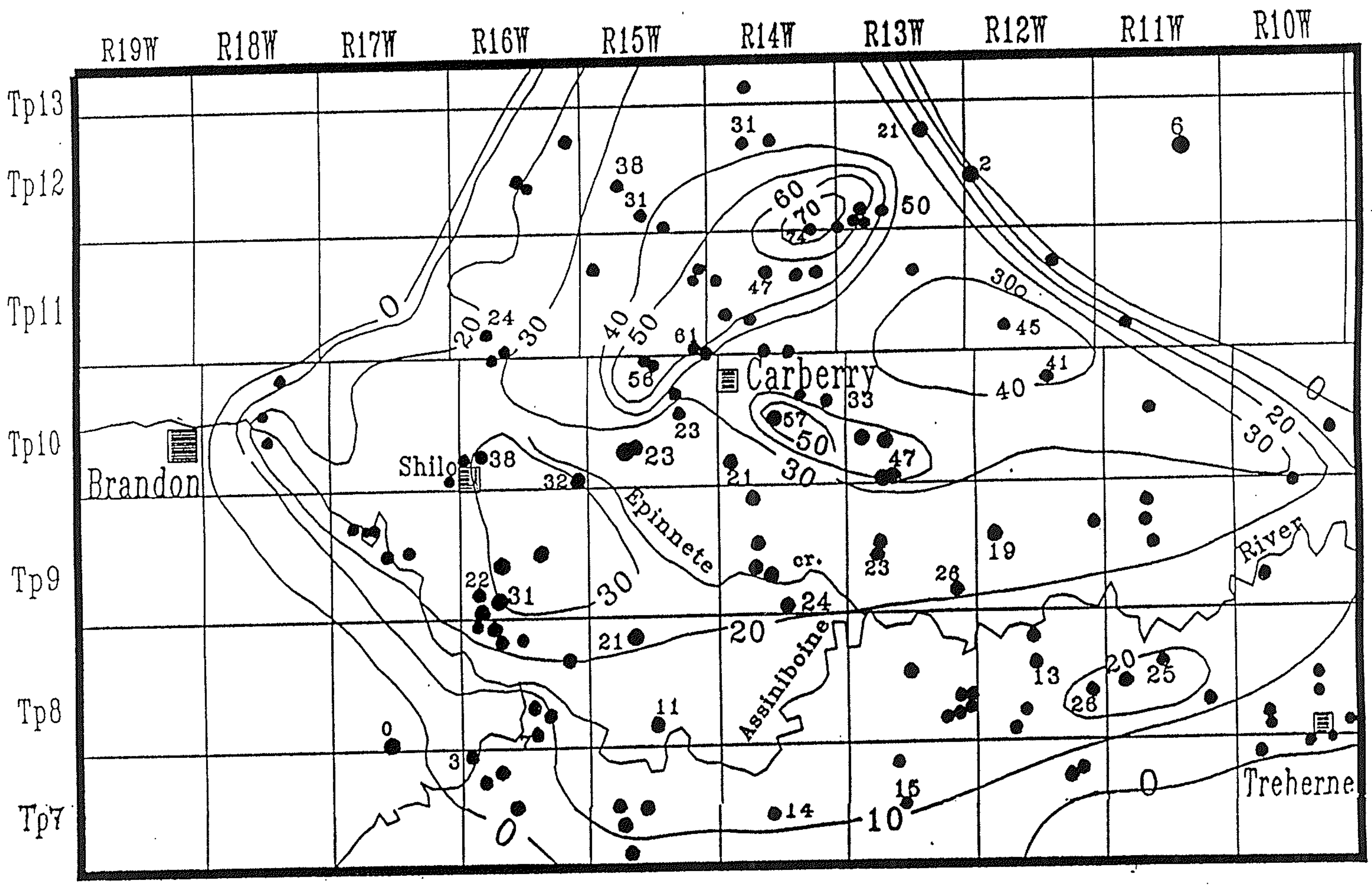


Fig. 3-6. Thickness (in m) of the Assiniboine fan delta. Selected numbers used to contour the map are shown. Dots represent deep wells drilled by P.F.R.A (1986, unpub. report) and Manitoba Water Resources Division. Contour interval=10m.

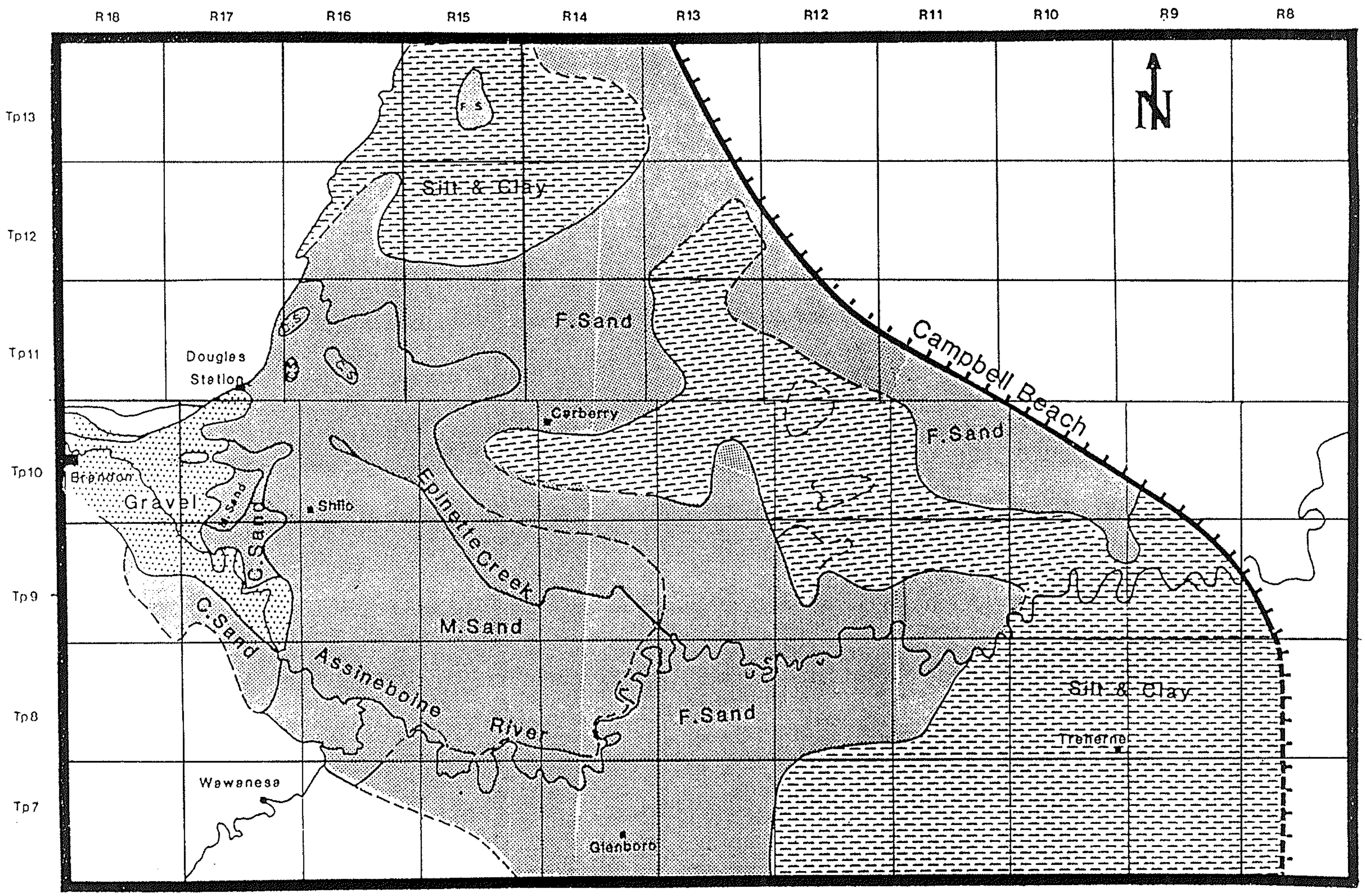


Fig. 3-7 Grain size distribution of the fan delta sediments near the surface. Data used to plot this map include 400 well logs, and outcrop exposures. Only the dominant grain sizes are shown in the figure.

and gravelly fan delta sediments in the study area. Where observed in outcrops and boreholes this silt/clay unit overlies till. The total thickness of the silt and clay unit under the fan delta ranges from a few m to 70 m (Fig. 3-8). The thickness of lacustrine silt and clay east of Range 13 is not clear because it is difficult to distinguish it from the younger fine grained distal fan delta fine sediments, especially in borehole descriptions by water well-drillers. Overall this unit thickens toward the northeast (Fig. 3-8).

To the west of the fan delta south of Kemnay and northwest of Brandon (Fig. 1-8), clay and clayey silt are exposed at the surface. In this area, the clay and clayey silt unit extends from the lower basin floor at an elevation of 405 m to higher land at an elevation of 442 m (Fig. 3-3). In the area close to the Assiniboine valley, where coarse grained sand and gravel is exposed at the surface, the underlying clay and silt unit is thin or absent. The vertical and lateral contacts between coarse grained and fine grained sediments are sharp (or erosional).

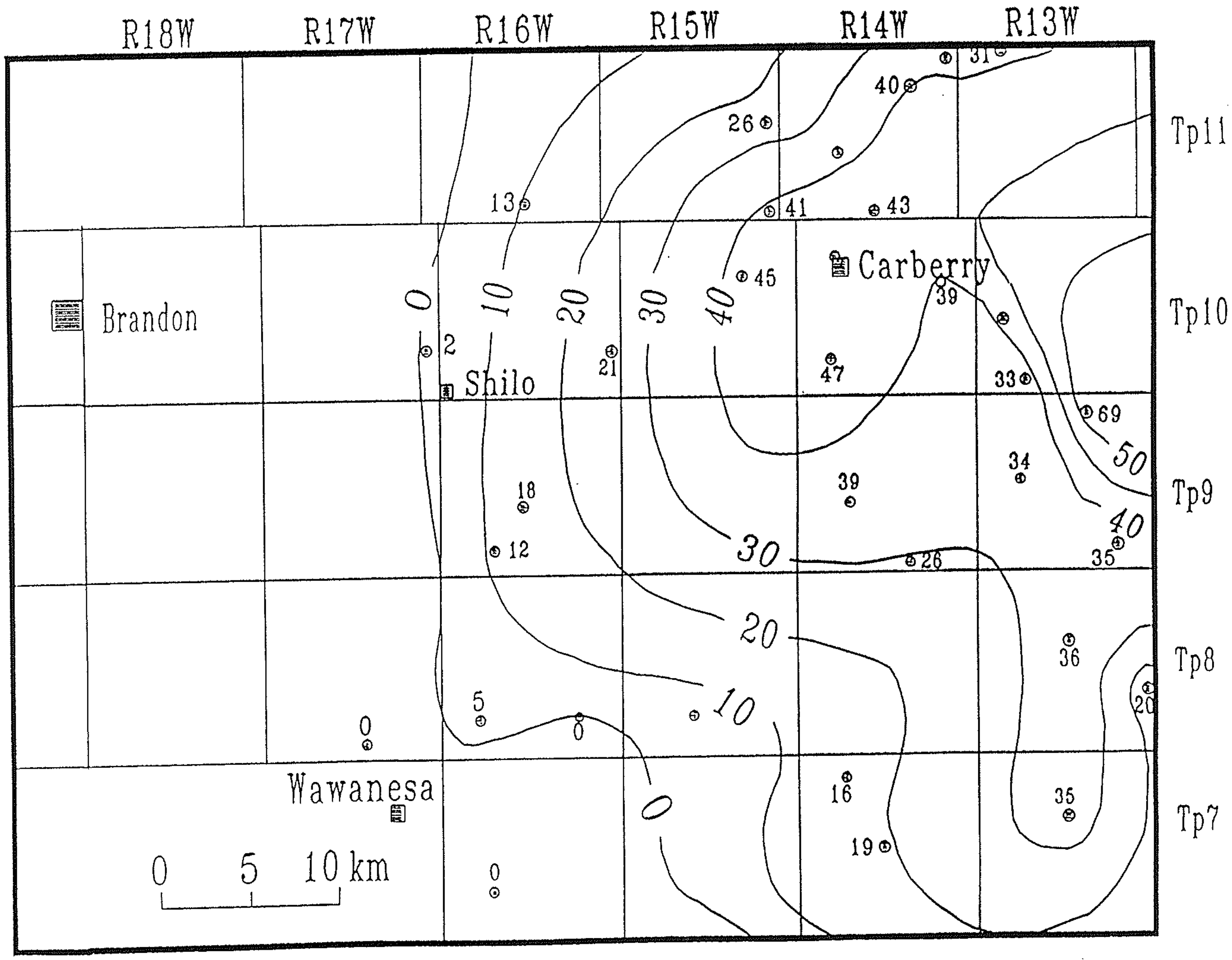


Fig. 3-8 Thickness (in m)of the lacustrine clay and silt unit below the coarser fan delta deposits. This isopach map was based on well logs, which were mostly drilled by the Department of Mines and Energy.

Chapter 4

SEDIMENT DESCRIPTION AND STRATIGRAPHY OF THE ASSINIBOINE FAN DELTA

Introduction

The stratigraphy of the Assiniboine fan delta has been constructed by studying 9 major outcrops, 4 river cuts, 6 radar profiles (a total about 1.5 km), and 2 new rotasonic cores. In addition hundreds of other well logs and outcrops were examined. The major outcrops, cores, and radar profiles are described in this chapter in the following order:

(1) Exposures in the river plain leading to the fan delta; (zone 1 in Fig. 3-1).

(2) Exposures and radar profiles in the proximal fan delta (zone 2 in Fig. 3-1).

(3) Exposures, river cuts, and cores in the middle fan delta (zone 3 in Fig. 3-1).

In order to facilitate the correlations of outcrops and cores, based on grain size measurements, stratigraphy, and studies of sedimentary structures, 8 sedimentary facies were established. Details of these facies and their depositional environments will be discussed in the following chapters.

Exposures in the River Plain Leading to the Fan Delta (zone 1)

Introduction

Exposures in the river plain near Brandon and west of Brandon include Exposure 1 (E-1 in Figs. 1-8 and 3-2), Exposure 2 (E-2 in Figs. 1-8 and 3-2), Exposure 3 (E-3 in Figs. 1-8 and 3-2), and Exposure 4 (E-4 in Figs. 1-8 and 3-2). These exposures are above an elevation of 380 m, and mostly consist of sand that is overlain by a thin unit of massive sandy gravel. In contrast, outcrops at lower elevations (360-380

m) are mostly composed of cross bedded sandy gravels overlain by 1-2 m of massive sandy gravels.

Exposure 1 (E-1).

This outcrop is located north the Assiniboine River west of Brandon at sec. 31 Tp10, R19W (E-1 in Fig. 3-2), and exposes an upper (gravel) and a lower (sand) unit (Fig. 4-1), which overlies silty clay just below the floor of the pit. The upper unit is composed of a massive sandy gravel (Facies Gm) (Fig. 4-1) that is non-graded, non-stratified, and supported by its sandy matrix; it contains 15 to 25 percent shale pebbles and high amounts of silt and clay. The lower contact is sharp.

The lower unit consists of 3 m of cross bedded gravelly sand (Facies St), and is dominated by coarse sand that contains 5 to 15 percent gravel (mostly shale pebbles and cobbles). Cross beds include gently inclined cross beds and steeply inclined cross beds, and are outlined by shale-deficient sand (brown colour) and shale-rich sand layers (grey colour). Cross beds dip about 80°E. The lower contact with the underlying pale brown (10YR 7/4) silty clay unit is very sharp.

Exposure 2 (E-2).

The Exposure 2 (Fig. 4-2, E-2 in Fig. 3-2) is a gravel pit 0.8 km east of Brandon at SW$\frac{1}{4}$ of NE$\frac{1}{4}$, Sect.18 Tp10, R18 at an elevation of 385 m. 3 m of pebbly sand were exposed in a north-south cut and a west-east cut. The north-south cut is composed of gravelly coarse grained sand (Facies St) that has large scale cut and fill structures, each about 10 to 20 m wide and 1 to 3 m thick. Gravel-sized clasts (mainly pebbles) make up 10 percent of the sediment, and are scattered throughout the section. In the west-east cut, sediments are dominated by cross bedded coarse grained sand (Facies St) (Fig. 4-2). Cross beds dip about 90°E. The lower contact is covered by slump.

Fig. 4-1 Exposure 1 (E-1 in Figs. 1-8 and 3-2) is located at the northern side of the Assiniboine River west of Brandon in middle of Sec.31, Tp10, R19W. This section is oriented at E80°N. Notice that the tabular cross bedded sand (St) is overlain by massive, non-graded sandy gravel (Facies Gm).

Fig. 4-2 Exposure 2 (E-2 in Figs. 1-8 and 3-2) is located east of Brandon at SW$\frac{1}{4}$ of NE$\frac{1}{4}$, Sect.18 Tp10, R18. Two photos show trough cross bedded (upper photo, north-south cut) and tabular cross bedded (lower photo, west-east cut) pebbly sand (Facies St).

Exposure 3 (E-3).

This exposure is located in the S$\frac{1}{2}$ of NE$\frac{1}{4}$, Sec.7, Tp10, R18W (E-3 in Fig. 3-2) at an elevation of 383 m. The upper 0.4 m of the section is composed of sandy gravel (Facies Gm) that is massive, non-graded, and nonstratified (Fig. 4-3). Gravels are dominated by cobbles and pebbles with less than 10 percent boulders. Elongated clasts are randomly oriented. The basal contact is sharp. The lower 1 m of the outcrop is composed of weakly horizontally bedded medium grained sand (Facies Sl), which is well sorted ($\sigma = 0.43\phi$) and contains less than 1 % silt and clay.

Exposure 4 (E-4).

This section is located in the SW $\frac{1}{4}$ of Sec.2, T10, R19W (E-4 in Fig. 3-2) at an elevation of 390 m, and is composed of two units (Fig. 4-4). The upper unit is a massive, non-stratification, very poorly sorted sandy gravel (Facies Gm) that is supported by a sandy matrix. It has a sharp lower boundary. Clasts in this unit are dominated by pebbles, with 10 to 20 percent cobbles. The lower unit is composed cross laminated silty very fine sand intercalating thin beds of coarse grained sand and fine gravels (Facies St).

Exposures and Radar Profiles in the Proximal Fan Delta (zone 2)

Introduction

The proximal fan delta is arbitrarily assigned to the area where sediments are dominated by gravels, specifically from 3 km east of Brandon to the middle part of Range 17 West (Figs. 1-8 and 3-2). Exposures in this area include gravel pits in the No.1 streamlined hill (E-5), the No.3 ridge (E-6), and the No.4 ridge (E-7); the elevation of the crests of these hills ranges from 374 to 380 m. Sediments in these hills, which are erosional remnants, are dominated by gravels, commonly comprised

Fig. 4-3 Exposure 3 (E-3 in Figs. 1-8 and 3-2) is located 0.5 km south and 0.5 km east of Brandon (south of NE$\frac{1}{4}$, Sect.12, Tp10, R18W). Notice that flat bedded clean sand (Facies Sl) is overlain by massive sandy gravels (facies Gm).

Fig. 4-4 Exposure 4 (E-4 in Figs.1-8 and 3-2) is located south of Brandon at SW$\frac{1}{4}$ of SW$\frac{1}{4}$, Sec.2, Tp10, R19W. The laminated silty sand (Facies Sl) is overlain by massive sandy gravels (Facies Gm).

of massive sandy gravel that is underlain by thick cross bedded gravel. The lower part of the fan delta sequence is not exposed in this area.

Exposure 5 (E-5), No.1 streamlined hill

The No.1 streamlined hill is located 6 km east and 1 km south of Brandon (E-5 in Figs. 1-8 and 3-2). Six units were identified in the No.1 streamlined hill from exposures and radar profiles, and are called, in ascending order, unit A, unit B, unit C, unit D, unit E, and unit F (Figs. 4-6 and 4-7, from lines B and D in Fig. 4-5). The upper 3 units (units D, E, and F) are exposed on the wall of a gravel pit located in the upstream part of the streamlined hill at S$\frac{1}{2}$ of Sec.10, Tp10, R18W (Figs. 4-8 and 4-9, from lines A and C in Fig. 4-5), while the lower 3 unit (units A, B, and C) can only be identified on radar profiles that were run across the floor of the gravel pit (Fig. 4-6). The lateral changes in units D and F can be seen by comparing sections across the stoss side of the hill (Fig. 4-8), middle of the hill (Fig. 4-9), and the lee side of the streamlined hill (Fig. 4-10, E in Fig. 4-5).

Unit A, identified in radar profile A (Fig. 4-6), is dominated by flat beds with an erosional upper contact. The grain size composition of this unit is unknown. An industrial well (Appendix III, Chemical plant well) at Sec.10, Tp10 R18W, 0.6 km north of the streamlined hill, revealed clayey till, overlain by 0.9 m of poorly sorted sand and gravel, which in turn, is overlain by 3.4 m of fine to medium sand (Appendix 3, Chemical Plant well). This poorly sorted gravel is thought to correlate with the unit 'A' (Facies Gm) because of the similar stratigraphic position.

Unit B, which is identified from the radar profile in Fig. 4-6, is about 0.5 m to 3 m thick, and has a channel fill geometry from 0 to 40 m along the profile. It may belong to Facies Gx or St. The upper and lower contacts are sharp.

Unit C is the lowest unit exposed on the floor of the gravel pit. It is about 2 to

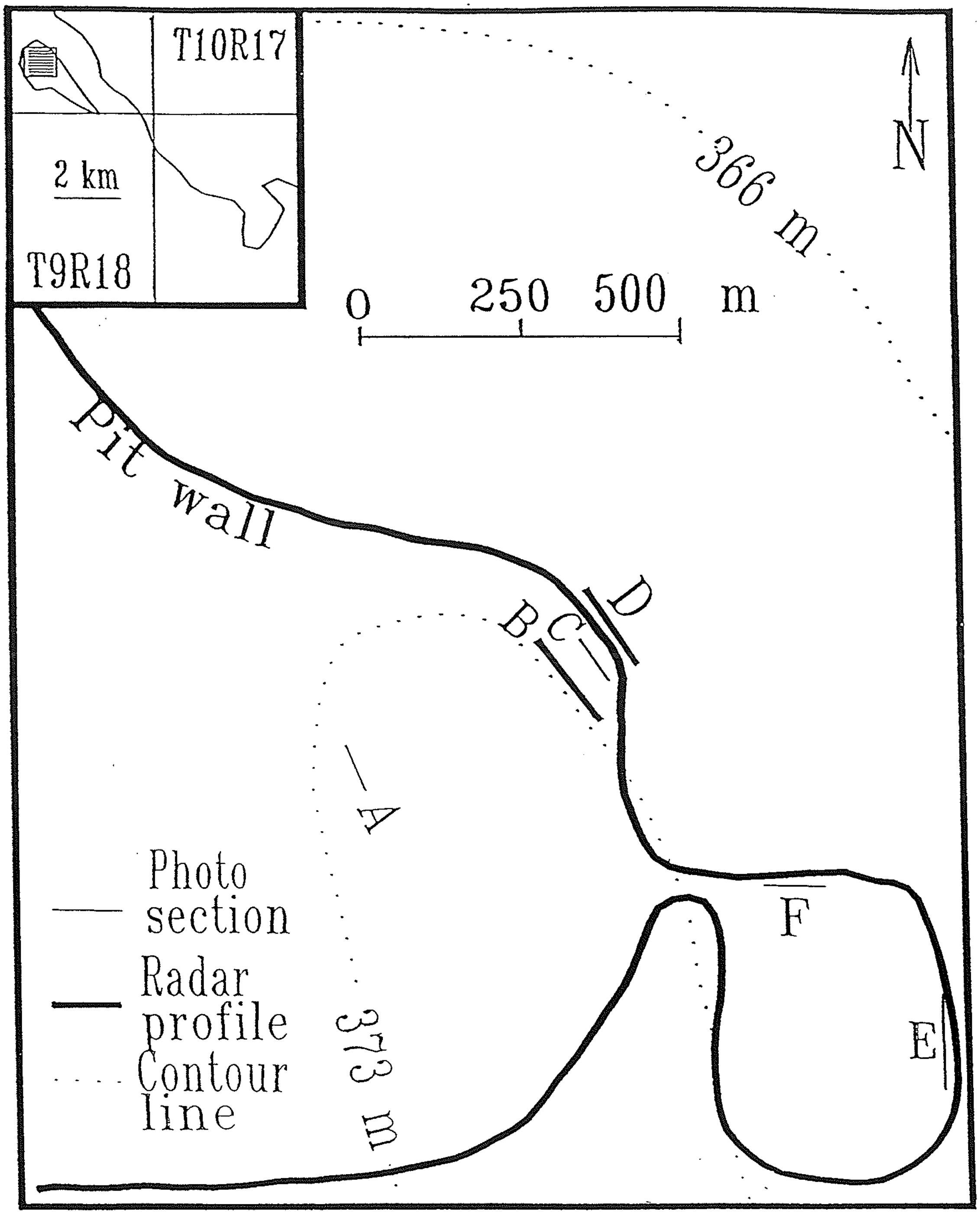


Fig. 4-5 Index of radar lines B and D and photos A, C, E, and F in the gravel pit located at the upstream portion of the No.1 streamlined hill (E-5 in Figs. 1-8 and 3-2). Dotted contour lines are those on the original landform, prior to the pit excavation.

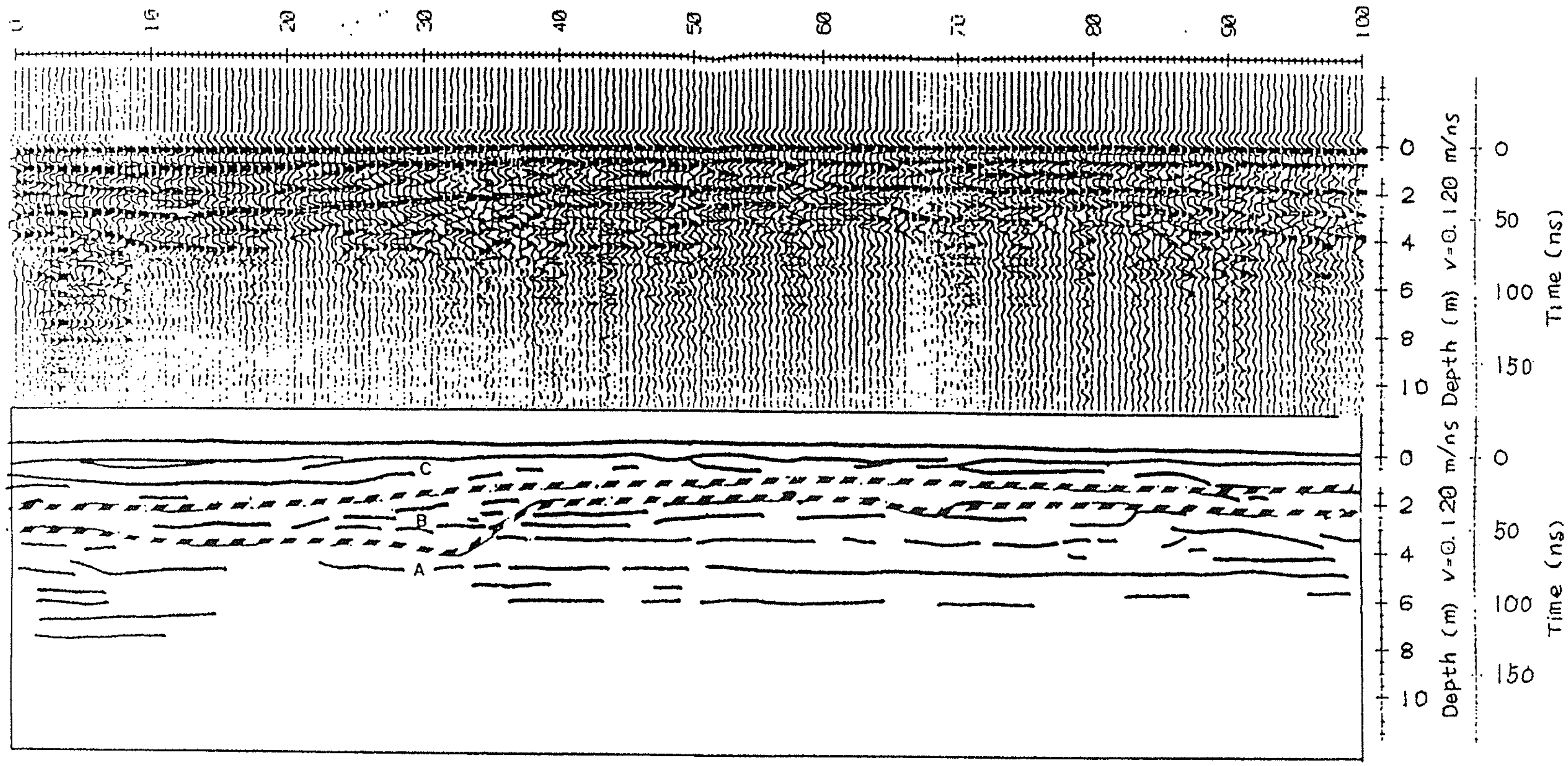


Fig. 4-6 Radar profile across the bottom of a gravel pit near the head portion of the No.1 streamlined hill (B in Fig. 4-5). The antennae used for this profile is 200 mega Hz. Note the outlined erosional contact between units B and A.

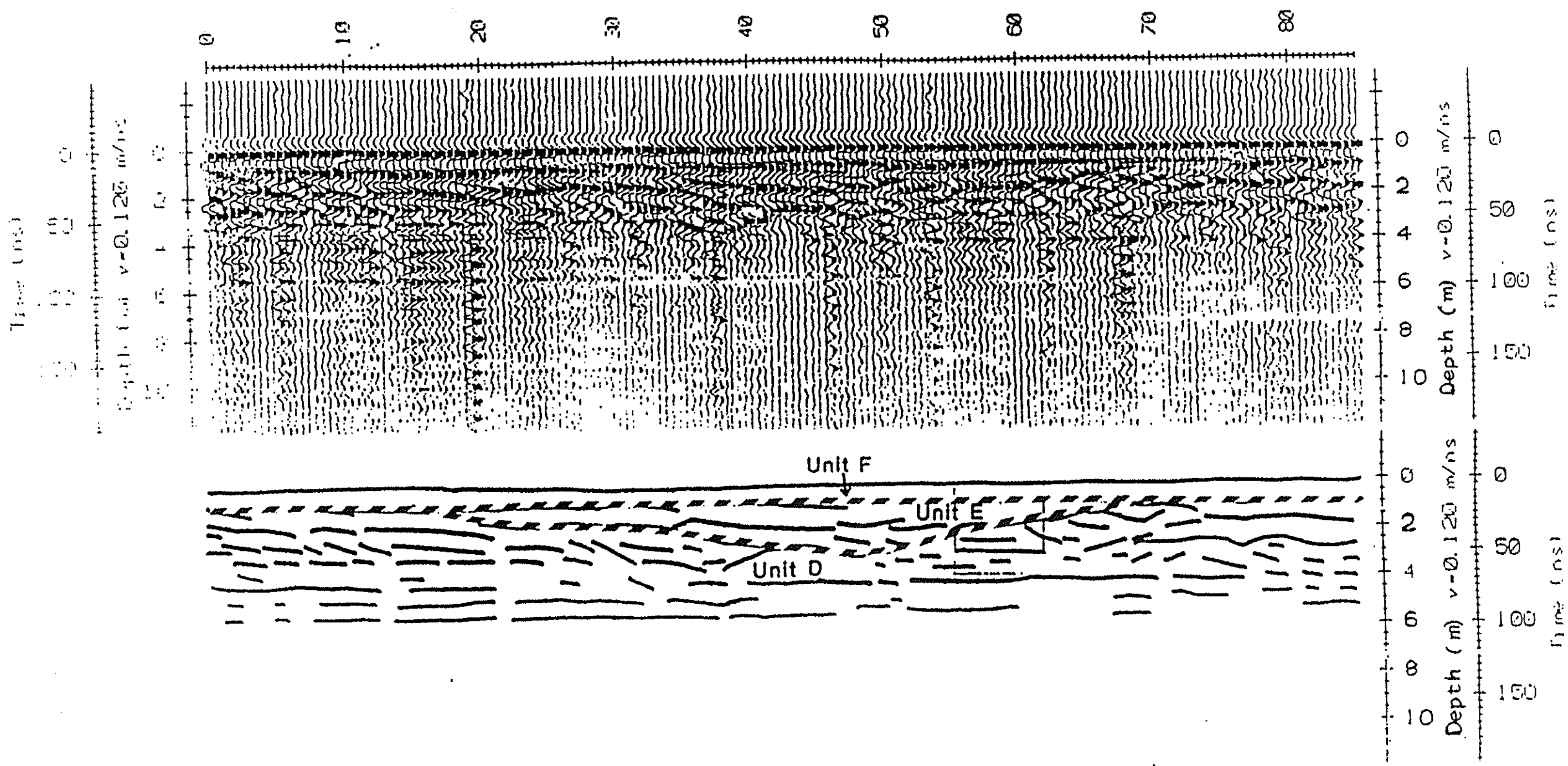


Fig. 4-7 Radar profile (D in Fig. 4-5) in the middle of the No.1 stream-lined hill and is oriented 130°. Note the cross beds in unit D and the erosional contact between units D and E.

Fig. 4-8 Cross bedded boulders and gravels (unit D, Facies Gx) overlain by massive, non-graded sandy gravel (unit F, Facies Gm) in the stoss side of the No.1 streamlined hill. Notice the sharp changes of the grain size in each cross bed from basal boulders to sandy pebbles. The exposure is about 7 m high. Maximum boulder size is 0.8 m. Location of this photo is shown as A in Fig. 4-5.

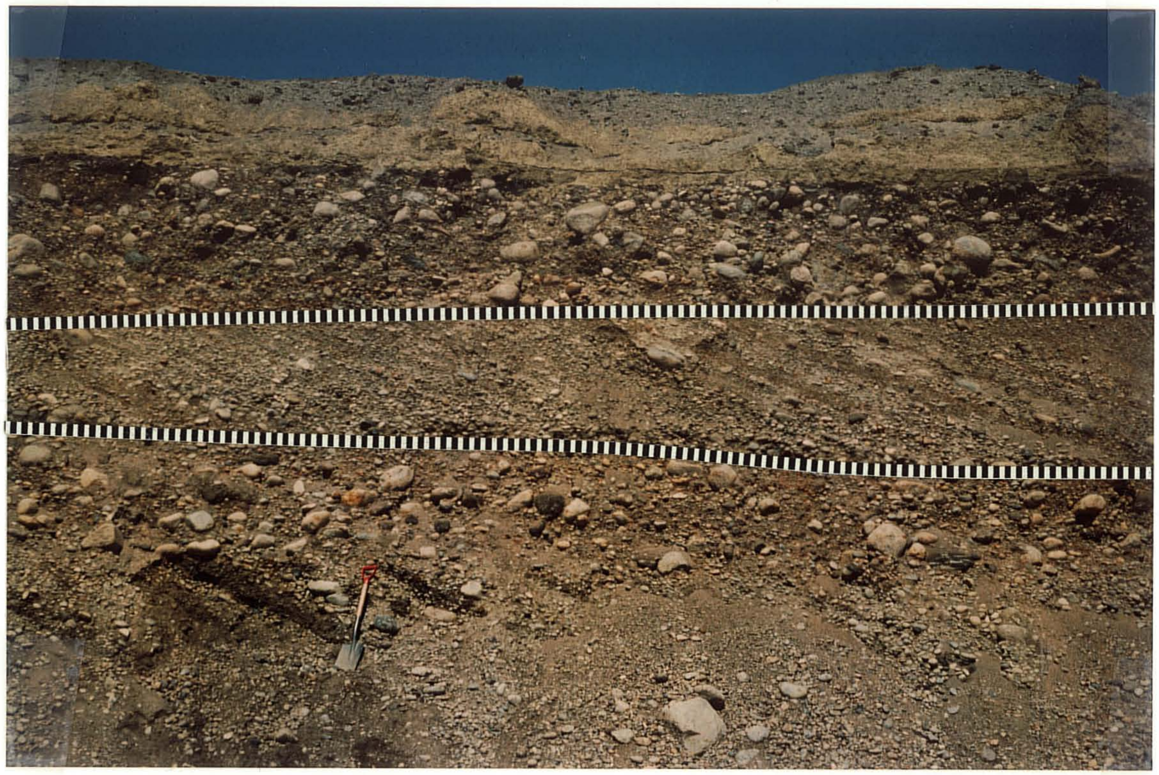


Fig. 4-9 A section near the central part of the no.1 hill (C in Fig. 4-5). Notice the poorly sorted and non-graded unit F (Facies Gm) on the upper part of the exposure underlain by the cross bedded and normally graded unit E (facies Gx), which in turn, is underlain by tabular cross bedded gravel of unit D (facies Gx).

Fig. 4-10 Gently inclined cross bedded to flat bedded gravels and sand of unit D (facies Gh), overlain by structureless sandy gravels (facies Gm), in the lee side of the No.1 streamlined hill. Location of this section is shown as E in Fig. 4-5.

3 m thick and consists of non-graded gravelly sand (Facies Sm). On the radar line (Fig. 4-6), this unit is dominated by flat beds or very low angle inclined beds.

Unit D is well exposed in the walls of the gravel pit (Figs. 4-8, 4-9, 4-10). The thickness of this unit ranges from 6 m on the stoss side section (Fig. 4-8, A in Fig. 4-5) to 4 m on the central section (Fig. 4-9, C in Fig. 4-5) and the lee side section (Fig. 4-10, E in Fig. 4-5). In the exposure at the stoss end, sediments are steeply cross bedded, with dips toward the lee end of the elongated hill. Individual cross beds are crudely graded (Facies Gx) and composed of large boulders and cobbles fining upward to poorly sorted pebbles and granules(Fig. 4-8). Boulders and cobbles at the base of each cross bed are clast supported, and contain boulders up to 0.8 m in diameter, with occasional fabric orientation to the flow. However, clasts in the upper part of each cross bed are dominated by granules and pebbles and supported by a sand matrix, with scattered cobbles and and occasionally rounded brown till balls of 0.25 to 0.6 m in diameter (Fig. 4-8).

In the central section (Line C in Fig. 4-5), sediments of unit D are cross bedded and crudely graded (Facies Gx) (Fig. 4-9). Clast supported, normally graded cobbles and pebbles with an open framework constitute the basal part of each cross bed, while poorly sorted, and matrix supported granules and pebbles constitute the upper portion of each cross bed. The upper contact of unit D is distinct (Figs. 4-7, 4-9).

On the lee side of the hill in a north-south exposure (location E in Fig. 4-5), Unit D is composed of flat bedded to gently inclined sandy cobbles and pebbles at the base, grading upward into pebbly granules and sand (Facies Gh) (Fig. 4-10). In a west-east section at the same site, both sandy gravel and gravelly sand are generally flat bedded.

Unit E is exposed locally in the central section of the hill (Fig. 4-9). Both on the radar profile and in exposure in the pit wall this unit consists of a lens (40 m long and 1.5 m thick) (Fig. 4-7) of cross bedded pebbles, granules, and sands, with some cobbles (Facies Gx) (Fig. 4-9). Sediments in each cross bed are graded, well sorted, and clast supported. The upper and lower boundaries of this unit are marked by zones of cobbles (Fig. 4-9).

Unit F is 1 to 2 m thick, and consists of massive cobbles, pebbles, granules, and sands (Facies Gm). It is poorly sorted, non-stratified, non-graded, matrix supported, with random fabric (Figs. 4-8, 4-9, 4-10). The characteristics of this unit are similar both on the stoss and the lee side of the hill, except that more till balls and un-deformed laminated clay balls are present locally in the lee side (Fig. 4-11, F in Fig. 4-5).

Exposure 6 (E-6), No.3 ridge section

Two sections were exposed by the mining of gravel in two pits located in a ridge (or lower hill) about 0.8 km east and 0.5 km south of the junction of Little Souris River with the Assiniboine River (Fig. 4-12, E-6 in Figs. 1-8 and 3-2). The surface elevation of the crest of this ridge is at 362 m.

Section A (A in Fig. 4-12) is located in the central zone of the ridge. It is oriented at N155°E, and exposes 5 m of cross bedded cobbles, pebbles, and granules (Facies Gx). Cross beds are generally symmetrical to a core zone and dip away from the core zone at 24°, similar to an 'anticline' (Fig. 4-13). Each cross bed is composed of cobbles grading across the cross-stratum to pebbles, granules, and coarse sand (Facies Gx). The basal cobbles are well sorted, graded, and clast supported, with an open framework. The maximum cobble size is 12 cm in diameter, with an average of 3.5 cm. Granules and pebbles in the upper part of each graded bed have an average

Fig. 4-11. Sediments of the unit F in No.1 hill (F in Fig.4-5). The upper photo shows till balls and rip-up clasts in unit F. The shovel is 0.92 m long. The lower photo shows a piece of clay with well preserved laminae. The compass is 0.22 m long.

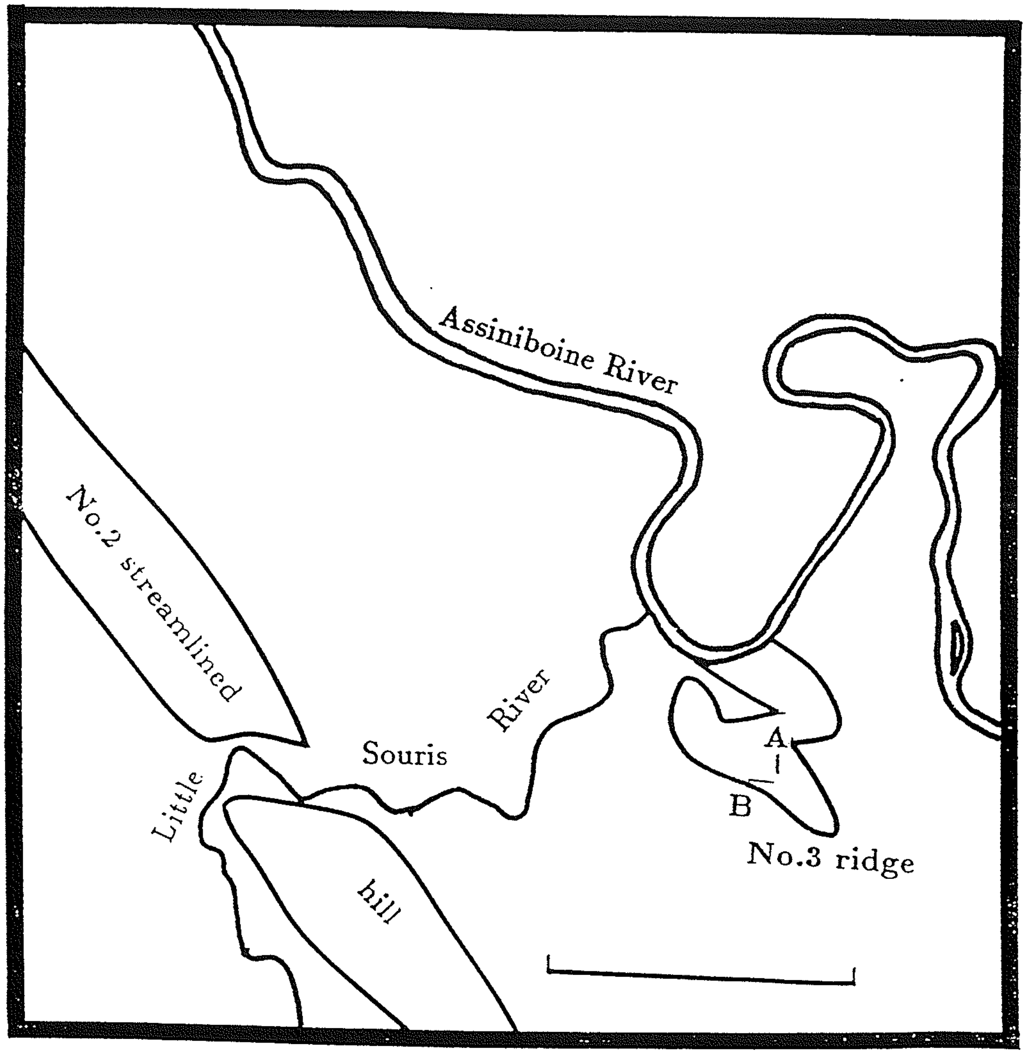


Fig. 4-12 Index for sections in the the No.3 ridge (hill). Section 'A' is oriented roughly N-S (Fig. 4-13), and section 'B' is oriented W-E (Fig. 4-14). Location of the no.3 ridge is shown as E-6 in Fig 1-8.

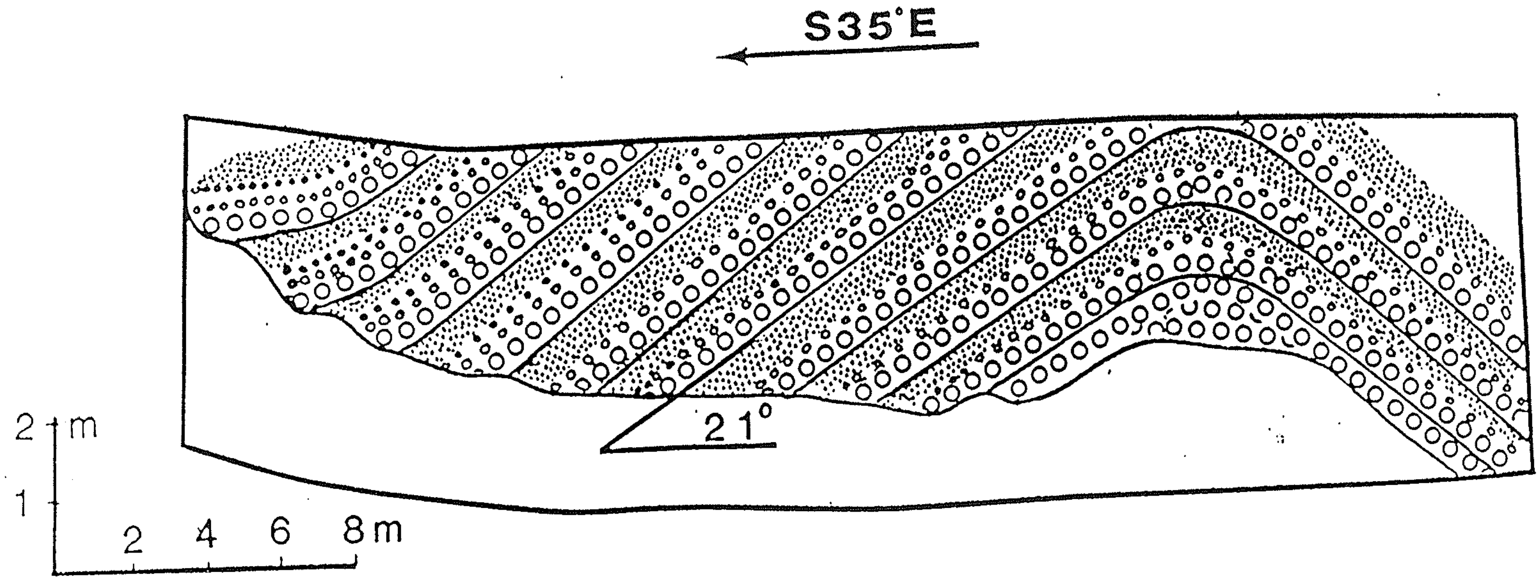


Fig. 4-13. A geological section exposed in pit 1 (A in Fig. 4-12) at SW of Sec.21, Tp.9, R.17W. Cross beds consist of well sorted boulders and cobbles in the lower portion overlain by poorly sorted pebbles and granules in the upper portion (Facies Gx) .

grain size of 1.5 cm, and are poorly sorted, non-graded, and matrix supported. The upper 0.5 m of this section has been removed by quarrying activities. The lower contact is concealed by slump.

Section B is located at the southern flank of the No.3 ridge (B in Fig. 4-12). It is oriented west-east and exposes 2 units. The upper unit is composed of massive sandy gravels that are poorly sorted, non-graded, matrix supported with random fabric (Facies Gm) (Fig. 4-14). The maximum gravel size is 15 cm and averages about 2 cm. The lower unit in this section consists of poorly sorted, crudely graded, and cross stratified pebbles, granules, and sand (Gx in Fig. 4-14). Granules and sand are the dominant clast components. Cross stratification is outlined by continuous laminae of sand and granules, while the basal cobble layer in each set of cross beds is very thin and discontinuous. Some cobbles float in sand and granules as outsized clasts.

Exposure 7 (E-7), No.4 ridge section

The No.4 ridge is located north of the Assiniboine River, 8 km west of Shilo (E-7 in Figs. 1-8 and 3-2). The ridge is oriented 100°E. The elevation of this ridge is about 372 m at the west end and rises slightly eastward. A gravel pit in the western end of the ridge exposes a 300-m-long and 4 to 6-m-high section. Four major sedimentary units have been identified from the exposed sections in combination with several radar lines.

The lowest unit (unit 1) can be seen on the radar profile E (Fig. 4-16, E in Fig. 4-15). This unit is exposed 0.5 m above the pit floor (3.5 m below the ground) and extends to a depth of at least 8 m below the pit floor. The upper 1.5 m of this unit was examined in the wall of the pit and in a small 1-m-deep hole in the floor of the gravel pit. Sediments are non-stratified, poorly sorted, consist of 70 to 75

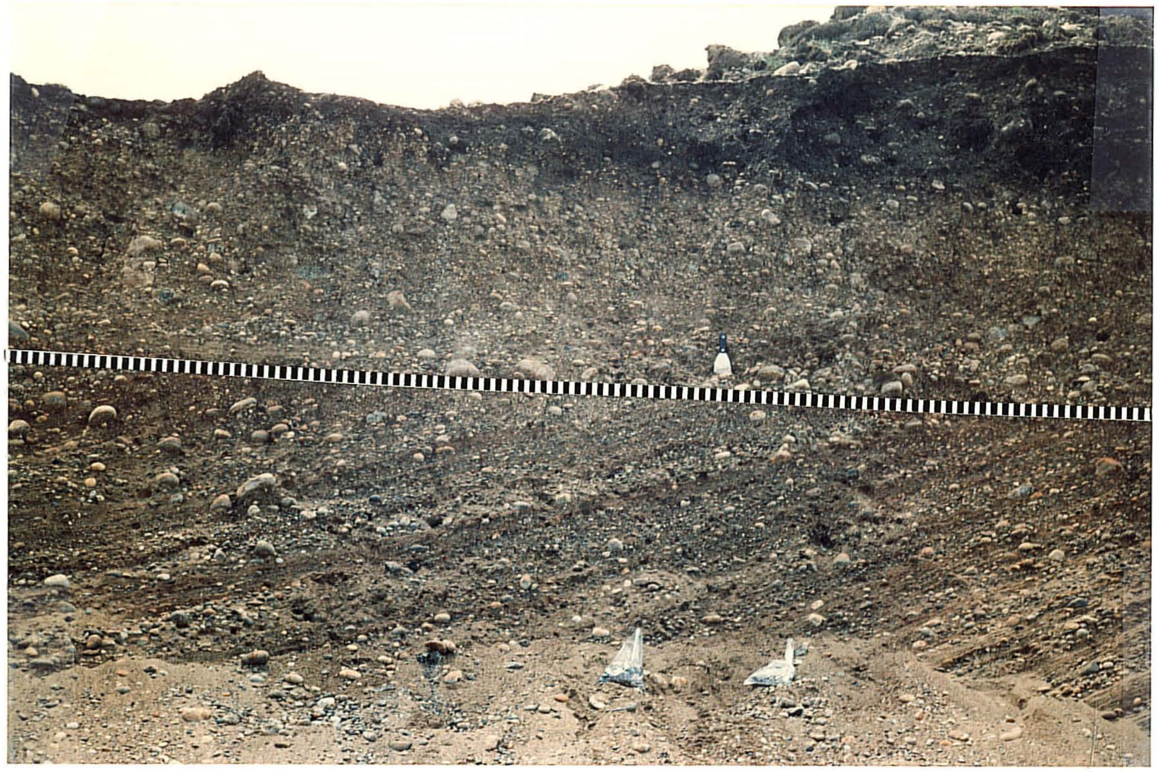


Fig. 4-14. A west-east oriented exposure at the southern flank of the No.3 ridge (B in Fig. 4-12). The exposure consists of massive, non-graded sandy gravels (Facies Gm) that is overlain by cross bedded granules and sand with some outsized cobbles (Facies Gx). The scraper is about 0.25 m long.

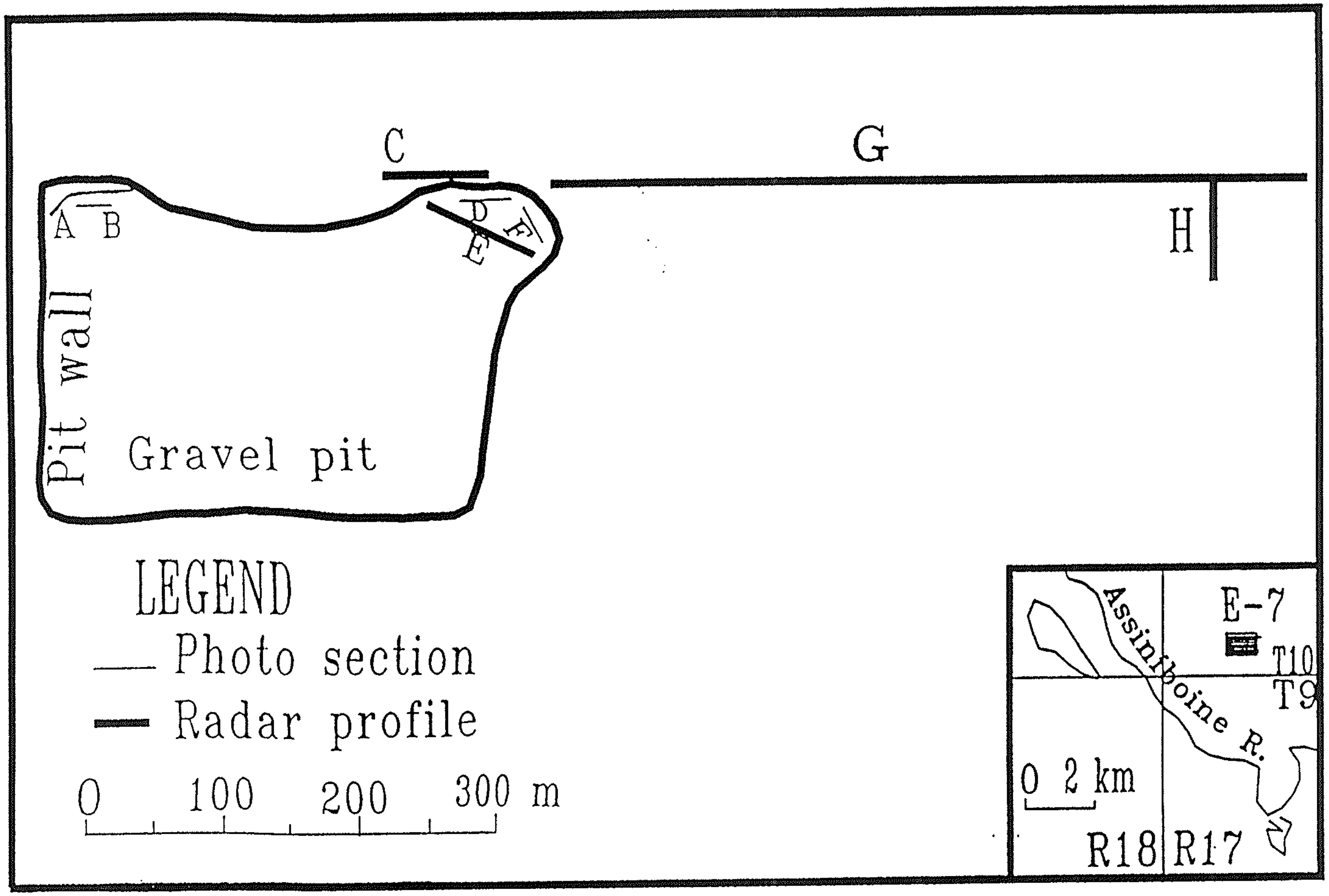


Fig. 4-15. Index of radar lines and photos in the gravel pit that is located at the No.4 ridge (E-7 in Fig. 1-8) at W$\frac{1}{2}$, Sec.5, Tp 10, R 17W.

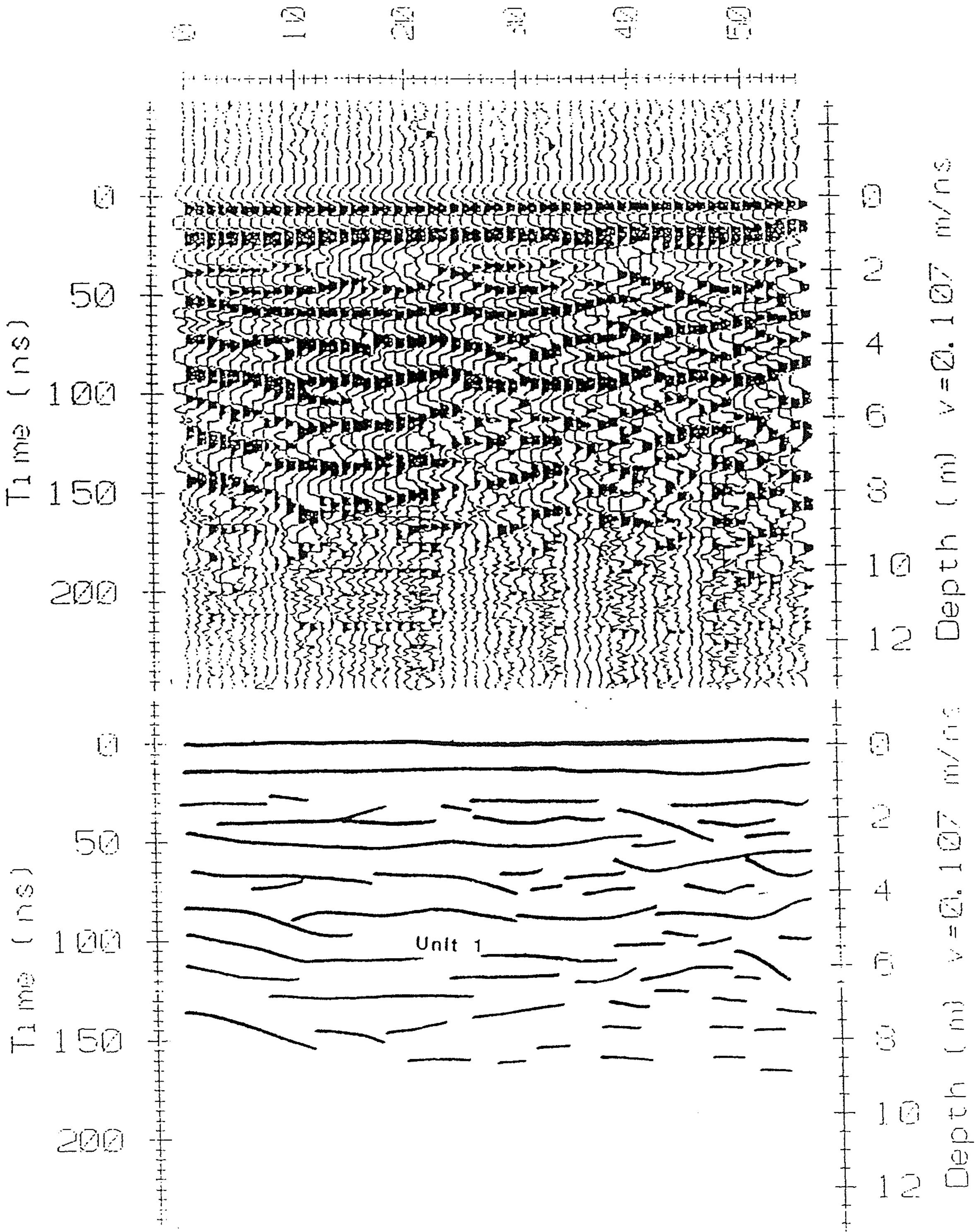


Fig. 4-16 Radar profile E across the floor of the eastern gravel pit in the No.4 ridge. This radar profile is oriented N80°E (E in Fig. 4-15).

percent sand, 15 to 20 percent pebbles and granules, and less than 2 percent lignite fragments. The maximum gravel size is 20 cm in diameter with an average of about 4 cm. On radar profile, this unit (unit 1) is at least 8 m thick, and is dominated by generally flat and relatively discontinuous internal reflections.

Unit 2 is about 4 to 6 m thick and is exposed in a 300 m-long exposure (west to east) in this gravel pit. This unit is cross bedded and graded, consists of cobbles to boulders grading upward to pebbles and granules in each cross bed (Figs. 4-17, 4-18, and 4-19; from lines A, B, and D in Fig. 4-15). The maximum size of clasts and thickness of the cross beds vary from the western end to the eastern end of the gravel pit. In the western end, the maximum clast size is 40 cm with an average of 8 cm in diameter. Each cross bed is about 2 to 6 m thick (Fig. 4-18). The basal boulder and cobble bed (1/3 to 1/2 the thickness of each cross bed) is very well graded, well sorted, and clast supported with an open framework (Facies Gx) (Fig. 4-18). In contrast the upper part of each cross bed is crudely stratified, poorly sorted, matrix supported, and contains outsized clasts (Fig. 4-18), but lacks a finer sand tail.

At the eastern end of this gravel pit, unit 2 is cross bedded, and consists of cobbles that fine across the cross-beds to pebbles and granules with sand matrix (Facies Gx) (Fig. 4-19). Cross beds change eastward gradually from steeply inclined (Fig. 4-19) to gently inclined (upper right in Fig. 21, F in Fig. 4-15). In order to compare with the outcrop in Fig. 4-19, a west-east radar line (Fig. 4-20, C in Fig. 4-15) was surveyed along the edge of the eastern gravel pit on the west (right) of the pile of gravel shown in Fig. 4-19. Major cross bedded units in the outcrop in Fig. 4-19 were detected by the radar survey (Fig. 4-20).

Unit 3, which overlies unit 2 sharply and is exposed in the pit wall (lower photo

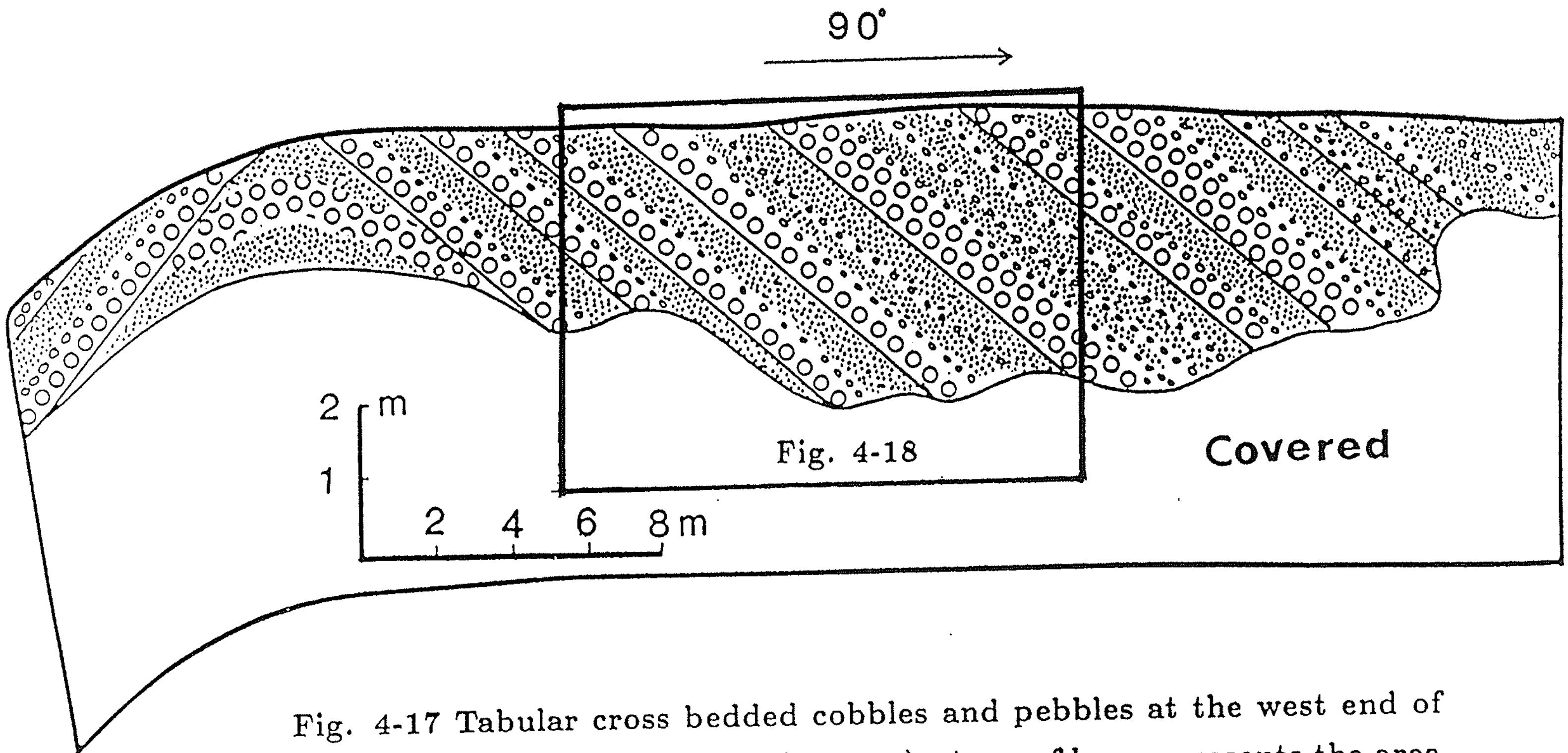


Fig. 4-17 Tabular cross bedded cobbles and pebbles at the west end of the No.4 ridge (A in Fig. 4-15). Area of box represents the area covered by Fig.4-18.

Fig. 4-18 Cross beds of unit 2 in west end of no.4 ridge (B in Fig. 4-15), showing graded nature of each bed (Facies Gx). Notice that cobbles in the lower portion of each cross bed are better sorted than pebbles and granules in the upper portion. The lower photo shows an detailed lower portion of a cross bed. The shovel is 0.74 meters.

Fig. 4-19 Cross bedded unit 2 at the eastern end of the gravel pit in the No.4 ridge (D in Fig 4-15), 300 m east of Fig. 4-18. a: Gravels become smaller and the basal gravels become thiner eastward (compare to that in Fig.4-18). b: a broad view of the eastern gravel pit.

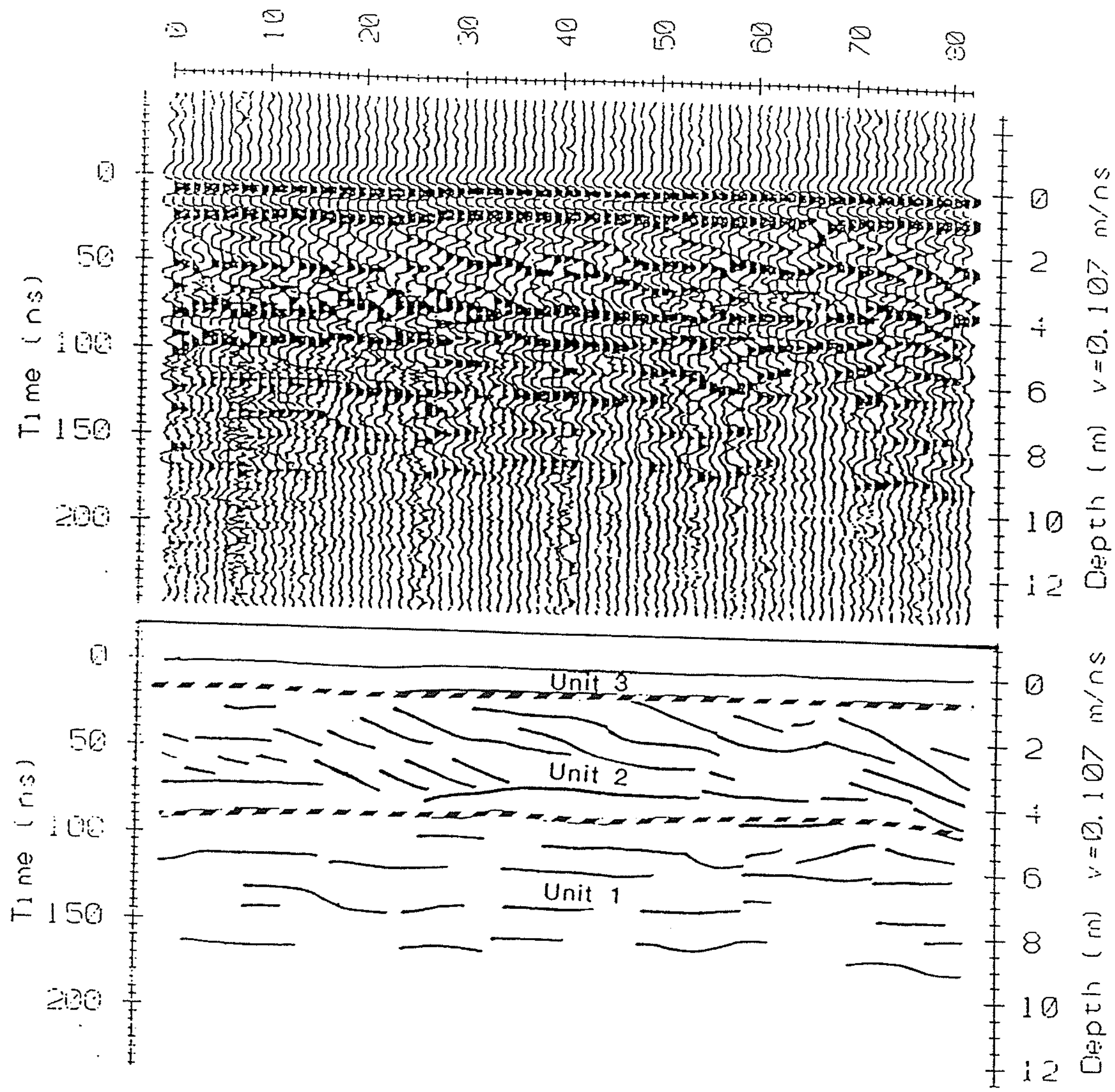


Fig. 4-20 A radar profile (C in Fig. 4-15) along the upper surface of the eastern gravel pit in the No.4 ridge. Unit 2 in this profile is indicated by steep cross beds. 40 to 80 m duplicated the left part (left of the van) of Fig. 4-19 b.

Fig. 4-21 A NW-SE exposure across the eastern gravel pit in the No.4 ridge (F in Fig 4-15). Notice that the gravel beds in the upper left are steep inclined, whereas a gravel bed in the upper right of the photo is gently inclined. See the gradational changes from tabular cross bedded unit to flat bedded unit in Fig 4-22 (40-60 m).

in Fig. 4-19) as well as in the radar profile (Fig. 4-20), is about 0.5 to 2 m thick and consists of poorly sorted and non-graded granules and sand.

To the east of the gravel pit, a radar profile (Fig. 4-22, in back pocket, G in Fig. 4-15) reveals the continuation of unit 2 from 0 to 60 m at the depth of 1.5 to 6 m. Farther to the east, the tabular cross bedded unit (unit 2, Facies Gx) is replaced by unit AA, which is overlain by unit BB. Unit AA consists of continuous horizontal beds (Figs. 4-22, 4-23). Unit BB overlies unit AA, and consists of small scale bi-directional cross beds in barrier shape geometry (BB1, BB2 in Fig. 4-22), unidirectional cross beds (BB3) with a channel shape geometry (Fig. 4-22), and small scale cross beds and flat beds in between (BB4). The contact between unit 2 and BB is gradational over 15 m (Fig. 4-22).

Exposure 8 (E-8), the Chater section

This exposure (E-8 in Fig. 3-2) is located at the southeastern corner, Sec.34 T10 R17, at an elevation of 375 m. The section, which was 3 m high and 100 m long, exposes 3 cross bedded fine gravel units that are separated by two cobble and boulder beds (Fig. 4-24). The cross bedded units are 0.3 to 1 m thick, and consist of fine pebbles and granules. The cobble and boulder beds are about one clast thick.

Exposures, River Cuts and Bore-hole Cores in the Middle Fan Delta (zone 3)

Exposure 9, north of Carberry section

Most exposures in the middle fan delta area are shallow, although Exposure 9 (E-9 in Fig. 1-8) is excellent. It is a road cut extending as a vertical face for about 100 m at a cross section road 12 km north of Caberry. The sediment sequence is described in Table 3.

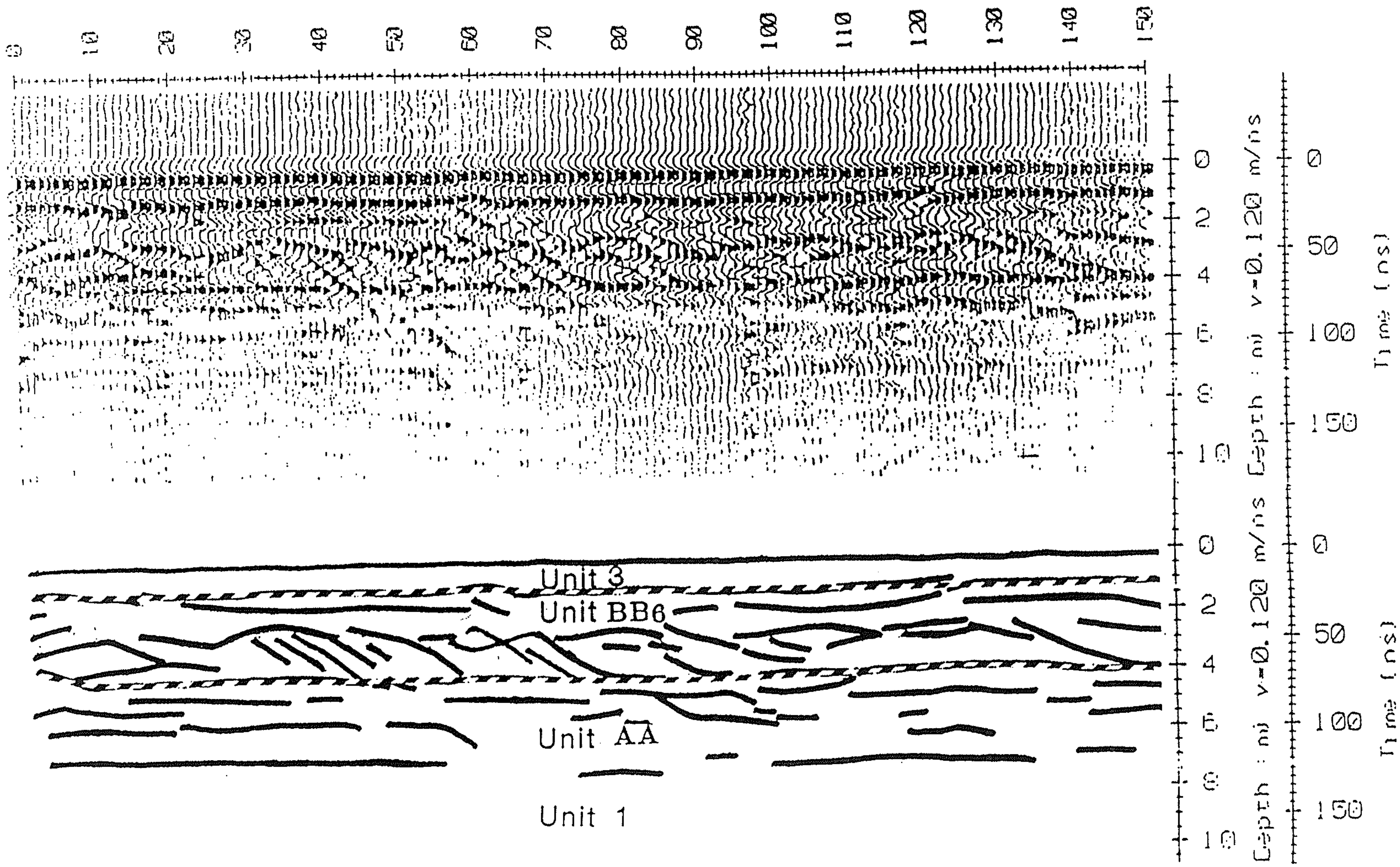


Fig. 4-23 A radar profile (H in Fig. 4-15) that is perpendicular to the radar profile in Fig. 4-22. Notice that cross beds (subunit BB6) overlie flat beds (unit AA) at the depth from 1 to 4 m.

Fig. 4-24 A section exposed 600 m north of Chater at $SE\frac{1}{4}$ of Sec.34, Tp10, R18W (E-8 in Fig. 1-8). Notice 3 sets of cross bedded sand are separated by 2 one clast-thick cobble layers. Is this a fan delta top set deposits.

Table 3, Stratigraphic description of north of Carberry section (Exposure 9) at the northeast corner of Sec.13 T12 R14 W (after Teller, 1990)

Depth (m)	Description
0 - 0.6	Vfg silty sand, black; non-laminated (reworked soil); abrupt flat basal contact.
0.6-1.0	Silty clay to silty vfg sand, tan-orange brown, no visible laminae (weathered); abrupt basal contact.
1.0-1.47	Silty clay, dark grey, no visible laminae, occasional less than 0.2 mm nodules $CaCO_3$; lower boundary gradational into the underlying unit.
1.47-1.81	Silty vfg sand, excellent scour and fill structures throughout (mostly less than 40 cm), becoming more silty and clayey in the basal 6 cm and more horizontally laminated; basal contact flat and abrupt.
1.81-2.91	Very fine sand to silty sand; occasional horizontal clayey silt laminae truncate excellent ripple cross stratified sub-units with some 1.5 to 2 cm high symmetrical ripples (each about 6.5 cm long) and climbing in drift (at 13°) ripples dipping south. Basal 17 cm is flat bedded vfg to fg sand with some clayey silt laminae(less than 1 mm thick); basal contact is sharp but with some relief.
2.91-3.81	Clayey silt to silty clay, laminated with some cross lamination (< 2 cm thick) dipping west, and tabular and lensoid unit in the lower half; basal contact flat and distinct with up to 1 cm of clay grain rip-ups in sandy silt.

3.81-4.31	Clayey silt to silty clay, mainly horizontally laminated; 4 cm thick clean fg-mg sand bed 12 cm up from the base of the unit; basal contact abrupt with laminated clayey silty over fg-mg rippled clean sand (14 cm spacing between the cress of the symmetrical ripples).
4.31-4.81	Sand, fg to mg; horizontally laminated, well sorted; basal contact not exposed.

River cuts

A series of river cuts occur downstream from the Souris River junction. Four of the river cuts will be described below. These four river cuts are: RC-1 (River Cut 1), the Souris River junction section; RC-2, Treesbank section; RC-5, Glenboro section; RC-6, Provincial Park section.

RC – 1, The Souris River junction section. This river cut, located along the northern bank of the Assiniboine River near the Souris River junction (RC-1 in Fig. 1-8), consists of a bed of sandy gravel overlain by interbeds of clay and sand that are capped by thick fine grained sand. The sandy gravel bed, which overlies till, is about 0.7 m thick, non-graded, and very poor sorted (Fig. 4-25). The upper and lower contacts are sharp. Overlying the sandy gravel bed are 3 clay beds interbedded with 2 sand beds. Each of the clay beds is about 0.5 to 0.9 m thick, and consist of nonbedded, dark blue clay or silty clay (Fig. 4-25). The sand beds are relatively thin (0.3 to 0.5 m thick) and become gradually finer upward from massive fine to medium grained sand ($Mz=1.7\phi$) to silty fine sand. Sediment above this is about 5 m thick, and consists of silty fine sand.

RC – 2, Treesbank section. A 2.5 m section (Fig. 4-26) is located on the north bank of the Assiniboine River valley, 500 m downstream from the Souris

Fig. 4-25. A river cut north of the the Souris River junction (RC-1 in Fig 1-8). The massive, poorly sorted sandy gravels (right) (Facies Gm) is overlain by interbeds of thin fine to medium sand and thick clay (left) (Facies Fs).

Fig. 4-26. A photo of the Treesbank River cut on the northern bank of the Assiniboine River 500 m down stream from the Souris River junction. The basal unit is composed of nonbedded coarse grained sand. The middle unit consists of nonbedded, and poorly sorted sandy gravel. The upper unit is composed of silty very fine sand with abundant wood fragments.

River junction section, about 3.2 km east and 3.2 km north of Treesbank. The top of this cut lies 7.5 m below the terrace at an elevation of 343 m (1125 ft.). The detailed description of this section is in Table 4.

Table 4, Stratigraphic description of the Treesbank section (RC-2) at SE$\frac{1}{4}$, Sec.22 Tp8 R16W

Depth (m)	Description
0 - 3.5	Fine grained eolian sand, slumping.
3.5-3.9	Silt to very fine grained sand that is flat bedded and is rich in wood (20 to 30%). Tree trunks up to 20 cm long, 5 cm wide and 2 cm thick lay horizontally in beds. The lower contact is gradational.
3.9-4.5	Silt and very fine sand that is rich in wood fragments (10-20%), no laminae. The lower contact is sharp. A bone (bison?) collected at the toe of the section is thought to be from this bed.
4.5-4.75	Massive, non-bedded, non-graded and poorly sorted sandy gravel, matrix supported. Gravel size ranges from 0.8 to 3 cm, average 1.5 cm. Lower contact is sharp and erosional.
4.75-6.0	Medium to coarse grained sand that is dominated by medium, grained sand ($Mz=1.3\phi$) that is moderately well sorted ($\sigma = 0.71\phi$), nonstratified. Lower contact of this unit is concealed by slump.
6.0-12.5	Slump covered, till is exposed near the river floor.
	The River floor is at an elevation of about 330.5 m.

Barendsen and others (1957) described a similar river cut from which a piece of wood was collected and dated to 9110 ±110 BP (Barendsen et al., 1957; Y-415)

> "Wood from base of 15-ft layer of silty sand containing bones (bison?) and wood, underlain by till and by 8 to 10 ft of gravel and sand deposited on eroded surface of till; overlain by about 25 ft of sand containing a fossil soil horizon and eolian at the surface. Section exposed at 1100 to 1125 ft altitude on the north bank of the Assiniboine River about 1 mi east of (downstream from) the mouth of the Souris River, NE$\frac{1}{4}$ Sec.15, T.8, R.16, W. Prin., 2 mi north and 2 mile east of Treesbank. The wood bearing sand was tentatively interpreted as part of the Assiniboine delta laid down in the Lake Agassiz."

The section described by Barendsen and others (1957) is believed to be the same as my section. However, Barendsen's location is confusing: NE$\frac{1}{4}$ of Sec.15, Tp8, R16W, which is 2 mi north and 2 mi east of the Treesbank, is 800 m from the Souris River mouth, not 1 mile (see Fig. 4-27). Furthermore, the Assiniboine River floor at NE$\frac{1}{4}$ Sec.15, Tp8, R16W lies at an elevation lower than 1100 ft.(335.3 m), while the terrace I described does lie at the elevation of 1125 ft (342.9 m). The section I studied is located at SE$\frac{1}{4}$ Sec.22, Tp8, R16W and is the only section that exposes wood fragments in that area today. Therefore, one possibility is that the section I described is the same section described by Barendsen and others (1957). The second possibility is that they are different sections and the section described by Barendsen and others (1957) was destroyed due to the northward shifting of the Assiniboine River. I think the first possibility (Barendsen's section is the same as my section) is high. If this is correct, then the dated wood comes from my unit at the depth of 3.5 to 4.5 m.

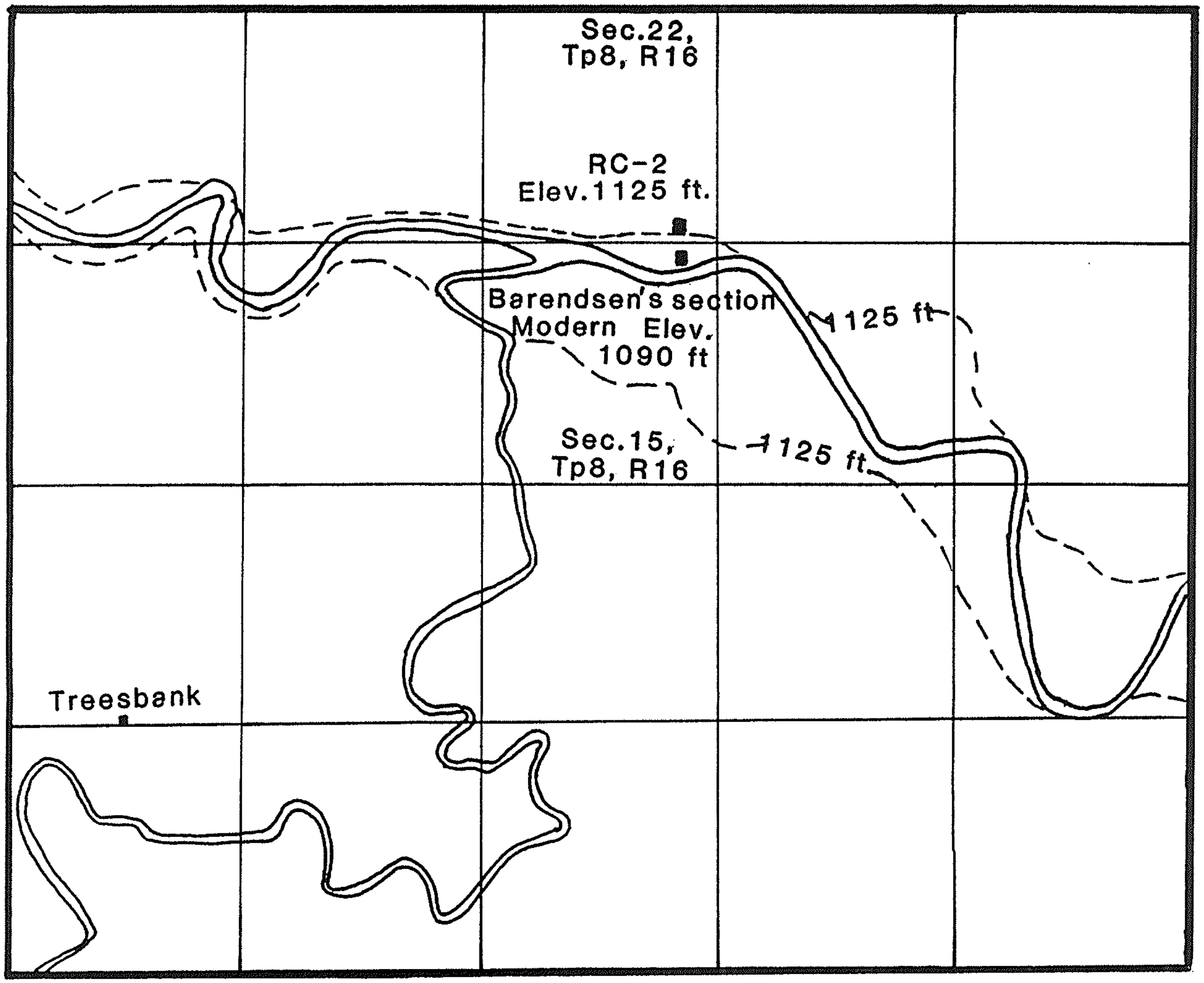


Fig. 4-27 Location of the Treesbank section (RC-2 in Fig 1-8).

RC – 3, West of Glenboro section. This section is located 6.5 km west and 6 km north of Glenboro at N$\frac{1}{2}$, Sec.30, Tp7, R14W (RC-3 in Fig. 1-8). The surface of this section is at the terrace level at an elevation of 343 m (1120 ft.). The importance of this section is that the delta sequence is a fining upward sequence that is underlain by a 4 m blue clay bed. The base of the delta sequence is dominated by medium sand (4 m thick) that fines upward to cross bedded, well sorted fine sand (1.5 m thick), and to silty clay that is hard and laminated (1.5 m thick).

RC – 5, Glenboro section. The Glenboro section, which was described by Teller in 1974 (Teller, 1980) (Table 6), is located 16 km north of Glenboro (RC-5 in Fig. 1-8). The top of this section lies at the terrace level at an elevation of 335 m.

Table 5, Stratigraphic description of the Glenboro section (RC-5) at S$\frac{1}{2}$, Sec.26, Tp8, R14W (after Teller, 1980)

Thickness	Stratigraphic Description
0.6-0.9m.	Sandy silt to silty sand, some pebbles (1.5cm max. diameter), some snail shells, massive, columnar jointing, heavy $CaCO_3$ precipitation in lower half (forming a nodular to undulating band), silty sand pockets (burrows) with pebbles throughout and down into underlying unit, lower boundary gradational over 0.3 m.
0.6-1.8m.	Sand, fine to medium grained, cross bedded in 8 cm to 11 cm thick units (dipping 30°E), coarse grained sand, granules and pebbles of gray shale oriented parallel with bedding, some flat beds, some cut and fill structures, abundant (irregular) iron staining, rare black organic material

and pieces of charcoal in lower 0.3 m, several thin (< 0.1 m) laminated and cross laminated clayey silt to sandy silt lenses with finely disseminated black organic material and broken mollusc shells in lower 0.6 m.

1.5 m. Sand, medium to very coarse-grained, rounded gray shale, carbonate, chalcedony, and granitic pebbles and granules abundant, bedded and cross bedded, several discontinuous thin sandy, pebble-rich beds containing predominantly granitic pebbles in sandier upper half, some cut and fill features, several laminated clayey silt and lenses with organic material inter-laminated in places; lower 1 meter coarser (gravelly and less distinctly bedded, some iron staining near base; 1.2 meter long (0.12 m in diameter) log 0.91 m from base of unit dated 9880 $\pm$ 225 YR BP (GX-3696).

10.7m. Clayey silt grading downward into silty clay in lower 3 m, some siltier zones in upper part (several cross laminated), laminated throughout (dark gray clay to light gray silt), several thin (< 0.5 cm) laminae with tan, silty grains and carbonate grains, scattered tan granules of silty to sandy material, rare limestone and granitic clasts and pebbles scattered throughout (more common in lower part), upper boundary sharp and flat with springs, distinct "units" across outcrop face but not recognized in detail, some laminae contorted (but overlain by horizontal laminae)

River At an elevation of 320 m

RC – 6, Provincial Park section

This section is located at the western Spruce Wood Provincial Park, within the Epinette Creek valley in the middle of Sec.14, Tp9, R14W (RC-6 in Fig. 1-8). The bottom of the cut is at an elevation of about 347 m (1140 ft). The upper part of the section is covered by slump, and only the lower 5 m are well exposed. The lower 3 m of the exposure is composed of well sorted and irregular rippled silty fine sand that changes upward to regularly rippled silty fine sand (facies Sl) (Fig. 4-28). The upper 2 m of this section is composed of sub-horizontally laminated silty very fine grained sand ($Mz=3.22\phi$) that is well sorted ($\sigma=0.5$), flat bedded, with short vertical joints filled by caliche from the soil zone (Fig. 4-29).

Borehole Sediments

Introduction. Two new wells (90AD-1, 90AD-2) were drilled in the central Assiniboine fan delta. The locations of the wells were chosen to obtain data from the Douglas Marsh area (along the abandoned Brandon-Chater-Douglas Station channel), and to obtain data from the center of the delta. Another well (ADA-50), which was drilled by P.F.R.A (1986), will also be described. These cores are important because they provide data about the sedimentary texture and structure at depth in the central fan delta area.

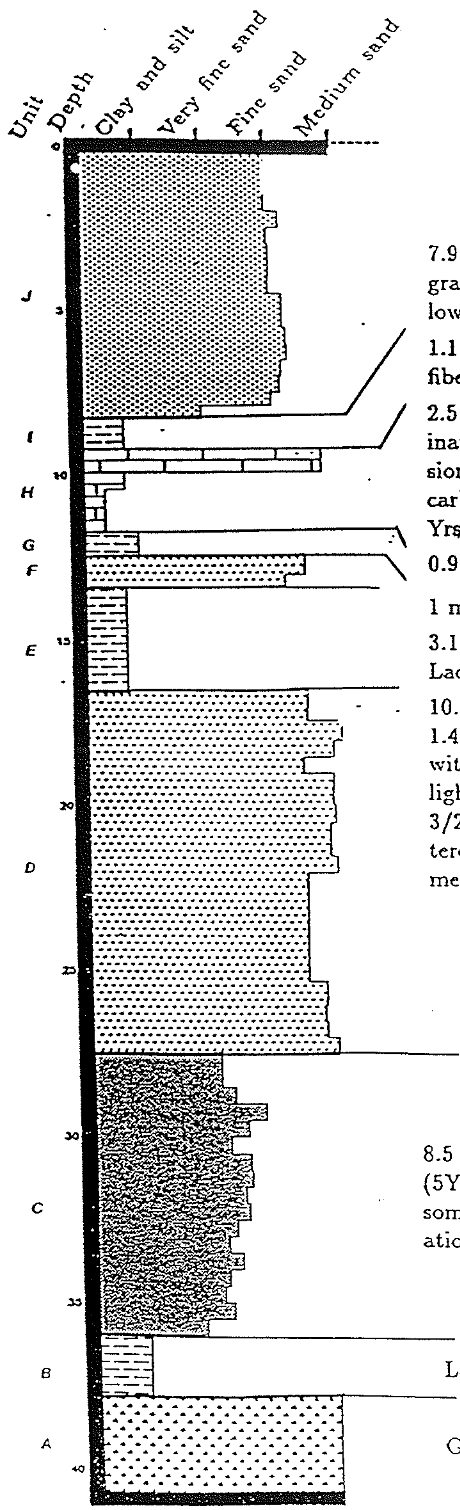
Well 90AD – 1. Well 90AD-1, which is located 9.1 km east and 4 km north of Shilo (90AD-1 in Fig. 1-8), consists of 10 units labeled A, B, C, D, E, F, G, H, I, and J. The full description of this core can be found in Appendix III (p.158-164), and an abbreviated description is presented in Fig. 4-30. Because the process of core recovery involved vibrating motion, the finer sedimentary structures (such as ripples and laminations) of the silty and sandy units were commonly destroyed.

The core of 90AD-1 is essentially composed of 4 parts.

1. Till is the lowest unit

Fig. 4-28 Rippled silty fine sands along the Epinette Creek. Notice asymmetrical ripples in the lower photo from the lower part of the cut and the symmetrical ripples (Facies Sl) in the upper photo from middle part of the cut. Location of this cut is shown as RC-6 in Fig 1-8.

Fig. 4-29 Flat-bedded silty fine sand (Facies Sl) in the upper part of the the Epinette Creek cut (RC-6 in Fig. 1-8), east of Highway 5. Notice white $CaCO_3$ deposits along bedding planes and joint or root channels in the upper part.



Stratigraphic Descriptions

7.9 m fine to medium grained sand, no distinct bedding, dark grayish brown (2.5Y 4/2) to olive brown(25Y 4/4), with sharp lower contact.

1.1 m clayey silts, slightly sandy with abundant plant fragments, fibers, small gastropods and shell fragments.

2.5 m calcium carbonate and marls intercalating with silt lamination, with abundant plant fibers, wood fragments and occasional gastropods and shell fragments. Radiocarbon dating on carbonate at 9.5m and on wood at 11.2 m yielded 9330 ± 160 Yrs BP (1253Cc) and 9340 ±70 Yrs BP (TO2197) respectively.

0.9 m Laminated sandy silt, clayey, organic-rich.

1 m fine to medium sand, moderate sorted ($\sigma > 0.8$).

3.1 m dark gray(2.5Y 4/1) silty clay, no visible lamination, Lacking organic material.

10.5 m massive coarse to medium grain sands ($Mz=0.9\phi$ to 1.45ϕ, facies Sm), well sorted to poorly sorted ($\sigma=0.46$ to 1.04), with scattered pebbles and granule, angular to sub-angular, light olive gray(5Y 6/2), olive(5Y 5/4) and dark olive gray(5Y 3/2). Shale fragments present throughout (15-30%), but scattered coal fragments (less than 2%) occur only in the upper 1 meters. Both the lower and upper contacts are sharp.

8.5 m laminated silty sands, olive gray (5Y 5/2) to pale olive (5Y 6/4), interbedded with fine sands and shale enriched sands, some small clay-balls. mean sizes=2.1ϕ to 3.1ϕ, standard deviations=0.45 to 1.1.

Laminated Lacustrine silty clay.

Gravelly sandy clayey till

Fig. 4-30 Stratigraphic column of 90AD-1. Location of this well is shown in Fig. 1-8. Detailed size analyses on sand units in Figs. 4-31 and 4-32.

2. The sequence overlying till is a coarsening upward sequence from lacustrine silty clay (Unit B, Facies Fm), to laminated fine sand (unit C, Facies Sl), and to massive, medium to coarse grained sand (unit D, Facies Sm) (Figs. 4-30, 4-31). The contact between laminated silty clay and stratified fine sand is gradational, whereas the contact between stratified fine sand and coarse sand is sharp.

3. Lacustrine silty clay and marl (Units E, G, H, and I), which are intercalated with a 0.9 m sand bed (unit F), abruptly overlie the coarse grained delta sand (Unit D), and are overlain by fine to medium grained sand (unit J). The organic content the in silty clay and marl increases upward from Unit E to Units H and I.

4. The top 7.9 m is fine to medium grained sand (unit J, Facies Sm) that has no visible lamination and is moderately well sorted.

5. Although unit D is coarser than unit J, and much coarser than unit C (Fig. 4-31), the standard deviation of the 3 units fall in the same range between 0.46 to 1.09 ϕ (Fig. 4-32).

Well 90AD – 2. Well 90AD-2 is located in SE$\frac{1}{4}$ of NE$\frac{1}{4}$, Sec.10, Tp10, R14W (90AD-2 in Fig. 1-8), about 5 km south and 5 km east of Carberry. Based on sedimentary structures and textures, the sequence can be divided into 8 labelled units: A, B, C, D, E, F G, and H (Fig. 4-33). This core is described in detail in Appendix III (Page 165-172) and its main features are described in Fig. 4-33. As in vibracore 90AD-1, many of the fine sedimentary structures (e.g. laminae and ripples) in the silt and sand sized units have been destroyed by the vibrating method of core recovery. A brief summary of the core 90AD-2 is as following:

1. This core consists of two major coarsening-upward sequences: The lower sequence

Fig. 4-31 Mean Size Versus Depth of Sand in

Core 90AD-1, Northeast of Shilo

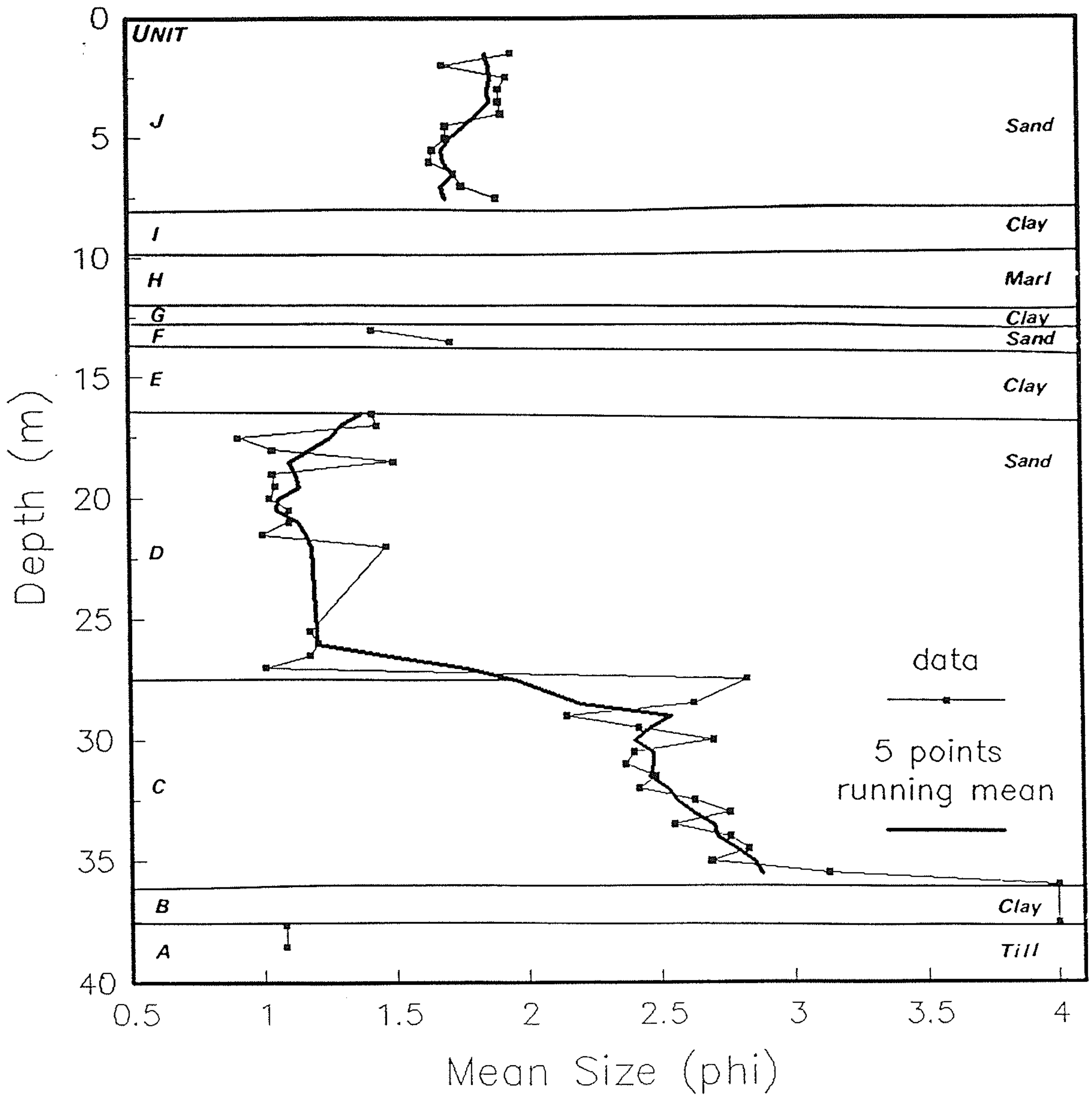
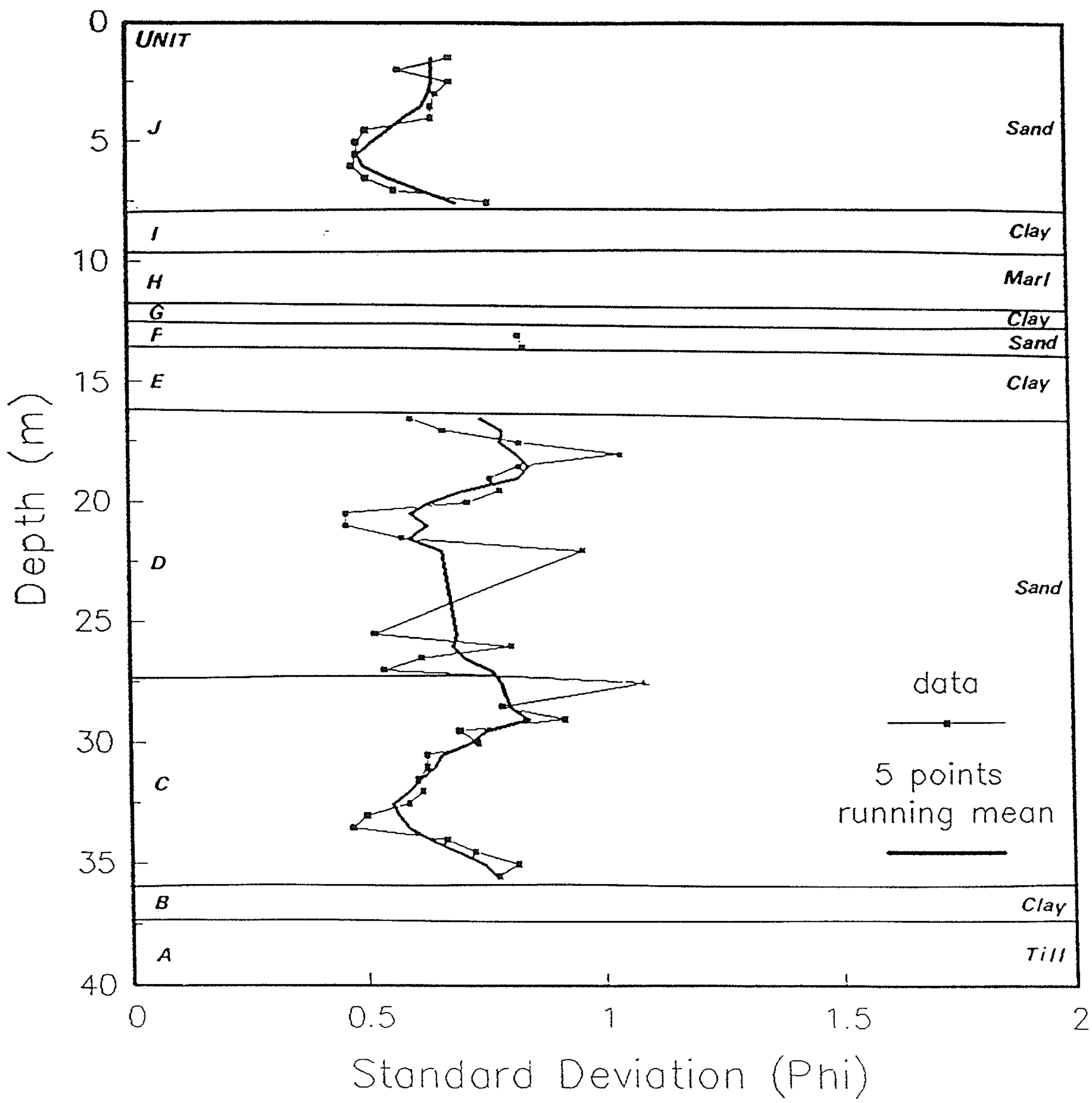


Fig. 4-32 Standard Deviation Versus Depth of Sand in Core 90AD-1, Northeast of Shilo



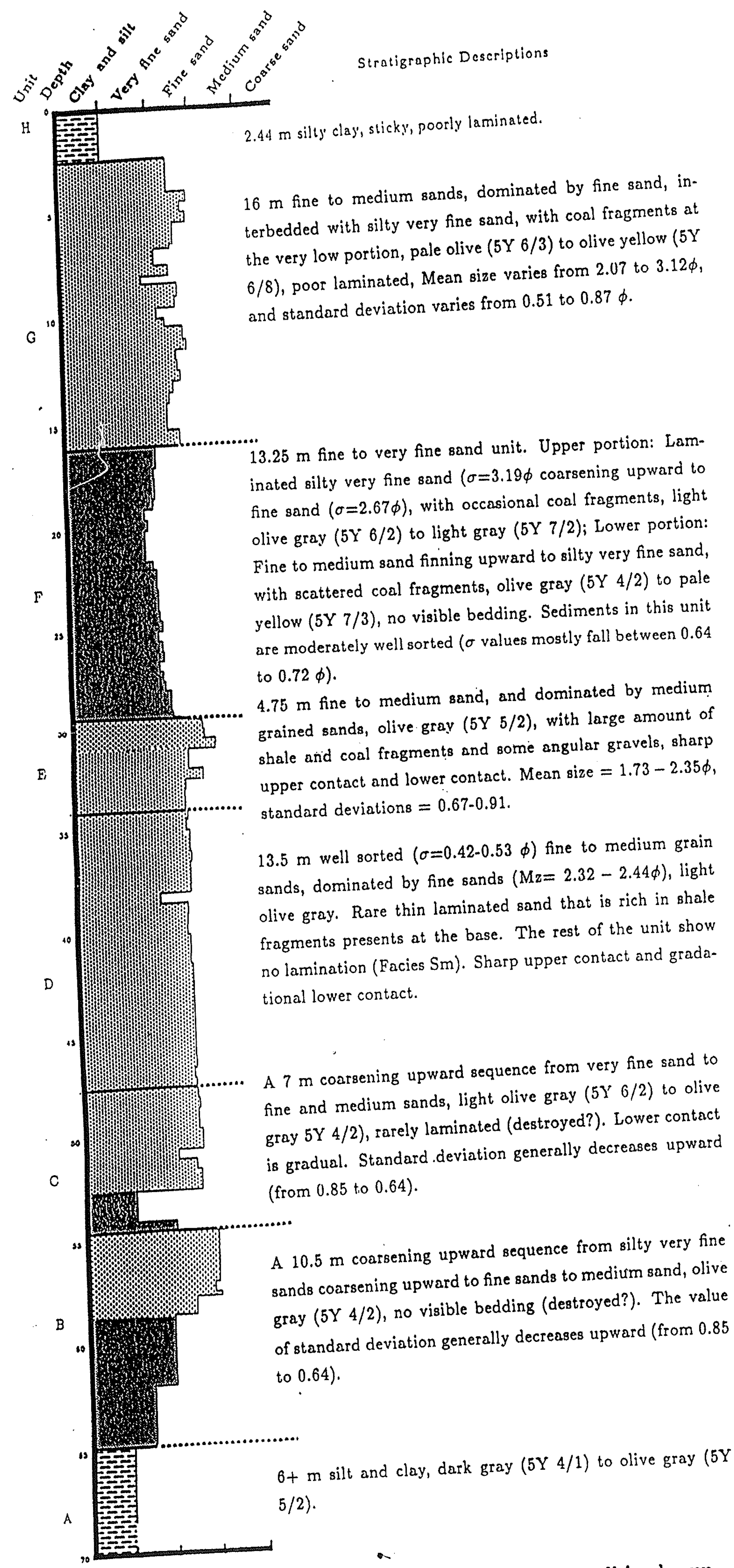


Fig. 4-33 Stratigraphic column of 90AD-2. Location of this well is shown in Fig. 1-8. Detailed size analyses on sand units in Figs. 4-34 and 4-35.

consists of at least 2 coarsening upward subcycles, one from unit A to unit B, the other one from unit C to E (Figs. 4-33, 4-34).

2. The second coarsening upward sequence is from the upper 1/3 of Unit F to the overlying Unit G (from fine-very fine sand to fine-medium sand). The lower 2/3 of Unit F is a fining upward sequence.

3. The uppermost unit (H) consists of silty clay that is poorly sorted.

4. Sediments are moderatly sorted in units B and C, and become well sorted in unit D (Fig. 4-35). Sediments in unit E and unit F are moderatly sorted (Fig. 4-35).

Well ADA – 50. Well ADA-50 was drilled by Prairie Farm Rehabilitation Administration at Sec.2, Tp10, R17W (ADA-50 in Fig. 1-8), and is described in detail in Appendix III (P.173). A brief summary is presented below.

1. Sandy gravel sharply overlies at least 1.2 m of silty clay at a depth of 14.25 m.

2. Core from 14.25 to 3 m consists of an overall fining upward sequence, but individual beds range from sandy gravels to fine-medium sand (in contrast to the coarsening upward sequence in cores of 90AD-1 and 2). Sediments are poorly sorted near the bottom of the core and get better sorted upward (see Fig. 4-36).

3. The top 3 m of the ADA-50 consists of fine-medium sand coarsening upward to gravelly medium-coarse sand.

Fig. 4-34 Mean Size Versus Depth of Sand in Core 90AD-2, South of Carberry

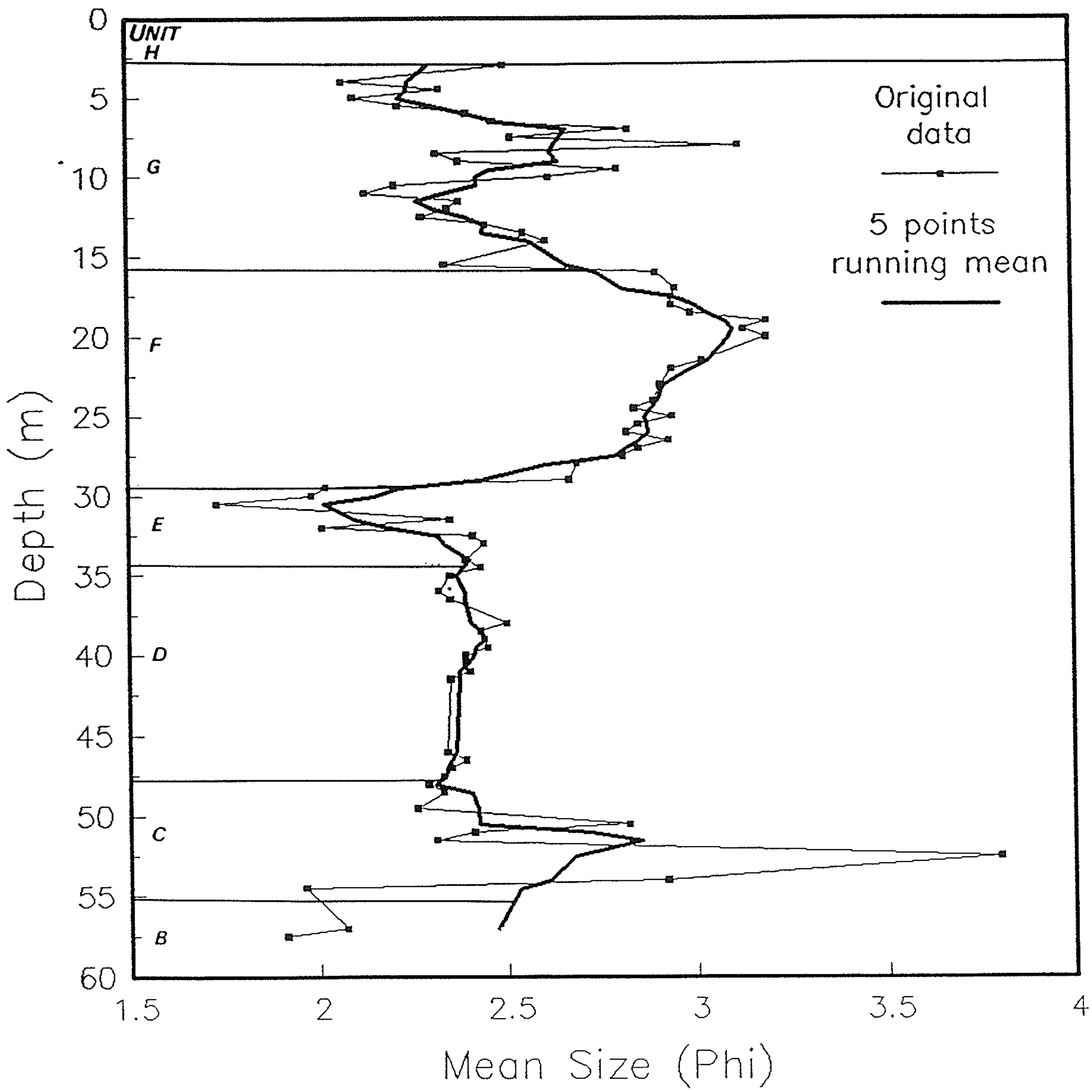


Fig. 4-35 Standard Deviation Versus Depth of Sand in Core 90AD-2, South of Carberry

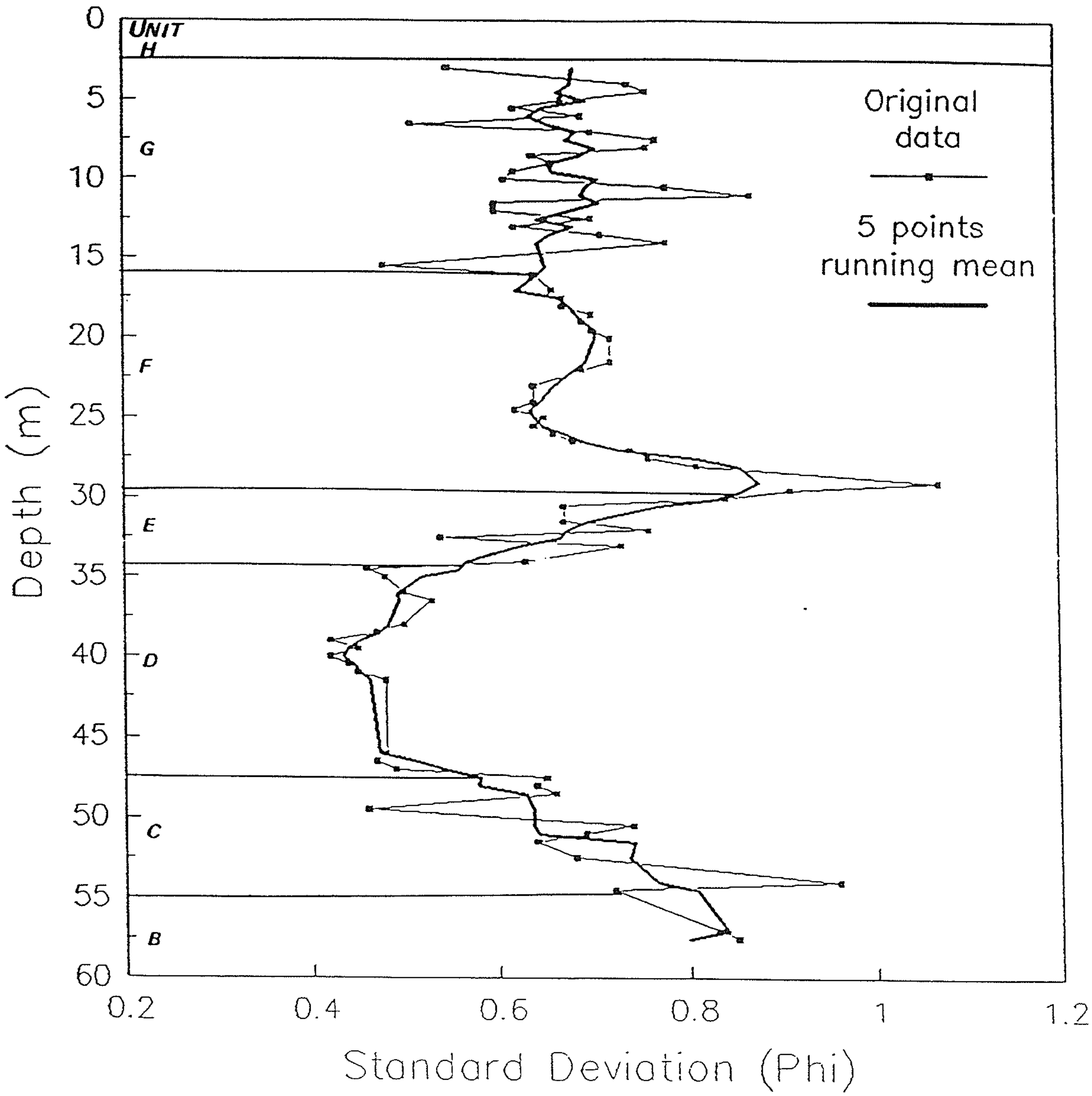
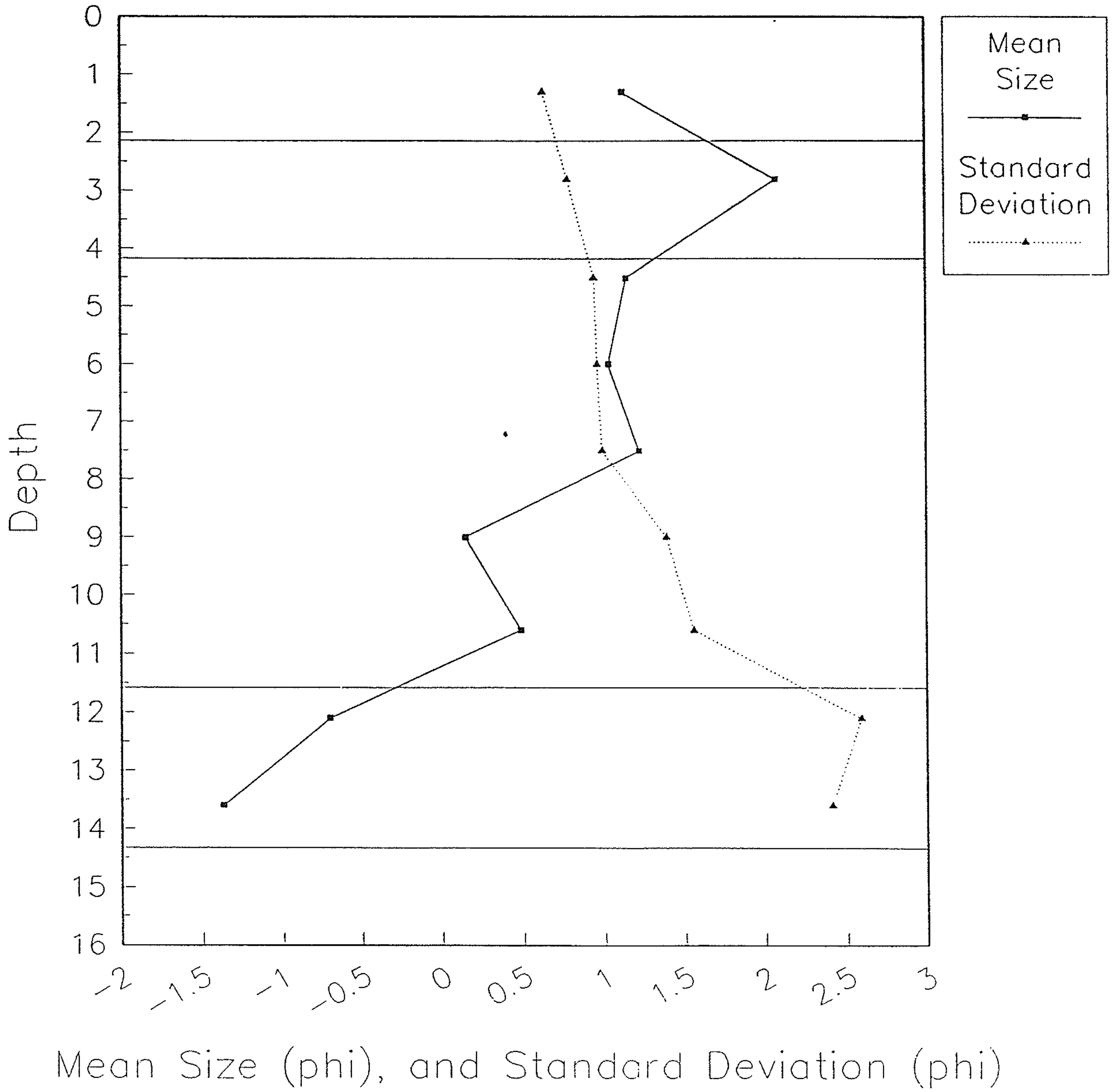


Fig. 4-36 Mean Size and Standard Deviation of Sand in Core ADA-50 Based on Size Analyses From Samples Provided by P.F.R.A.



Chapter 5

DEPOSITIONAL FACIES AND HYDRODYNAMICS OF SEDIMENTATION

Introduction

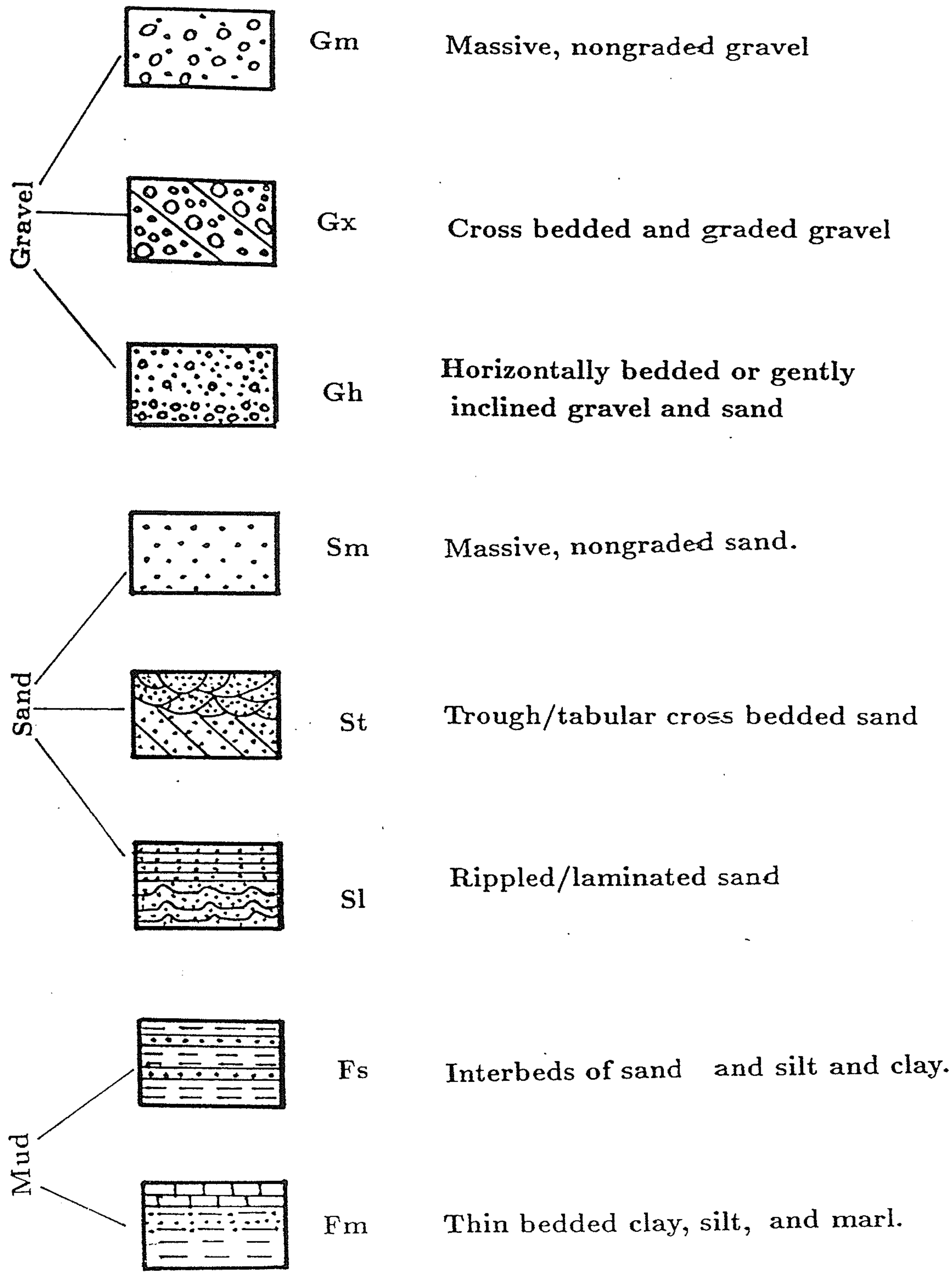
Based on descriptions of 13 outcrops/river cuts and 3 cores discussed in Chapter 4, 8 sedimentary facies are identified in the Assiniboine fan delta (Fig. 5-1). The names of these facies are modified from Rust and Koster (1984), Hwang and Chough (1990), and Postma (1990). The sedimentary sequence has been grouped into three broad categories based on grain size :(1) gravel facies, which include massive gravel facies (Facies Gm), cross bedded and graded gravel facies (Facies Gx), and horizontally stratified sandy pebbles and pebbly sand facies (Facies Gh); (2) sand facies, which is subdivided into massive sand facies (Facies Sm), trough and tabular cross bedded sand facies (Facies St), and laminated (horizontally bedded) and rippled sand facies (Facies Sl); and (3) mud dominated facies, which consist of interbeds of silty clay and sand facies (Facies Fs) and mud facies (Facies Fm). The purpose of the classification of facies is to simplify the interpretation of depositional mechanism, process, and environments. Sediments deposited by similar process and mechanism are classified into one facies regardless their relative chronology. The relationship between facies is shown in Fig. 5-2. The depositional environments of facies will be discussed in the next chapter.

I. Gravel Facies

Facies Gm, massive sandy gravel

Description This facies consists of matrix supported, non-graded, massive gravel

Fig. 5-1 Facies classification



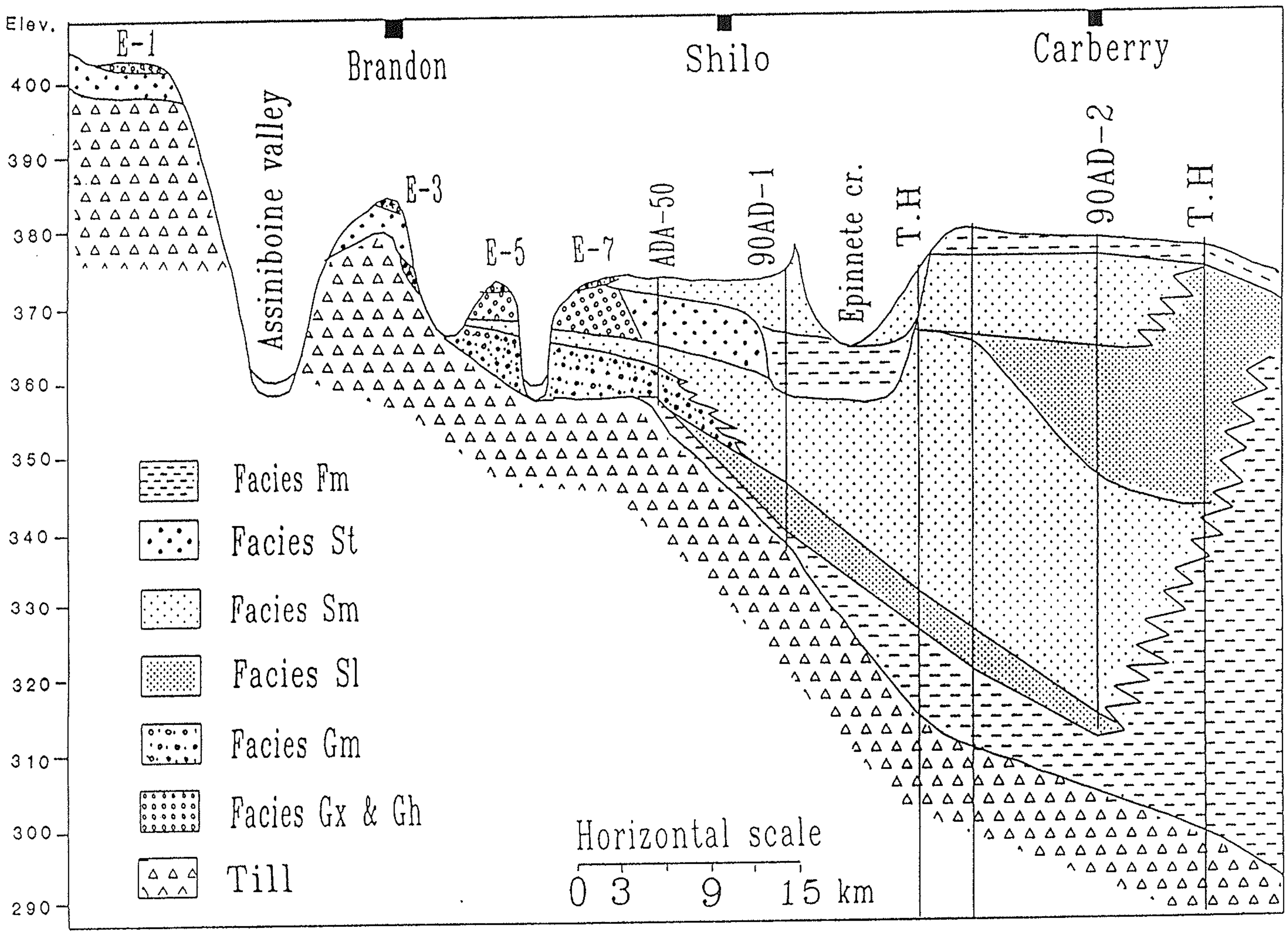


Fig. 5-2 Spatial distribution of major depositional facies in the Assiniboine fan delta. This cross section is based on 4 exposures (E-1, E-3, E-5, and E-7) and 2 new cores (90AD-1, 90AD-2), one P.F.R.A. core (ADA-50), and 3 GSC test holes (T.H).

with random fabric orientation. Clasts are mainly comprised of pebbles to cobbles with some boulders. The matrix consists of silt and clay to granules and is dominated by medium to coarse grained sand. Rip-up clasts (silty clay) with internal fine laminae are well preserved in some places and dip slightly upstream (Fig. 4-11b).

This facies has a sheet-like geometry with a sharp lower contact to the underlying formation. It occurs as the uppermost unit in the braided river plain leading to the fan delta, and in the proximal fan delta close to the Assiniboine channel. The thickness of this facies ranges from 1 m to 2 m. In Chapter 4, this facies is described in Exposure 1 (Figs. 4-1), Exposure 3 (Fig. 4-3), Exposure 4 (Fig. 4-4), Exposure 5 (Figs. 4-8, 4-9, 4-11), Exposure 6 (Fig. 4-14), and RC-2 (Fig. 4-26). In one case this facies overlies till, and is the lowest unit of the fan delta (RC-1 in Fig. 4-25).

Interpretation In facies Gm, lack of grading, absence of sedimentary structures, random fabric orientation, and matrix support suggest that cohesive strength played an important role in supporting clasts during transportation. The well preserved rip-up clasts suggest a steady laminar flow of a dense and viscous slurry of sediments where clasts were supported mainly by buoyance and cohesive forces. Lack of grading and stratification suggest that, during deposition, the deposition rate was too fast to allow the formation of grading and stratification (Arnott and Hand, 1989). The sheet-like geometry and wide distribution of the facies indicates deposition by a major event. Therefore, this facies represents a rapid deposition of cohesive debris flow during a major flood.

This facies resembles Surly's (1984) facies 1 (non-graded, matrix supported conglomerate with random fabric), which was interpreted as having been deposition by inertial sandy debris flow, and Lowe's (1982, Fig. 12) type 1 and 3 deposition, which represents deposition from cohesive (debris) flows.

Facies Gx, cross bedded and graded gravel

Description This facies consists of beds of boulders and gravel that are graded and normally sandy but clast supported, and which occur in cross stratified units. Individual cross bedded units are typically 2-6 m thick. In the No.1 hill area (Fig. 3-2) this facies contains boulders up to 80 cm in diameter, fining across the cross stratum to cobbles, pebbles and granules (Fig. 4-8). Eastward in the fan delta, this facies consists of mainly cobbles or pebbles fining across the cross stratum to granules and coarse sand (Figs. 4-18, 4-19). Within each cross bed, boulders, cobbles and pebbles in the lower portion are generally better sorted than pebbles and granules in the upper portion (Fig. 4-18). Coarse gravels and boulders in each cross bed are evenly distributed along the cross bedding plane.

This facies is about 4 m to 6 m thick, with a sharp upper contact to the facies Gm (massive sandy gravel). It occurs on the stoss side of the No.1 hill (Exposure 5, Fig. 4-8), No.3 ridge (Fig. 4-13), and No.4 ridge (Fig. 4-18).

Interpretation Clast support and random fabric orientation in this facies suggest that dispersive pressure was developed but deposition was too fast to allow fabric orientation. The normally graded, well sorted nature, and open framework suggest that turbulent flow was an important supporting mechanism during transportation, and indicate deposition from suspension. Large boulders at the base of each cross bed were probably deposited from traction carpets.

This facies is similar to the facies 11 of Surlyk (1984, graded and clast supported conglomerate with random fabric), and the facies R3 of Lowe (1982). It also is similar to the deposition from fully-turbulent debris flow described by Nemec and Steel (1990). Thus this facies is interpreted as deposition directly from gravelly high density turbidity currents or from fully-turbulent debris flows.

Facies Gh, horizontally bedded sandy pebbles

Description This facies can be divided into two subfacies: Gh1 and Gh2. Facies Gh1 comprises cobbles and pebbles in a sand matrix grading upward to coarse sand with large floating clasts scattered throughout. Gravels in this facies make up about 40 to 60 percent of the sediments, and are more concentrated in the lower portion of the stratified bed than the upper portion. The sedimentary structures are dominated by flat beds and gently inclined beds, which are outlined by layers of gravels in a sand matrix. This facies is different from the cross bedded and graded gravel facies (Gx) because of the fine clast size, flat/gently inclined bedding, and stratification within individual beds. This facies occurs in the center and lee side of the No.1 hill (Exposure 5, Fig. 4-10), and possibly in the lee side of the No.4 ridge (unit AA in Fig. 4-22).

Facies Gh2 consists of couplets of fine gravel and cobble beds. The cobble beds are about one clast thick, with elongated clasts dipping upstream. Fine gravel beds are generally 0.5 to 2 m thick. The internal structures in each fine gravel bed consist of intermediate dipping, parallel (unidirectional) bedded foresets. This facies is exposed in the Chater section only (Fig. 4-24).

Interpretation In facies Gh1, the high clast concentration in the basal beds suggests that dispersive pressure developed in the flow. Sand with larger floating clasts in the upper portion of each cross bed suggests rapid deposition from a density flow. Therefore sediments in this facies are likely being deposited by "freezing" of the traction carpets, followed by direct deposition of fine materal from suspension due to a rapid deceleration of sandy high density turbidity flow. This facies is similar to facies 15 of Surlyk (1984, stratified pebbly sandstone), and facies S2 of Lowe (1982).

Facies Gh2 was more likely formed in a wide fluvial channel. The thin cobble beds and the imbrication of elongated gravels are indicative of deposition from traction (Blatt et al., 1980) or erosion (lag). The sheet-like fine gravel with cross beds is interpreted as the result of migration of gravel bars over the channel floor.

II. Sand Facies

Facies Sm, massive and non-graded sand

Description This facies consists of massive coarse to medium grained sand that is poorly sorted to moderately sorted, non-stratified, contains scattered lignite particles, and contains less than 20 percent pebbles and cobbles (up to 20 cm, with an average of 4cm in diameter) as floating clasts in sand matrix. This facies is different from the massive gravel facies (Gm) because of the low gravel content, low silt and clay content, and fine clast size. It is different from horizontally bedded pebbly sand facies (Facies Gh) because of the lack of internal stratification and lack of grading. This facies occurs below the foreset beds in the No.1 hill (Unit C), and in the No.4 ridge (upper part of Unit 1). It also occurs in the Treesbank section (the basal non-bedded sand unit Fig. 4-26), in 90AD-1 (Fig. 4-30), and in 90AD-2 (Fig. 4-33).

Interpretation The deposition of lignite (low density coal particles) and gravels together with coarse to medium grained sand suggests rapid deposition. The massive and non-graded nature may result from rapid deposition during the last stage of deposition by a turbidity flow because less rapid deposition may permit more gradual compaction of the bed with some movement by traction and formation structures such as plane lamination and ripple-drift (Blatt et al., 1980). Absence of water escape structures was due to the coarse grain size (Surlyk, 1984). This facies is similar to the non-graded sandstone of Surlyk (1984, facies 16) and the S1 of Lowe (1982). It was deposited from high-density turbidity current.

Facies St, cross bedded sand

Description This facies includes tabular cross bedded and trough cross bedded sand. Sediments in this facies are dominated by coarser grained sand with some pebbles and granules (< 10 percent). Tabular cross bedded sand occurs in the west Brandon exposure (Fig. 4-1). Both the cut and fill structures and tabular cross beds are observed in Exposure 2 (Fig. 4-2), in which pebble trains are present at the base of each trough cross bed.

Interpretation. The occurence of both the tabular cross beds and trough cross beds in coarse grained sand are indicative of fluvial deposits by migration of sand dunes and diagonal bars over the river floor. Lack of grading in sediments suggest rapid deposition (Blatt et al., 1980) or high sediment concentration. Therefore this facies indicates fluvial deposition with high sediment concentration, possibly a braided river environment.

Facies Sl, laminated and rippled sand

Description This facies includes horizontally bedded sand and rippled sand that is well sorted to moderately well sorted. It occurs in the Provincial Park section (RC-6, Figs. 4-28, 4-29), where well sorted silty very fine sand with asymmetrical ripples is overlain by silty very fine sand with flat laminae or small symmetrical ripples. Stratified and laminated sand also occur in 90AD-1 (Unit C) and 90AD-2 (Fig. 5-2), and climbing in-drift ripples occur at Exposure 9.

Interpretation The absence of water-escape structures reflects either coarse size or slow rate of deposition (Hwang and Chough, 1990; Surlyk, 1984; Blatt et al., 1980). In this facies, the fine grained nature and absence of fluid-escape structures suggests slow rate of deposition. Flat laminae and small symmetrical ripples (with no indication of ripple migration) suggests deposition from suspension in lower shoal areas where waves were too weak to generate large ripples. The occurrence of climbing in-drift ripples and unidirectional cross lamination (Exposure 9) are indicative of near shore current flows and high sediment supply.

III. Mud Dominated Facies

Facies Fs, interbeds of mud and sand

Description This facies is exposed in the Souris River junction section (RC-1, Fig. 4-25). It consists of alternating beds of sand and clay. Each of the clay beds is about 0.55 to 0.9 m thick (Figs. 4-24, 4-25) and beds are composed of massive dark blue clay to silty blue clay. Sand beds are 0.3 to 0.5 m-thick, and are moderately sorted (σ=0.8). Both the upper and lower boundary of these sand beds are very distinct (Fig. 4-25).

Interpretation The massive, non-stratified dark blue clay suggests deposition in an anoxic environments. The presence of poorly sorted and massive thin sand between clay and silty clay beds may reflect the influx of coarse grains to the lake floor during storms or periods of high runoff.

Facies Fm, mud facies

Description This facies consists of laminated silty clay, clayey silt, and marls. Plant fibers, wood fragments and occasional gastropod and mollusc shell fragments can be abundant in some beds. This facies occurs in the lower unit of the Glenboro section (RC-5), where it overlies till and underlies channel sand. It is also present in well 90AD-1 (units B, E, G, H, I, Fig. 4-30), and 90AD-2 (Fig. 4-33).

Interpretation The laminated silty clay and clayey silt were deposited in a water body that was far from source supply. Silty clay in RC-5 was probably deposited in a deep lake environment, whereas units E, G, H, and I in 90AD-1 were deposited in a small lake in Epinette Creek valley after the valley was abandoned. Marls with abundant fossils were deposited in an oxygenated environment (shallow lake). Some of the marls could be deposited by plants (Blatt et al., 1980). Marl with abundant plant fibers and wood fragments were deposited in a bog environment transitional to shallow lake environment.

Vertical Facies Sequences

The sedimentary sequence in the southern fan delta area shows a fining upward trend, as evidenced in core ADA-50 (Appendix III, ADA-50, Fig. 5-2) and river cut 3 (RC-3). The fining upward sequence starts with sandy gravel (cobble to pebble) in the area near the Souris River junction, and medium to coarse sand in the area near Glenboro, then fines upward to fine sand and silty sand. In some outcrops, more than one fining upward cycle was found.

In the northern fan delta, the sedimentary sequence is dominated by coarsening upward cycles, such as in cores 90AD-1 and 90AD-2. The coarsening upward trend normally starts with laminated silty clay coarsening upward to fine sand and to medium or coarse grained sand (Figs. 4-30, 4-33).

Chapter 6

DEPOSITIONAL ENVIRONMENTS

Introduction

There are 2 types of river mouth accumulations in the Assiniboine fan delta: 1) Gilbert-type delta; and 2) underflow fan delta. The main difference between a Gilbert-type delta and an underflow fan delta is the occurence of foreset beds. A Gilbert-type delta has steep foresets, built to lake level by progradation; it is characterized by a coarsening upward sequence from prodelta mud at the base to large scale foreset beds to topset beds (Miall, 1984). An underflow fan delta, however, does not have steep foreset beds. Sediments in underflow fan deltas are moved by turbidity flows, waves, and currents.

For convenience, the depositional environments in this chapter will be approached in the order of: 1) the river plain leading to the fan delta, 2) subaerial debris flow leading to the fan delta, 3) Gilbert-type delta, and 4) underflow fan delta. Because of the lack of biological evidence, which is the case for most fan deltas built into glacial lakes by meltwater, the depositional environments are studied based on the sedimentary textures and structures only.

I. Braided River Plain

Occurrence : Sediments in this setting include fine to coarse sand and less than 10% gravels. Clay beds are generally absent. Sedimentary structures include trough cross beds, tabular cross beds (Facies St), flat beds and rippled laminae (Facies Sl). The surface of this setting is characterized by ancient/relict braided scoured channels and low (<6 m) diagonal or longitudinal bars (hills) at an elevation from 383 in

Exposure 3 (Fig. 3-2) to 410 m near the Little Saskatchewan River junction. The Assiniboine River flows across this area and the modern river floor is 20 to 40 m lower than the base of sand and gravel.

Paleoenvironment : Lack of clayey units in these facies suggests deposition from a high energy system without development of a flood plain, i.e. a braided river system. The braided surficial channel pattern supports this interpretation. Tabular cross beds and trough cross beds were produced by migration of diagonal bars or dunes on the river floor (Blatt et al., 1980). Sediments were apparently deposited during high lake level, and were subsequently entrenched by the Assiniboine River after the lake fell to a lower level.

II. Subaerial debris flow deposits

Occurrence : Facies Gm occurs in the eastern braided river plain (up to an elevation of 404 m) and in the southern proximal fan delta (in the vicinity of the modern Assiniboine valley), but is absent from the northern proximal fan delta (Fig. 6-1; see also Chapter 5, Facies Gm). It overlies both the braided river deposits and the Gilbert-type delta sediments. Sediments in facies Gm are very poorly sorted, nongraded, massive gravels that are supported by a sand matrix.

Paleoenvironment : The massive, nongraded gravels are indicative of debris flow deposits. Lack of mud/silt interbeds and sand dikes suggest it may have been deposited as a subaerial debris flow. Other evidence for a subaerial origin comes from the stratigraphic position. Since it overlies both the braided river deposits and the Gilbert-type delta deposits, sediments in facies Gm are not older than the main fan delta. The occurrence of facies Gm sediments in the vicinity of the modern Assiniboine River valley suggests deposition after the excavation of the southern Assiniboine River valley. But the level of Lake Agassiz water never reached above

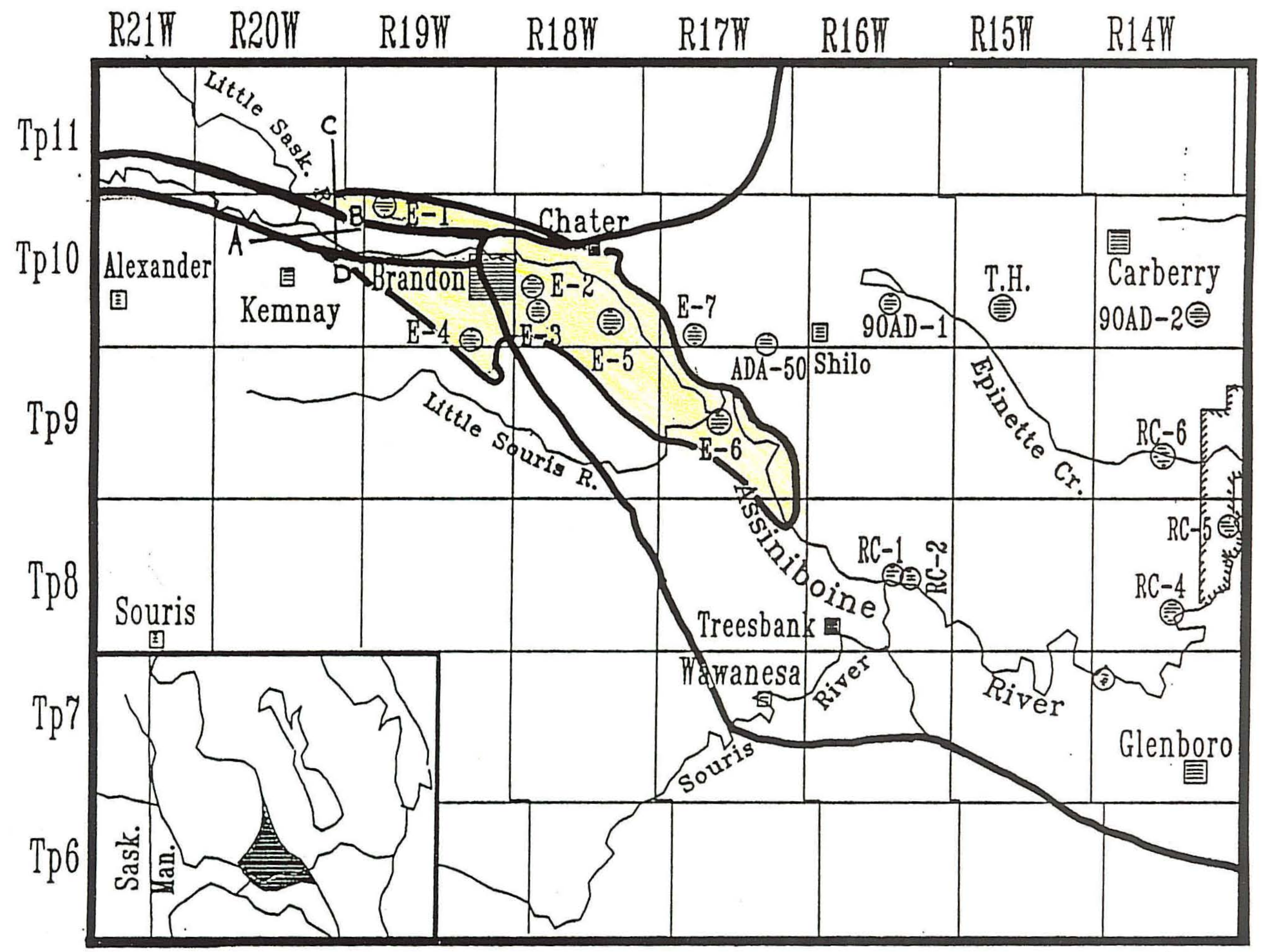


Fig. 6-1 Aerial distribution of the subaerial debris flow deposits (the colored area).

the delta surface after the deposition of the fan delta, suggesting that the massive and nongraded gravel (facies Gm) was deposited by a subaerial debris flow, not by subaqueous debris flows.

III. Gilbert-type delta

Introduction

The Gilbert-type delta occurs in the area from Brandon to at least 15 km east of Brandon. The foresets of the delta are exposed in several sites (E-5, E-6, E-7 in Fig. 1-8) but the topsets are exposed in the Chater section only (E-8 in Fig. 1-8), and the toesets were not exposed at all. The toeset of the fan delta may be present in the radar profile in Fig.4-22 (unit AA). But because there are at least two explanations for this radar profile, which will be presented below, it is difficult to describe the sedimentary structures and textures of the toeset.

Gilbert-type topsets

Occurrence : The Gilbert-type topset beds are absent in the vicinity of the Assiniboine valley, where most of the exposures occur. The only Gilbert topset exposure is in the Chater section (Exposure 9), where the Gilbert-type topset is represented by alternating, one clast-thick, horizontal to sub-horizontal beds of cobbles, and sets of cross bedded fine gravels (Facies Gx) (Fig. 4-24). These cross bedded fine gravels have a sheet like geometry. They are different from the braided river plain because they lack large scale channel geometry and cut and fill structures. They differ from the delta foresets because of a lack of large scale foreset beds (each cross bedded unit in Fig. 4-24 is only 0.3 to 1 m). Facies Gm, which occurs in most of the exposures in the proximal fan delta area, may be part of the topset sequence, but it is important enough to be discussed separately.

Interpretation : The cross bedded and subhorizontal fine gravels were deposited by migration of gravel bars. Lack of cut and fill structures and the sheet-like geometry suggest deposition in broad channels, or by sheet flows over the delta surface. The absence of the Gilbert-type topset in the vicinity of the Assiniboine valley is probably due to erosion, as evidenced by the lower elevation of the delta surface in the vicinity of the valley.

Gilbert-type foresets (Facies Gx and Gh)

Occurrence : The fan delta foresets occur in the area from Brandon (the paleo-river mouth during the deposition of the Assiniboine fan delta) to at least 15 km east of Brandon, where the surface sediment is dominated by gravels (Figs. 5-2; zone 2 in Fig. 3-1). Sediments in this region consist of thick tabular cross bedded units; each bed of the set typically consists of a sequence of graded boulders, cobbles, pebbles, and granules (Facies Gx). Flat bedded units (Facies Gh) also occur. Changes in sedimentary character from upstream to downstream is evident:

(1) From west to east (downstream), sedimentary structures appear in at least two cycles, each cycle consists of several 2 to 6 m tabular cross beds changing downstream to gently inclined beds and flat beds

(2) Maximum and average clast size generally decrease from tabular cross bedded sediments to flat bedded sediments.

Some of the characteristics in both the west and east are identical:

(1) Cross beds are all straight and steep (24 to 31°).

(2) The thickness of boulders and cobbles beds in each cross-stratum is the same in both the upslope and downslope portions.

(3) There are no traces of sand, silt, or clay being deposited between cross beds.

Paleoenvironment : The tabular cross bedded gravels were deposited in a shallow lake environment by floods or catastrophic floods, and the gently inclined to flat bedded gravels and sand were deposited during waning stages of floods or by surging flows after floods. The reason for this interpretation is discussed below.

The exposed delta sequence is only about 4 to 6 m thick. Its surface describes a generally flat fan delta plain (rising northeastward due to isostatic rebound), suggesting that the fan delta was graded to the lake level when the water was shallow. On the other hand, the normal graded and well sorted cobbles and pebbles suggest deposition from a strong flow (see Facies Gx, Chapter 5). Therefore, the sediments were possibly deposited by floods into a shallow lake. Lack of top set deposits, if it was not due to subsequent erosion, suggests deposition in a continuous high velocity flow, which generated strong flow separation and expansion in the lee side of dune, or delta (Fig. 6-2, IA).

How did cross beds form? Cross beds can be produced by: (1) avalanching of grains down the lee slope of advancing dunes, bars, fans or deltas (Blatt et al., 1980); (2) the combination of collective settling of bedload and grainfall deposition from suspension in flow separation and expansion (Schenk, 1987).

Avalanching commonly forms inversely graded cross beds (Blatt et al., 1980). It can not form cross bedded and normally graded sediments that fine across the cross stratum. Blatt et al. (1980, p.135) explained the sedimentary structures formed by different avalanching rates:

> "At low rate of bed load movement, avalanching occurs at more or less regular intervals. In the intervals between avalanches, finer-grained sediment accumulates by settling on the lower part of the lee slope (foreset). Many individual cross-laminae are coarser toward the base and show a slight reverse grading across the lamina. At highest rate of bed load movement, sediment avalanches more or less continuously down the lee slope and the coarsest particles are found at the top, not the bottom, of cross-laminae."

Allen (1980) and Walker (1984) suggested that the formation of cross beds may be controlled by three factors: (1) the fluctuation time and velocity pattern, (2) water depth, and (3) bed-material calibre. Under a certain water depth, according to Allen (1980) and Walker (1984), a continuous strong unidirectional flow would generate strong flow separations in the lee slope of a delta (or dune), and form steep dipping foresets without the development of topsets (Fig. 6-2 Ia). When it is a highly fluctuating multidirectional flow (affected by river flows and waves), sedimentary structures should be dominated by small scale, multidirectional cross beds (Fig. 6-2 IVa). Schenk's (1987) experiments show that large foresets are formed by settling of bedload and suspension load (grainfall) in strong flow separation zones. Therefore, a possible explanation for the foresets in this fan delta is that cross beds were formed during catastrophic floods by the combination of collective settling of bedload and grainfall deposition from suspension in strong flow separation and expansion zones. Deposition from grainfall (suspension) probably was dominant, as indicated by the presence of normal grading in cross bedded boulders and gravels. According to Lowe (1982) and Nemec and Steel (1990), graded bedding in cross

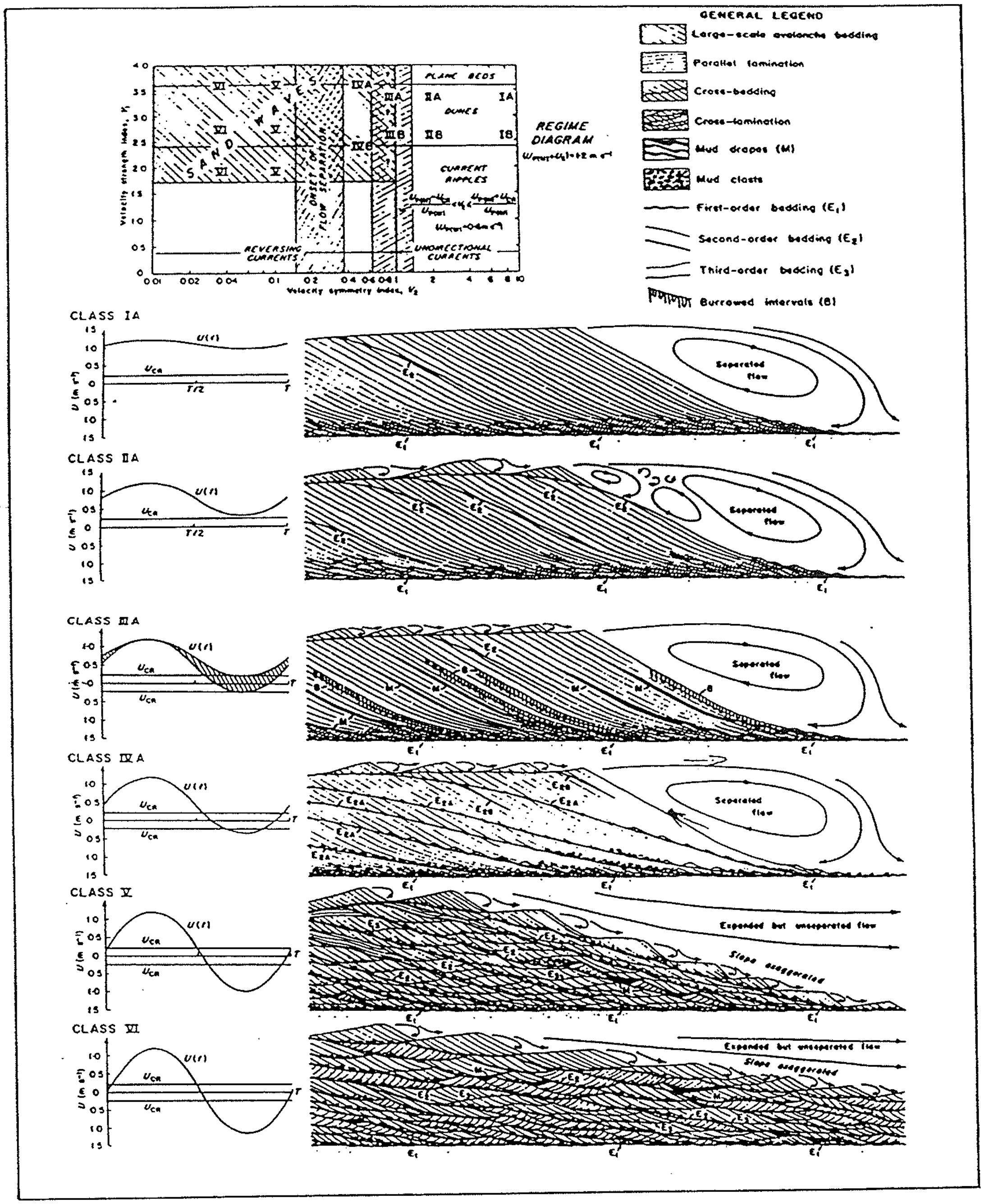


Fig. 6-2 Diagram showing the internal structures of sediments deposited in different type of high velocity flows. Note that Class IA is deposited by continuous (symmetrical to time) unidirectional (asymmetrical to velocity) flows that generate strong separated flow zone in the lee side of foresets. In class VI, discontinuous (asymmetrical to time) multiple-directional flows (effected by river and valves) can not form large foreset beds (from Allen, 1980).

beds is the result of deposition from full turbulent debris flows. This means that sediments in the fan delta foresets were deposited mostly from suspension load due to rapid deceleration in flow separation zones. Deceleration caused particles to leave the flow in the order of size.

The next question is whether one flood deposited one of the graded foreset beds only or whether pulsing flood flow may have resulted in many foresets beds in a series (Teller, 1992, personal commun.). If one flood deposited one foreset bed, there should have been deposition of finer grained beds between the foreset beds, unless the time was brief (e.g. daily fluctuating flood) or the current did not subside much. Because of the absence of finer units between the foreset beds in the fan delta, it is believed that one flood could have deposited the repetitive series of graded foresets, with each foreset bed reflecting a pulse in the flood current. The fluctuation of a flood may be related to irregular damming of flood water by accumulation of ice blocks or landslides in constricted parts of the Assiniboine valley, or due to the different arriving time of peak waves in different drainage basins along the Assiniboine and Qu'Appelle Rivers during a major storm.

Interpretations to the radar profile in Fig.4 – 22. The reason to interpret the radar profile in Fig. 4-22 is because this profile provides some indication about whether or not barriers existed offshore from the advancing foreset beds, and how far the gravel foresets extend. There are at least two possible interpretations.

First version : The profile can be interpreted as a delta foresets-barrier-interbarrier complex. As shown in Fig. 6-3A, The large scale cross bedded unit (unit 2 in Fig. 4-22) is interpreted as part of the Gilbert-type foreset sequence because it is the continuation of the cross bedded unit (facies Gx) in Figs. 4-19 and 4-20. The horizontally bedded unit (unit AA) in the radar profile (Figs. 6-3,4-22, 4-23) is

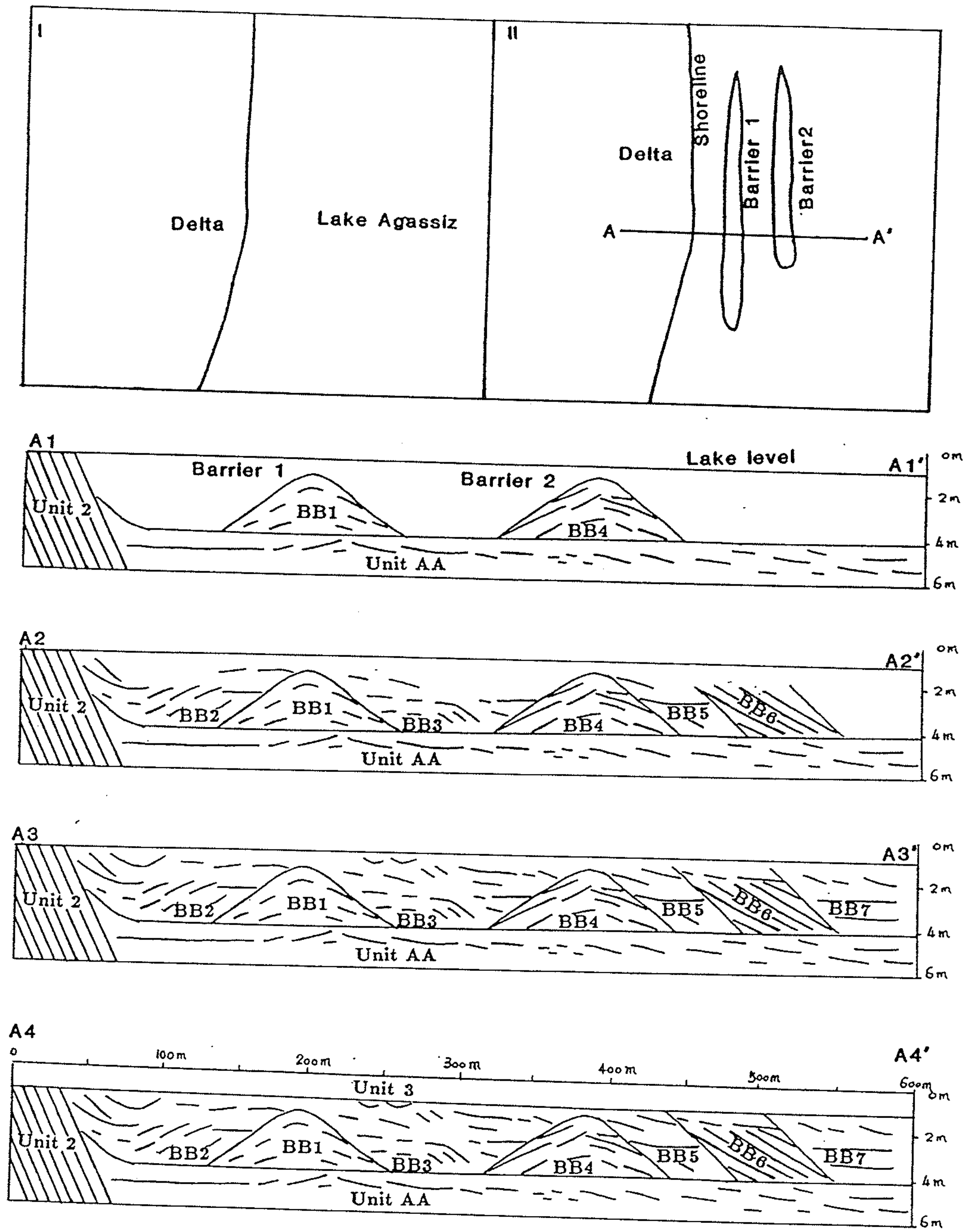


Fig. 6-3A Diagram interpreting the Radar profile in Fig.4-22. This is a foreset-barrier-interbarrier setting. A1: Deposition of barriers (BB1 and BB4) after the deposition of delta foresets; A2: Deposition of interbarriers (units BB2, BB3, BB5, and BB6); A3: Deposition of flat bedded lacustrine sediments (BB7); and A4: Deposition of unit 3.

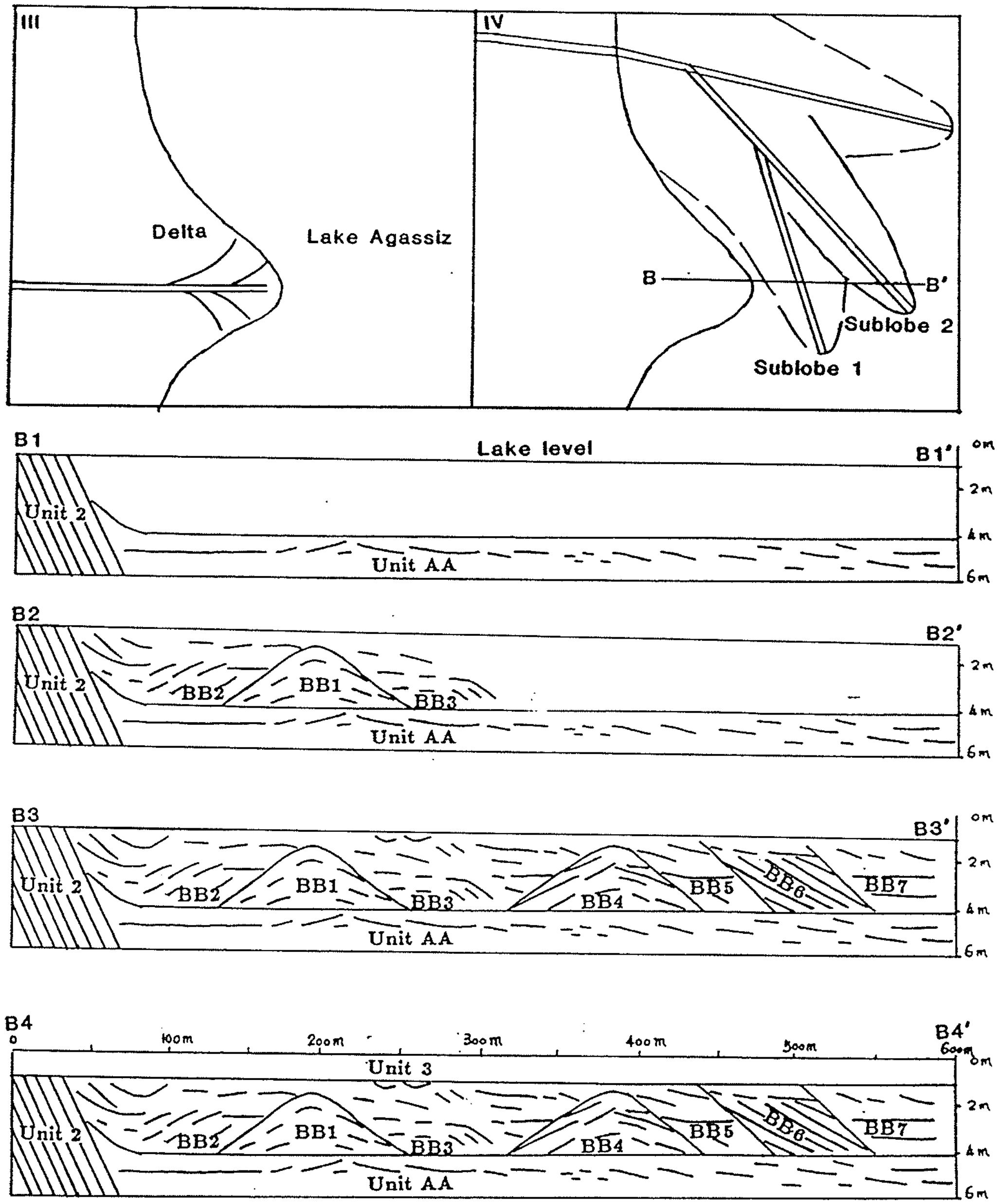


Fig. 6-3B Diagram interpreting the Radar profile in Fig.4-22. This is a delta-sublobe complex. B1: Deposition of Gilbert-type foresets and toesets (units 2 and AA); B2: Deposition of delta sublobe 1 (units BB1, BB2 and part of BB3); B3: Deposition of the second sublobe in the the order of B4, BB5, BB6, and BB7; B4: Deposition of unit 3.

interpreted as finer grained lacustrine deposits, but could be the delta toeset. Unit BB is interpreted as having been deposited in the nearshore zone. Bi-directional cross beds with bar shape geometry (BB1 and BB2) are barriers, and discontinuous flat beds and small scale cross beds (BB3) between barriers are inter-barrier deposits, probably accumulated shortly after the bar was deposited. The chronology of deposition is shown in Fig.6-3A.

Second version :. The profile can be also interpreted as a delta foreset-sublobe/crevasse spray complex, as shown in Fig. 6-3B. Unit AA is interpreted as the delta toeset because of its flat beds (Figs. 6-3B, 4-22 and 4-23). Unit 2 was a Gilbert-type foresets advancing from west to east. Whereas unit BB was deposited by multiple delta sublobes/crevasse splays which accumulated from north/northwest to south/southeast. Units BB1 and BB2 are the core of those lobes/crevasse splays, and unit BB3 was deposited as the splays continued to grow southward. Alternatively, BB1 and BB2 may have been deposited by southward growing spits, with interspit sediments of BB3 deposited as the shoreline gradually encroached into this protected embayment beyond the barrier spit (Teller, 1992, personal commun.).

IV. Underflow fan delta (Facies Sm, Sl, and Fm)

Occurrence : There is no trace of foreset beds in the middle and distal fan delta region. River cuts, exposures, and cores reveal massive fine to coarse sand (Facies Sm) (Figs. 4-25; 4-30, Unit D; 4-33, Units B, C, D, E; E-9, Chapter 4), symmetrically rippled and horizontally laminated fine to very fine sand (Facies Sl) (Figs. 4-28; 4-29; 4-30, Unit C), and sand with climbing ripples and asymmetrical ripples (see Exposure 9, chapter 4).

Paleoenvironment : Sediments in the middle and distal fan delta (Zone 3 in Fig. 3-1) are mostly underflow fan delta deposits rather than Gilbert type of delta deposits because of the absence of foresets, and the co-existance of massive sand and sand with symmetrical ripples and climbing in-drift ripples. The thick, massive fine to coarse sand facies indicates rapid deposition from suspension, or deposition from very highly concentrated sediment dispersions (see Blatt et al., 1980, p.136). This means that the incoming meltwater flows had a high sediment load, probably flowing beneath the lake surface as continuous turbidity flows. The co-existance of massive sand and sand with wave ripples and current ripples indicate subaqueous deposition. The overall coarsening upward cycles in the massive sand facies suggest lakeward progradation of the fan delta.

Sand with climbing in drift ripples with internal cross laminations suggests deposition by current in near shore zone, whereas sand with symmetrical ripples and rounded crests indicates oscillatory flow above wave base.

V. Pro-delta (Facies Fm)

Occurrence : Sediments deposited in the pro-delta environment include laminated, dark gray silty clay and clayey silt. This facies is mainly exposed in river cuts near Glenboro (RC-3, Table 5; RC-5, Table 6), and is revealed in the core of 90AD-1 (Unit B in Fig. 4-30) and 90AD-2 (Unit A Fig. 4-33).

Paleoenvironment : The dark gray clay reflects deposition in a deep lake environment or deposition in the distal region of the fan delta. Laminated silty clay and clayey silt reflects either seasonal, daily, or episodic events, perhaps related to storms, or to pulsive nature of river discharge.

Summary of depositional environments

(1). The River plain west of Brandon was deposited by braided river flows during a high lake level.

(2). The Assiniboine fan delta was deposited mainly by meltwater discharge, with a high concentration of sediments at the western margin of Lake Agassiz, and by the subaqueous extension of the discharge eastward into the lake.

(3). The Gilbert-type delta, which extends from Brandon to at least 15 km east of Brandon (Zone 2 in Fig. 3-1), was constructed mainly in shallow water by catastrophic floods when the lake level was at about 380 m.

(4). The rest of the fan delta (Zone 3 in Fig. 3-1) was mostly constructed by meltwater underflows, turbidity flows, waves and currents after the deposition of the Gilbert delta, although the low part may be deposited by the extension of subaerial floods that deposited the Gilbert delta.

(5). After deposition of the fan delta, and excavation of the modern Assiniboine channel across the southern fan delta, a subaerial debris flow deposited the massive and nongraded gravels over the braided river plain and the proximal fan delta.

Chapter 7

DEGLACIAL HISTORY AND LAKE AGASSIZ

Introduction

This chapter will discuss the currently accepted chronology of Lake Agassiz presented by Teller (1985, 1987), Fenton et al.(1983), Teller et al. (1980), and Teller and Fenton (1980). Another chronology by Klassen (1975, 1983) will also be discussed. The author intends to use this discussion to narrow the time span for construction of the Assiniboine fan delta, and to relate its history to the water level of Lake Agassiz.

The Currently Accepted Chronology of Lake Agassiz

Introduction

During the late Wisconsinan, ice from the Labradorean center (Hudson Bay area) advanced southwestward to about the western edge of the Lake Agassiz basin (Fenton et al., 1983). As the Labradorean ice began to waste, the Keewatin center west of Hudson Bay expanded southeastward into the area formerly occupied by the Labradorean ice (Fenton et al., 1983).

Case-Lockhart phase

After the maximum advance of late Wisconsinan ice into central Iowa at 14,000 BP (Ruhe, 1969, Teller, 1987), the ice began to retreat northward toward the Lake Agassiz basin. The retreat was interrupted many times by the surging of Red River ice southward (Clayton et al., 1985). One of the surging events into South Dakota and Iowa occurred at about 12,300 BP (known by dating wood buried by glacial deposits) (Teller et al., 1980; Clayton and Moran, 1982; Fenton et al., 1983). By

11,700 BP, Lake Agassiz was established in the southern end of the basin (Fig. 7-1) (Teller, 1987; Fenton et al., 1983, Clayton, 1983).

The Souris Lobe of ice, a sub-lobe of the Keewatin ice which lay to the west of the fan delta, retreated to form glacial Lake Souris in the southern Souris Basin (Fenton et al., 1983). Initially this allowed the Souris River, glacial Lake Souris, and the southern Tiger Hills (south of Brandon, Fig. 1-8) to drain through the Souris River valley to the Pembina Spillway and into Lake Agassiz in North Dakota (Conley, 1986), bypassing the still glaciated Assiniboine fan delta area to the north until shortly after 11,300 BP.

Fenton et al. (1983) and Conley (1986) speculated that there was a readvance of the Red River lobe at 11,400 BP to the Edinburg moraine in North Dakota, and the Darlingford moraine between the Pembina and Assiniboine valleys (Fig. 2-1), which is the northern extension of the Edinburg moraine in southwestern Manitoba. West of Brandon, the ice advanced to the Alexander moraine, which is essentially the northwestern extension of the Darlingford moraine (Figs. 2-1, 7-2). It blocked waters in the Souris River drainage system and formed glacial Lake Souris (mainly in North Dakota) and glacial Lake Hind (in southwestern Manitoba).

As the ice retreated, Lake Agassiz expanded northward. By 11,500 to 11,300 BP, Lake Agassiz had expanded at least to the International Boundary. By 11,000 BP, Lake Agassiz expanded northward into central Manitoba (Teller, 1987). The approximate ice positions are shown in Fig. 7-3. Notice in Fig. 7-3 that the Assiniboine fan delta is located between the ice margin positions at 11,300 and 11,000 BP. Therefore, the Assiniboine fan delta might have begun to form shortly after 11,300 BP. This did not start until the Alexander moraine was breached along the Assiniboine bedrock valley because the Alexander moraine blocked the major

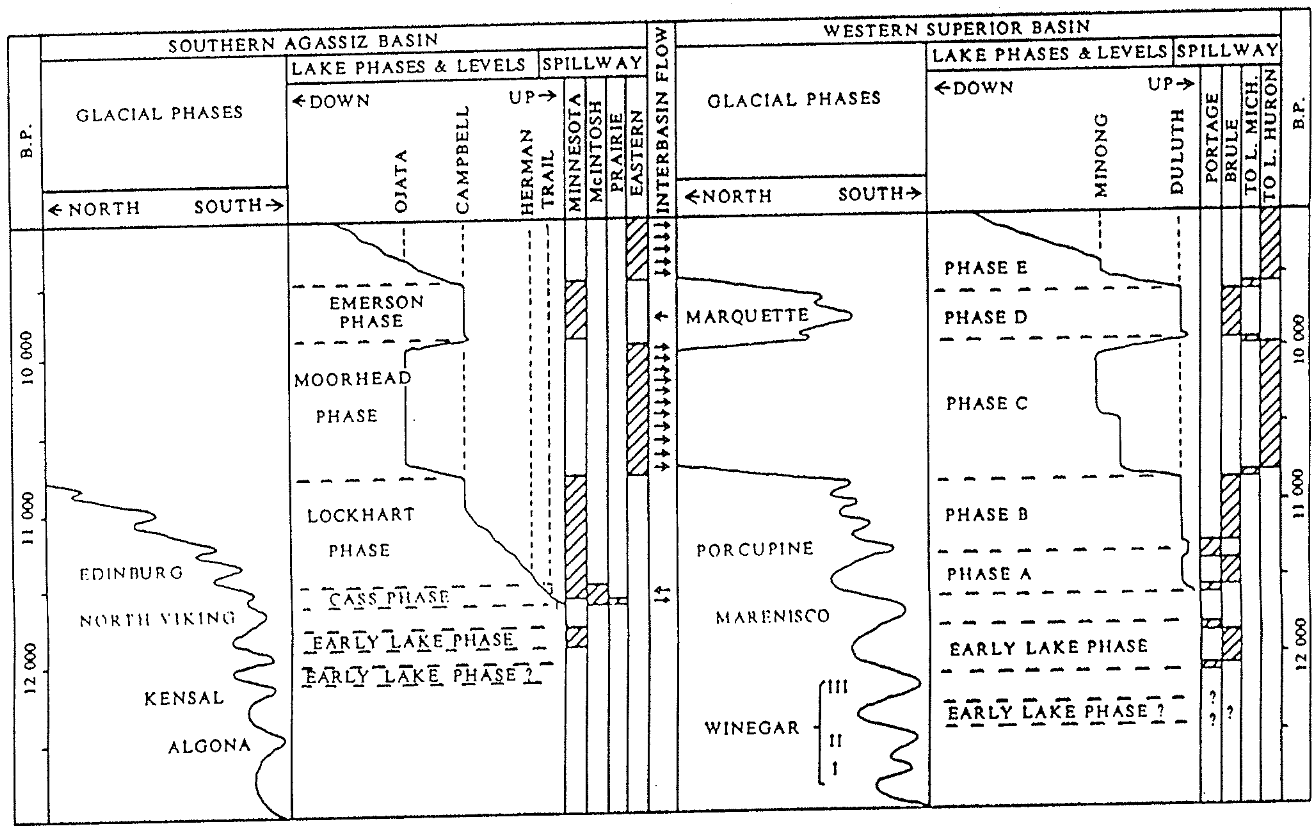


Fig. 7-1 Chronology of Lake Agassiz, and its dainage to the eastern outlets and the southern outlets (after Clayton, 1983, p.259).

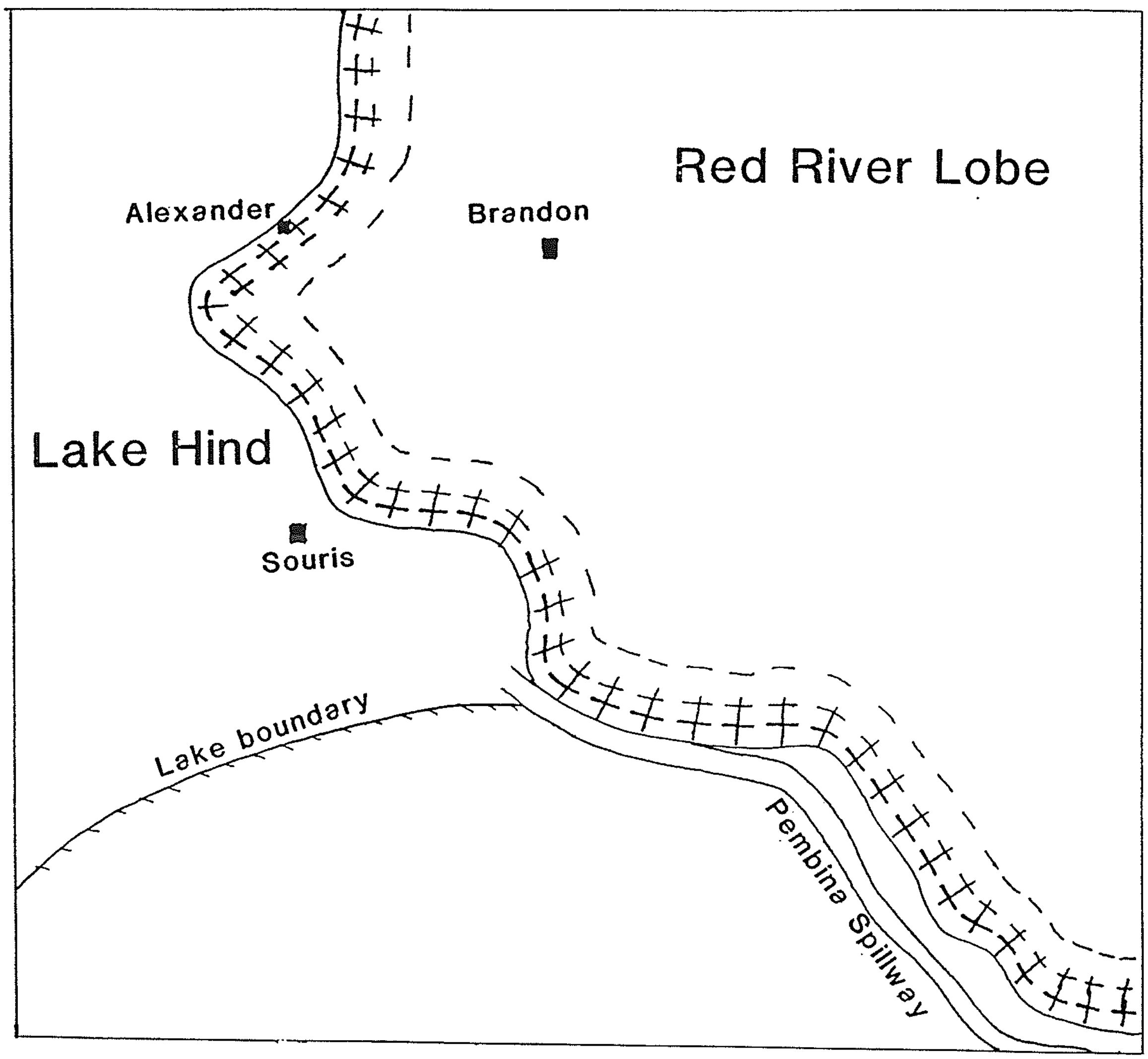


Fig. 7-2 Formation of the Darlingford moraine and Alexander moraine by advance of the Red River Lobe at about 11,400 BP.

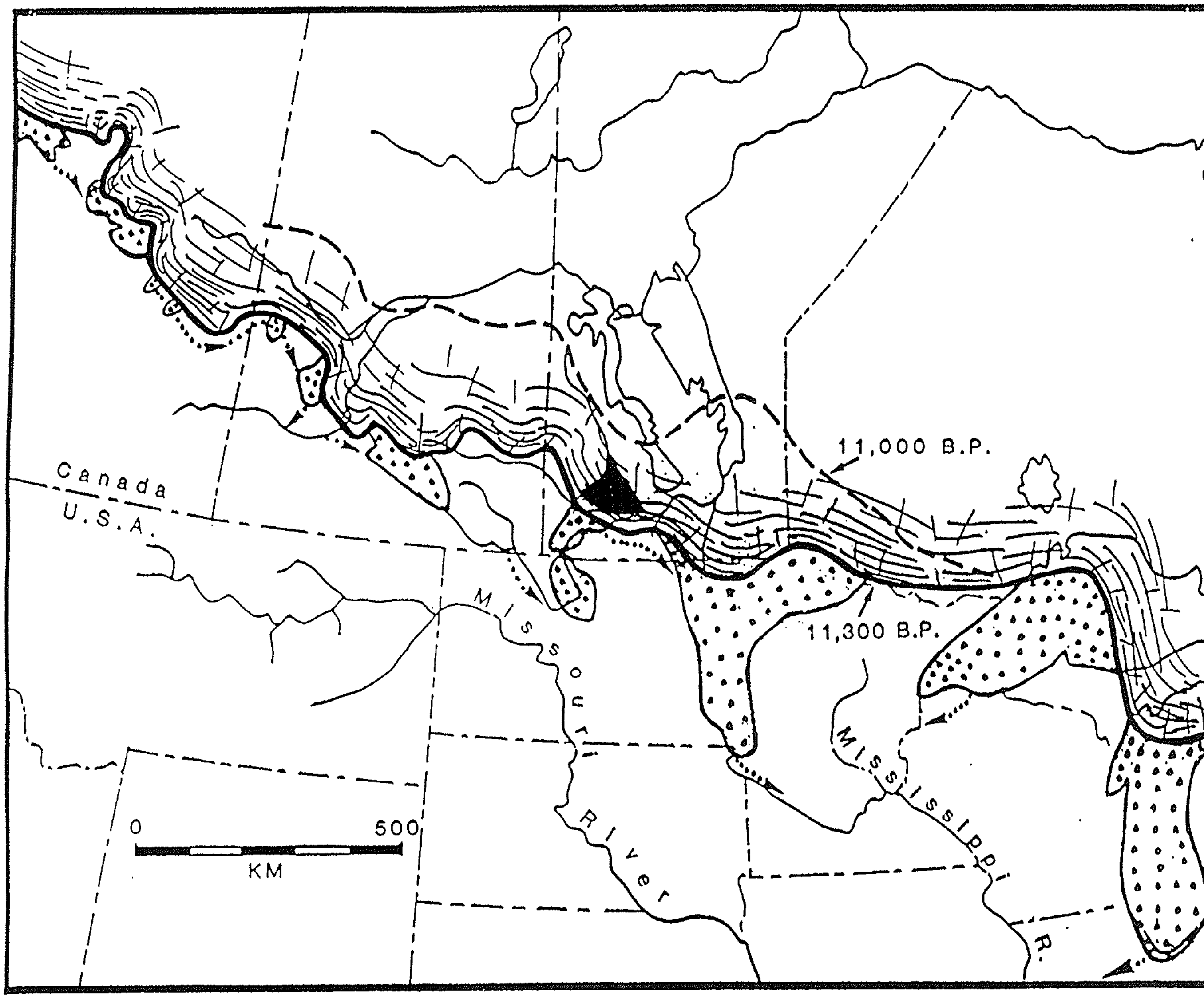


Fig. 7-3 Ice marginal position at 11,300 BP and 11,000 BP (from Teller, 1987, p.55). The Assiniboine fan delta is the shaded area.

drainage system into the Assiniboine fan delta.

The water level of Lake Agassiz was probably at about 400 m (an upper Herman Beach, Upham, 1890) when the first stage of the fan delta began, as indicated by the deposition of braided river deposits west of Brandon (E-1 in Fig. 3-2). The lake level fell to 380 m during the construction of the Gilbert-type delta. At the end of Lockhart phase the water level of Lake Agassiz had declined to the Campbell beach level (Figs. 7-1, 7-4).

Moorhead phase

"Shortly after 11,000 BP, the Lockhart phase came to an end when the ice margin wasted north of Thunder Bay, Ontario, allowing Lake Agassiz to overflow eastward into Lake Superior and drop below the level of the southern outlets" (Teller, 1985, p.4) (Figs. 7-1, 7-4). As one after another lower eastern outlets were opened, the Lake Agassiz water level declined rapidly, exposing large parts of the lake floor in the southern basin. Fluvial sediments, formed during the Moorhead low water stage and now buried by later Agassiz sediments, occur as far north as the international boundary (Arndt, 1977). In addition there is a pedogenic horizon in the Lake Manitoba basin sediments at an elevation of about 248 m (Teller and Last, 1981, 1982). Teller (1985) estimated that the lake could not have been deeper than 20 m at Winnipeg during the lowest level of the Moorhead phase. The implication is that the Assiniboine fan delta was exposed, and possibly was trenched by the Assiniboine River during the Moorhead phase. Importantly, this means that most of the Assiniboine fan delta was deposited between about 11,300 BP and shortly after 11,000 BP! Since then no major deposition has occurred in the fan delta except within the Assiniboine valley itself.

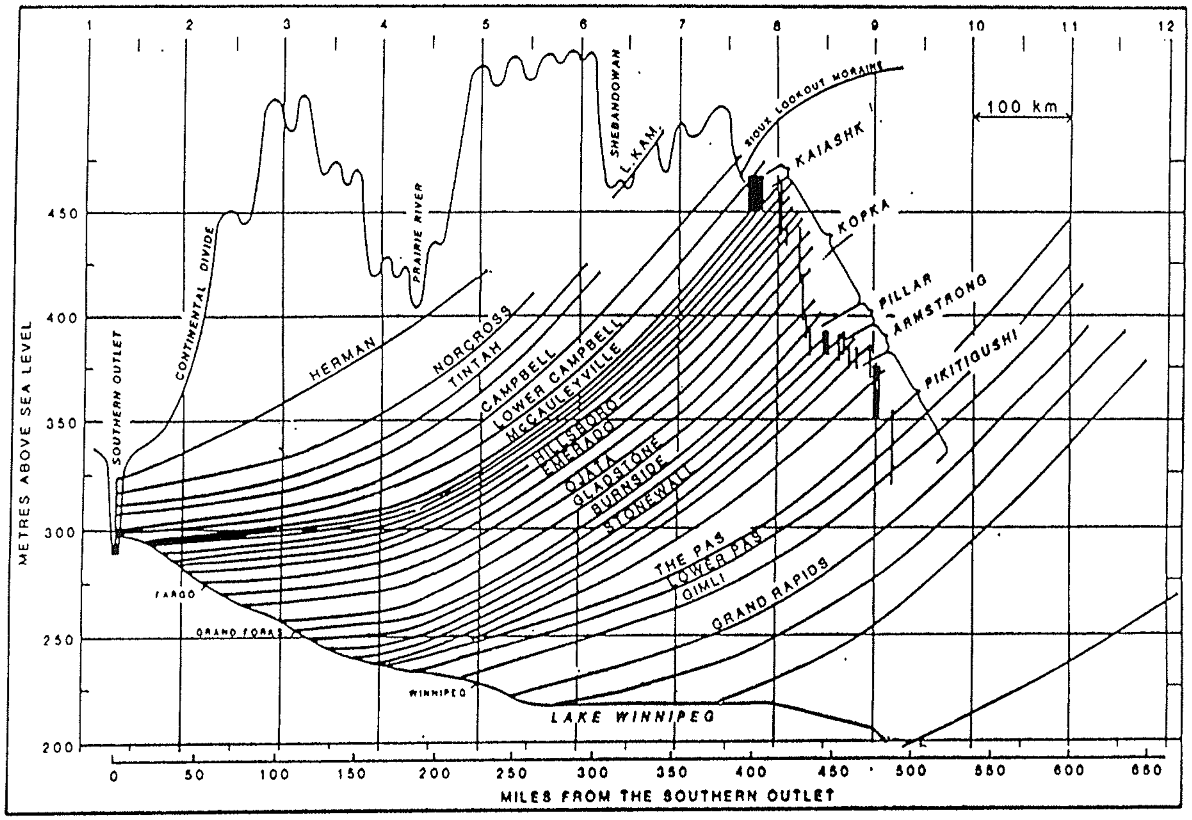


Fig. 7-4 Lake Agassiz strandline diagram (from Teller and Thorleifson, 1983, p.265).

Emerson phase

A major readvance of ice across the Superior Basin (the Marquette advance) dammed the eastern Lake Agassiz outlets (Teller, 1987, 1985; Teller and Thorleifson, 1983). The time of this advance has been placed at 10,000 BP by the dating of wood buried by glacial sediments of this advance in the Lake Superior basin (Drexler et al., 1983; Clayton, 1983), and wood and peat on the Moorhead phase unconformity (Elson, 1967; Clayton and Moran, 1982; Fenton et al., 1983, Teller, 1985). Lake Agassiz rose at least to the Campbell level at this time (Teller, 1985) and again began to overflow through the southern outlet. Warman (1991) and Thorleifson (1983) argued that the lake rose to the Norcross level (above the Campbell and slightly lower than the Herman level, see Fig. 7-4). If the later is true, part of the Assiniboine channel across the southern fan delta would have become an estuary, and river terraces should have been graded to lake level unless the lake level stood at the Norcross beach level for only a short time before it dropped to a lower level. Klassen's (1975, Table 3) study along the Assiniboine valley shows that the highest terraces were graded to the Tintah level (350 m), which is slightly lower than the Norcross level and higher than the Campbell level. This means that the main part of the fan delta surface, which is at an elevation of between 373 to 388 m, remained above the level of Lake Agassiz during the Emerson phase.

Nipigon phase

The Emerson phase came to an end by about 9500 BP (Teller, 1985, 1987). The level of Lake Agassiz fell rapidly from the Campbell level to the Gimli level, dropping 200 m in 1000 years (from 9500 to 8500 BP). The decline in lake level was due to the undamming of the lower eastern outlets of Lake Agassiz to Lake Nipigon. The Nipigon phase marked the end of most deposition in the Assiniboine fan delta, except for fluvial and aeolian reworking of the sediments during the Holocene.

Ojibway phase

The Nipigon phase came to an end because ice retreated north of the Nipigon basin and Lake Agassiz overflow was then directed toward the east into proglacial Lake Ojibway at about 8500 BP (Teller, 1987, 1985). By about 7500 BP, Lake Agassiz had drained into the Tyrrell Sea (Klassen, 1983).

The 'Older' Chronology of Lake Agassiz

Introduction

Klassen (1975, 1983) presented another version of deglaciation chronology for Lake Agassiz. Klassen's chronology will be discussed below because it was specifically about the chronology of the Assiniboine fan delta. The rational to reject Klassen's chronology will be presented after the summary of Klassen's version of Lake Agassiz. But it is important to emphasis that the *sequence* of events described by Klassen (1975, 1983) for the fan delta may still be used as an guideline to reconstruct the history of the Assiniboine fan delta, even if the absolute chronology is rejected.

Summary of Klassen's chronology of Lake Agassiz and the Assiniboine fan delta

Klassen's chronology included 3 major phases: (1) Formation of the Assiniboine delta; (2) Post-Assiniboine delta and pre-Campbell interval; (3) Formation of the Campbell beaches.

Formation of Assiniboine delta According to Klassen (1975), Lake Agassiz came to exist between 15,000 and 14,500 BP. Assiniboine fan delta building began more than 13,000 BP and was completed before 12,000 BP (Klassen, 1983). During the beginning of the delta building, the Keewatin Lobe stood along the northern margin of the delta (Klassen, 1983, Figure 5). Lake Agassiz level was at an elevation about 400 m.

This phase is equivalent to the Lockhart phase in the Teller (1985, 1987) and Fenton et al. (1983) version of Lake Agassiz, which they dated at 11,700-11,000 BP.

Post – Assiniboine delta and pre – Campbell interval : "this interval began from 12,500 BP and ended at 11,000 BP" (Klassen, 1983, p.222). During this period, the Keewatin and Hudson Lobes retreated northward, and Lake Agassiz fell to a low level that resulted in the excavation of the Assiniboine valley. A subsequent major readvance of the Hudson Lobe southward about 11,000 BP resulted in a rise in the level of Lake Agassiz to at least the Campbell level (Klassen, 1983, p.225, Figure 7). At this time Lake Agassiz water flooded the Assiniboine valley that had been incised in the delta during the previous low lake level, and deposited the oldest parts of the valley fill in the Assiniboine fan delta area.

The same kind of events have been described by Teller (1985, 1987) and Fenton et al. (1983) for the Moorhead low water phase (11,000-9900 BP, low lake level) and Emerson phase when the lake rose to at least the Campbell level of Lake Agassiz.

Formation of the Campbell beaches : "Between about 11,000 to 9700 BP the Campbell beach formed." (Klassen, 1983). That means that the lake stood at the Campbell level at about 11,000 BP and again at about 9700 BP, with a lower lake level in between. During this time the young alluvial fill in the delta area was deposited. It seems that this interval is equivalents to Teller's (1985, 1987) Moorhead and Emerson phases of Lake Agassiz (11,000-9500 BP).

Problems with Klassen's chronology Klassen's classification of phases for the Assiniboine fan delta, and therefore for Lake Agassiz, was mainly based on radio-

Table 6 Radiocarbon dates used by Klassen to establish the chronology of the Assiniboine fan delta (from Klassen, 1983, p.215-216).

Years BP	Lab No.	Material	Lat.	Long.	References	Comments
Dates from deposits younger than main part of Assiniboine delta but older than Campbell Beach						
13 900 ± 240	I-3476	organic detritus	49°50′N	99°35′W	Ritchie, 1976, p. 1799	Brandon; dated sediment was from a post-glacial lake within a swampy flat on the upper part of Assiniboine delta at *ca.* 375 m a.s.l.
12 400 ± 420	Y-165	peat	49°47′N	98°35′W	Preston *et al.*, 1955	Rossendale; alluvial fill *ca.* 4 m below surface of abandoned channel on delta at *ca.* 328 m a.s.l.
12 100 ± 160	GSC-1319	peat	49°47′N	98°35′W	Lowdon *et al.*, 1971, p. 282	Duplicate of Rossendale sample Y-165; ident. by M. Kuc as peat-moss (*Scorpidium scorpioides*).
Dates from deposits related to Campbell Beach						
10 600 ± 150	GSC-902	plant detritus	49°45′N	98°39′W	Lowdon and Blake, 1970, p. 65	Rossendale; fluvio-lacustrine sediment fragments *ca.* 18 m below Campbell terrace at 320 m a.s.l.
10 300 ± 200	BGS-617	bone	52°11′N	101°26′W	Teller, 1980, p. 6	Swan River, water worn bison bone fragment between upper and lower Campbell beach at *ca.* 344 m a.s.l.
10 200 ± 80	GSC-1909	organic detritus	49°06′N	96°14′W	Lowdon and Blake, 1976, p. 7	Sundown; *ca.* 8 m below surface at Campbell strandline at *ca.* 326 m a.s.l.
10 000 ± 150	GSC-870	wood	49°45′N	98°45′W	Lowdon and Blake, 1970, p. 65	Rossendale; alluvium, *ca.* 8 m below surface of Campbell terrace in Assiniboine valley across delta at *ca.* 320 m a.s.l.
9900 ± 160	GSC-391	wood	49°00′N	95°14′W	Lowdon *et al.*, 1967, p. 10	Buffalo Point; *ca.* 2 m below Campbell terrace surface at *ca.* 323 m a.s.l. lower Campbell beach
9700 ± 140	GSC-797	wood	49°45′N	98°39′W	Lowdon and Blake, 1979, p. 65	Rossendale; alluvium, *ca.* 4 m below surface of Campbell terrace at *ca.* 320 m a.s.l. in Assiniboine valley across delta

carbon dates list in Table 6. The validity of the first 3 key dates (13,900 ±240, I-3476; 12,400 ±420, Y-165; 12,100 ±160, GSC-1319) has been challenged by Teller (1989).

The organic detritus, which yielded a date of 13,900 ±240 BP, was comtaminated (Klassen, 1989; Teller, 1989). The other two old dates (12,400 ±420 BP, Y-165; and 12,100 ±160 BP, GSC-1319), which have been used by Klassen to establish the 'old' chronology, have been discussed by Teller (1989). These two dates are both from the subaquatic moss *Scorpidium scorpioides* (Lowden et al., 1971), and may be subject to the hard-water effect (Teller, 1989). Subsequent dates on wood from the same zone gave two younger dates, 9600 ±70 BP (TO-534) and 9510 ±90 BP (GSC-4490) (Teller, 1989).

Since all of Klassen's dates for an "old" chronology have been rejected, there are no grounds for an old chronology for the Assiniboine fan delta. However, the relative sequence of events described by Klassen (1975, 1983) for the fan delta is correct, and will be used as a guideline for reconstructing the Assiniboine fan delta history.

Summary of Deglacial History

(1). By 11,700 BP, Lake Agassiz was established in the southern end of the Lake Agassiz basin due to the northward retreat of the Red River Ice Lobe, which was a sublobe of the Labradorean Ice.

(2). The Lake Hind basin (southwestern Manitoba), the Souris River drainage system, and part of the Assiniboine drainage system became ice free due to northward retreat of the Keewatin Ice Lobe.

(3). By 11,400 BP, the Red River Ice Lobe readvanced southward to the Darlingford moraine and southwestward to the Alexander moraine.

(4) The southern Assiniboine fan delta area became ice free due to northeastward retreat of the Red River Ice Lobe, shortly after 11,300 BP. Deposition of the Assiniboine fan delta started after the Alexander moraine failed along the Assiniboine bedrock valley. By this time, the water level of Lake Agassiz was at about 400 m. During the next several hundred years, the lake level fell to the Campbell level.

(5) The Moorhead phase began shortly after 11,000 BP when the level of Lake Agassiz fell below the Campbell beach at 320-327 m. Most parts of the fan delta were exposed during this phase.

(6) The water level rose again to at least the Campbell level at about 9900 BP, and stood at the Campbell level until about 9500 BP. During this time (the Emerson phase), river terraces across the southern fan delta probably were graded to the lake level.

(7) Lake level fell below the Campbell level during and after the Nipigon phase (9500-8500 BP).

Chapter 8

INTEGRATION OF THE ASSINIBOINE FAN DELTA HISTORY WITH THE LAKE AGASSIZ HISTORY

Pre-Lake Agassiz Phase

After the maximum advance of Late Wisconsinan ice into central Iowa at 14,000 BP (Ruhe, 1969; Teller, 1987), the ice began to retreat north toward the Lake Agassiz basin. At about 12,300 BP, the ice margin was in South Dakota and across northern Iowa (Fenton et al., 1983). By 11,700 BP, the Red River Lobe retreated into the Lake Agassiz basin, and therefore marked the birth of Lake Agassiz (Fig. 7-1).

Case-Lockhart Phase of Lake Agassiz

Pre-Assiniboine fan delta (11,700 to 11,300 BP)

As ice continued to retreat, Lake Agassiz expanded northward. By 11,500 to 11,300 BP, Lake Agassiz had expanded to the International Boundary (Teller, 1987). At the same time, Lake Souris formed in the Souris basin to the west of Agassiz, as the result of northward retreat of the Keewatin Lobe (Klassen, 1975). Lake Souris received runoff from much of the newly deglaciated region to the west. Initially Lake Souris overflowed into the southern end of Lake Agassiz through the Sheyenne Spillway of North Dakota (Teller, 1987; Fenton et al., 1983). As the Keewatin Lobe wasted northward to open the Souris River valley and as the southern Tiger Hills gradually became ice free, drainage from Lake Souris abandoned the Sheyenne spillway and flowed through the Souris River valley to Lake Hind which, in turn, overflowed through the Pembina Spillway into Lake Agassiz (Kehew and Clayton, 1983). Kehew and Clayton (1983) presented evidence to show that the flow from

Lake Souris to Lake Hind was a catastrophic flow, and was part of a domino-like series of floods from Lake Regina in central Saskatchewan via Lake Souris and Lake Hind, to Lake Agassiz.

Reactivation of the Red River ice about 11,400 BP pushed the ice margin southward to the Edinburg moraine in North Dakota (Fenton et al., 1983) and the Darlingford moraine (Conley, 1986) and Alexander moraine (Fig. 7-3) in southwestern Manitoba. Retreat of the Keewatin Lobe from this area allowed the formation of glacial Lake Hind in southwestern Manitoba, which was ponded by the topographic higher area to the west and south, and the Red River Ice Lobe to the east (Fig. 7-3).

After this advance, perhaps shortly after 11,300 BP (see Fig. 7-2), as the Red River ice margin retreated northeastward, a proglacial lake named Lake Brandon by Elson (1967) came to exist between the retreating ice margin and the Alexander moraine to the west (Fig. 8-1A). The evidence for this lake is the deposition of silty clay to an elevation from 405 to 442 m northwest of Brandon and east of the Alexander moraine (Fig. 3-3). After a short period, this lake merged with Lake Agassiz when the ice margin retreated farther. But deposition of the Assiniboine fan delta did not begin because there was no major sediment feeding system. The Assiniboine River and water in the Lake Hind basin were still prevented by the Alexander moraine from flowing to Lake Agassiz througth the Brandon area (Fig. 8-1A).

Early fan delta phase

Deposition of the Assiniboine fan delta started when Lake Hind water, impounded to the west of the Alexander moraine, flooded into Lake Agassiz after the water breached the Alexander moraine along the Assiniboine River bedrock valley.

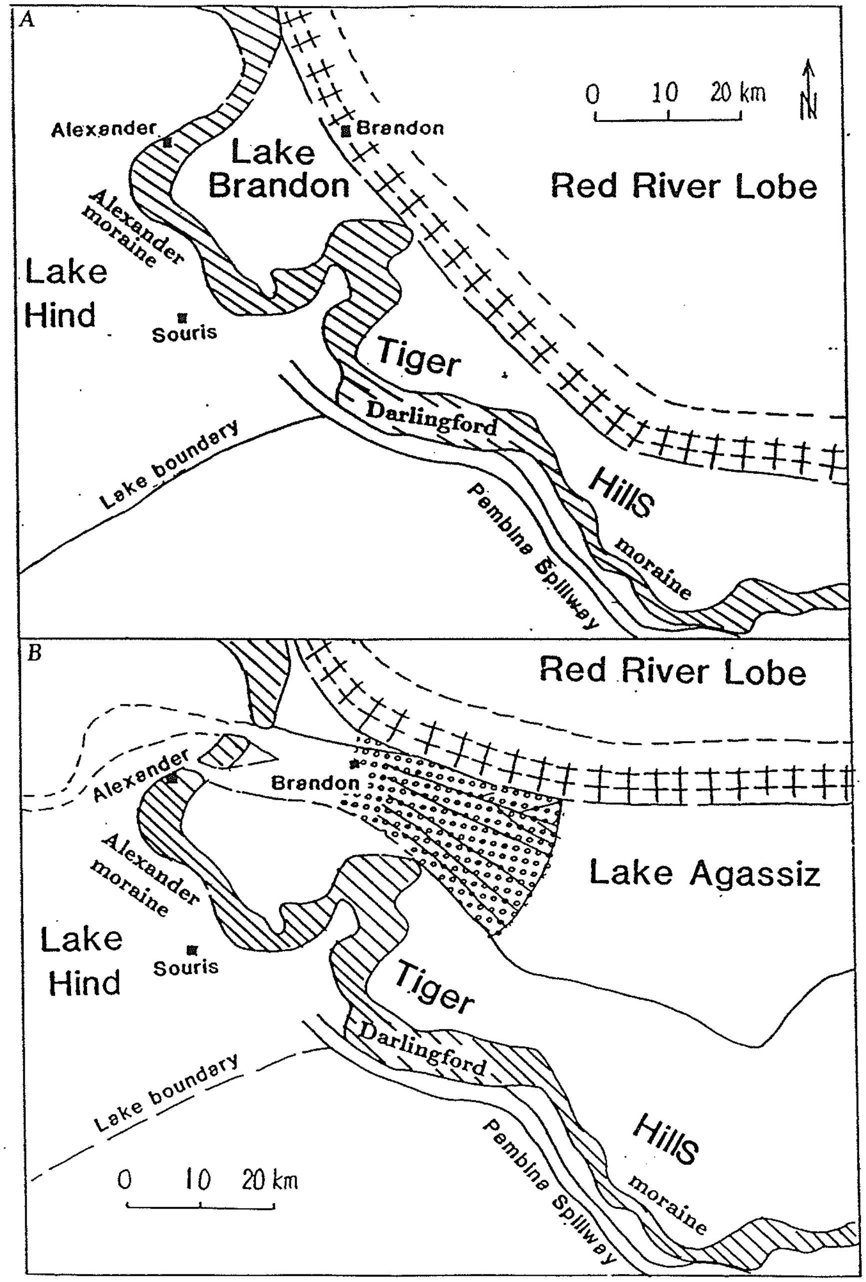


Fig. 8-1, A: Formation of the glacial Lake Brandon between the Alexander moraine and the retreating Red River Lobe.

B: Early Assiniboine fan delta phase. A Catastrophic flood, which occurred after the breach of the Alexander moraine, eroded the area it passed (Zone 1 and Zone 2 in Fig. 3-1), and deposited gravels in the proximal fan delta area. The ice was probably stood in the northern fan delta area (after Klassen, 1983, p.223).

The result was probably a catastrophic flood, which produced many scoured channels in till and left patches of coarse gravels and boulders in scoured channels west of Brandon (see Chapter 3). Silty clay sediments deposited during the Lake Brandon time were scoured away in the river plain (west of Brandon) and in the proximal fan delta area (Figs. 8-1B and 8-2B), which resulted in deposition of poorly sorted, massive cobbles, pebbles, and some boulders at the base of the fan delta, overlying the till (Figs. 5-2 and 8-2B). It is not certain where the margin of the ice was at that time. Klassen (1983, Figure 5) proposed an ice margin across the northern fan delta area from west to east in the early Assiniboine fan delta phase. This explain why there are no gravels at the base of the northern fan delta area. In contrast, coarse gravels overlie till (see chapter 4, RC-1) or silty clay (see chapter 4, ADA-50) in the southern fan delta area. Thus it is likely that the ice was in the northern fan delta area, which directed flood water to the southern fan delta (Fig. 8-1B).

Braided river plain phase

Most scoured channels produced by the initial flood now contain fluvial braided channel sand and fine gravels (Facies St). As cuts west of Brandon show (Figs. 4-1, 4-2, 4-3, 4-4, 3-2), these sand and fine gravels are thin and contain cut and fill structures. For these reasons, and because there are no coarse-grained beds (all are smaller than cobble size), normal meltwater runoff seems to have been dominant during the early post flood period. It seems likely that the braided river plain, which is graded to about 400 m at Brandon (E-1 and E-2 in Chapter 4), was deposited during the time when Lake Agasssiz was at the high Herman beach level (about 400 m, Upham, 1890) (Fig. 8-2B)

Fan delta progradation phase

When the level of Lake Agassiz fell to an elevation of 380 m, discharge to the

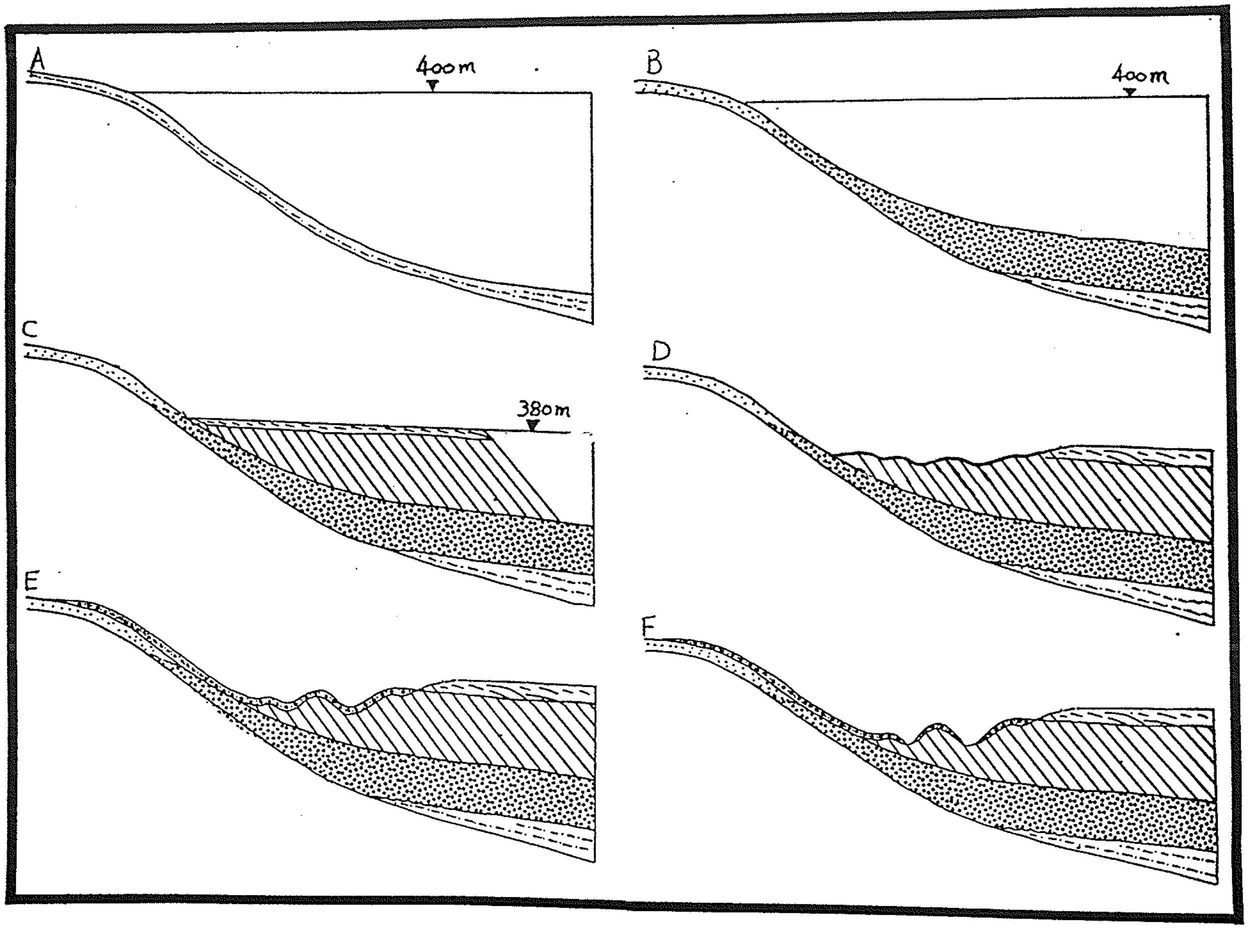


Fig. 8-2 Cross section showing the evolution of the proximal fan delta.

A. Pre-fan delta phase, water level = 400 m.

B. Early fan delta phase, water level = 400 m.

C. Gilbert fan delta phase, water level = 380 m

D. Post gilbert delta phase, lake level below 380

E. Deposition of debris flow deposits

F. Erosion of the debris flow deposits

Assiniboine fan delta strengthened, evidenced by the large clast size in the proximal fan delta as compared to that in the braided river plain to the west. Several flood bursts from the upper reaches of the Assiniboine valley (Wolfe, personal comm. 1992) and the Qu'Appelle valley (Teller, personal comm. 1992) deeply eroded the Assiniboine and Qu'Appelle valleys, and generated flows highly charged with sediment eroded along the valleys. Gravels by these flood bursts, which are exposed in the No.1 hill and No.3 and 4 ridges, show large scale cross bedding with normal grading across the cross-stratum (Facies Gx, Figs. 4-8, 4-9, 4-13, 4-14, 4-18, 4-19), which are believed to have been formed by fully turbulent debris flows that generated strong flow separation and expansion in the lee slope of the advancing foresets. It is believed that each flood was responsible for the deposition of many foreset beds, because there is no evidence of lacustrine clay beds or sand partings deposited between foreset beds, and the contacts between foresets are sharp and straight, with no traces of erosion. After a series of flood pulses, the gravelly Gilbert delta extended at least 15 km east into the Lake Agassiz (Figs. 8-2C, 8-3). This period of delta construction probably began about 11,200 BP and ended about 11,100 BP.

After deposition of the Gilbert-type foresets, lake level fell slightly, which resulted in the erosion of the delta foresets along the broad and shallow Brandon-Chater-Douglas station channel (Fig. 8-2D, 8-4) (but formation residual hills were not complete until a later event). For an unknown period of time during the Lockhart phase, meltwater discharged through this channel into the deep water of Lake Agassiz. Because of the heavy sediment load, the incoming flows became underflows once they entered the lake, depositing sediments onto the lake floor as the flow dispersed. As the sediment supply continued, the underflow fan delta migrated offshore and resulted in an overall coarsening upward sequence in the central and

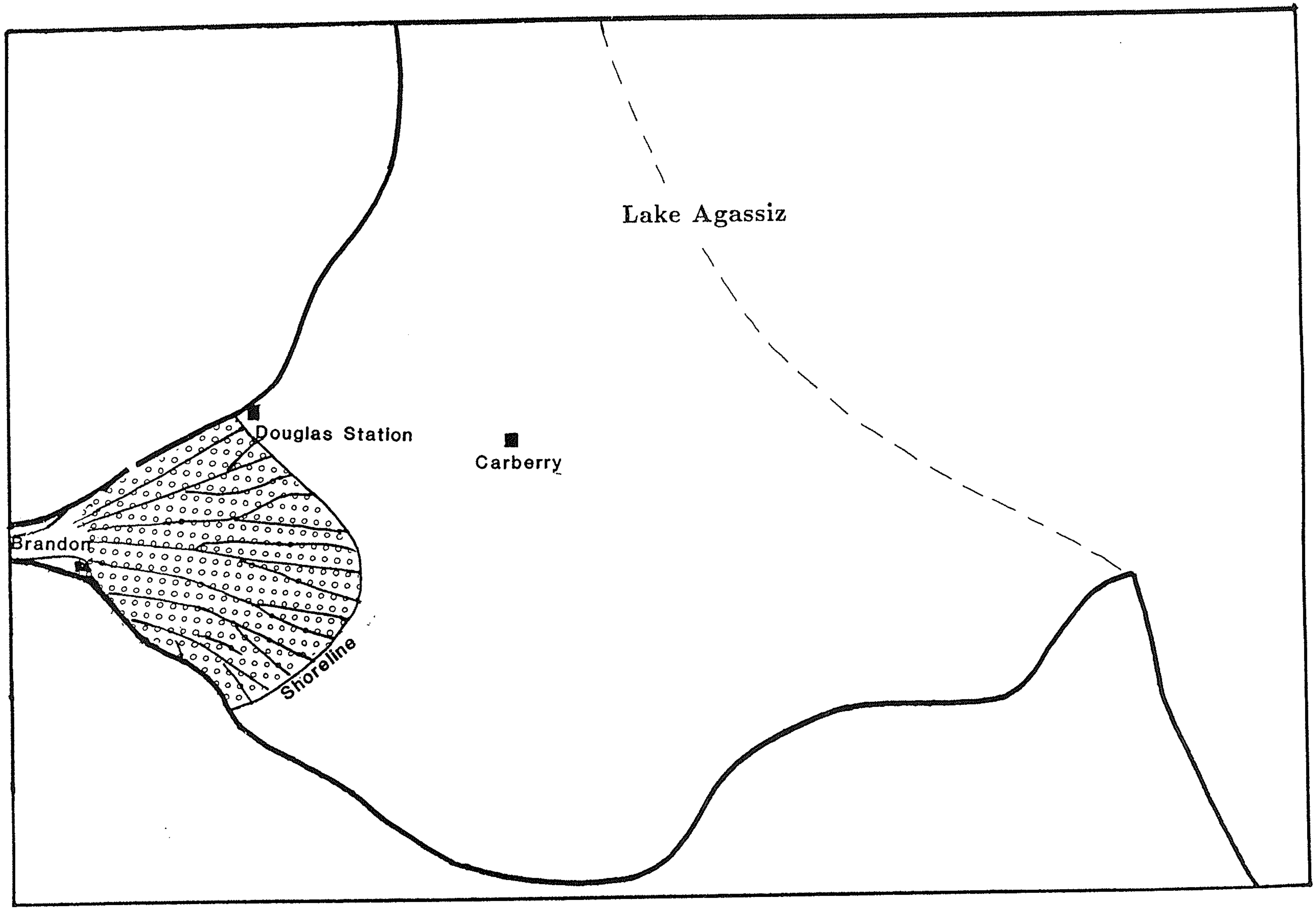


Fig. 8-3 Fan delta progradation-Gilbert type delta phase. Lake level was at an elevation about 380 m.

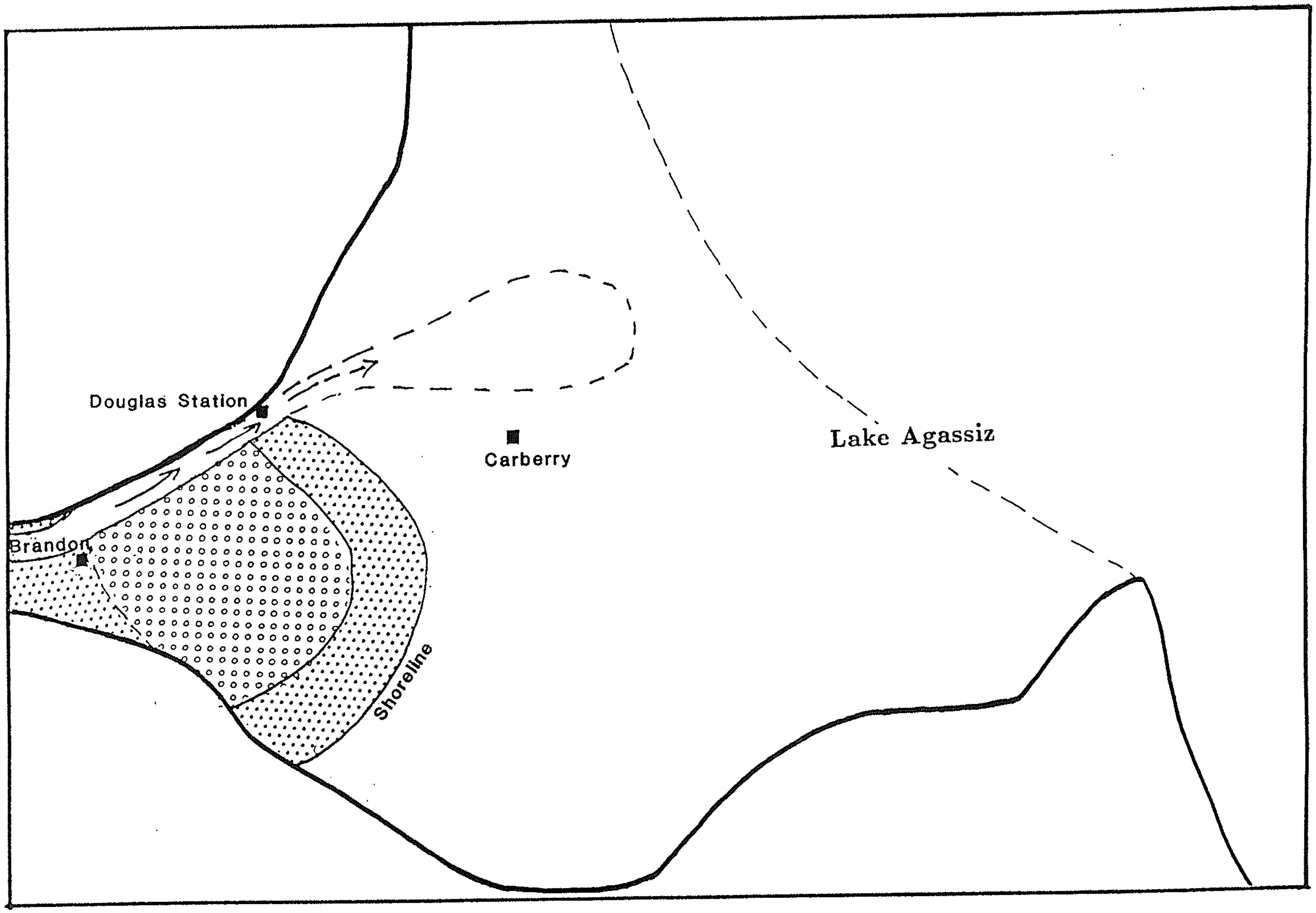


Fig. 8-4 Fan delta progradation-underflow fan delta phase. The Assiniboine River entrenched abroad and shallow channel across the northern proximal fan delta. Lake level was at about 370 m.

northern Assiniboine fan delta (see Chapter 4, 90AD-1 and 90AD-2). The gradual northward shift of the delta lobe resulted in a fining upward sequence in the southern fan delta (see Chapter 4, ADA-50, RC-3). This period of subaqueous fan accumulation ended shortly after 11,000 BP, when the level of Lake Agassiz fell during the Moorhead phase.

The lake level may have fluctuated during the late Lockhart phase of Lake Agassiz. This is indicated by a 5 point running mean of grain size from Units F to G in 90AD-2 (Fig. 4-34). Elson (1967) suggested, based on his study of the Treesbank Ferry section, that Lake Agassiz briefly discharged into the Lake Superior basin due to the opening of the eastern outlets. This was followed by a brief rising of lake level due to closing of the eastern outlets by advancing ice. After that, Lake Agassiz got back to its steady falling course. The brief rising of lake level may have resulted in a fining upward sequence, whereas the subsequent steady falling of lake level may result in a coarsening upward sequence overlying the fining upward sequence. Therefore, grain size from Units F to G in 90AD-2 (Fig. 4-34) may reflect the transgression to the norcross level (Elson, 1967) during the Lockhard phase, which was followed by a regressive sequence (a coarsening upward cycle) that ended with the abrupt decline in lake level at the start of the Moorhead phase about 11,000 BP.

It is not certain when the massive, non-graded, and poorly sorted sandy gravels (Facies Gm) were deposited on top of the braided river plain and the proximal fan delta near the Assiniboine channel (Figs. 8-2E, 6-1). This subaerial debris flow must have been deposited after the southern route of the Assiniboine valley was established across the fan delta because they occur in the vicinity of the modern Assiniboine River valley. The same event may also have deposited coarse fluvial sediments in the lower reaches of the valley. Medium to fine grained, fluvial pebble gravels in the Steels Ferry section north of Glenboro (99°14′, 49°42'), described by Elson (1967), are overlain by about 2.5 m interbedded sand and silty clay, which in turn, is overlain by marl. Marl, believed to be the same age as that in the Steel Ferry section (Elson, 1967), overlies about 2.5 m of silt and sand in the Laveham section and gave a radiocarbon age of 10,600 ± 150 BP (GSC-383, Lowdon et al., 1967). Wood in the sand and silt bed, 1.5 m below GSC-383, yielded an age of 10,550 ± 200 BP (Y-411, Barendson et al, 1957). This suggests that the Assiniboine River valley across the southern fan delta was eroded before 10,600 BP, possible at the start of Moorhead stage of Lake Agassiz.

All residual hills and scoured channels occur in the southern fan delta (Fig. 3-2), in the same area as the subaerial debris flow deposits (Figs. 3-2, 6-1). In the northern fan delta area where there are no residual hills, scoured channels, or grooves, there are no subaerial debris flow deposits. The massive, nongraded gravels (debris flow deposits) draped over residual hills, suggesting that, after the delta was exposed and after the northern fan delta became higher than the southern area due to isostatic rebounding, floods (possibly catastrophic) prefered to flow along the lower southern fan delta and deposited massive, poorly sorted sheet of gravels over the eroded delta surface (Fig 8-2E). Later events may have remoulded the residual hills, eroded the debris flow deposits in channels, and exposed tills on the floors of the scroured channels (Fig. 8-2F).

Moorhead Phase of Lake Agassiz

The opening of the eastern outlets caused the water level to drop below the level of the southern outlet and the Campbell beach, thus initiating the Moorhead phase of Lake Agassiz (10,800-9900) (Fig. 7-1). During this period, the entire upper fan delta area west of the Campbell beach was exposed, and nearly all sedimentation ended except along the northern and southern route of Assiniboine valley. Two radiocarbon dates from the Lavenham section, one on marl (10,600 $\pm$ 150 BP, GSC-383), the other one on wood (10,550 $\pm$ 20 BP, Y-411) (Elson, 1967), and the similarity between the Lavenham section and the Steels Ferry section that is located along the modern (southern) route of the Assiniboine River near Glenboro (Elson, 1967), suggest that the modern (southern) route of the the Assiniboine River valley was incised (Fig. 8-5) near the end of Lockhart phase or in the early Moorhead phase. At the same time the broad and shallow channel in the northern fan delta (along the Epinette Creek) was abandoned.

Emerson Phase of Lake Agassiz

The Moorhead phase ended and the Emerson phase started when the Marquette advance of ice in the Lake Superior basin again blocked the eastern outlets of Lake Agassiz (Fig. 7-1). The closure of the eastern outlets caused the lake to rise, and because of that, the lower reaches of the Assiniboine valley across the fan delta became an estuary. This led to deposition of a fining upward sequence of channel fill sediments, such as at the Glenboro section (RC-5, Table 5 in Chapter 4), which consists of sand and gravel that contains cut and fill structures and is overlain by sandy silt and silty sand that contains snail shells. The maximum elevation of river terraces along the Assiniboine channel across the fan delta reaches an elevation of 350 m, which is about the Tintah level (Klassen, 1975, Table 3). The river terrace north of Glenboro at RC-5 section is at an elevation of 335 m, which is about the Upper Campbell level (Fig. 8-6). A date on wood in these gravel of 9880$\pm$225 (GX-3696) indicates deposition during the Emerson phase of Lake Agassiz. This suggests that the lake level in the early Emerson phase (9900-9500 BP) rose to the

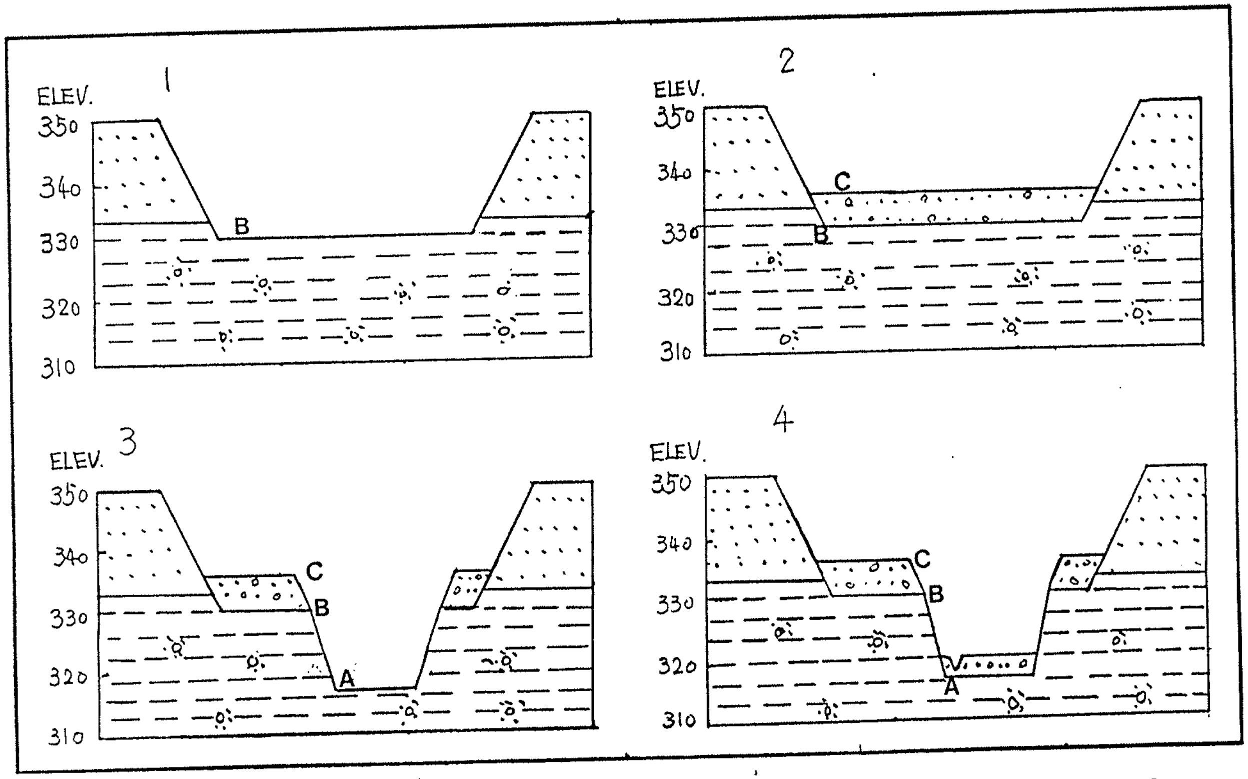


Fig. 8-6 Evolution of river terrace at RC-5. 1) the Assiniboine channel was entrenched into level 'B' during the early Moorhead phase. 2) Coarse channel fill sediments were deposited mainly in early Emerson phase after the lake level rose to a higher level (Tintah level). After several hundreds years, sediments were built to level 'C'. 3) Subsequent falling of lake level during the Nipigan phase led to the incision of the channel into channel fill and lacustrine clay (to Level A). 4) Subsequent deposition has formed the modern floodplain.

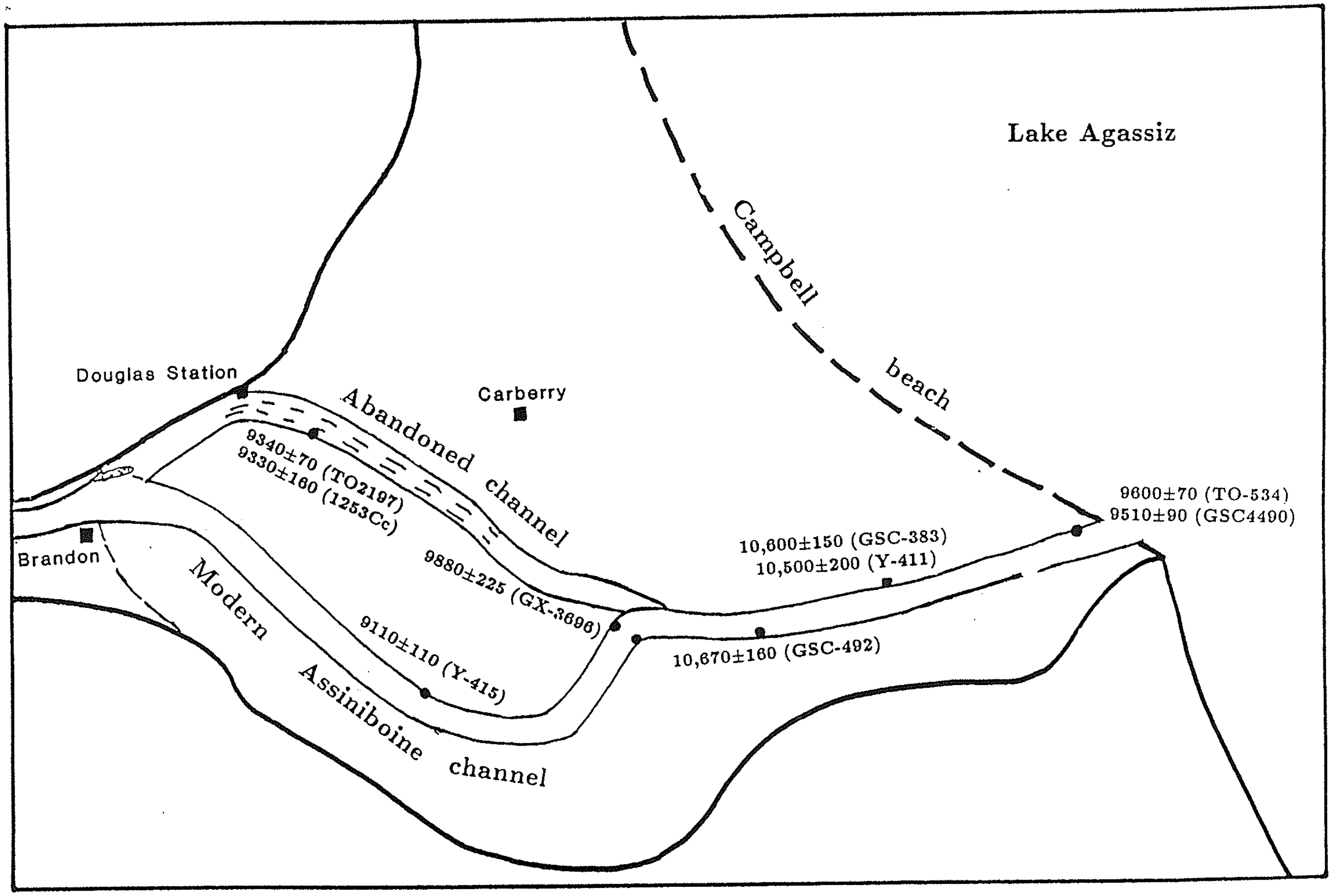


Fig. 8-5 Key radiocarbon date on wood in the Assiniboine fan delta area. Black dots indicate location of samples. Notice the oldest radiocabon date along the Assiniboine River channel is 10,670 ±160 (GSC-492), suggesting that the Assiniboine River channel was excavated before before 10,670 BP.

Tintah level for a short period of time. It fell to the the Campbell level shortly afterward.

Nipigon Phase of Lake Agassiz

The lake level fell during the Nipigon phase of Lake Agassiz (9500-8500 BP) due to the opening of lower eastern outlets (Teller and Thorleifson, 1983). The Assiniboine River cut a deeper channel into the former flood plain (Emerson phase deposits) and the pre-delta silty clay (Lockhart phase deposits) of the southern route (Fig. 8-6). Along the abandoned northern Assiniboine channel, silty clay and marls with abundant organic material was deposited, Two radiocarbon dates, one on wood in silty clay (9330±160 BP, 1253Cc), another on marls (9340±70 BP, TO2197), suggest that the silty clay and marls were deposited during the Nipigon phase of Lake Agassiz.

SUMMARY OF THE LATE WISCONSINAN HISTORY IN THE STUDY AREA

Lake Agassiz formed in the southern end of the basin at about 11,700 BP as ice of the Red River Lobe retreated northward. The lake expanded into southern Manitoba by about 11,500 BP.

Readvance of ice at about 11,400 pushed the ice margin southward to the Darlingford moraine and Alexander moraine in southwestern Manitoba. Retreat of ice from the last advance formed Lake Brandon between the retreating ice margin and the Alexander end moraine. Lake Brandon merged into Lake Agassiz as the ice retreated farther.

Deposition of the Assiniboine fan delta began shortly after 11,300 by a catastrophic overflow from Lake Hind and/or the Assiniboine River which produced many scoured channels into till west of the fan delta. Because the ice margin at that time lay in the northern Assiniboine fan delta area, the flood water was directed by the ice to the southern fan delta area, depositing in the proximal region poorly sorted cobbles, pebbles, and some boulders at the base of the fan delta sequence there.

Following the initial flood, the incoming flows from the Assiniboine River were dominated by normal meltwater flows, which deposited tabular cross bedded and trough cross bedded sand by migration of dunes and bars across the braided channel. When the lake level fell to 380 m, the Herman beach level of Lake Agassiz, a series of floods, possibly catastrophic, deposited a gravelly Gilbert-type delta. Each foreset bed of the Gilbert-type delta consists of normally graded, well sorted boulders and cobbles grading across the cross-bed to pebbles and granules. A pulsing flood flow is thought to have been responsible for these growing, graded foresets.

A further decline in lake level led to the formation of elongated erosional residual hills near the apex of the fan delta, and to the erosion of a broad and shallow channel across the northern fan delta. Meltwater with high sediment load flowed through the northern (Epinette) channel to the middle fan delta area, where the flows became underflows and deposited the main part of the underflow fan delta into Lake Agassiz.

Shortly after 11,000 BP, at the start of the Moorhead phase, Lake Agassiz fell far below the Campbell beach level because of the opening of the eastern outlets. The Assiniboine River abandoned the northern channel and and flowed along the southern route. Shortly after that, a subaerial debris flow deposited sheet-like, massive, nongraded gravels over the braided river plain and along the southern Assiniboine channel. Subsequent floods remoulded the residual hills.

The Moorhead phase ended and the Emerson phase started at about 9900 BP when the eastern outlets were again blocked by ice, which caused the lake to rise. As a result, the lower reaches of the now-entrenched Assiniboine valley became an estuary. The surface of the channel fill sediments indicates that the lake reached a maximum at the Tintah level.

Falling of the lake level in the Nipigon phase caused the entrenchment of the Assiniboine River channel into the channel fill deposits, and formed paired and unpaired terraces along the channel. In the abandoned Assiniboine channel across the northern fan delta, silty clay with abundant organic material was deposited, together with marls. A portion of the abandoned valley was filled later by migrating sand dunes.

REFERENCES CITED

Allen, J.R.L. 1980, Sand waves: a model of origin and internal structure: Sediment Geology, v.26 p.281-328.

Arndt, B.M., 1977, Stratigraphy of offshore sediment of Lake Agassiz, North Dakota: North Dakota Geological Survey Report of Investigation 60, 58p.

Arnott, R.W.C., and Hand, B.M., 1989, Bed-forms, primary structures and grain fabric in the presence of sediment rain: Journal of Sedimentary Petrology, v.59, p.1062-1069.

Barendsen, G.W., Deevey, E.S. and Gralenski, L.J., 1957. Yale natural radiocarbon measurements III: Science, v.126, p.908-919.

Blatt, H.; Middleton, G.; Murray, R., 1980, Origin of sedimentary rocks, Prentic-Hall, Inc., Englewood Cliffs, New Jersey, 782p.

Clayton, L., 1983, Chronology of Lake Agassiz drainage to Lake Superior: in Teller, J.T. and Clayton, L. (eds): Glacial Lake Agassiz. The Geological Association of Canada, Special Paper 26, p.291-307.

Clayton, L., and Moran, S.R., 1982, Chronology of lake Wisconsinan glaciation in middle north America: Quaternary Science Reviews, v.1, p.55-82.

Clayton, L., Teller, J.T., and Attig J.W., 1985, Surging of the southwestern part of the Laurentide Ice Sheet: Boreas, v.15, p.235-241. v.1, p.55-82.

Conley, G.G., 1986, Surficial geology and stratigraphy of the Killarney-Holmfield area, southwestern Manitoba: MSc Thesis, University of Manitoba, Winnipeg, Manitoba, 168p.

Davis, J.L. and Annan, A.P., 1989, Ground-penetration radar for high resolution mapping of soil and rock stratigraphy: Geophysical Prospecting, v.37 p.531-551.

Drexler, C.W., Farrand, W.R., and Hughes, J.D., 1983, Correlation of glacial lakes in the Superior Basin with eastward discharge events from Lake Agassiz: in Teller, J.T. and Clayton, L. (eds): Glacial Lake Agassiz. The Geological Association of Canada, Special Paper 26, p.309-329.

Dreimanis, A., and Lundquist, J., 1984, What should be call till?: in Konigsson, L.K. (ed), 10 years of Nordic research, Striae, v.20, p.5-10.

Elson, J.A., 1956, Surficial geology of the Tiger Hills regions: Unpubl. Ph.D. Thesis, Yale University. New Haven, Connecticut, 316p.

Elson, J.A., 1960, Surficial geology, Brandon, Manitoba: Geological Survey of Canada, Map 1067.

Elson, J.A., 1967, Geology of glacial Lake Agassiz: in Mayer-oakes, W.J., (ed), Life, land and Water, University of Manitoba Press, Winnipeg, p.36-95.

Fenton, M.M., Moran, S.R., Teller, J.T., and Clayton, L., 1983, Quaternary stratigraphy and history in the southern part of the Lake Agassiz Basin: *in* Teller, J.T. and Clayton, L. (eds): Glacial Lake Agassiz. The geological Association of Canada, Special Paper 26, p.49-74.

Folk, R.L., 1966, A review of grain size parameters: Sedimentology, v.6, p.73-93.

Friedman, G.M., 1961, On sorting, sorting coefficients, and the lognormality of the grain-size distribution of sandstones: American Journal of Science, v.260, p.737-753.

Halstead, E.C., 1959. Ground-water resources of the Brandon map-area, Manitoba: Geological Survey of Canada Memoir 300, 67p.

Hwang, I.G., and Chough, S.K., 1990, The Miocene Chunbuk Formation, southeastern Korea: marine Gilbert-type fan-delta system: *in* Colella, A., and Prior, D.B., (eds): Coarse-grained delta, Blackwell Scientific Publications, p.235-254.

Johnston, W.A., 1934, Surface deposits and groundwater supply of Winnipeg area: Manitoba, Geological Survey of Canada, Bulletin 7, 20p.

Kehew, A.E. 1982, Catastrophic flood hypothesis for the Souris Spillway, Saskatchewan and North Dakota: Geological Society of America Bulletin, v.93, p.1051-1058.

Kehew, A.E. and Clayton, L. 1983, Late Wisconsinan floods and developments of Souris-Pembina Spillway system in Saskatchewan, North Dakota, and Manitoba: *in* Teller, J.T. and Clayton, L. (eds): Glacial Lake Agassiz. The geological Association of Canada, Special Paper 26, p.187-210.

Kehew, A.E., and Lord, M.L., 1986, Origin and large-scale erosional features of glacial lake spillways in the northern great plains: Geological Society of America Bulletin, v.97, p.162-177.

Klassen, R.W., 1972, Wisconsin events and the Assiniboine and Qu'Appelle valleys of Manitoba and Saskatchewan: Canadian Journal of Earth Sciences, v.9, p.544-560.

Klassen, R.W., 1975, Quaternary geology and geomorphology of Assiniboine and Qu'Appelle valleys of Manitoba and Saskatchewan: Geological Survey of Canada, Bulletin 228, 61p.

Klassen, R.W., 1979, Pleistocene geology and geomorphology of The Riding Mountain and Duck Mountain areas, Manitoba- Saskatchewan: Geological Survey Canada Memoir 396, 52p

Klassen, R.W., 1983, Assiniboine delta and the Assiniboine-Qu'Appelle valley system-implications concerning the history of Lake Agassiz in Southwestern Manitoba: *in* Teller, J.T. and Clayton, L. (eds): Glacial Lake Agassiz. The Geological Association of Canada, Special Paper 26, p.211-230.

Klassen, R.W., 1989, Quaternary geology of the southern Canadian interior plains: *in* Fulton, R.J. (ed), Quaternary geology of Canada and

Greenland, Geological Survey of Canada, Geology of Canada, no.1, p.138-173.

Klassen, R.W., Wyder, J. and Bannatyne, B.B., 1970, Bedrock topography and geology of southern Manitoba: Geological Survey Canada Paper 70-51.

Lowden, J.A., Fyles, J.G., and Black, W.Jr., 1967, Geological Survey of Canada radiocarbon dates VI: Radiocarbon, v.9, p. 156-197.

Lowden, J.A., Robertson, I.M., and Black, W.Jr., 1971, Geological Survey of Canada radiocarbon dates XI: Geological Survey of Canada Paper 71-7, p.255-324.

Lowe, D.R., 1982, Sediment gravity flows: II. Depositional model with special reference to the deposits of high-density turbidity currents: Journal of Sedimentary Petrology, v.52, no.1, p.279-297

Manitoba Mineral Resources Division, 1979, Geological Map of Manitoba: Map 79-2, Scale 1: 1,000,000.

Manitoba Mineral Resources Division, 1980, Quaternary geology map, Southern Manitoba 62G, Brandon, Manitoba Mineral Resources Division, Geological Map AR807, Scale 1:250,000.

Manitoba Water Resources Branch, 1970, Ground-water Availability in the Carberry area: Ground-water Availability Study Report No.2, Water Resources Branch, 26p.

Manitoba Water Resources Branch, 1977, Groundwater availability in the south Cypress area: Groundwater Availability Studies Report No.6, Department of Mines, Resources and Environments Management, Water Resources Branch, 7p.

Manitoba Water Resources Division, 1978, Groundwater resources in the Cypress planning district: Water Resources Division, 10p.

McCabe, H.R., 1971, Stratigraphy of Manitoba, an introduction and review: in Turnock, A.C., (ed.), Geoscience studies in Manitoba: Geological Association Canada Special Paper 9, p.167-187.

Miall, A.D., 1984, Deltas: in Walker, R.G. (ed), Facies Models, Geoscience Canada Reprint Series 1, p. 105-118.

Michalyna, W., Podolsky, G.P., Gardinar, W., 1976, Soils of the Brandon region study area: Soil Report No.30, Manitoba Department of Municipal Affairs, Municipal Planning Branch, 251p. Manitoba

Moss, A.J., 1962, The physical nature of common sandy and pebbly deposits. Part I. American Journal of Science, v.260, p.337-373.

Nemec, W., and Steel, R.J., 1984, Alluvial and coastal conglomerates: their significance features and some comments on gravelly mass-flow deposits: in Koster, E.H., and Steel, R.J. (eds), Sedimentology of gravels and conglomerates, Memoir of Canadian Society of Petroleum Geology, No.10, p.1-31.

P.F.R.A., 1986, Aquifer Capacity Investigation Reports: Unpublished report, Prairie Farm Rehabilitation Administration.

Posta, G., 1990, Depositional architecture and facies of river and fan delta: a synthesis: *in* Colella, A., and Prior, D.B., (eds): Coarse-grained delta, Blackwell Scientific Publications, Cambridge Massachusetts etc., p.13-28.

Ritchie, J.C., 1976, The Late-Quaternary vegetational history of the western interior of Canada: Canadian Journal of Botany, v.54, p.1793-1818.

Ruhe, R.V., 1969, Quaternary landscapes in Iowa: Ames, Iowa, Iowa State University Press, 255p.

Rust, B.R., and Koster, E.H., 1984, Coarse alluvial deposits: *in*: Walker, R.G.,(ed): Facies models, Geoscience Canada Reprint Series 1, p.53-71.

Schenk, C.J., 1987, Primary textures in experimentally formed oolite cross-strata: U.S. Geological Survey Bulletin 1670, 11p.

Surlyk, F., 1984, Fan-delta to submarine fan conglomerates of the Volgian-Valanginian Wollaston Foreland Group, eastern Greenland: *in* Koster, E.H. and Steel, R.J.(Eds): Sedimentology of gravel and conglomerates. Canadian Society of Petroleum Geologists Memoir 10. p.359-382.

Teller, J.T., 1980, Radiocarbon dates in Manitoba: Mineral Resources Division Geological Report GR80-4, 61p.

Teller, J.T., 1985, Glacial Lake Agassiz and its influence on the Great Lakes: in Karrow, P.F., and Calkin, P.E., Quaternary evolution of the Great Lakes, Geological Association of Canada special paper 30, p. 1-16.

Teller, J.T., 1987, Proglacial lakes and the southern margin of the Laurentide Ice Sheet: in Ruddiman, W.F., and Wright, H.E.,Jr., (eds), North America and Adjacent Oceans During the Last Deglaciation. The Geological Society of America, v. K-3, p.39-69.

Teller, J.T., 1989, Importance of the Rossendale site in establishing a deglacial chronology along the southwestern margin of the Laurentide Ice Sheet: Quaternary Research 32, p.12-23.

Teller, J.T., 1990, Volume and routing of late-glacial runoff from the southern Laurentide ice sheet: Quaternary Research 34, p.12-23.

Teller, J.T., and Bannatyne, B., 1976, Geology and topography of the buried bedrock surface of southern Manitoba: Manitoba Mineral Resource Division Surficial Map 76-3, Scale 1:500,000.

Teller, J.T., and Bluemle, J.P., 1983, Geological setting of the Lake Agassiz region: in Teller, J.T., and Clayton, L., (eds)., Glacial Lake Agassiz, Geological Association of Canada Special paper 26, p.355-374.

Teller, J.T., and Fenton, .M., 1980, Late Wisconsinan glacial stratigraphy and history of southern Manitoba: Canadian Journal of Earth Sciences, 17, p.19-35.

Teller, J.T., and Last W.M., 1982, Pedogenic horizons in lake sediments: Earth surface processes and landforms, v.7, p.367-379.

Teller, J.T., and Last W.M., 1981, Late Quaternary history of Lake Manitoba, Canada: Quaternary Research, v.16, p.97-116.

Teller, J.T., and Last W.M., 1981, Late Wisconsinan glacial stratigraphy and history of southeastern Manitoba: Canadian Journal Earth Sciences, v.17, p.19-35.

Teller, J.T., Moran, S.R., and Clayton, L., 1980, The Wisconsinan deglaciation of southern Saskatchewan and adjacent areas; Discussion: Canadian Journal of Earth Sciences, v.17, p.539-541.

Teller, J.T., Sun, S., and Wolfe, B., 1990, Catastrophic flooding into Lake Agassiz: Programme and Abstracts, Canadian Quaternary Association and American Quaternary Association first joint meeting, Waterloo, Ontario, Canada, p.32.

Teller, J.T., and Thorleifson, L.H.,1983, The Lake Agassiz-Lake Superior connection: *in* Teller,J.T. and Clayton L. (Eds): Glacial Lake Agassiz, Geological Association of Canada Special Paper 26, p.261-290.

Thorleifson, L.H., 1983, The eastern outlets of Lake Agassiz: M.Sc. Thesis, Winnipeg, University of Manitoba, 87p.

UMA Group (Underwood McLellan and Associater Limited), 1977, Sand and gravel resources of the Brandon region, A report to the Department of Mines, Resources and Environmental Management, Manitoba, 167p.

Upham, W., 1895, The glacial Lake Agassiz: U.S. Geological Survey Monograph, v.25, 658p.

Walker, R.G., 1984, Turbidities and associated coarse clastic deposits: in Walker, R.G.(Ed): Facies models. Geoscience Canada Reprint Series 1., p.171-188.

Warman, T. A., 1991, Sedimentology and history of deglaciation in the Dryden, Ontario area, and their bearing on the history of Lake Agassiz: M.Sc. Thesis, Winnipeg, University of Manitoba, 256p.

Welsted, J., Young H.R., 1980, Geology and origin of the Brandon Hill, southwest Manitoba: Canadian Journal of Earth Sciences, v.17, p.942-951.

Wickenden, R.T.D., 1945, Mesozoic stratigraphy of the eastern plains, Manitoba and Saskatchewan: Geological Survey of Canada Memoir 239.

Appendix I. Carbonate Content of Till

Location	$CO_3\%$ of Whole Sample	$CO_3\%$ of Silt and Clay Portion
$25-07-17W$	26.8	19.6
$02-08-17W$	21.4	20.3
$34-07-14W$	39.6	27.7
$16-10-16W$	25.5	29.6
$25-10-19W$	25.9	26.0
$06-09-18W$	24.1	19.3
$11-08-20W$	24.6	14.2
$25-08-21W$	30.6	28.2
$16-09-21W$	34.3	25.5
$19-09-21W$	23.5	28.9
$06-10-21W$	36.1	36.2
$19-10-20W$	35.6	35.0
$27-10-20W$	37.5	24.5
$28-11-20W$	34.1	30.2
$14-11-21W$	17.2	17.0
$07-11-20W$	16.9	12.4
$15-11-20W$	20.4	12.8
$30-10-26W$	-	19.3

Appendix II-1. Sieving Result of 90AD-1

Depth (m)	$< -1\phi$	$-1-0\phi$	$0-1\phi$	W% $1-2\phi$	$2-3\phi$	$3-4\phi$	$> 4\phi$	Mz (ϕ)	σ (ϕ)	Sk_I
1.5	0.3	0.8	6.0	53.8	34.3	3.3	1.5	1.96	0.69	0.07
2.0	-	0.04	6.33	61.9	29.07	2.1	0.56	1.7	0.58	0.09
2.5	-	-	3.5	52.5	39.4	3.8	0.7	1.94	0.69	0.17
3.0	-	0.4	6.96	49.05	39.99	3.2	0.75	1.91	0.66	-0.04
3.5	-	-	3.3	56.3	36.8	2.3	1.2	1.91	0.65	0.16
4.0	-	-	1.5	55.5	40.4	1.2	1.3	1.92	0.65	0.08
4.5	-	-	7.1	66.0	26.18	0.54	0.18	1.71	0.51	-0.05
5.0	-	-	4.2	64.4	30.35	0.65	0.5	1.71	0.49	0.06
5.5	-	-	4.3	68.7	25.6	0.9	0.5	1.66	0.49	0.06
6.0	-	-	5.0	69.6	24.3	0.75	0.4	1.65	0.48	0.06
6.5	-	0.08	6.32	64.49	27.76	0.97	0.37	1.74	0.51	0.15
7.0	-	-	2.75	59.7	36.22	.6	0.8	1.77	0.57	0.13
7.5	-	0.1	7.0	55.5	28.9	8.2	0.2	1.9	0.77	0.24
9.0	Clay									
10.0	Marl									
12.0	Clay									
13.0	-	0.77	23.35	58.12	10.31	3.95	3.65	1.42	0.83	0.25
13.5	-	0.2	17.71	54.11	14.5	7.84	5.53	1.72	0.84	0.28
14.0	Clay									
16.0	Clay									
16.5	2.33	1.86	14.47	71.33	8.26	1.02	0.7	1.42	0.6	-0.24

17.0	2.16	3.12	13.68	68.24	10.15	1.54	1.1	1.44	0.67	-0.27
17.5	4.2	6.78	43.95	37.64	6.15	0.71	0.52	0.91	0.83	-0.16
18.0	-	7.38	26.84	45.95	11.09	0.72	0.5	1.04	1.04	-0.26
18.5	1.84	3.99	18.84	53.12	17.92	2.46	1.8	1.5	0.83	-0.11
19.0	0.88	5.9	43.24	40.09	8.61	0.83	0.39	1.04	0.77	0.12
19.5	0.62	5.37	42.54	38.94	10.66	1.33	0.54	1.05	0.79	0.13
20.0	3.74	2.67	37.85	46.87	6.95	1.49	0.42	0.13	0.72	-0.04
20.5	-	0.88	33.49	60.96	3.68	0.69	0.3	1.1	0.46	-0.06
21.0	0.51	1.66	35.05	59.15	2.85	0.69	0.3	1.1	0.46	-0.05
21.5	1.34	4.9	43.67	45.93	2.9	0.74	0.51	1	0.58	-0.04
22.0	0.22	2.23	26.94	49.65	11.7	4.8	4.45	1.47	0.96	0.28
25.5	0.53	1.46	29.03	59.93	3.67	0.61	0.22	1.18	0.52	-0.03
26.0	1.54	6.15	24.53	56.56	8.95	1.54	0.72	1.21	0.81	-0.23
26.5	0.16	1.59	37.67	52.56	5.57	1.28	1.2	1.18	0.62	0.1
27.0	1.32	4.18	43.45	47.96	2.52	0.34	0.23	1.01	0.54	-0.03
27.5	-	-	3.2	16.7	43.5	20	16	2.83	1.09	0.06
28.5	-	-	-	21.8	52.53	20	5.6	2.63	0.79	0.17
29.0	-	-	0.7	44	38.2	10	7	2.15	0.92	0.31
29.5	-	-	1.05	22.6	59.78	13.8	2.8	2.42	0.7	0.06
30.0	-	-	-	19.7	56.8	13.8	2.8	2.7	0.74	0
30.5	-	-	-	22.2	65.1	8.4	4.3	2.4	0.63	0.14
31.0	-	-	0.46	26.7	61.5	6.9	4.4	2.37	0.63	0.15
31.5	-	-	0.6	17.2	66.6	11.2	4.4	2.48	0.61	0.14
32.0	-	-	1.07	16.2	68.55	9.68	4.38	2.42	0.62	0.09
32.5	-	-	-	8.4	72.9	14.3	4.3	2.63	0.59	0.22
33.0	-	-	-	3.5	73.3	16.7	6.4	2.76	0.5	0.05
33.5	-	-	-	10.1	74	1.3	3	2.55	0.47	0.13
34.0	-	-	-	6.1	61.1	20.5	5.7	2.76	0.67	0.3
34.5	-	-	-	7.2	59.9	26.6	6.4	2.83	0.73	0.28

35.0	-	-	-	21.8	48.3	22	7.9	2.69	0.82	0.11
35.5	-	-	-	3.4	54.7	25	16.8	3.13	0.78	-0.18
36.0	Clay							4		
37.0	Clay							4		
38.5	13.86	14	14.15	21	20.8	705	9.1	1.08	1.9	0.05

Appendix II-2. Sieving Result of 90AD-2

				W%						
Depth (m)	$< -1\phi$	$-1-0\phi$	$0-1\phi$	$1-2\phi$	$2-3\phi$	$3-4\phi$	$> 4\phi$	Mz (ϕ)	σ (ϕ)	Sk_I
3.0			-	13	74.9	9.7	2.4	2.5	0.55	-0.01
4.0			1	49.4	42.23	4.2	3.2	2.07	0.74	0.19
4.5			4.1	29	58.3	5.2	3.5	2.33	0.76	-0.11
5.0			1.3	44.4	48.4	3.9	2.0	2.1	0.67	0.02
5.5			-	32.3	62.8	3.8	1	2.22	0.62	-0.21
6.0			-	22	66.5	7.4	4.2	2.4	0.69	-0.11
6.5			-	14.3	78.5	6.1	1.1	2.47	0.51	-0.16
7.0			-	4.1	66.5	22.6	6.8	2.83	0.7	0.34
7.5			0.4	53.1	35.6	7.3	3.5	2.52	0.77	0.24
8.0			-		45.4	41.9	12.7	3.12	0.76	0.08
8.5			0.5	26.9	61.1	8.5	2.9	2.32	0.64	0.05
9.0			-	25.5	62	8.8	3.6	2.38	0.66	0.1
9.5			-	-	73.07	21	5.14	2.8	0.62	0.28
10.0			-	11	69.2	17.2	2.6	2.62	0.61	0.19
10.5			-	38	52.9	5.9	3.2	2.21	0.78	0.05
11.0			-	46.5	39.6	7.9	6	2.13	0.87	0.2
11.5			0.1	23.3	66.46	8.6	1.6	2.38	0.6	0.03
12.0			0.1	26.6	63.6	8.3	1.4	2.35	0.6	0.07
12.5			-	33.8	53.6	10.6	2	2.28	0.7	0.06
13.0			-	21.5	63.2	13.2	2.3	2.45	0.62	0.09
13.5			0.1	18.0	62.5	15.22	4.1	2.55	0.71	0.15
14.0			-	19.1	58.4	16.7	5.7	2.61	0.78	0.16
15.0			-	48.3	46.4	3.8	1.4	-	-	-

15.5		0.7	23	71.6	3.8	0.8	2.34	0.48	0
16.0		-	0.7	60.8	35.2	3.2	2.9	0.64	0.21
17.0		-	1.1	55.1	39.1	4.7	2.95	0.66	0.15
17.5		-	1.5	57	36.6	5.1	2.94	0.67	0.19
18.0		-	2.1	55.2	36.3	6.43	2.94	0.67	0.19
18.5		-	1.2	53.5	38	7.4	2.99	0.7	0.18
19.0		-	0.3	41.3	48.9	9.5	3.19	0.69	0.09
19.5		-	-	35.8	52.7	11.5	3.13	0.7	
20.0		-	3.68	31.6	52.6	12.1	3.19	0.72	-0.15
21.5		-	1.5	51.5	37.4	9.9	3.02	0.72	0.15
22.0		-	1.3	58.1	33.5	7.1	2.94	0.69	0.23
23.0		-	-	59.4	37.6	3	2.91	0.64	0.16
24.0		0.08	0.7	62	34	3.01	2.89	0.64	0.2
24.5		-	0.6	67.1	30	2.3	2.84	0.62	0.24
25.0		-	0.4	56	39.9	3.7	2.94	0.65	0.14
25.5		0.4	2.2	63.14	30.9	3.3	2.85	0.64	0.23
26.0		0.7	4.6	62.1	28.5	4.1	2.82	0.66	0.26
26.5		0.4	5.2	51.7	37.6	5.2	2.93	0.68	0.14
27.0		-	9.6	52.5	33	4.8	2.85	0.74	0.13
27.5		-	12.5	51.7	32.9	4.2	2.81	0.76	0.1
28.0		0.7	20.5	47.5	26.5	4.8	2.69	0.81	0.13
29.0		-	32	24.5	34.1	9.5	2.67	1.07	-0.04
29.5		2.47	67.5	13.4	12.4	4.3	2.02	0.91	0.49
30.0		3.7	51.8	34.7	5.1	4.6	1.98	0.84	0.27
30.5	0.07	6.5	67.4	21.3	2.4	2.3	1.73	0.67	0.22
31.5		0.2	27	59.2	10.4	3.2	2.35	0.67	0.1
32.0		0.2	56.4	36	5.7	1.7	2.01	0.76	0.26
32.5		-	19.8	72.4	5.7	2.09	2.41	0.54	0.07
33.0		0.3	24.5	58.5	11	5.7	2.44	0.73	0.15

34.0		-	22.8	65.1	8.7	3.4	2.39	0.63	0.08
34.5		-	15.5	78.8	4.8	0.86	2.43	0.46	-0.04
35.0		-	23	72.9	2.8	1.3	2.35	0.48	-0.12
36.0		0.2	27.7	66.8	3	2.26	2.32	0.5	-0.07
36.5	0.4	0.4	24.6	69.14	3.56	2.63	2.35	0.53	0
38.0		-	-	51.31	34.89	13.76	3.05	0.75	0.21
38.5		-	15.7	78.2	5.04	1.03	2.43	0.47	-0.02
39.0		0.08	13.3	81.5	4.2	0.8	2.44	0.42	-0.06
39.5		-	12.9	81.2	4.9	1	2.45	0.45	-0.02
40.5		0.06	16.6	78	4.2	1.1	2.39	0.44	-0.07
41.0		0.5	18.2	76.8	3.3	1.6	2.4	0.45	-1
41.5		-	23.1	75	3.8	1.1	2.35	0.48	-0.07
46.0		-	22.6	73	3.2	1.2	2.34	0.48	-0.11
46.5		-	18.8	77	3.4	0.8	2.39	0.47	-0.13
47.0		0.3	22.7	71.92	3.2	1.5	2.35	0.49	-0.14
47.5		-	24.6	67.2	4.1	4	2.33	0.65	0.04
48.0		0.2	29.2	63.2	3.9	3.5	2.29	0.64	0.06
48.5		0.5	24.5	66.4	4	4.6	2.33	0.66	0.08
49.5		-	22.8	71	3.5	2.6	2.26	0.46	-0.24
50.5		-	11.6	47.9	38.9	1.6	2.82	0.74	0
51.0		-	21.2	66.3	6.1	6.3	2.41	0.69	0.13
51.5		-	28.8	62.93	5.1	3.2	2.31	0.64	0.06
52.5		-	3.3	11.9	34.9	50	3.8	0.68	-0.52
54.0		-	16.5	51.2	10.7	21.5	2.92	0.96	0.29
54.5		5.9	47.6	40.9	3.4	2.1	1.96	0.72	0.04
57.0		0.9	46	43.2	4.9	4.9	2.07	0.83	0.17
57.5		2.8	57.6	29.7	4.9	4.9	1.91	0.85	0.43

Appendix II-3. Selected samples from exposures

Depth (m)	$<-1\phi$	$-1-0\phi$	$0-1\phi$	W% $1-2\phi$	$2-3\phi$	$3-4\phi$	$>4\phi$	Mz (ϕ)	σ (ϕ)	Sk_I
The lower sand unit in Exposure 3, east of Brandon at Sec.7, Tp 10, R 18W										
1.2m			0.48	66.74	31.08	1.42	0.27	1.9	0.43	0.18
The lower sand bed in RC-1, the Souris River junction section at SW$\frac{1}{4}$, Sec.22, T8, R16W										
	1.16	1.65	12.22	55.06	25.5	3.21	1.21	1.7	0.8	0.18
The lower sand unit from RC-2, 1 km from the Souris River junction section										
	0.67	1.64	23.57	63.6	6.29	3.51	0.7	1.3	0.71	0.12
A sample from RC6 along the Epenette Creek at Sec.14, T9, R14W										
				0.49	36.45	55.98	7.49	3.22	0.5	0.09

APPENDIX III

Well 90AD – 1

Core 90AD-1 is located 9.1 km east and 4 km north of Shilo at NE$\frac{1}{4}$ of SE$\frac{1}{4}$, Sec.16, Tp10, R16W (see Fig. 1-8). General description presented in Fig. 4-30 with specific grain size and standard deviation shown in Figs. 4-31, 4-32.

Depth (m)	Thickness (m)	Stratigraphic Description
Unit J		
0-1.3	1.3	Soil in gravelly coarse sand
1.3-4.2	2.9	Sand, fg-mg, well sorted to moderate well sorted ($\sigma = 0.48$ to 0.69), color laminae from 2.5 to 3.5m, others no distinct bedding, olive brown (2.5Y 4/4).
4.2-7.9	3.7	Sand, fg-cg, moderately well sorted ($\sigma = 0.77$) no distinct bedding, dark greyish brown (2.5Y 4/2), lower contact is sharp.
Units I		
7.9-8.1	0.2	Sand, vfg-fg, silty, organic rich with plant fragments and fibers, black (2.5Y 2/0). Lower boundary is gradational.

8.1-8.2	0.1	Peat, silty, slightly sandy (vfg), black (2.5Y 2/0). Plant fragments and fibers are up to 1 cm in diameter, and some of them look like cottail leaves. Lower boundary is distinct.
8.2-9.2	1.0	Clayey silt with some slightly sandy zone (vfg), greyish brown (2.5Y 3/2) but very dark grey (5Y 3/1) in the upper 6cm, organic rich, poorly laminated to non visible lamination. There are abundant small gastropods (most 1 mm or less) and shell fragments, occasional plant fragments and plant fibers (roots). Basal contact is abrupt.

Units H

9.2-9.9	0.7	Calcium carbonate fragments (marl) with occasional gastropods and shell fragments, black to brown organic fragments and quarts ($<$ 1%) grains. Carbonate pieces are mainly less than 1 mm, irregular to twig(bone?)-shaped, and the center of some elongated pieces are hollow. There is no visible bedding or lamination. Radiocarbon dating on calcium clast from the depth of 9.5 m yielded an age of 9330 $\pm$160 yrs BP (1253Cc). Lower boundary is distinct.

9.9-10.4	0.5	Silt, clayey, poorly laminated, dark olive grey (5Y 3/2). Some gastropods (high spired type), occasional shell fragments and wood fragments present, and become abundant in the upper 5cm. Lower boundary is indistinct.
10.4-10.75	0.35	Marl, very fine grained, laminated light colored (olive 5Y 5/4) and dark color (dark olive 5Y 5/2), occasional gastropod shells (some¿3mm) and wood fragments. Lower boundary is distinct.
10.75-10.8	0.05	Sandy clayey silt, organic-rich, laminated with light sandy marl like the underlying unit. One large pieces of wood is 10 cm long. Lower boundary grades into underlying unit.
10.8-10.95	0.15	Marl, light olive grey (5Y 5/2) with abundant small sized gastropods and plant fragments, well laminated to weakly laminated. Lower boundary is gradational.
10.95-11.7	0.75	Peaty marl, abundant plant fragments but rare shells, poorly laminated to non-laminated. It colored dark greyish brown (2.5Y 4/2) with some more $CaCO_3$-rich laminae, light olive grey (5Y 5/2). A pieces of wood at the depth of 11.2 m yielded a radiocarbon date of 9340 $\pm$70 yrs BP (TO2197).

Unit G

11.7-12.4	0.7	Laminated sandy silt, clayey, greyish brown (2.5Y 3/2) to black (10YR 2/1). It is organic-rich, rare molluscs, occasional plant fibers, with abundant plant fibers near 11.9 and 12m, including acorns. Lower boundary is distinct.

Unit F

12.4-13.5	1.1	Fg to mg sand to clayey sand, moderately sorted ($\sigma = 0.84$), occasional thin ($< 2mm$) clay laminae, greyish brown (2.5Y 3/2) to black (10YR 2/1). Lower boundary is distinct.

Unit E

13.5-16.3	2.8	Clayey sandy silt, no visible laminae, dark grey (5Y 4/1).

Unit D

16.3-17.4	1.1	Cg to mg sand, some granules and rare pebbles, with abundant shale fragments and coal (lignite) fragments (30%). Pebbles and granules are subrounded to angular, moderately well sorted ($\sigma =$ 0.6 to 0.67). Lower boundary graduate.

17.4-25.3	7.9	Cg sand, olive (5Y 5/4), mostly moderately sorted (σ = 0.72 to 1.0), some are well sorted (σ = 0.46), occasional shale and quarts pebbles up to 2.5 cm in diameter. There are abundant shale fragments but rare coal fragments. Ther are no visible laminae. Lower boundary gradational over 3 cm.
25.3-25.4	0.1	Fg-cg sand, some granules and fine pebbles, clayey, dark olive (5Y 3/2), carbonate cemented and hard. Lower boundary distinct.
25.4-27.0	1.6	Cg sand, olive (5Y 5/4), mostly moderately sorted (σ = 0.7 to 0.79), occasional pebbles. There are abundant shale fragments but rare coal fragments. Ther is no visible laminae. Lower boundary gradational.
27.0-27.3	0.3	Fg sand, silty, with some shale particle, olive (5Y 5/3).

Unit C

27.3-28.7	1.4	Laminae of silt to very fine sand, clayey, moderately sorted (σ =0.79 to 1.09), olive grey (5Y 5/2), intercalating laminae of fg sand. Laminae are most deformed, possiblly by drilling.Lower boundary is distinct.

28.7-30.1	1.4	Fg to very fine sand, silty, pale olive (5Y 6/4), with abundant elongated clay balls up to 4cm in diameter, and shale fragments, no visible lamination, moderately well sorted ($\sigma = 0.74$).
30.1-33.2	3.1	Fg-very fine sand, pale olive (5Y 6/4), moderately well sorted ($\sigma = 0.5$ to 0.62), intercalating with beds of sand that is shale-rich, olive grey (5Y 5/2), and lens of pale yellow sand (5Y 7/3), no visible bedding. Lower boundary is distinct.
33.2-34.4	0.8	Silt to very fine sand, some shale particles, pale olive (5y 6/4), no visible lamination. Lower boundary is indistinct.
34.4-34.7	0.3	Laminae of olive grey (5Y 4/2) very fine sand and pale olive (5Y 6/4) very fine sand. Lower boundary is indistinct.
34.7-36.0	1.3	Clayey silt to silty very fine sand, olive gray (5Y 5/2), moderately well sorted ($\sigma = 0.73$ to 0.82) no visible laminae. Lower boundary is indistinct.

Unit B

36.0-36.3	0.3	clayey silt to silty clay, fine laminated, olive grey (5Y 4/2). Lower boundary is indistinct.
36.3-37.3	1.0	Silty clay, light gray (5Y 7/2), irregular laminated (deformed during drilling?). Lower boundary is indistinct.

37.3-37.8	0.5	silty clay, olive (5Y 5/4), no visible lamination. Low boundary is distinct.

Unit A

37.8-40.8	3.0	Till, gravelly, sandy and clayey, light olive grey (5Y 6/2).
40.8		End of the well.

Well 90AD – 2

Core 90AD-2 is located 5 km south and 5 km east of Carberry at NE$\frac{1}{4}$ of SE$\frac{1}{4}$, Sec.10, Tp10, R14W (see Fig. 1-8). General description presented in Fig. 4-33 with specific grain size and standard deviation data shown in Figs. 4-34, 4-35.

Depth (m)	Thickness (m)	Stratigraphic Description
Unit H		
0-1.26	1.26	Soil in fg sand.
1.26-2.44	1.18	Clayey silty, sticky, weakly laminated, light olive brown(2.5Y 5/4), distinct lower boundary.
Unit G		
2.44-4.54	2.1	Fine sand to very fine sand, a few clayey silt laminae (0.1 cm each), and cg sand beds (2 cm each) moderately well sorted ($\sigma = 0.55$ to 0.74), pale olive (6Y 6/3), no visible bedding.
4.54-5.34	0.8	Laminae of olive yellow (5Y 6/8) fg-mg sand and pale olive (5Y 6/3) fg-mg sand, moderately well sorted ($\sigma = 0.62$ to 0.67), with a few laminae of olive grey (5Y 5/2) fg-silty sand. Each laminae is about 0.5-1 cm thick. Lower boundary is indistinct.

5.34-6.71	1.37	Fg sand, occasional beds of silty vfg sand, pale olive (5Y 6/3), no visible bedding, moderately well sorted (σ = 0.51 to 0.69), indistinctive lower boundary.
6.71-8.57	1.86	Silty fg t vfg sand, olive yellow (5Y 6/6), moderately well sorted (σ = 0.64 to 0.77), poorly laminated, with a distinct lower boundary.
8.57-9.5	0.93	fg-vfg sand, pale olive (5Y 6/3), moderately well sorted (σ = 0.66), no visible bedding.
9.5-10.37	0.87	Vfg silty sand, olive yellow (5Y 6/8), moderately well sorted (σ = 0.61 to 0.62), no visible laminae. Lower contact is distinct.
10.37-15.25	4.88	Fg-mg sand, pale olive (5Y 6/3), moderately well sorted to moderately sorted (σ = 0.6 to 0.87), no visible bedding. Lower boundary is distinct.
15.25-16.0	0.75	Mg-fg sand, dominated by mg sand, olive gray (5Y 5/2), abundant coal fragments, no visible bedding. Lower boundary is distinct.
Unit F		Sand in this unit are all moderately well sorted (σ= 0.62 to 0.76).
16.0-17.0	1.00	Silty fine to very fine sand, olive grey (5Y 5/2), no visible laminae, with indistinct lower boundary.

17.0-19.0	2.0	Laminae of fg-vfg sand, gray (5Y 6/2), and thin laminae of clayey silt. The fg-vfg sand is colored light olive gray (5Y 6/2), while the clayey silt contains coal fragments, 0.2-0.5cm thick and is colored olive gray (5Y 5/2). Lower boundary is distinct.
19.0-20.9	1.9	Laminae (0.2-2.5cm) of silty very fine sand, light olive gray (5Y 6/2), with thin laminae (0.2-0.5) of clayey silt, olive gray (5Y 5/2). Lower contact is indistinct.
20.9-21.5	0.6	Laminae of olive gray (5Y 4/2) silty vfg sand and light olive gray (5Y 6/2) vfg sand, containing lens of dark gray (5Y 3/2) clayey sand and coal fragments. Lower boundary is indistinct.
21.5-22.6	0.9	Distinctly laminated fg-vfg sand, light gray (5Y 7/2), and clayey silt, olive gray (5Y 5/2). Each laminae of the fg-vfg sand is about 2 to 5cm thick, while each laminae of clayey silt is about 0.2cm to 2cm. Lower boundary is distinct.
22.6-23.5	0.9	Fg-vfg sand, light olive gray (5Y 6/2), no visible laminae, with indistinct lower boundary.
23.5-28.45	4.95	Fg-vfg sand, olive gray (5Y 4/2), no visible laminae, rare coal grains, with indistinct lower boundary.

28.45-28.65	0.2	Laminae (each 1cm) of vfg-fg sand, olive gray (5Y 5/2), and laminae (about 2mm) of silty very fine sand, dark olive gray (5Y 3/2). Lower boundary is distinct.
28.65-29.0	0.35	Mg-fg sand, silty, olive gray (5Y 5/2), sandwiched with a 4cm laminae of mg sand, pale olive (5Y 7/3). Coal fragments up to 2.5cm are spreaded throughout. No visible fine laminae. Lower boundary is distinct.
29.0-29.1	0.1	Mg-fg sand, pale olive (5Y 6/3), some pieces of coal fragments, no visible bedding. Lower boundary is distinct.
29.1-29.25	0.15	Fg-vfg sand, gray (5Y 5/1), no visible laminae. Lower boundary is distinct.

Unit E

29.25-29.5	0.25	Mg-fg sand, pale olive (5Y 6/3), some pieces of coal fragments, no visible bedding. Lower boundary is distinct.
29.5-29.9	0.4	Fg-mg sand, moderately sorted (σ = 0.91), clayey, olive gray (5Y 4/2), no visible laminae, distinct lower boundary.
29.9-30.7	0.8	Fg-mg sand, moderately sorted (σ = 0.84), pale olive (5Y 6/3), no visible laminae. There is a few coal and shale fragments. Lower boundary is indistinct.

30.7-31.0	0.3	Fg-mg sand, olive gray (5Y 6/3), with abundant coal and shale fragments, no visible laminae. Lower boundary is indistinct.
31.0-32.94	1.94	Fg-mg sand, dominated by fg sand, olive gray (5Y 5/2), with lens of olive gray sand (5Y 4/2), moderately well sorted ($\sigma = 0.54$ to 0.76), poorly laminated. Lower boundary is indistinct.
32.94-34.0	1.06	Fg sand, olive gray (5Y 5/2), intercalating with laminae of light olive gray (5Y 6/2) sand and one 1cm lamina of olive gray (5Y 4/2) sand.
Unit D		Mostly well sorted ($\sigma = 0.42$ to 0.53).
34.0-34.8	1.86	Fg sand, olive gray (5Y 5/2), intercalating with laminae of light olive gray (5Y 6/2) sand Lower boundary is indistinct.
34.8-36.2	1.4	Fg sand, silty light olive gray (5Y 6/2), containing one lens (5cm long and 2cm wide) of pale olive (5Y 6/4) fg sand. Lower boundary indistinct.
36.2-36.43	0.23	Mg-fg sand, pale olive (5Y 6/3), no visible laminae.
36.43-40.2	3.77	Fg sand, silty, light olive gray (5Y 6/2), no visible laminae.
40.2-40.6	0.4	Vfg sand, light olive gray (5Y 6/2), no visible laminae.
40.6-42.15	1.55	Fg sand, light olive gray (5Y 6/2), no visible laminae, with distinct lower boundary.

42.15-42.26	0.11	Silt, gray (2.5Y 6/), carbonate cemented, with distinct lower boundary.
42.26-42.4	0.14	Mg sand, olive gray (5Y 5/2), abundant shale fragments, with an distinct lower boundary.
42.4-47.5	5.1	Fg sand, light olive gray (5Y 6/2), no visible bedding, with an indistinct lower boundary.
Unit C		Sediments are mostly moderately sorted to moderately well sorted ($\sigma = 0.64$ to 0.96).
47.5-49.4	2.1	Fg sand, light olive gray (5Y 6/2), no visible bedding, with an indistinct lower boundary.
49.4-50.8	1.42	Fg-mg sand, light olive gray (5Y 6/2), no visible laminae. Lower boundary is distinct.
50.8-52.5	1.7	Clayey silt with mixed vfg sand to fg sand: sand probably mixed in by drilling.
52.5-53.1	0.6	Clayey silt to vfg sand, olive (5Y 5/3), laminated. There are 2 clay balls (olive gray, 5Y 4/2) that are 2cm long and 0.4mm thick, 1.5cm wide. Lower contact is distinct.
53.1-53.4	0.3	Laminae of clay, dark gray (5Y 4/1) and clayey silt, olive gray (5Y 5/2). Several carbonate clasts (one 4 cm long; Lower boundary si distinct.
53.4-53.85	0.45	Silt to vfg sand, olive gray (5Y 5/2), no visible bedding. One dolomite gravel in the bed is 4cm in diameter and is colored pale yellow (5Y 8/3). Lower boundary is distinct.

53.85-54.3	0.48	Vfg-fg sand, white (5Y 8/2), no visible bedding.
54.3-54.45	0.15	Silt-vfg sand and an 2cm fg sand bed.

Unit B

54.45-56.6	2.15	Fg-vfg sand, pale yellow (5Y 7/3), no visible laminae. Lower boundary distinct.
56.6-57.0	0.43	Fg-mg sand, dominated by fg sand, olive gray (5Y 5/2), with balls of pale yellow (5Y 7/3) very fine sand. Lower boundary is indistinct.
57.0-57.7	0.7	Fg-vfg sand, pale yellow (5Y 7/3), no visible laminae. Lower boundary distinct.
57.7-57.9	0.2	Vfg-fg sand, olive gray (5Y 5/2), Lower boundary is distinct.
57.9-58.0	0.1	Fg-mg sand, pale yellow (5Y 7/3), no visible laminae. Lower boundary distinct.
58.0-58.6	0.6	Vfg-fg sand, clayey and silty, olive gray (5Y 5/2). Lower boundary is indistinct.
58.6-59.1	0.5	Fg sand mixed with silt and clay, no bedding, poorly sorted. Lower boundary is indistinct.
59.1-63.8	4.7	Silt-vfg sand, olive gray (5Y 4/2), containing laminae of light olive grey (5Y 6/3) silt and clay zone about 0.1-0.5cm wide. Lower boundary is distinct.

Unit A

63.8-64.3	0.5	Sticky silty clay (5Y 4/2), one 4mm angular grains and a few coarse grains of sand. Lower boundary is indistinct.
64.3-64.6	0.3	Silt to silty clay, dark gray (5Y 4/1), no visible laminae.
64.6-64.7	0.1	Silty clay to clay, soft, sticky, dark gray (5Y 4/1), no visible laminae.
64.7-68.9	4.2	Silty clay, dark gray (5Y 4/1) and soft when wet, olive gray (5Y 5/2) and hard when dried. A few poor laminae.
68.9		End of the well

Well ADA-50

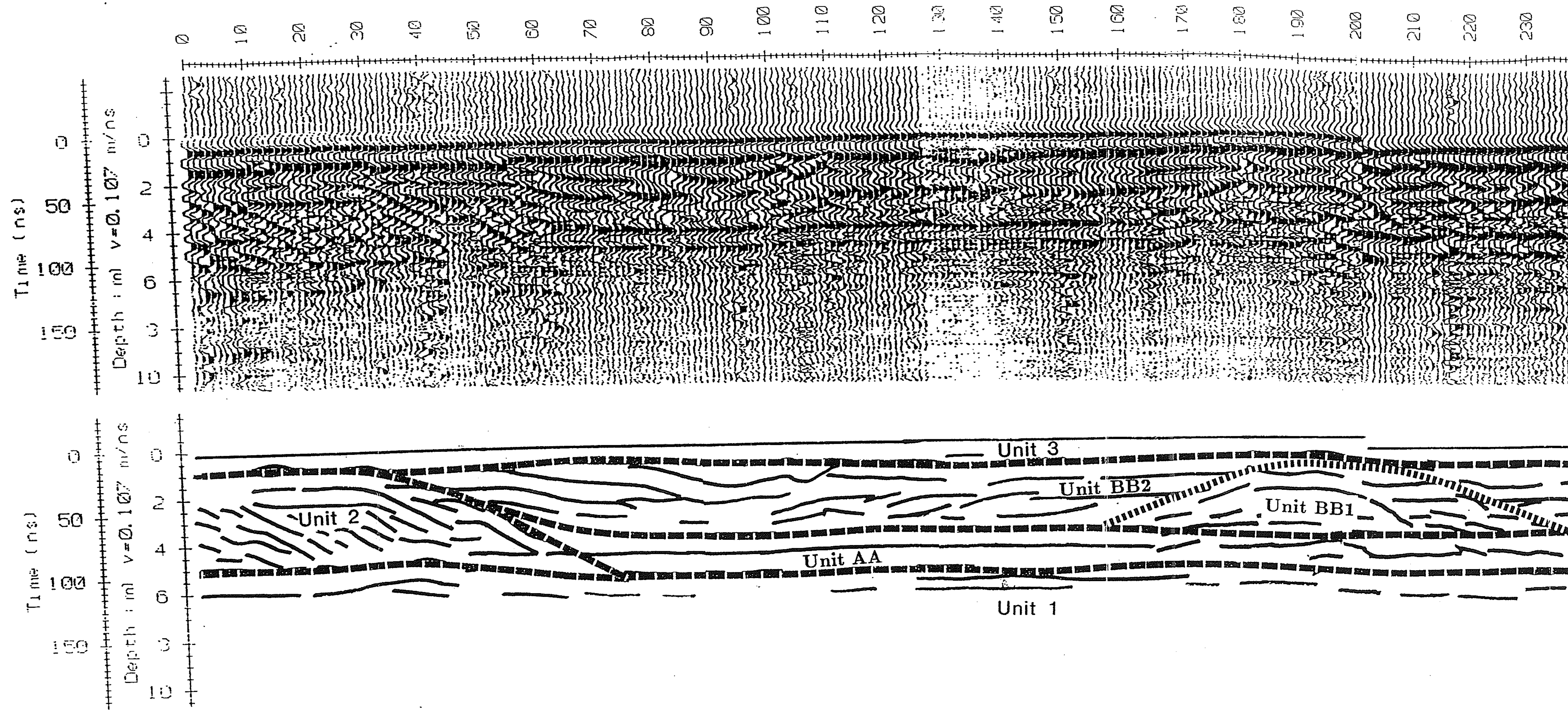
Well ADA-50 is located 6.4 km west of Shilo, at NW$\frac{1}{4}$ of SW$\frac{1}{4}$, Sec.2, Tp10, R17W. It was drilled and described by P.F.R.A.(P.F.R.A., 1986, umpub. report). Specific grain size data are shown in Fig. 4-36.

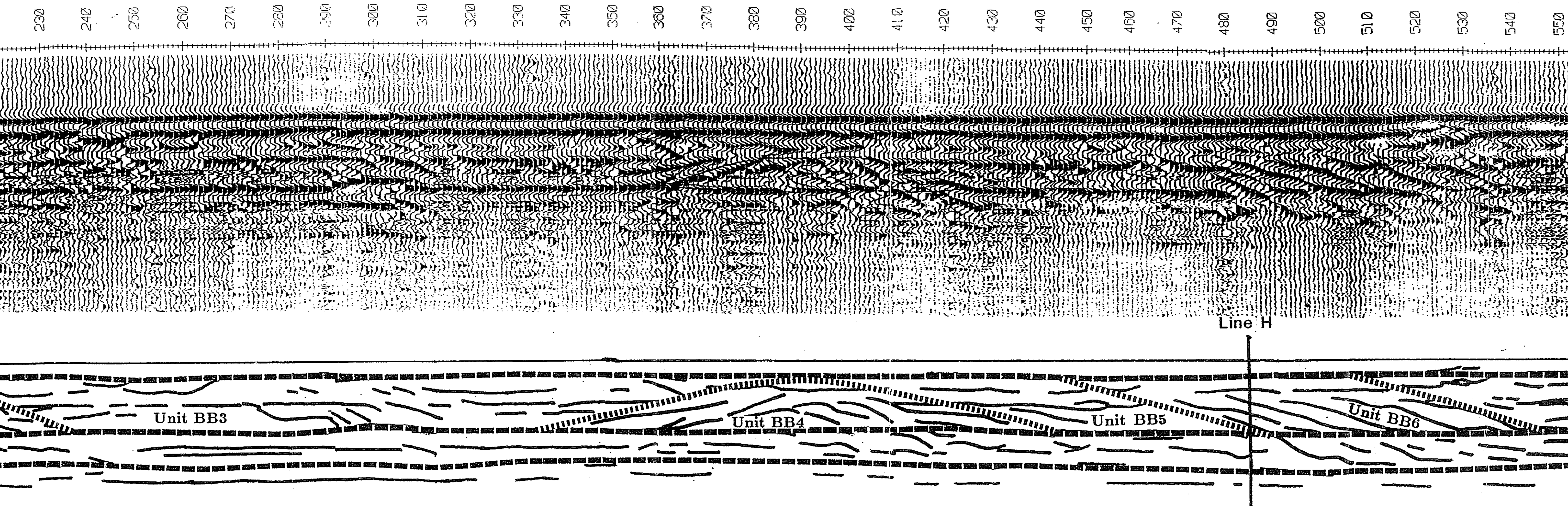
Depth (m)	Thickness (m)	Stratigraphic Description
0	0.46	Top soil, Black organic rich sand.
0.46	1.68	Fg-mg sand, mg sand is dominant, moist brown, 5% fines($<$ 0.074mm).
2.14	2.13	Fg-mg sand, fine is dominant, moist brown, 6% fines($<$ 0.074mm).
4.27	4.42	Fg-cg sand, mg sand predominant, moist brown, less than 2% gravel and 4-5% fines($<$ 0.074mm).
8.69	2.89	Fg-cg sand, mg sand predominant, gravelly, wet brown, 6-7% gravel and 3-4% fines($<$ 0.074mm).
11.58	0.61	Gravelly fg-cg sand, mg sand predominant, wet grey, 23% gravel and 7% fines($<$ 0.074mm).
12.19	2.14	Fg-cg sand and gravels, wet grey. Gravels and cg sand are predominant, 42% gravel and 17% cg sand, 35% fg to mg sand and 6% fines($<$ 0.074mm).
14.33	1.21	Clay and silt.
15.54		End of the well

Chemical Plant Well

This well is located at an Chemical Plant at NE$\frac{1}{4}$, Sec.10, Tp10, R18W, about 0.6 km north of the No.1 streamlined hill.

Depth (m)	Thickness (m)	Stratigraphic Description
0	7.0	Gravel
7.0	0.3	Gravels and boulders.
7.3	6.1	Soft sandy clay mixed and pebbles.
13.4	1.4	Gravel.
14.8	25.9	Hard sandy clay mixed with Gravel (till?).
40.7	1.7	Soft silty, with some boulders.
42.4	7.9	Sand.
50.3	13.7	Gravel.
64		End of well.


Time (ns)
Depth (m) v=0.107 m/ns
Unit 3
Unit BB2
Unit 2
Unit BB1
Unit AA
Unit 1



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Unit BB5
Unit BB6

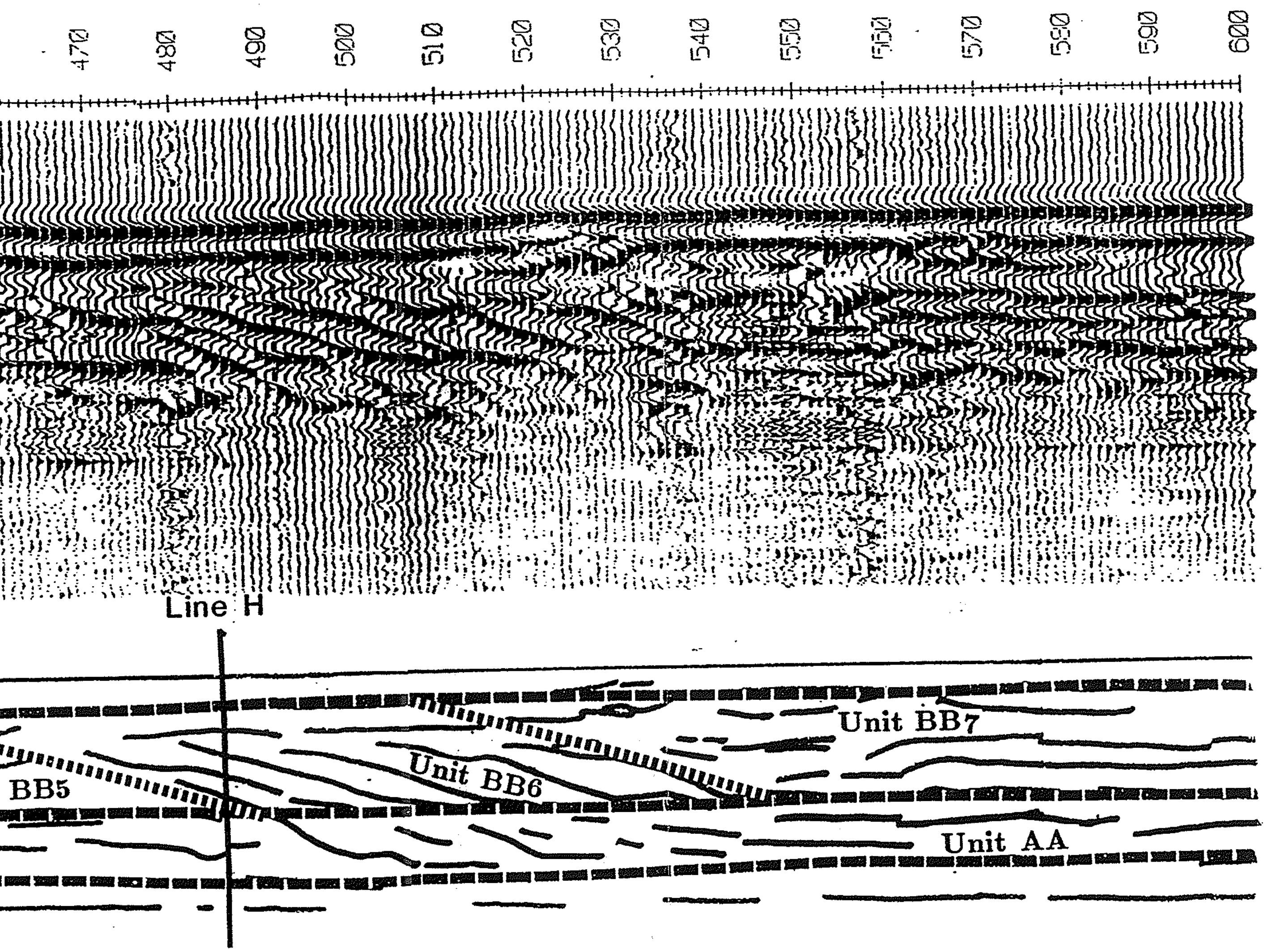

Fig. 4-22 Radar profile G extends eastward from the eastern end of the gravel pit in the No.4 ridge at Sect.5, Tp10, R17W (Fig. 4-15). This profile is located 40 m east of the profile in Fig. 4-20. The steeply cross bedded unit, which is the eastward continuity of unit 2 in Fig. 4-20, terminated 60 m from the start point, and is replaced laterally by unit AA, and unit BB. Unit AA (60-550 m) is dominated by flat beds. Unit BB, which overlies unit AA, consists of cross beds (multiple directional and unidirectional) and discontinuous flat beds. Subunits BB1 and BB4 consist of bi-directional cross beds (in bar shape), and is flanked by subunits BB2, BB3, and BB5 that are dominated by small scale cross beds and discontinuous flat beds. Subunit BB6 consists of large scale unidirectional cross beds, whereas subunit BB7 is flat bedded.