WINNIPEG '96

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FIELD TRIP GUIDEBOOK

SEDIMENTOLOGY, GEOMORPHOLOGY AND HISTORY OF THE CENTRAL LAKE AGASSIZ BASIN (FIELD TRIP B2)

by

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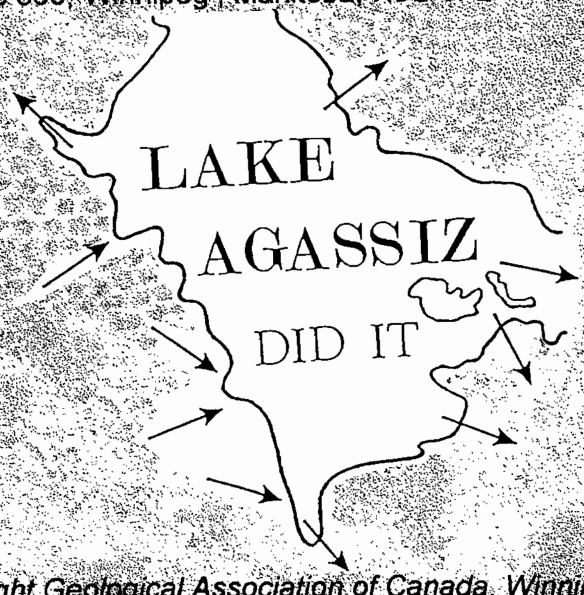

WINNIPEG '96

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SEDIMENTOLOGY, GEOMORPHOLOGY, AND HISTORY OF THE CENTRAL LAKE AGASSIZ BASIN

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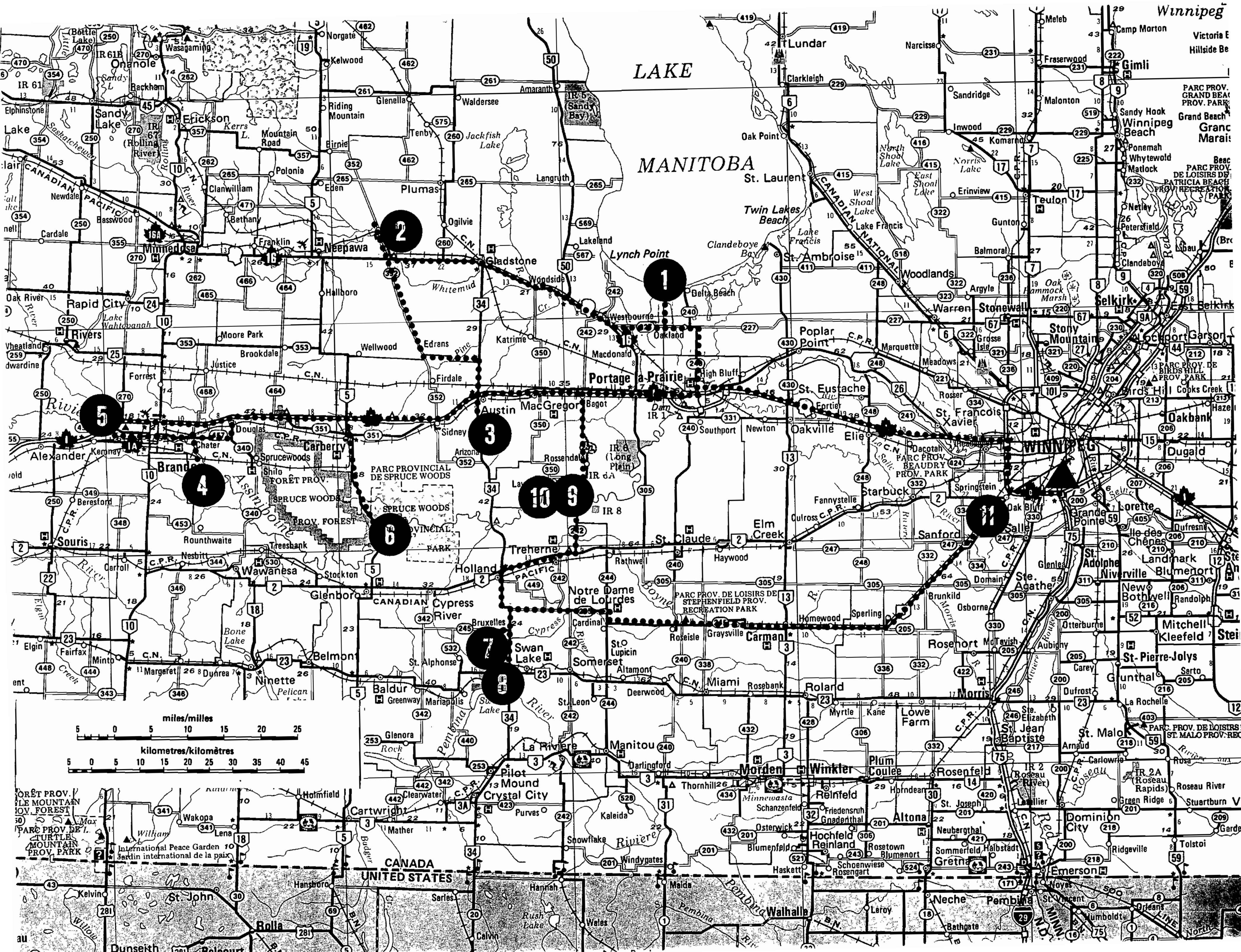
LAKE MANITOBA
Winnipeg
WINNIPEG
Portage la Prairie
Brandon
Neepawa
Minnedosa
Gladstone
Carberry
Austin
MacGregor
Treherne
Holland
Notre Dame de Lourdes
Carman
Morris
Morden
Winkler
Altona
Emerson
Selkirk
Stonewall
Gimli
Souris
Killarney
Boissevain
PARC PROVINCIAL DE SPRUCE WOODS
SPRUCE WOODS PROVINCIAL PARK
FORET PROV. SPRUCE WOODS PROV. FOREST
PARC PROV. DE LOISIRS DE STEPHENFIELD PROV. RECREATION PARK
International Peace Garden
Jardin international de la paix
CANADIAN PACIFIC
CANADIAN NATIONAL
CANADA
UNITED STATES
miles/milles
kilometres/kilomètres
1
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11

THE WINNIPEG AREA

The Quaternary sediments in the Winnipeg area (population 650,000) overlie Ordovician and Silurian bedrock, dominated by dolomite and dolomitic limestone. The bedrock geology is shown on the "Manitoba Geological Highway Map" (separate folded map). This map also describes and discusses a number of "geological highlights" of the Precambrian, Paleozoic, Mesozoic, and Cenozoic geology and history of the province, including several maps and block diagrams about the late Pleistocene.

If you arrived in Winnipeg by air, you may have noticed (if the wheat has not obscured them) large (> 1 km long) curved and straight, light-toned features on the surface (see Figure 1). These are low (mostly less than 1 m in relief) curvilinear features formed in Lake Agassiz silty clay that are better drained and, therefore, the soils have lower organic content in them; there are many thousands of these and they extend for hundreds of kilometres from the southern end of the Lake Agassiz basin to well north of Winnipeg. Clayton *et al.* (1965) concluded that they are the result of iceberg (or winter ice) drag/plough marks with adjacent pushed ridges in the offshore Lake Agassiz sediment, and therefore date from the period 12-9 ka. Recent studies in Manitoba (Woodsworth-Lynas & Guigne, 1990; Nielson & Matile, 1982) attributed the ridges to differential compaction between ice-scoured troughs filled by silty sediment and the adjacent clays. These ideas are more fully developed at one of our stops (STOP 11).

Surface sediments at Winnipeg and in the Red River Valley are composed mainly of silty clay ("gumbo") deposited during the last (Emerson) phase of glacial Lake Agassiz, 10-9.2 ka. Ice-rafted boulders are scattered across the surface or occur within the lacustrine clay but, overall, the upper half of the clayey sediment is free of ice-rafted material, whereas the lower part is rich in this detritus. In places in the Winnipeg area and west to Portage la Prairie, a thin (< 1 m) silt-rich bed lies near the surface, which poses major engineering problems because of its instability; this deposit may be related to Holocene flooding of the Assiniboine River between Portage la Prairie and Winnipeg, perhaps after the re-diversion of this river from the Lake Manitoba basin (see discussion at STOP 1), or to density underflows on the floor of Lake Agassiz during its late stages (see STOP 11).

The stratigraphic sequence of Lake Agassiz silty clay is < 20 m thick at Winnipeg and overlies carbonate-rich silty clay (loamy) till, which overlies carbonate bedrock. The lacustrine deposits are poorly to non-laminated and quite uniform in character from top to bottom. Little pollen and few microfossils have been found. As Figure 2 shows, this sequence thickens toward the south and west; westward the Lake Agassiz sediment becomes coarser grained.

A few summary comments on the "Environmental geology of Winnipeg" (groundwater, Red River floodway, engineering problems, etc.) are presented in the separate reprint.

Figure 1. Iceberg scour marks on floor of Lake Agassiz southeast of Winnipeg. The lighter tones are more silty, better drained, and relatively low in organic matter.

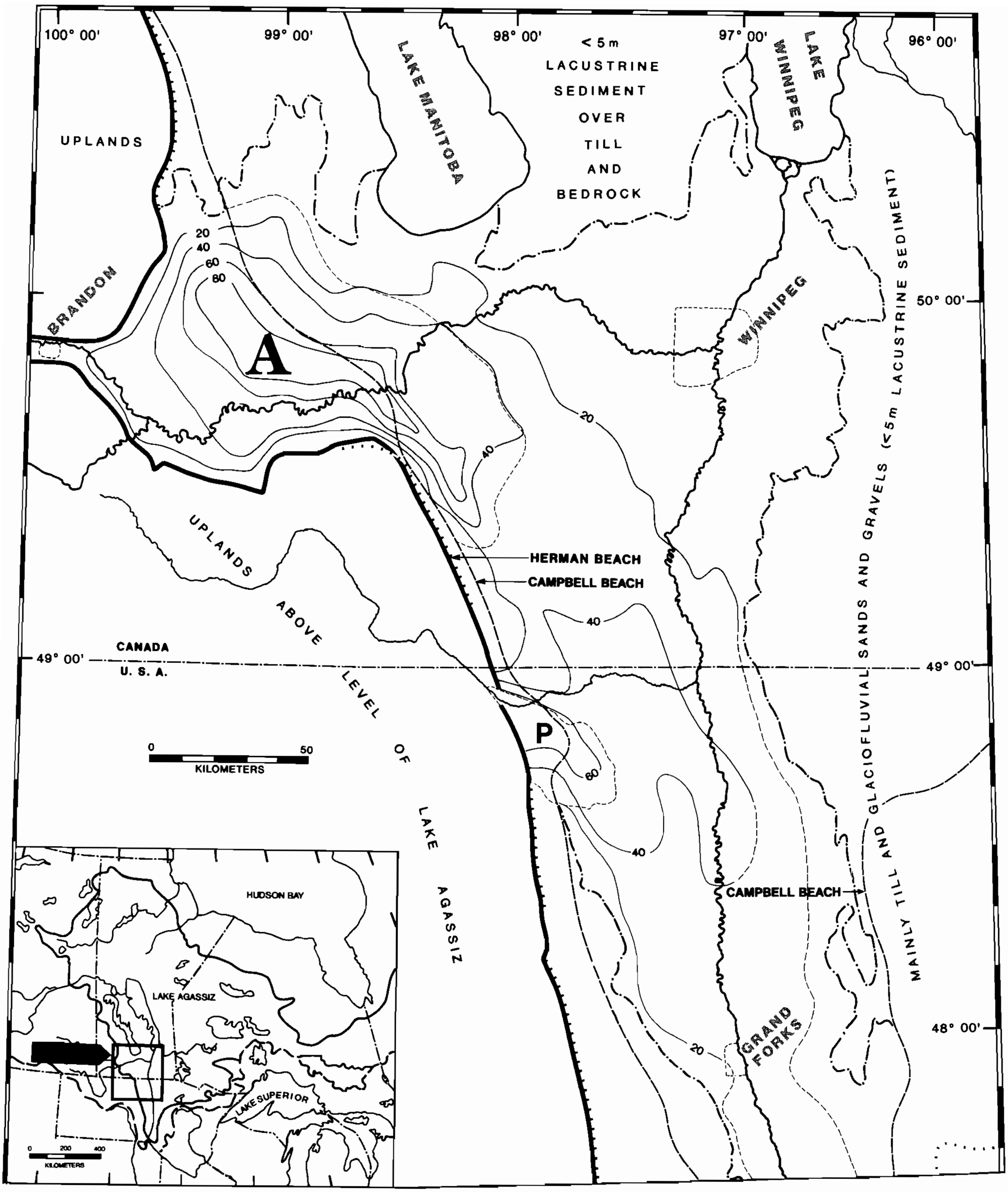

Figure 2. Isopach map (in m) of lacustrine clay, silt, and sand in southern Lake Agassiz basin, showing the highest regional beach (Herman), the most extensive beach (Campbell), the outline of the Assiniboine (A) and Pembina (P) fan-deltas, and other sediment types and thicknesses (J. Teller, unpublished).

GENERAL INTRODUCTION TO LAKE AGASSIZ

Glacial Lake Agassiz not only was important in the late glacial history of this region, but also helped control the routing of meltwater from the western Laurentide Ice Sheet (as well as influence the retreat and readvance/surging of this ice sheet) (see, for example, Teller, 1987,1990, 1995; Clayton *et al.*, 1984). Hydrological systems "downstream" were significantly influenced by the highly variable overflow from Lake Agassiz, and the record of its overflow into the Mississippi River valley, Great Lakes, St. Lawrence valley, and Mackenzie valley is only now being evaluated. In turn, the impact of water variably routed from the Agassiz basin (Figure 3) has been implicated in altering late glacial ocean circulation and climate (e.g. Broecker *et al.*, 1988,1989, 1990; Keigwin *et al.,* 1991).

Therefore, the history of Lake Agassiz and its overflow (Figure 3) is important to North American deglaciation and to considerations of global change. For many of the changes it may be true that "LAKE AGASSIZ DID IT".

The route of this fieldtrip is shown on the photocopied highway map, with the stops numbered. There are a number of key components in the Lake Agassiz system that are essential for an understanding of how this huge glaciohydrological system worked, including the location of the ice margin, the lake's outlets, and isostasy. Data on all of these come from the sediments and morphology of the Lake Agassiz basin and from various spillways, lakes, and oceans outside of its 2,000,000 km^2 drainage basin.

We will visit a selection of sites that, hopefully, will convey an idea of what is in the central Lake Agassiz basin, and about how the system operated during its 4000-5000 year life during the last deglaciation. There are a number of controversial aspects, and the history of this giant lake and its impact on systems "downstream" are today being investigated with greater rigour than ever before. Part of the difficulty relates to the size of the lake (and the remoteness of the bulk of its evidence) – the lake spans about 15° of latitude and 20° of longitude. "New discoveries" in one corner of the basin, for example in the Northwestern Outlet area, cannot be treated in isolation from "evidence" and interpretations in other areas, for example in the Southern Outlet (1500 km away), and even in the Gulf of Mexico, because the whole system is interactive.

A section in the Appendix to this Fieldguide, "Review of Lake Agassiz history" provides background on the evolution of ideas about the history of the lake and describes the main components used to make these interpretations; a revised history of the lake is also presented, mainly as a series of maps and diagrams with expanded captions (Figures 27-41). No doubt some aspects of this history will be controversial, and there are alternative scenarios to some parts of the reconstruction. Further refinements and revisions no doubt will occur, and we encourage all researchers to contribute to our understanding of this important lake.

The fieldtrip will look at 1) offshore and nearshore sediments and morphology (STOPS 1, 3, 9, 10, 11), 2) shoreline sediments and morphology (STOPS 1, 2), 3) fluvial morphology and deposits of inflow spillways (STOPS 4, 5, 8), 4) aeolian deposits (STOP 6), and 5) glacial sediments (STOPS 7, 10).

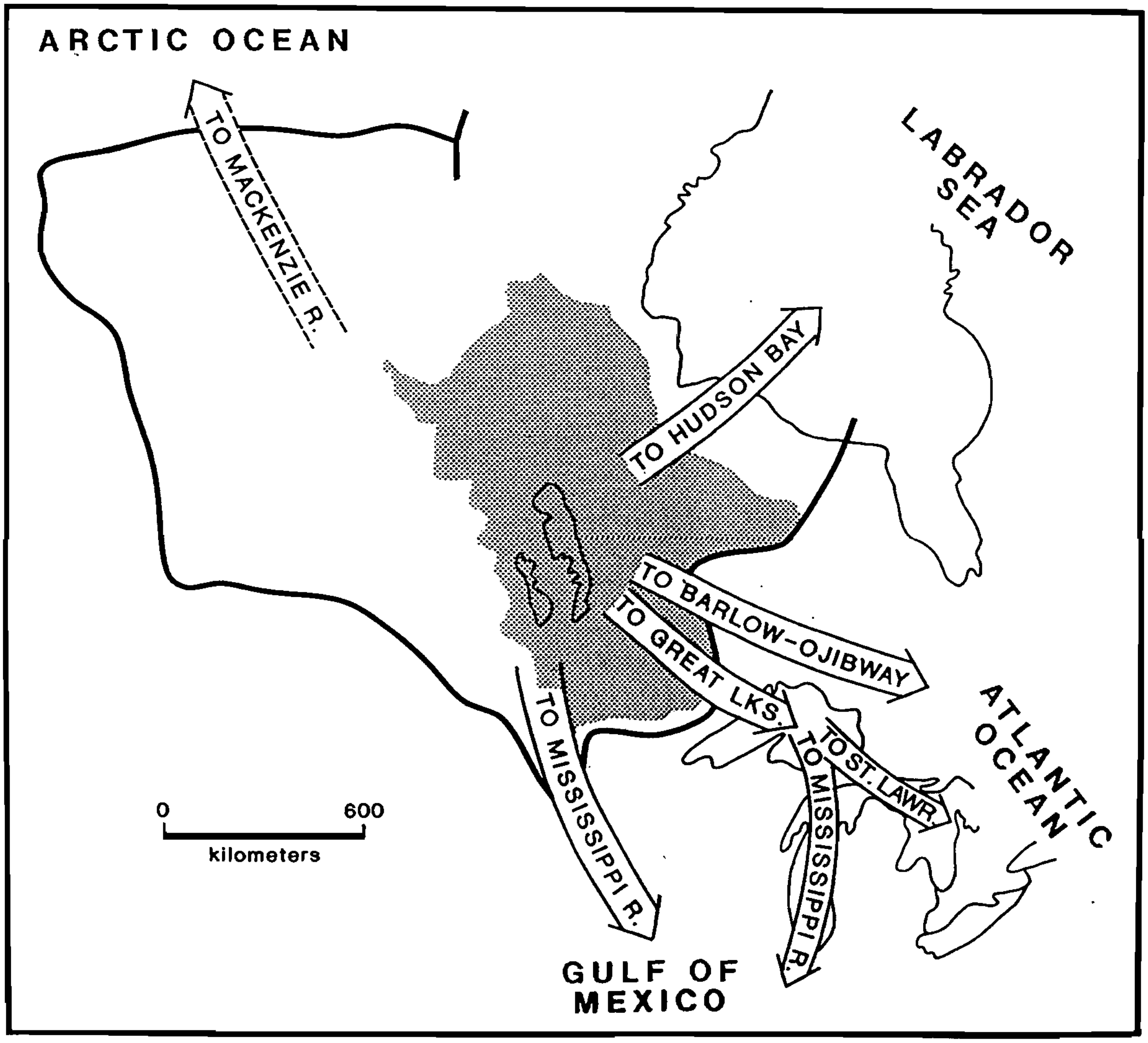

Figure 3. Major overflow routes from the Lake Agassiz basin, showing total area ever covered by the lake (stippled) and the outline of its drainage basin (J. Teller, unpublished).

DAY 1

WINNIPEG TO DELTA MARSH FIELD STATION

We travel west around the city of Winnipeg on the Perimeter Highway to the Trans Canada Highway (No.1) and head west toward Portage la Prairie. There is almost no relief on the Lake Agassiz plain, except for a few metres of incision by postglacial rivers (mainly the Red and Assiniboine rivers and its small tributaries) and for the subtle (1-3 m) almost imperceptible swells and swales related to ice-scour marks and bedrock relief (see Figure 1 and STOP 11). Surface materials are dominantly silty clay with local veneers of silt.

As the elevation rises toward the west (by about 30 m in the 80 km between Winnipeg and Portage la Prairie), surface and subsurface Lake Agassiz sediments become somewhat more silty because the source for much of the offshore sediment in this region was the Assiniboine Spillway, which entered the basin along the western side of the lake.

On the eastern side of the town of Portage la Prairie (population 13,500) is the Fort la Reine Museum and Pioneer Village. Some of the buildings are original structures moved to this site from nearby areas, others are replicas. The village and the artifacts in each building are intended to show the way of life in the eastern Prairie region of Canada from the 18th through the early 20th century. There is an old fort and trading post, log homestead, country church and school, barn and machinery, fire hall, railway display, and more.

Today the Assiniboine River Diversion, which we cross en route to the Field Station (Figure 4), is the only link between the Assiniboine River and Lake Manitoba; this channel carries excess spring melt from the Assiniboine River to the lake for several months each spring in order to alleviate the flood threat in this very flat region. During the Holocene, the natural terminus of this river has fluctuated between Lake Manitoba and the Red River at Winnipeg, 100 km to the east, where it is today (see Figures 5,6, and 7). This influx of clastic sediment had a major impact on the sedimentology of the basin, which we will discuss at STOP 1. Relicts of the old Assiniboine River distributaries to the lake are scattered throughout the region (see Figure 6) between the modern river (Portage la Prairie area) and the present shoreline. Figure 4 shows one of these old distributary channels (Blind Channel) and the modern Assiniboine Diversion that links the Assiniboine River and Lake Manitoba. We cross both of these en route to the Field Station, which is located on the modern beach that lies between the lake and the truncated end of Blind Channel (Figure 4).

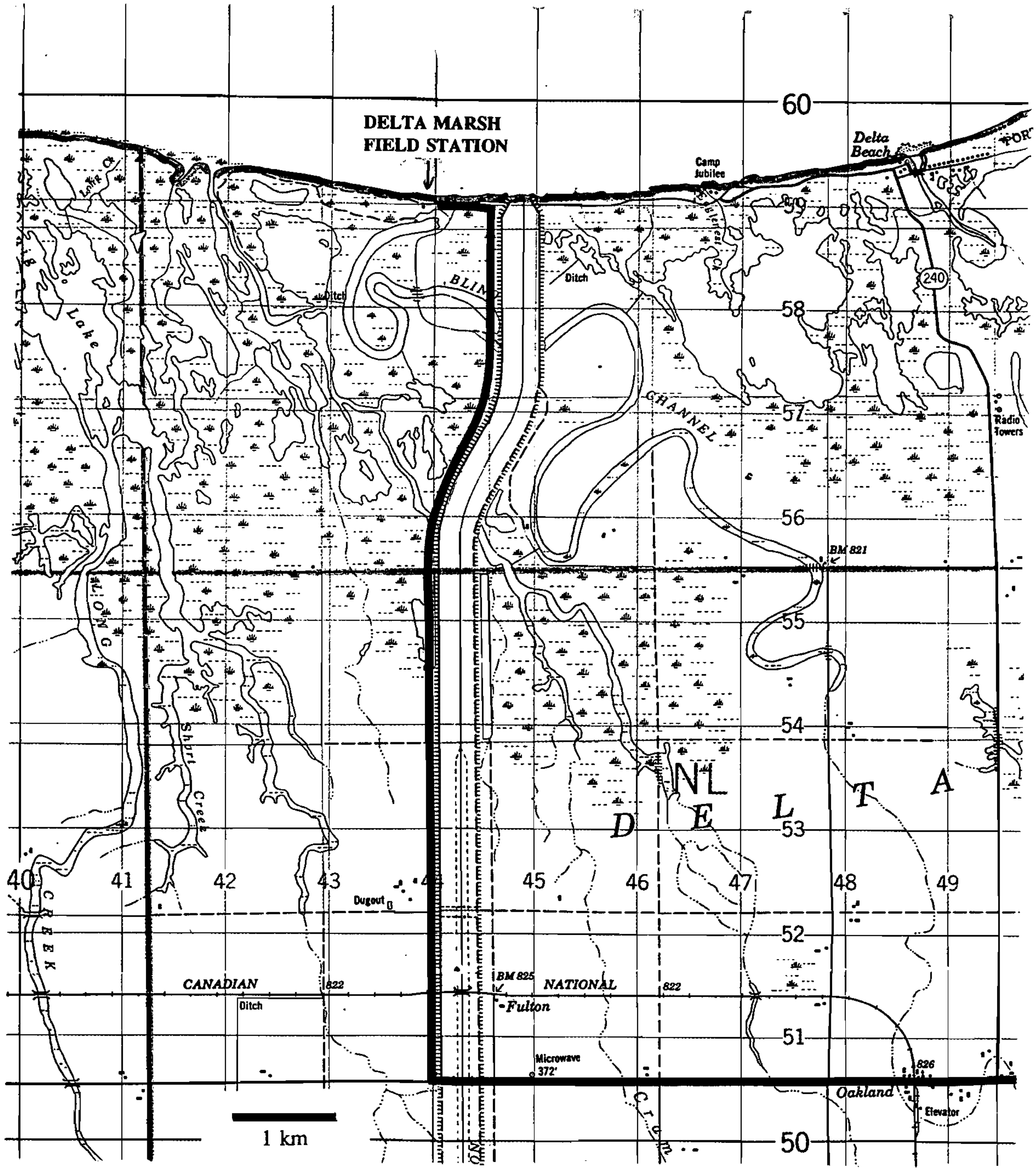

Figure 4. Map showing Delta Marsh Field Station on barrier beach between Lake Manitoba and the marsh. Blind Channel, one of the mid-Holocene distributaries of the Assiniboine River can be seen, as can the Assiniboine River Diversion channel that diverts excess spring runoff from the modern Assiniboine River (near Portage la Prairie) north into Lake Manitoba. Our route, which is highlighted, crosses Blind Channel several times.

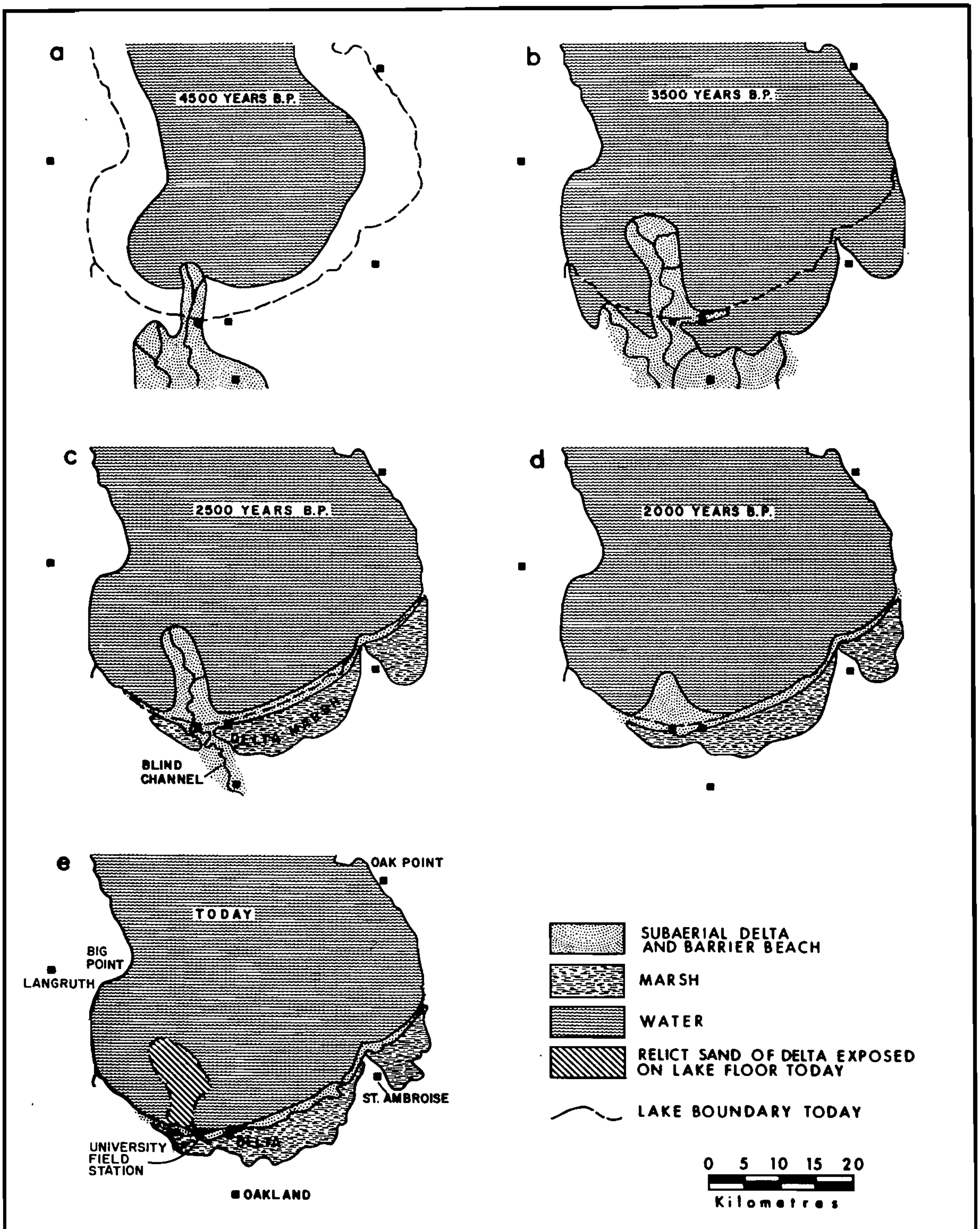

Figure 5. Maps showing the chronology of development of the small Assiniboine fan-delta in Lake Manitoba during the Holocene, and subsequent reworking by wave action and longshore drift to form a barrier beach and Delta Marsh (Teller & Last, 1981).

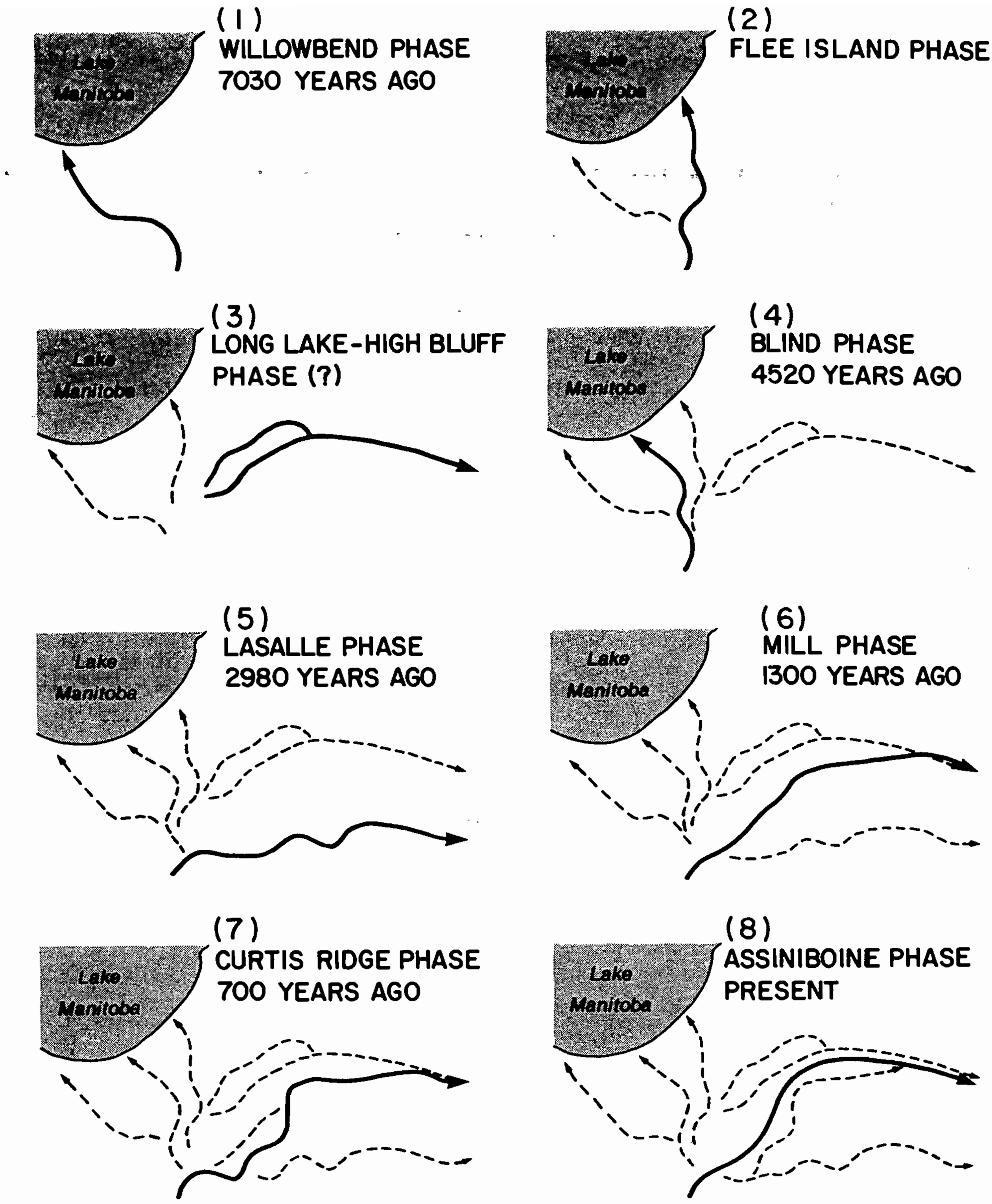

Figure 6. Schematic evolution of Assiniboine River distributaries during the Holocene (Rannie et al., 1989).

STOP 1 and OVERNIGHT
DELTA MARSH FIELD STATION AND OVERVIEW OF LATE QUATERNARY SEDIMENTATION

The University of Manitoba Delta Marsh Field Station was established in 1966, originally being a private lodge for duck hunters. Today, it serves mainly as a base for biological research activities in the region.

The Field Station lies on a sand barrier, constructed during the Holocene by the reworking of old fluvio-deltaic sediments (Figure 5) at the south end of Lake Manitoba. Behind the barrier is Delta Marsh, one of the largest waterfowl staging areas in North America.

As discussed at this stop, the level of Lake Manitoba has fluctuated dramatically since its inception following the lowering of Lake Agassiz below its rim, probably about 8200 BP. Ostracodes, diatoms, pollen, stable isotopes, and sediment characteristics in the lake reflect these changes, as does the history of the Assiniboine River to the east, and were the result of several interacting factors:

(1) differential isostatic rebound, which shifted the water mass from north to south in the basin,
(2) erosion of the northern outlet,
(3) variable influx of the Assiniboine River,
(4) climate,
(5) variable contribution (composition and relative volume) of groundwater.

Articles by Teller & Last (1981, 1982), Last & Teller (1983), Last (1982, 1984), Rannie *et al.* (1989), Nambudiri *et al.* (1980), Nambudiri & Shay (1986), Last *et al.* (1994), and others elaborate on various aspects of sedimentation and history in this basin. Current studies by Risberg, Curry, Ito, Last, and Teller will be discussed at this stop.

The early history of Lake Manitoba is largely the history of Lake Agassiz in this region, because it formed a sub-basin on the floor of Lake Agassiz for several thousand years. Once Agassiz levels fell below the rim of Lake Manitoba, the lake became largely independent of its "parent" lake, although regional climate and hydrology, including groundwater flow, must have been influenced by Lake Agassiz for some time after this. After Lake Agassiz fell to about the Gimli beach level, its waters were confined by till-covered Paleozoic bedrock to the north, east, and west (its basin is in a glacially-scoured trough) and by a now-buried ridge of till just south of the modern Delta Marsh area (Fenton and Anderson, 1971; Teller & Last, 1981); overflow at this time, as today, was eastward through the Fairford River channel toward Lake Winnipeg (Figure 7).

Sediments deposited in the early stages of Lake Manitoba (i.e. during its early Agassiz phase) are silty clays containing ice-rafted clasts (bedrock, till, and silt fragments). Above this, there are only a few obvious lithological differences in the sequence except in the southern end of the Lake Manitoba basin where sandy sediment from the Assiniboine River forms a Holocene fan-delta (Figure 5).

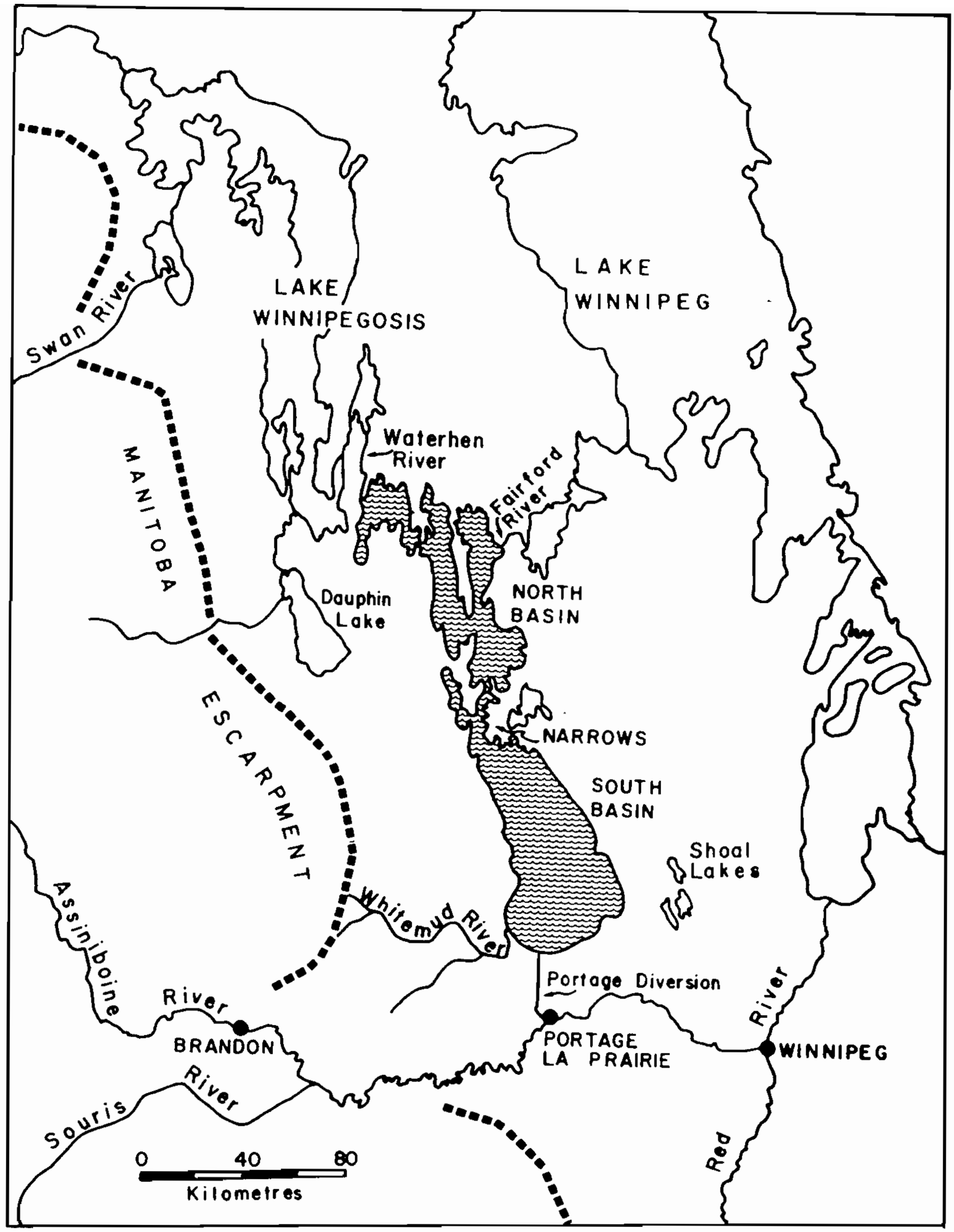

Figure 7. Lake Manitoba (patterned) and major lakes and rivers of the region, including the Fairford River outlet from the lake. The dashed line represents the Manitoba Escarpment which separates the low and relatively flat area to the east (i.e. the Agassiz basin) from the elevated area to the west (Teller & Last, 1981).

Within the lower few metres of cores from drill holes that have penetrated the entire late Quaternary lacustrine sequence, there are almost no Quaternary organic remains. Investigations for diatoms (by Risberg), ostracodes (by Forester and Curry), and molluscs (by Pip) show the oldest sediments to be almost barren of fauna and flora; the occasional presence of the ostracode *Candona subtriangulata* in this lower zone (Last *et al.*, 1994; Curry *et al.*, 1995) indicates a cold dilute lake, which has been interpreted to represent the Lake Agassiz phase (*ca.* 11-8.2 ka) of Lake Manitoba.

Dating of the lower (Lake Agassiz) part of the sequence has been fraught with problems. Samples of dark-colored muds are known to have been contaminated by finely-disseminated old carbon, based on the presence of high percentages (up to 80%) of pre-Quaternary palynomorphs (Teller *et al.*, 1981; Nambudiri *et al.*, 1980; Teller & Last, 1981); although dates on these samples were "corrected" by using the percentage of old palynomorphs as an indicator of total old carbon, they remain at best general chronological indicators. Because of the scarcity of organics in the Agassiz part of the sequence, further resolution may not be possible; a new AMS date of 4040±40 (Woods Hole #OS-02659; S.M. Colman, 1994, written commun.) on ostracode valves from this lower zone has added confusion and probably relates to a labelling error.

$\delta^{18}O$ values on ostracode valves in the Agassiz part of the sequence are relatively low at -15 to -17 ‰ PDB (Last *et al.*, 1994).

Although no dates clearly define the end of the Agassiz phase of Lake Manitoba, it is interpreted to have occurred around 9.2 ka (Teller & Last, 1981). This change is marked by an abrupt increase in diatoms and ostracodes and by changes in some other measured parameters such as magnetics, moisture content, and $MgCO_3$ content in calcite.

The history of Lake Manitoba after it became independent from glacial Lake Agassiz is complex. Diatoms (Risberg), ostracodes (Curry), stable isotopes (Ito), and various physical and mineralogical parameters (Last and Teller) are being used to interpret the varying hydrological conditions in the basin, which were a function of isostasy, outlet erosion, variable influx of the Assiniboine River, climate, and variable contribution of groundwater. Lake levels and the chemistry of this shallow lake ($Z = 4.5$ m; $Z_{max} = 6.3$ m) fluctuated substantially.

The Holocene history of Lake Manitoba has been studied from 48 short (< 4 m) cores and 6 long cores collected by Livingstone coring and by Shelby tubes in hollow-stem augers from winter ice in the South Basin. More than 15,000 analyses have been made, including grain size, moisture content, clay mineralogy, carbonate mineralogy, quartz and feldspar content, pH, Eh, magnetics, organic matter content, organic typing, trace element analyses (K, Na, Ca, Mg, Fe, Mn, P, Zn, Cd, Cu, Pb, Hg, Cl, S), and stable isotopes. Detailed studies of the pollen, ostracodes, and diatoms (and other siliceous microfossils) also have provided important paleoenvironmental insight.

Major problems (uncertainties) have occurred in resolving the chronology of events in the basin, because most of the vegetal organic material is microscopic and some has been contaminated by pre-Quaternary organics; mollusc and ostracodes shells are either too few in number to be dated (in the early, Agassiz part of the sequence) or are potentially contaminated by the "hard-water effect".

Diatoms and ostracodes suggest that waters were fresh to brackish in the early Holocene. The combination of warming temperatures, differential isostatic rebound, and (possibly) changes in groundwater discharge are reflected by diatoms, phytoliths, ostracodes, oxygen isotopes, and calcite geochemistry in sediments of this large but very shallow lake. Evidence within and to the south of the basin is that the Assiniboine River, which today by-passes the lake, played an important role during the mid-Holocene (Figure 5; Teller & Last, 1981; Rannie *et al.*, 1989). Water depths varied during this period as reflected by several low-moisture, pedogenic-like zones in the sediment (Teller & Last, 1982).

Another dramatic change in sediment parameters, especially in diatoms, ostracodes, $\delta^{18}O$, calcite, and moisture content occurred at 4-5 ka; these changes overlie a widespread low moisture zone found in every core that penetrated to a depth of 3.5 m in the sediment, and represents a period of prolonged low (or no) water in the basin when P/E ratios were low and the Assiniboine River was diverted away from the basin. The final diversion of the river away from the basin, higher P/E ratios, continuing differential isostatic rebound, and (possibly) changes in groundwater discharge have resulted in varying brackish water conditions over the past several thousand years.

Abstracts for this GAC meeting by Curry and Ito *et al.* can be consulted for further elaboration.

The barrier beach on which Delta Marsh Field Station lies formed largely by the reworking of Assiniboine River fluvio-deltaic sediments, and has been transgressing south over marsh sediments for at least 3000 years, in part because of differential isostatic rebound. In places along the face of the beach, these marsh sediments are being re-exposed, along with the organic-acid stained bones and teeth of bison.

Because new data from a new core is being studied as this Guidebook is being prepared, a more elaborate, and possibly revised, history of the lake will be presented and discussed at this stop.

DAY 2

DELTA MARSH FIELD STATION TO BRANDON TO BRUXELLES, RETURN

Seven Stops

EN ROUTE TO NEXT STOP

West from Delta Marsh Field Station we traverse the flat silty clay surface of the Lake Agassiz plain. Although the surface rises, and sediment becomes more silty and sandy, it is not until we reach the best-developed beaches of Lake Agassiz, the Campbell beaches (Figure 2 & 8) that the topography changes. This beach extends from the southern outlet of Lake Agassiz, 500 km to the south in North Dakota, irregularly along both the eastern and western lake margins, (see separate coloured map "Maximum extent and major features of Lake Agassiz"), and is composed of a complex of barrier beaches, lagoons, and wave-cut strandlines. The northern end of the beach has been differentially uplifted more than 150 m above the southern end (see Figure 28 in Appendix), as have the many other Lake Agassiz strandlines associated with the interaction of isostasy, outlet erosion, and fluctuations of the Laurentide ice margin.

Interpretations of the age of the Campbell beach vary, and some have considered it a composite beach, formed during equilibrium levels both as the lake fell during the Moorhead low water phase 11-10 ka (~Younger Dryas) and again about 9.7 ka during the last fall from the Emerson phase of Lake Agassiz. Dates confirm beach (and lagoonal) sedimentation did occur during the Emerson phase.

Directly west of Portage la Prairie, the Campbell beach strandlines were eroded from and cut into sediments of the old Assiniboine River fan-delta (Figure 8; STOPS 3, 4, 9), which was deposited into Lake Agassiz during the early (Lockhart) phase of the lake, about 11.3-10.8 ka. See Figure 9. West of the Campbell beach escarpment in that area the surface is sandy, and the thick stratigraphic sequence (see Figure 2) was deposited rapidly by density underflow currents (off the mouth of the Assiniboine Spillway) and by nearshore (wave) processes. In areas where the fan-delta surface is comprised of very fine to medium grained sand, irregular "blowouts" and associated downwind dunes have periodically formed by the wind; nearly all of these dunes are stabilized by vegetation today. The separate 1:250,000 "Quaternary Geology Map" of Brandon shows these dunes and the nature of surface deposits in this region.

STOP 2
CAMPBELL BEACH RIDGE AT ARDEN

Arden is situated on the upper Campbell strandline which is locally known as the Arden ridge (Figure 10). The upper Campbell beach is the best developed and the most continuous shoreline of Lake Agassiz. As Figure 28 shows, the strandline defines a parabolic surface whose slope decreases toward the south. Locally the strandline is at an elevation of 333 m and slopes at approximately 0.25 m/km (Johnston, 1946). The upper

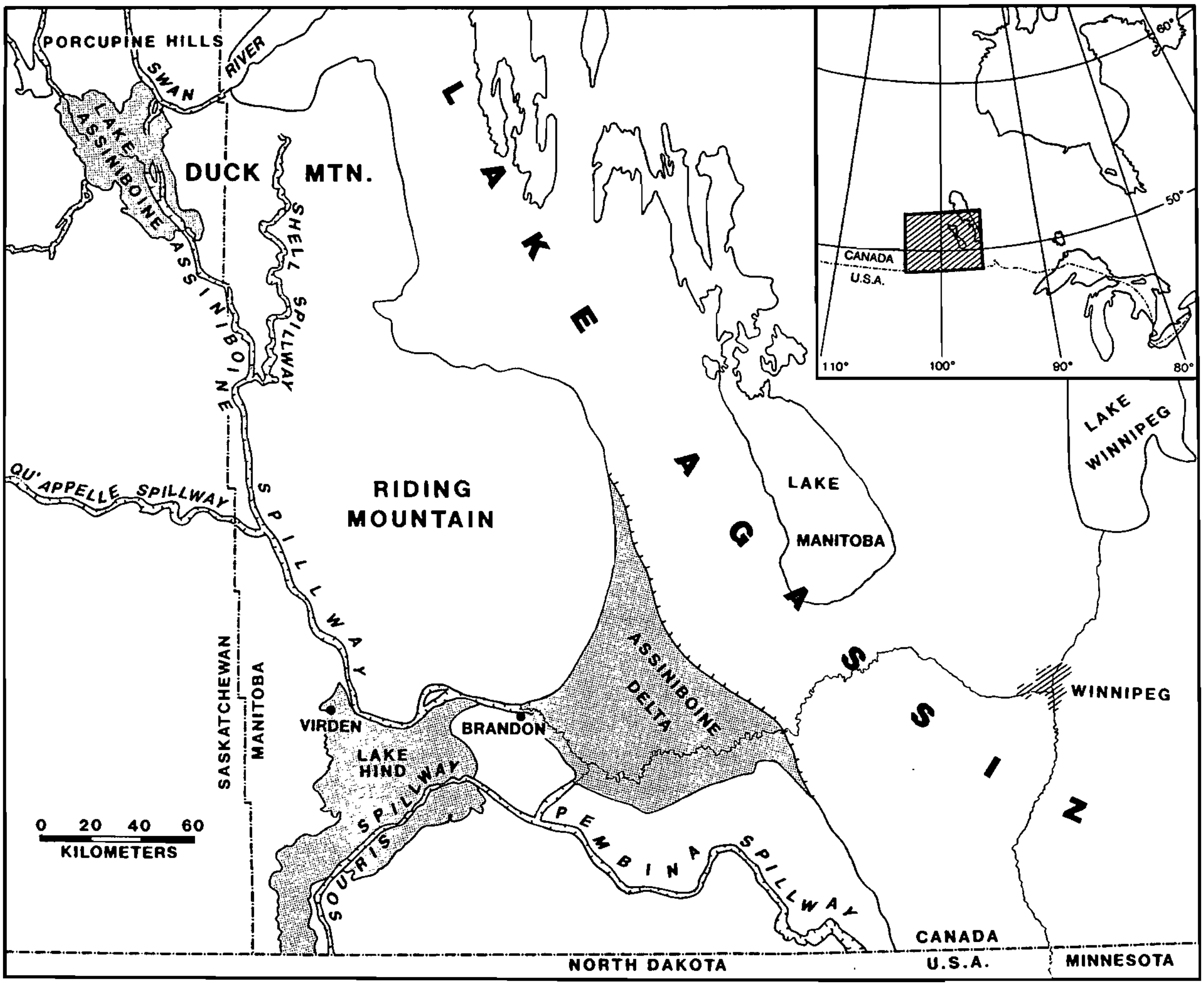

Figure 8. Major features in the central Lake Agassiz basin, including the fan-delta and major glacial lakes and spillways leading to the lake. The Campbell Beach marks the western limit of the lake during its later stage and is indicated by the hachured line (and its continuation) across the delta (Wolfe & Teller, 1993).

Campbell strandline is approximately 98 m above Winnipeg.

Because the Arden area lies along the Manitoba Escarpment, there is lots of local relief and, therefore, a number of strandlines nearby. A flight of seven beaches, located 10-15 km west of Arden, range in elevation from 360 to 378 m and approach the upper limit of Lake Agassiz in the area. These beaches represent the Herman strandlines and were built during the Lockhart phase of Lake Agassiz. To the east, between Arden and Gladstone, Johnston (1946) measured the elevation of eleven strandlines, which range from 333 to 269 m. This second flight of beaches, which includes the Upper and Lower Campbell, were built during the Emerson and Nipigon phases of Lake Agassiz and represent the final regression of the lake.

In addition to their importance in establishing the history of Lake Agassiz, beaches were vital to paleo-Indians and early settlers. They are 'high and dry' and continuous for many kilometres; as such they provided excellent transportation routes. This was especially important in spring when the surrounding lower areas were wet. Also, because they are relatively easy to dig, they provided good burial sites. Beaches are still considered to be an excellent choice for road building and cemetery plots.

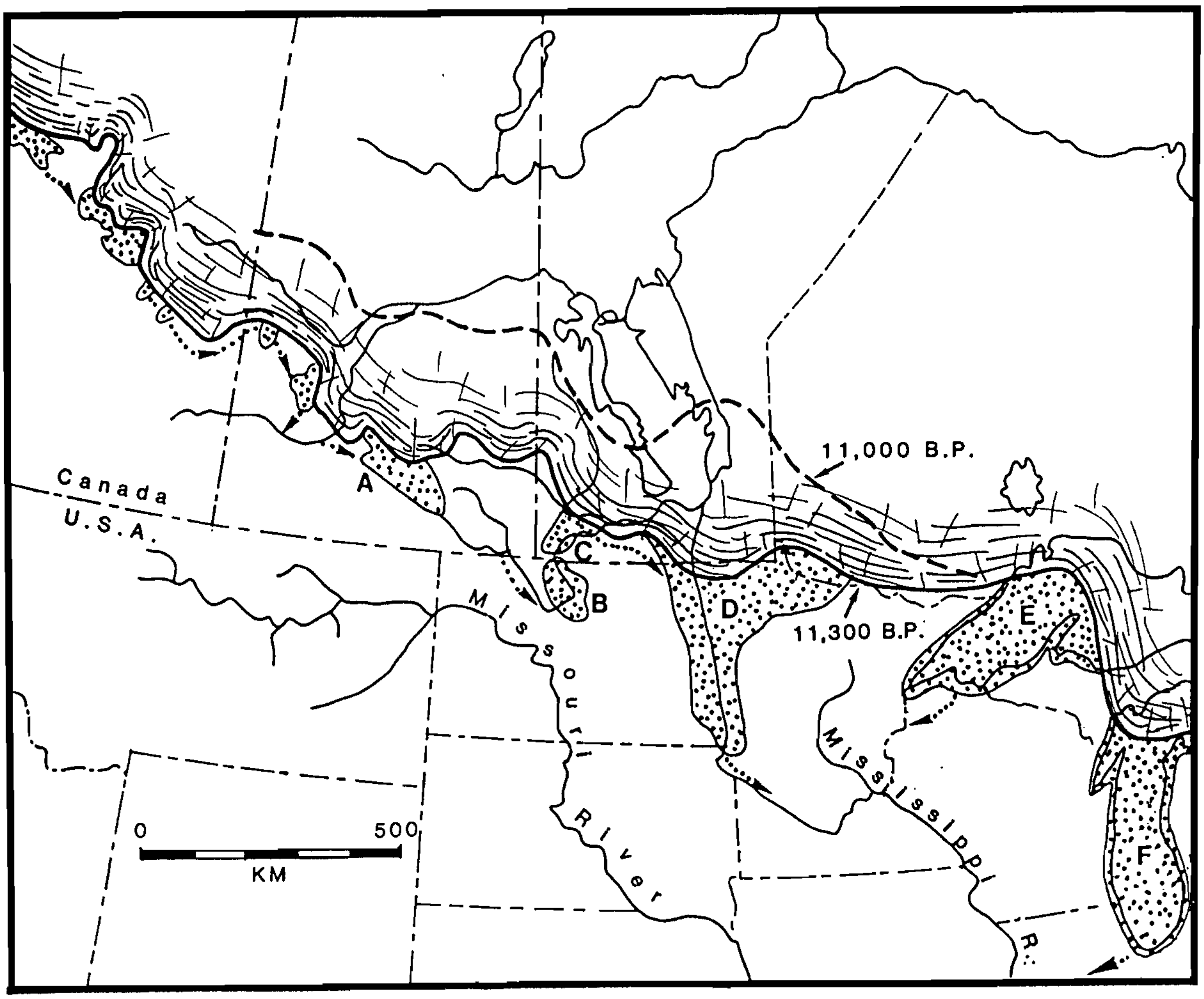

Figure 9. Proglacial lakes along the western Laurentide ice margin about 11.3 ka; the glacial boundary and additional area of Lake Agassiz at 11 ka are shown by heavy dashed line. All lakes in this region overflowed south into the Mississippi River at this time. A = Lake Regina, B = Lake Souris, C = Lake Hind, D = Lake Agassiz (Lockhart Phase), E = Lake Superior (Duluth Phase), F = Lake Michigan (Calumet Phase) (Teller, 1987).

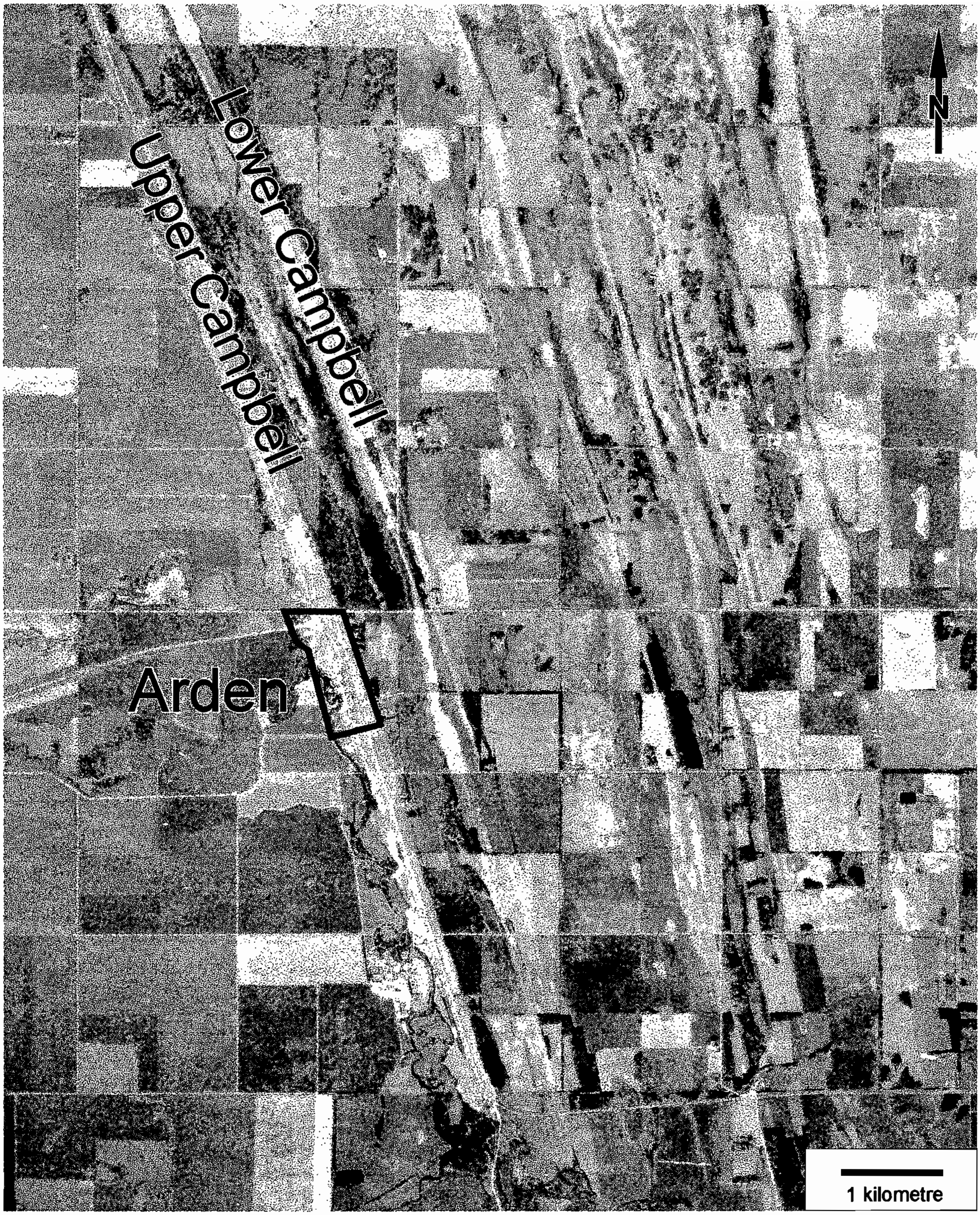

Figure 10. Aerial photo of Arden area showing Campbell beaches.

Section Description:

In the fall of 1995, a sand and gravel pit, 1 km east of Arden, provided a good example of the sediments formed in the lower Campbell beach in this area. The lower Campbell is 10 m below the upper Campbell strandline in this area. As sand and gravel pits tend to be active on a somewhat random basis and a clean pit face is not certain, a substitute pit may be chosen. As a result, the section description is purposely generalized.

The deposit consists of >2 m gently dipping sand and pebble gravel with low angle truncations and 10 cm of cross-bedded sand units (paleocurrent = 130°, although multiple directions were observed).

EN ROUTE TO NEXT STOP

For about 8 km south of Arden, the Campbell beach continues as a barrier island, composed of sandy to gravelly, flat-bedded sediment, with a marshy paleolagoon just to the west; Highway 352 is built on this barrier beach. Farther south, the road lies mainly along the base of a small escarpment, cut by Lake Agassiz waves into older fine sands and silts. These fine sediments were deposited by density underflow currents into the lake when it was at higher levels during the Lockhart phase, and are part of the distal accumulation of the Assiniboine River fan-delta complex. Therefore, the bulk of the Campbell beach along this stretch is composed of these wave-eroded sands, and only occasionally is a coarser Campbell beach distinguishable. To the west of this wave-cut shoreline, the surface of the Lockhart underflow fan has been dissected by spring sapping, because these sands and silts overlie clayey impermeable offshore sediment. In some areas, as discussed later, this surface has been reworked into dunes during the Holocene.

Shortly after we reach the Trans Canada Highway we turn south on Highway 34 at the town of Austin. Just after this, the road rises up over the Campbell Beach, which is a wave-eroded escarpment into the older Lockhart underflow fan sands and silts that we see at STOP 3. Note the old steam tractors and farm machinery on the shoreline at the Manitoba Agricultural Museum.

About 4 km farther south is a cut into the distal fine-grained sediments of the Assiniboine fan-delta, whose apex lies 70 km to the west at Brandon (see Brandon 1:250,000 "Quaternary Geology Map").

STOP 3
DISTAL SEDIMENTS OF ASSINIBOINE FAN-DELTA

Exposed here are 9 m of well-laminated and bedded silt and very fine sand that display a wide variety of sedimentary structures and deformational (loading) structures. All of these features suggest rapid sediment influx to this site, probably by density underflow currents entering Lake Agassiz from > 50 km to the west.

These sediments were deposited during the early stage of Lake Agassiz (*ca.* 11.3-10.8 ka in this region), and are typical of the distal part of the Assiniboine fan-delta. The thickness of the Assiniboine fan-delta exceeds 50 m in places (see Figure 2) and covers an area of about 6400 km^2 west of the Campbell beach escarpment. From bottom to top there is a gradation from more distal to slightly more proximal sediments.

The general stratigraphy at this site is:

4 m Silt to clayey silt, interbedded with silty very fine sand; well-laminated and cross laminated with ripples and climbing ripples, several intervals contain highly deformed bedding and occasional faults (mobilized during rapid accumulation over high water content zones), light olive brown (2.5 Y 5/4).

5+ m Silty clay grading up into silt, scattered variably coloured silt clasts in basal part decreasing upward; some deformational structures, poorly laminated to massive in lower part becoming laminated in upper part, upper boundary gradational, dark olive grey (5Y 3/2) at base becoming light yellowish brown (2.5 Y 6/4) in upper part.

These sediments are typical of those found throughout this part of the Assiniboine fan-delta, and were deposited in waters up to 70 m in depth; we will see a similar sequence at STOP 9. They are interpreted as representing a period of rapid sediment influx to the western side of the lake through the Assiniboine Spillway, whose mouth lay at least 50 km to the west. A long core collected 28 km to the west of STOP 3, 7 km southeast of Carberry, contained 68 m of offshore clay (near the base) that grades upward into silty fine sand (near the top). The sequence overlies till. Some of this sediment influx was related to catastrophic bursts of water from other ice-marginal lakes to the west of Lake Agassiz (e.g. Kehew & Teller, 1994a; Sun, 1993); this topic will be discussed at STOPS 4 and 5.

Rapid sedimentation is indicated by the climbing ripples and deformational structures. Some wave reworking may have taken place, and the top of the sequence may have been eroded away as Lake Agassiz fell to the Moorhead low-water phase after 11 ka. Subsequently, when the lake rose again during the Emerson phase, wave erosion occurred around the periphery of this underflow fan, resulting in formation of the Campbell strandline (see Figure 2).

EN ROUTE TO NEXT STOP

We return to the Trans Canada Highway and proceed west over the Assiniboine fan-delta surface, which rises slowly toward its apex at Brandon. In some places where the surface sediment is dominantly very fine to medium grained sand, wind deflation has occurred when the surface vegetation has been disturbed. Blowouts and associated dunes (only a few are nicely parabolic) have developed over a large area (see 1:250,000 "Brandon Quaternary Geology Map"). Today the only active area is the Bald Head Hills, about 22 km south of the town of Carberry (Figure 17).

In some of the roadcuts through the aeolian dunes in the Assiniboine fan-delta, you can see dark zones within the sand dunes. These are weakly-developed soils, formed on ancient sand dune surfaces and later buried by advancing dunes. There are many such soils throughout this area, some which have been radiocarbon dated. The dates indicate that periods of relative stability (soil development) alternated with active dune migration throughout at least the past 3700 years (see cross section, Figure 11). Some of these paleosols may be (or may have been) extensively developed over the region (as is the case today), others may only have been locally formed in the wetter interdune areas. Likewise, active dune migration probably occurred at times over a widespread area, while at other times activation occurred only locally (as temporary blowouts or sub-regional activation, as is occurring today in the Bald Head Hills).

As we get close to the town of Brandon, one of the now-abandoned routes of the late-glacial Assiniboine River can be seen to the south of the Trans Canada Highway. This broad shallow feature was eroded into gravels that were deposited during the deep-water (Lockhart) phase; this erosion may have occurred after Lake Agassiz fell to the Moorhead low-water phase. This channel, named the Epinette Creek Channel, was abandoned in favor of a more southerly route across the Assiniboine fan-delta, perhaps because of differential isostatic rebound but possibly just as a result of the normal process of distributary shifting in this delta system. The Brandon 1:250,000 "Quaternary Geology Map" clearly shows this abandoned route and the subsequent (modern) route.

A long core through the delta, at the edge of Epinette Creek Channel where now-stabilized dunes have invaded the channel (5 km northeast of Shilo), passed downward through 8 m of aeolian sand, 9 m of organic-rich alluvial (distributary channel) sediment, 19 m of Assiniboine fan-delta sand, 2 m of prodelta deep-water silty clay, and then till. Radiocarbon dates on wood and gastropods in the organic-rich alluvial sediment yielded ages of 9330 ±160 yrs BP (1253 Cc) and 9340 ± 70 yrs BP (TO 2197), respectively.

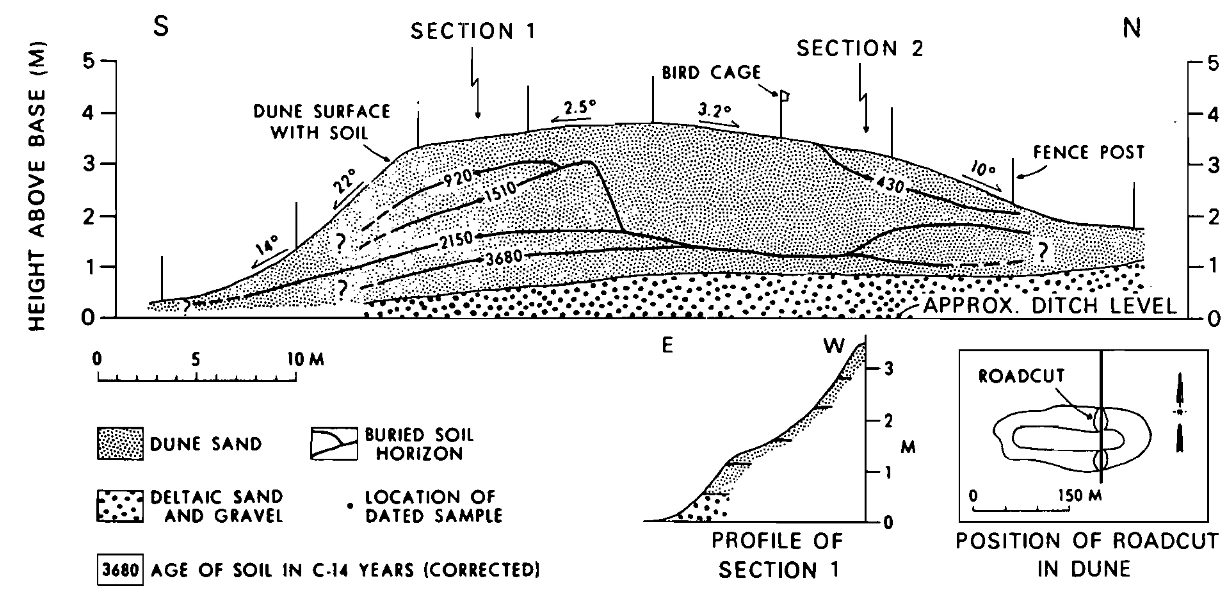

Figure 11. Diagrammatic cross section through dune in Carberry Sand Hills (Brookdale Road section), showing buried soils and their radiocarbon ages (David, 1971).

The city of Brandon (population 39,400) lies adjacent to the Assiniboine River valley which has entrenched itself into the apex of the old Assiniboine fan-delta. This is an underfit river; its valley was cut mainly during late glacial and early postglacial time by meltwater. The valley (and the area for about 10 km on either side) occupies one of the broad gaps eroded through the Cretaceous shale escarpment along the western boundary of Lake Agassiz; this Cretaceous sequence is capped in places by resistant (siliceous) Odanah shale and forms a nearly continuous east-facing escarpment for > 1000 km, from North Dakota, across Manitoba, and into Saskatchewan.

The last advance of ice into the southern Lake Agassiz basin of Manitoba pushed westward through this broad gap before 11 ka (Figure 12), depositing the Darlingford-Alexander end moraine (see this end moraine on Brandon and Virden 1:250,000 "Quaternary Geology Maps"). After this ice wasted away, Lake Brandon formed between the moraine and ice margin (Figure 13A).

Subsequently, meltwater from the Souris and Qu'Appelle valleys, which drained much of the glaciated and just-deglaciated region of the southern Canadian Prairies at this time, topped the Darlingford-Alexander moraine in the Brandon region along a broad (> 15 km) front (Figure 13B), scouring shallow channels into the surface, which are today abandoned and armoured in many places by cobbles and boulders. We will see these upland channels en route to STOP 5. This eventually led to establishment of the present course of the Assiniboine River through the city of Brandon and deposition of the Assiniboine fan-delta. This glacial river provided the largest flow of meltwater to glacial Lake Agassiz during its early history. Klassen (1975) and Christiansen (1960; 1961) describe the nature of this spillway, its sediments and history, and of the Qu'Appelle valley, which is its westward extension across the Prairies. Kehew and Teller (1994a) discuss the nature and history of all major spillways to the west of Lake Agassiz, and part of this paper is included in the Appendix to the Guidebook.

To the west of the Darlingford-Alexander moraine lies the glacial Lake Hind basin, a flat surface of silty to sandy clay, capped in places by small aeolian dunes and blowouts. This former lake basin covers 4900 km^2 and is connected to the glacial Lake Souris basin south along the Souris River in North Dakota. Water was impounded in this low area until the Pembina spillway was entrenched (Figure 12, STOP 8) and overfow through the new Assiniboine River Spillway at Brandon was established. This entrenchment and the drainage of Lakes Hind and Souris may have occurred as a result of catastrophic flooding through the Souris River valley from glacial Lake Regina (e.g. Kehew, 1982; Kehew and Clayton, 1983).

The nature of the flow of water through the Assiniboine Spillway and other late glacial channels of the Prairies will be discussed at STOPS 4 and 5. Part of the classic paper by Kehew and Lord (1987) on glacial lake outburst foods and their impact on downstream valleys and lakes is included in the Appendix to this Guidebook.

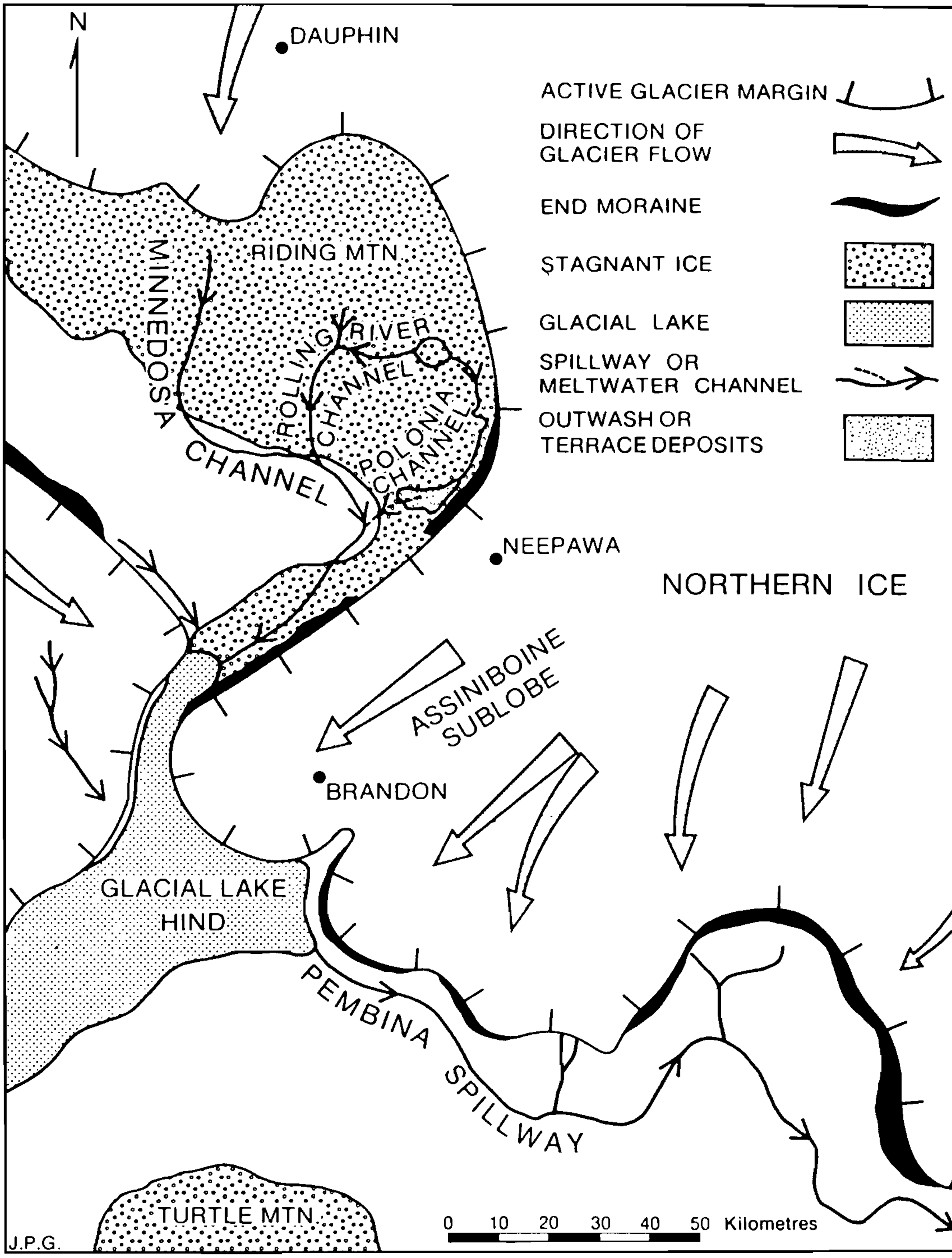

Figure 12. Position of active ice margin during its last advance into Tiger Hills (Aber et al., 1987, after Klassen, 1975, and McGinn & Giles, 1987). Overfow from Lake Hind into the Pembina Spillway included inflow to that lake through the Souris River Spillway located to the southwest (not shown).

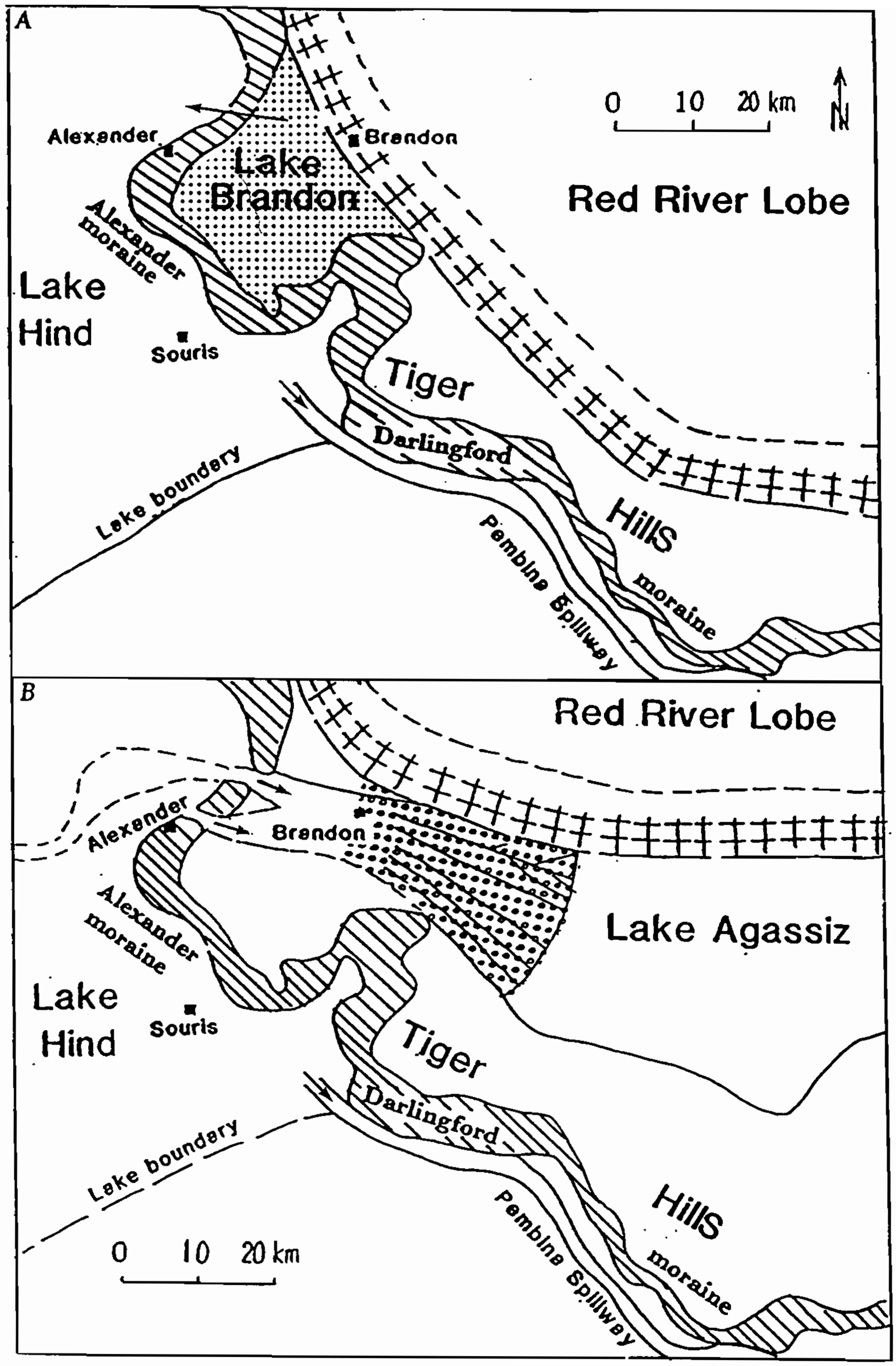

Figure 13. ***A.*** *Formation of glacial Lake Brandon between the Alexander end moraine and the retreating Red River Lobe.*
B. *Early Assiniboine fan-delta formation after the moraine was breached, initiating the Assiniboine Spillway through Brandon (Sun, 1993).*

STOP 4
FLOOD GRAVELS AT HEAD OF ASSINIBOINE FAN-DELTA

Between the towns of Shilo and Brandon lie coarse gravels deposited in the apex of the Assiniboine fan-delta. A number of pits expose eastward-dipping foreset beds believed to have been deposited by rapidly accumulating Gilbert-type delta foresets during meltwater floods (catastrophic?) from the Canadian Prairies.

Either **A** or **B** will be visited: only general descriptions are given because of the rapidly changing development of pits in this area.

A. A pit in section 5, T.10, R.17 (located at E-7 of Figure 14) exposes thick (4-5 m) foresets composed of cobbly gravel. These foresets have been traced for about a kilometre to the east in continuous cuts and by ground-penetrating radar (see radar profile in Figure 15).

Overlying the foresets is a metre of gravel. Underlying the forsets (only occasionally exposed) are horizontally bedded gravels and, below that, a unit of fine gravel in 1-m-thick forsets which contain an abundance of coal pebbles. Because it seems that coal can only have been derived from the region to the southwest (in the Souris River basin *vs.* the Assiniboine-Qu'Appelle River basin), these lower gravels may have been carried to this site by a catastrophic flood through the Souris spillway.

B. This gravel pit is in an elongated (streamlined) erosional remnant of the Assiniboine fan-delta, located at E-5 of Figure 14. The forsets dip eastward and extend across this NW-SE oriented erosional residual, which was formed after the forsets were deposited. Ground-penetrating radar confirms the eastward extension of this dip in the hill.

The thick (up to 6 m) coarse gravel forsets are crudely graded and contain large cobbles and boulders (up to 0.8 m in diameter). Some of these clasts are made of till, some of shale. The presence of soft clasts like these is typical in deposits left by high discharge, hyperconcentrated flows (Lord and Kehew, 1987) that rapidly erode channels. Various estimates for maximum flows through Prairie spillways, based on various methods of paleohydraulic flow calculation, range from 1 x 10^4 to 1 x 10^6 m^3/sec.

The thick cross-beds overlie > 4 m of gravel, which is flat bedded or has a broad channel-fill geometry. Till underlies the fluvial sediment in the Assiniboine spillway and on the scoured uplands south of Brandon, and is exposed at the surface in places between the elongated hills (see Brandon 1: 250,000) "Quaternary Geology Map".

EN ROUTE TO NEXT STOP

Although the Assiniboine Spillway west of Brandon may roughly follow a pre-last glaciation bedrock valley (Klassen, 1975), its present form is largely the result of meltwater events associated with the last deglaciation. As described by Sun & Teller (in press), the valley between glacial Lake Hind and Lake Agassiz was first eroded when ice wasted back far enough to allow part of the overflow from Lake Hind to take this route (Figure 12). This outflow, which may have been related to the catastrophic outburst through the Souris

Spillway from glacial Lake Regina (Kehew & Clayton, 1983; Kehew & Lord, 1986, 1987; Sun & Teller, in press), scoured the uplands west of Brandon forming the subupland channels we cross en route to STOP 5. Residual "islands" in this broad zone of early Assiniboine Spillway development are elongated, and coarse gravel as well as local lags of boulders are present; all of this suggests that discharge may have been abrupt and catastrophic (see Kehew & Lord, 1987, in Appendix to this Guidebook).

Flow through this part of the Assiniboine spillway was supplemented by runoff from the upper Assiniboine River Spillway, which included catastrophic flood bursts from glacial Lake Assiniboine (Wolfe & Teller, 1993, 1995). Eventually, all water from the upper Assiniboine and its "tributary" Qu'Appelle Spillway (Figure 16) were routed through the Assiniboine Spillway at Brandon.

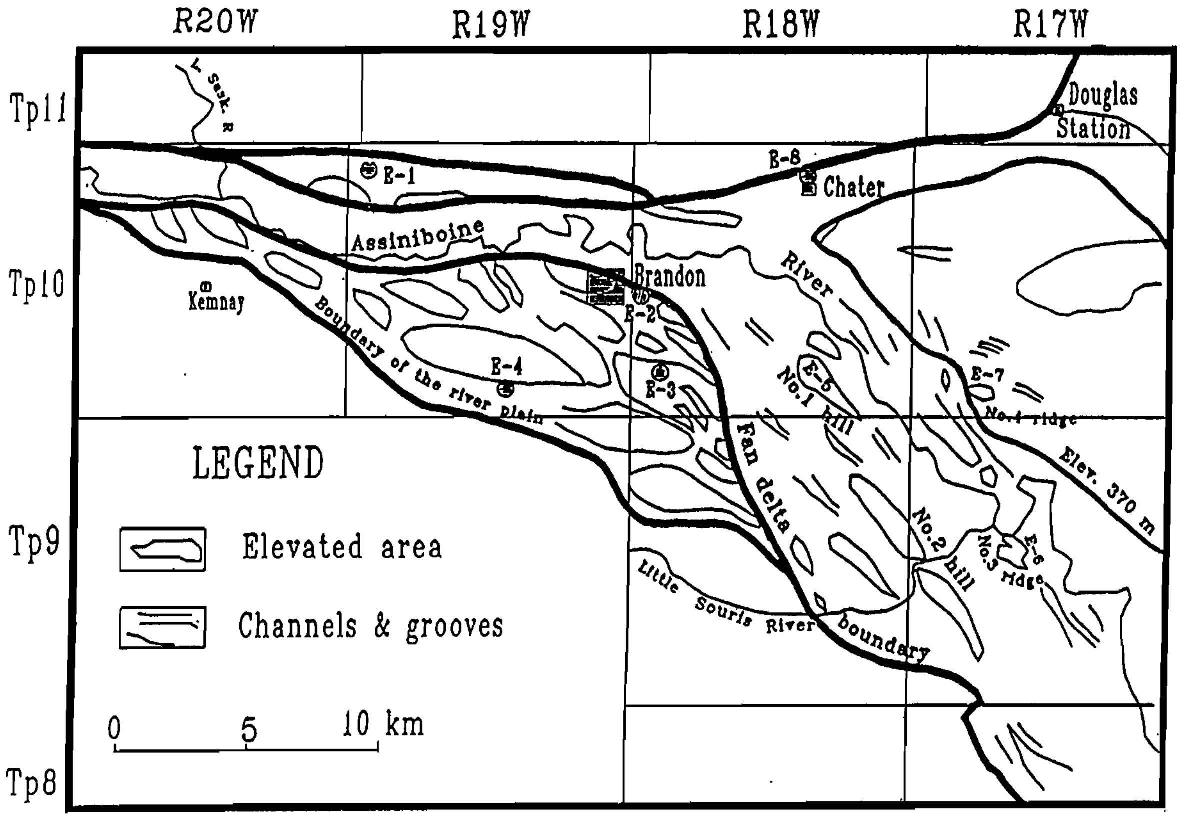

Figure 14. Geomorphology of the western Assiniboine fan-delta, showing streamlined (elliptical) erosional hills and flood grooves in subupland areas (mainly south and west of Brandon) and in the main spillway (Sun, 1993).

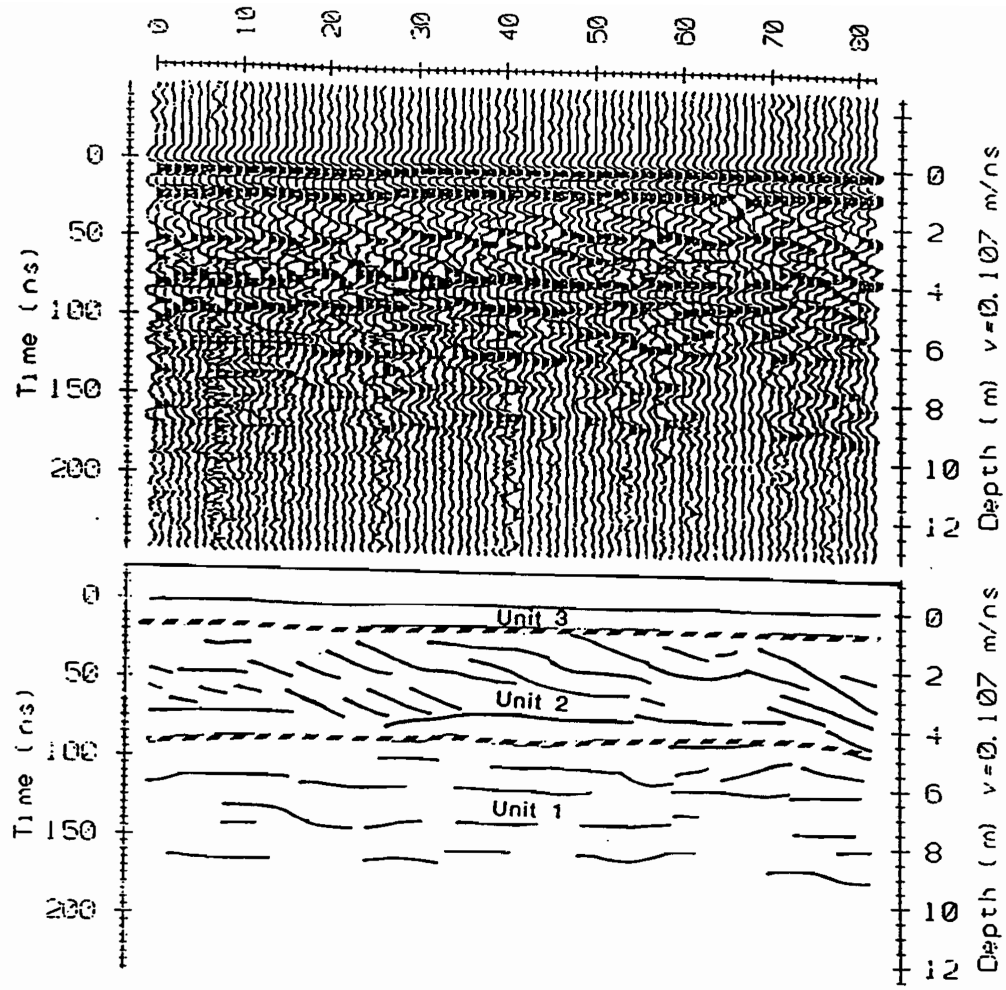

Figure 15. Ground-penetrating radar profile through residual streamlined hill (E-7 of Figure 14), showing thick foresets of Assiniboine fan-delta (Sun, 1993, after data by H. Jol & D. Smith).

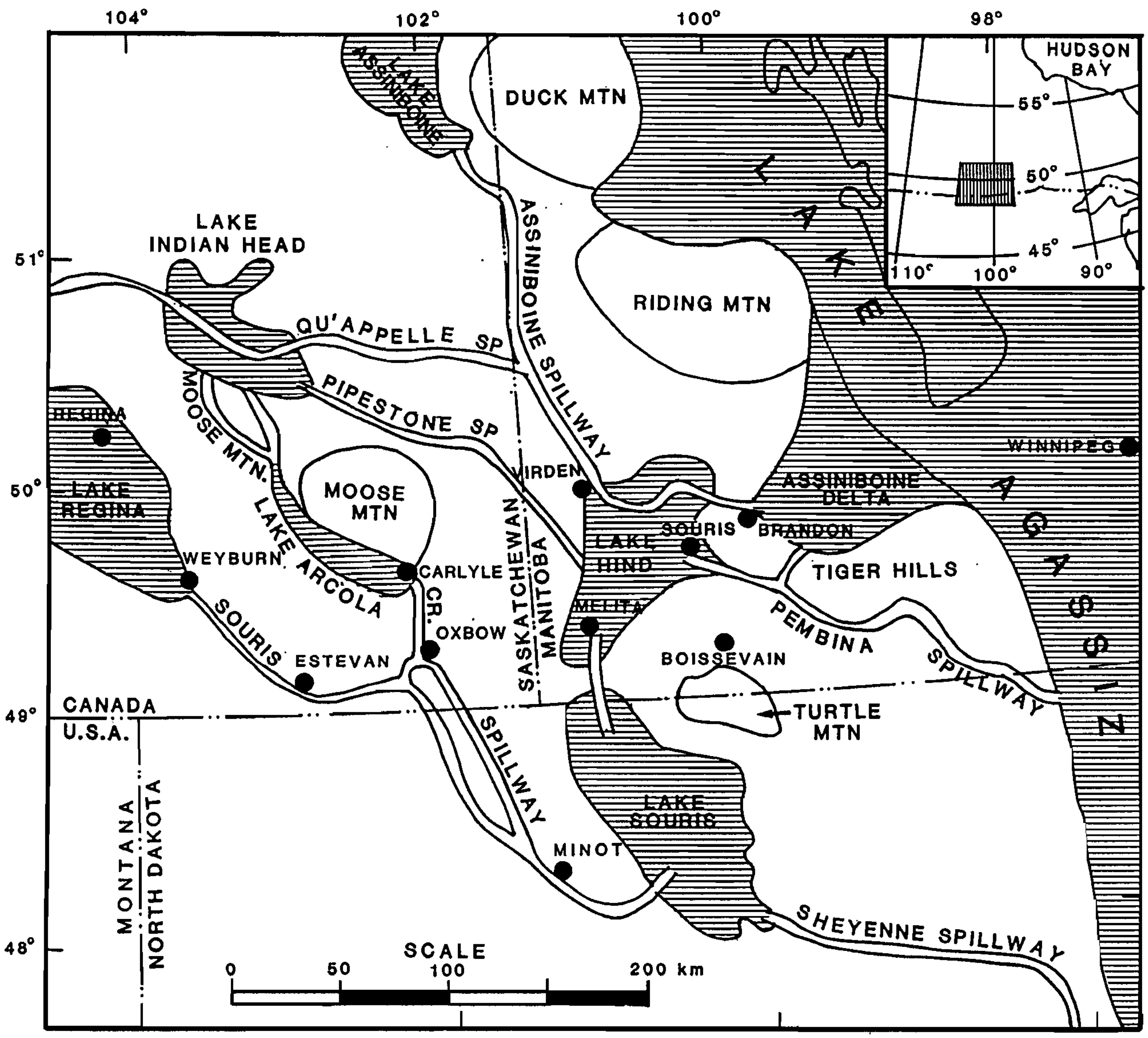

Figure 16. Glacial spillways, lake basins, and elevated areas of the region west of the central Lake Agassiz basin (Sun & Teller, in press, after Kehew & Teller, 1994).

STOP 5
STOTT BUFFALO JUMP AND ASSINIBOINE SPILLWAY
LUNCH STOP

A lunch break will be taken at the Stott Buffalo Jump stop, located within the Assiniboine spillway. The archaeological importance of the Stott site was originally recognized by the land owner, Frank Stott, in the early 1940's, upon the discovery of artifacts in freshly plowed fields. Major archaeological excavations were conducted in 1947-1952 and again in 1982, uncovering burial mounds and large pits full of bison bones.

Blackduck people have utilized the area for at least 1200 years. Buffalo were stampeded along mile-long driving lines into the spillway, where they would be seriously injured by the steep dropoff and could easily be killed en masse and butchered nearby. The paleo-Indians used virtually all of the carcass; the meat was made into jerky and pemmican in an attempt to preserve it; the bones were made into tools, ornaments, and 'bone butter' (fat produced from boiling), while the hides were made into shelter, clothing, and containers.

Most spillway channels in the Canadian Prairies are deep, steep-sided trenches that have few large tributaries. Scoured subupland valleys with streamlined erosional residuals and coarse boulder lags commonly lie along these spillways (see Kehew & Lord, 1987, in Appendix to the Guidebook). Erosion of the trench progresses from the broad subupland zone of erosion ('outer zone") to the development of longitudinal grooves, and eventually to entrenchment of one of these grooves. The size and depth of the resulting flood trench is a function of the rate and total volume of the flow. The present-day size of the many catastrophically-eroded spillways in Saskatchewan, North Dakota, and Manitoba is the result of several floods (see Kehew & Teller, 1994, in Appendix to this Guidebook); terraces within these spillways may reflect these multiple flood events.

In many places below the floor of the lower Assiniboine Spillway, including that portion between the Stott site and Brandon, coarse gravel (waning flood deposit) is present (Klassen, 1983). In the Stott-Brandon area, Klassen (1983) states that gravel is overlain by clay which may have been deposited when Lake Agassiz levels rose and drowned the lower Assiniboine Spillway; in turn, these clays are capped by Holocene alluvium.

EN ROUTE TO NEXT STOP

The road between Stott and Brandon lies within the Assiniboine Spillway. Its walls are steep-sided with occasional small "terraces" armoured by boulders visible on the north side; boulder "lines" are also occasionally visible along the valley wall, and probably represent the remnants of flood-eroded levels. The uplands along the southern margin of the valley contain residual streamlined hills, scoured zones, and boulder lags.

About 40 km east of Brandon our route turns south off the Trans Canada Highway at Carberry and again crosses the Assiniboine fan-delta, which is covered by stabilized sand dunes that roughly coincide with the very fine to medium grained fan-delta sediment from which they were derived. As noted before, these dunes have been periodically active during the Holocene (David, 1971). Near the town of Carberry, our route passes across a potato growing area, which is irrigated by the Assiniboine Delta Aquifer. To the south, the road crosses a small shallow vestige of the Epinette Creek distributary channel that we saw just east of Brandon; here, the former channel has been obscured by dune activity. Near the modern Assiniboine River valley, with its cutoff meander loops and oxbow lakes, we pass by Spruce Woods Provincial Park, which has many trails through the dunes, including the only area with active dunes, the Bald Head Hills (Figure 17).

STOP 6
SPRUCE WOODS

This stop is in Spruce Woods Provincial Park and will highlight:

1) the sandy sediments of the mid-fan which have been extensively reworked by eolian processes;
2) the well-developed meandering form of the Assiniboine River;
3) gradual incision of the Assiniboine River valley into the fan;
4) slope failures in sandy sediments along the Assiniboine River;
5) distinctive modern biota of the Spruce Woods region.

In contrast to the coarse sediments at the apex of the fan at Brandon, the mid-fan is dominated by fine to medium sand. In Spruce Woods Provincial Park these deposits have been reworked into dunes (Figure 17). Sets of parabolic dunes migrating southeastward emanate from ridges which may represent relict longitudinal dunes. The notes of surveyors who traversed this area in the 1870's, which are available in the Manitoba Archives, include descriptions of dunes that are comparable to the present form of these features.

The Assiniboine river has eroded a valley over 50 metres deep in this portion of the fan. Information regarding the timing of this downcutting has been obtained from dated raised alluvial deposits. A 5 m sequence of fossiliferous fine grained sediments will be examined at this STOP located near Highway 5, 2 km north of the Ernest Thompson Seton Bridge.

Sediment samples collected along the river were examined for macro-fossils by R.E. Vance (GSC). Several complete gastropod fragments and a *Chara* oogonium at 3 m, as well as wood fragments at 2 and 3 m, were recovered. Sediments from the site also were examined for ostracodes by C. Rodrigues (University of Windsor). Ostracodes were not observed in samples from 1.5 and 2.0 depth, but *Candona* sp. and *Cyclopris* sp. were recovered from samples taken at 2.5 and 3.0 m. In a sample from 4.0 m depth, low numbers of etched and probably reworked valves of the ostracodes *Candona obtusa* and *Ilyocypris gibba* were recovered. Delorme (1989) noted that *Ilycopris gibba* is mainly confined to moving water.

Wood from slumped sediments similar in appearance to the *in situ* sediments were dated at 2600 ± 70 (BGS-1816). Wood in overlying sand, of probable eolian origin, was dated at 2500 ± 70 (BGS-1817). Similar sediments, including fossiliferous silt and peat, outcropping on the north bank of the Assiniboine River 6 km southwest of the Ernest

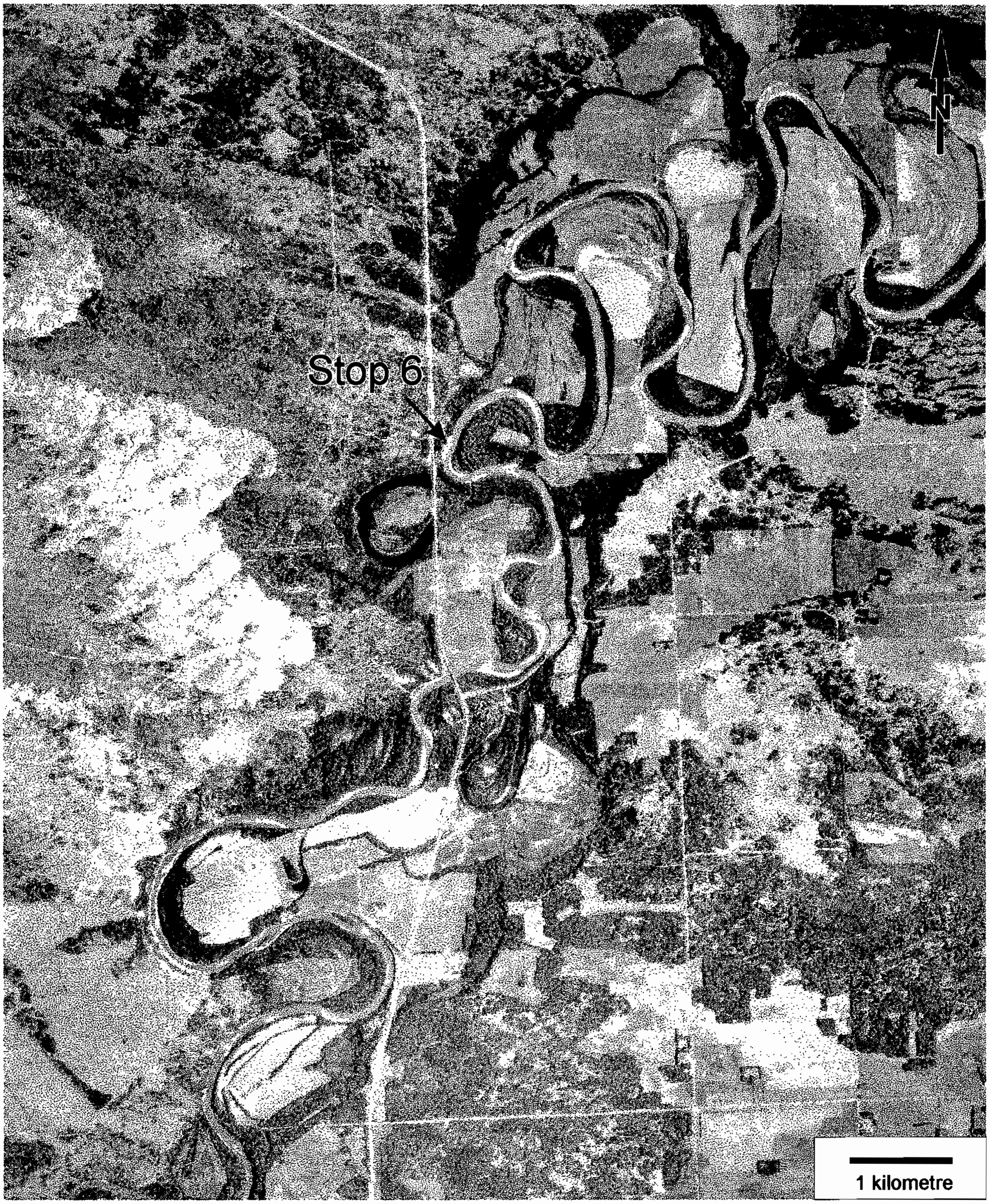

Figure 17. Aerial photo of Spruce Woods area, showing the Bald Head Hills active dunes (west of the highway) in the largely stabilized dune field; note the meanders, oxbows, and meander scrolls along the Assiniboine River, which has been entrenched below the surface of the Assiniboine fan-delta and its capping dunes.

Thompson Seton Bridge yielded a more diverse biota. Abundant insect remains were recovered by D. Schwert of North Dakota State University. Numerous *Scirpus, Carex,* and other Cyperaceae seeds, as well as *Typha, Potomageton*, and *Zannichellia* seeds were identified by R.E. Vance. Wood from this site yielded a radiocarbon age of 6400 ± 90 (BGS-1821). Wood from another similar site located 5 km northwest of the Highway 34 bridge yielded an age of 2330 ± 70 (BGS-1822).

These sediments are interpreted as alluvium, and the elevation of the top of these sediments precludes deposition by the modern river. Gradual downcutting of the river since retreat of Lake Agassiz is indicated, with incision on the order of 5 metres in the late Holocene. Slow incision probably relates to the fact that much of the river bed through the fan rests on till, as well as Cretaceous bedrock north of Holland.

EN ROUTE TO NEXT STOP

Our route takes us to Glenboro, where we turn east on Highway 2 along the south side of the Assiniboine fan-delta. To the south we can see the more elevated and irregular topography of the drift-mantled shale bedrock. Many hills in this region are composed of shale; some of these hills as well as many in North Dakota and Saskatchewan, consist of ice-thrust masses of shale bedrock, and Bluemle & Clayton (1984) show that there commonly is a socket (depression) today up-flow from the hill that reflects where the shale (and drift) mass came from (Figure 18).

After we turn south on Highway 34 at Holland, we pass by a large hill (>75 m high, >5 km^2) named Verdigen Hill that probably is composed of one or more large masses of glacier-thrust shale and glacial drift. Our route passes along the south side of this hill, where there is an exposure of displaced shale, dipping much more steeply than in-place bedrock does.

These uplands (the eastern Tiger Hills) are characterized by disordered terrain consisting of irregularly distributed hills and depressions. However, many of the hills have relief in excess of that expected in terrains typical of glacial stagnation. Although natural exposures are few, road cuts, borrow pits, and water well drill logs indicate surficial geology consistent with the ice thrusting of bedrock, followed by stagnation. Road cuts on the flanks of large hills have exposed folded or tilted Cretaceous Riding Mountain Formation (Odanah Member), suggesting that such hills are partly composed of glacially-thrust bedrock blocks (e.g. Verdigen Hill, 4 km east of Bruxelles). Water well drill logs also suggest the repetition of bedrock and till to depths of several tens of metres. The borrow pit at Bruxelles (STOP 7) provides a more detailed look at these conditions.

The conditions for thrusting of bedrock by ice have been well documented by Aber *et al.* (1989) and others. These authors have interpreted ice thrusting of older stratified drift accompanying the last advance into Manitoba to account for the origin and geology of the Brandon Hills to the northwest. It appears that similar thrusting, in this case involving Cretaceous shale, may have occurred throughout the Tiger Hills. The Tiger Hills thrusting may have been promoted by the presence of the Manitoba Escarpment, in a manner similar to that shown for thrusting in the Turtle Mountains and along the Missouri Escarpment in North Dakota (Bluemle and Clayton, 1984) (Figure 18).

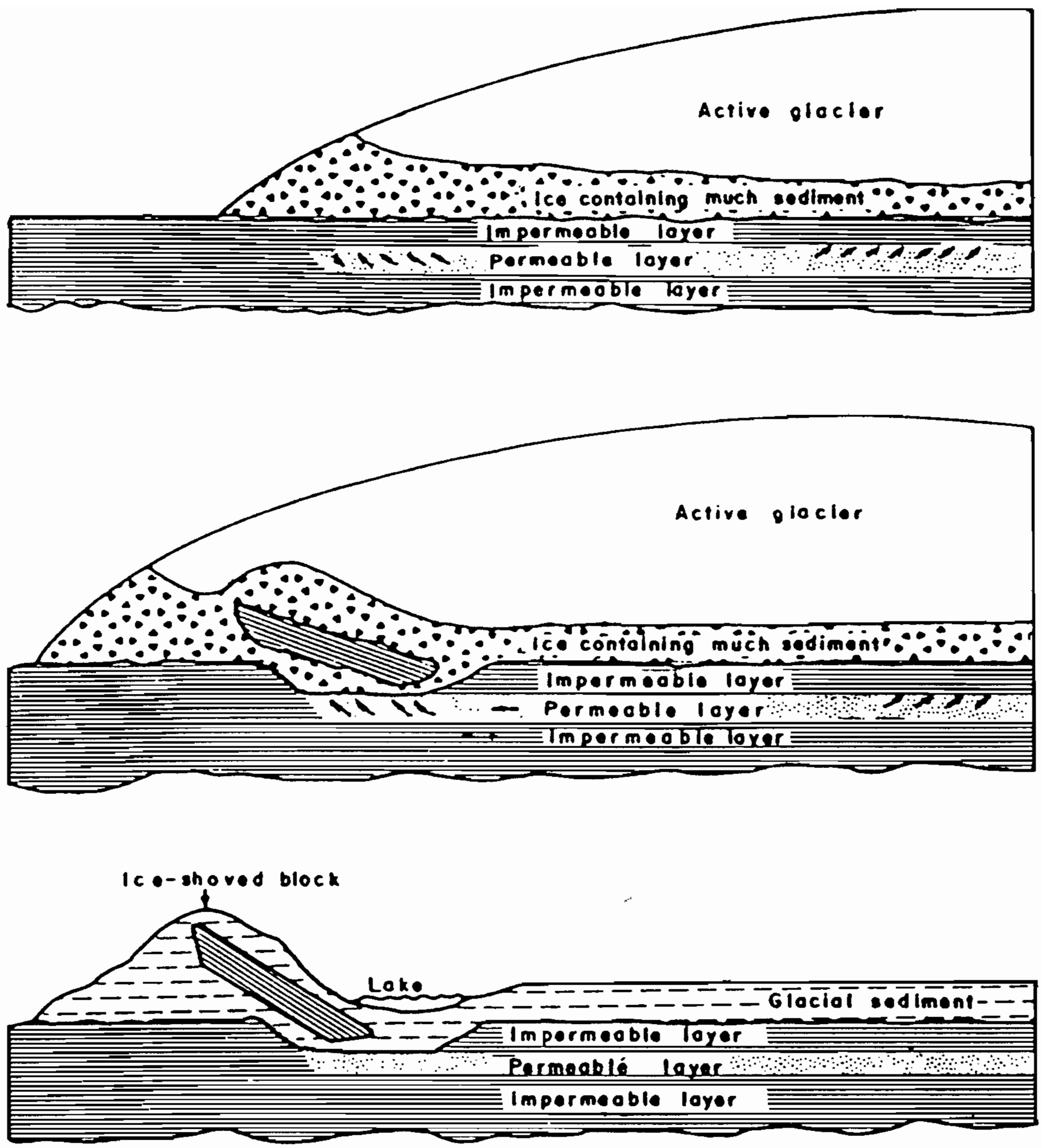

Figure 18. Schematic three-part diagram showing how ice thrusts may form. The top sketch shows how water under cryostatic pressure may move outward and force overlying impermeable and weak strata such as shale upward into the active flow path of the glacier (middle sketch). This ice thrust bedrock and associated drift may be left as a hill, commonly with a lake-filled depression upflow (Bluemle, 1992); Bluemle and Clayton (1984) refer to these as hill-depression forms.

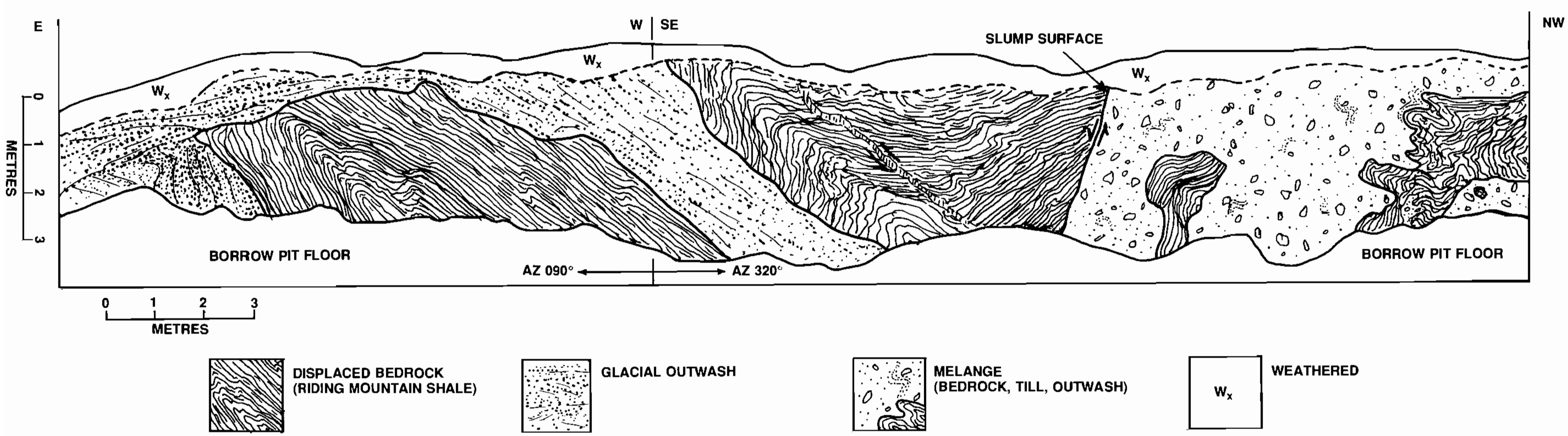

Figure 19. Sketch of surficial geology exposed in the borrow pit behind Dan's Service Station, Bruxelles, Manitoba (W. Brisbin, unpublished). Note the change in orientation of the pitface in the section at W/SE.

STOP 7
BRUXELLES GLACIOTECTONIC CUT

This stop is located in the borrow pit behind Dan's Service station in Bruxelles, Manitoba. A portion of the main excavation face in the pit, as depicted in Figure 19, is the subject of this stop. Water wells within the town limits suggest that horizontal Riding Mountain shales (Cretaceous) are at a minimum depth of 15 m. The cut reveals evidence of the complex combination of processes involved in the development of Tiger Hills surficial geology.

The pit face comprises two differing geologic domains separated by a younger displacement (slump) surface. The easterly domain consists of two large shale bedrock slabs enveloped in poorly stratified drift; the northwesterly domain is a melange consisting of a chaotic mixture of till, outwash, and bedrock fragments.

The large shale enclaves of the easterly domain are interpreted as glacially thrust blocks from the underlying Riding Mountain Formation. Bedding in the blocks has been folded flexurally and reveals the presence of small incipient thrust faults. These structures testify to some history of horizontal compression before the blocks became detached and incorporated within the ice. The present margins of the blocks are fracture surfaces indicating that thrust slabs, already containing complex folds, were probably much larger and were broken up during glacial transport. The material in which the shale blocks rest is stratified and cross bedded. This domain, consisting of blocks and stratified drift, can best be accounted for best by outwash envelopment of let-down slabs of displaced shale.

The melange at the northwest end of the pit face (to the right of the slump surface, Figure 19) is characterized by the mixing of three components, namely till, stratified drift, and shale bedrock. The mixing process is manifest in the overall jumbled appearance of the assemblage, in the contorted and discontinuous stratification of the outwash, and by the convoluted margins of shale blocks which still retain folds inherited from the ice thrusting. The origin of the melange is problematic; mixing by glaciotectonic shearing with elevated fluid pressure could be responsible; alternatively landsliding of a saturated mass could yield a similar product.

The displacement surface separating the two domains has been interpreted as a late slump structure, along which the hanging wall has moved down. If this interpretation is correct, the displacement surface is probably listric and the slump has resulted in a clockwise rotation of the hanging wall block. Such a scenario implies some surface relief with unstable slopes, and infers that the melange unit underlies the assemblage of outwash and shale blocks.

STOP 8
PEMBINA SPILLWAY VIEW

We approach the Pembina River Spillway from the north, along Highway 34, near Swan Lake. The spillway connects the Lake Souris and Hind basins to Lake Agassiz (Figure 16), where the Pembina fan was deposited. Elson (1955) describes a complex trench which increases incrementally in size downstream, from 20-35 m deep by less than 3 km wide above the Souris River trench, to 135 m deep by 6.5 km wide near Lake Agassiz. The Pembina Spillway is clearly shown on the separate Brandon 1:250,000 "Quaternary Geology Map".

Portions of the spillway were initiated as meltwater steams drained the ice margin to the north and east into an early phase of Lake Agassiz (Elson, 1955). An advance of ice around 11.4 ka from the north within the Lake Agassiz basin, which formed the Edinburg moraine (Fenton *et al.*, 1983), partially infilled the channel.

Although the valley may have initially been eroded by meltwater from earlier glaciations, as is possible for other large bedrock valleys in the Prairies, the record known in the Pembina Spillway has been related to the last deglaciation. Initially, this spillway served as an overflow route from proto Lake Hind and for runoff from the margins of the Souris and Red River ice lobes in Manitoba (Sun & Teller, in press). As ice retreated, meltwater from a large part of the southern Prairies, as far west as the Rocky Mountains, was diverted from its original route across North Dakota (which was through the James and, later, Sheyenne spillways) into the Pembina Spillway, via Lakes Souris and Hind (see Figures 5-8 *in* Kehew & Teller, 1994, in Appendix to this Guidebook). Catastrophic flooding from glacial Lake Regina before 11 ka contributed to the erosion of the Pembina Spillway (Kehew & Teller, 1994), as did an earlier flood from glacial Lake Arcola and glacial Lake Moose Mountain (Sun & Teller, in press). Terraces formed in response to successively lower levels of Lake Agassiz (Elson, 1955). Today a series of shallow lakes, including Swan Lake, have been dammed behind Holocene alluvial fans on the floor of the spillway.

DAY 3

DELTA MARSH FIELD STATION TO LATIMER TO WINNIPEG

Three Stops

EN ROUTE TO NEXT STOP

The trip from Delta Marsh to Portage la Prairie again crosses the complex of old Assiniboine River alluvium and abandoned distributary channels (Figures 4-6). West of Portage la Prairie, our route turns south onto Highway 242. Near the town of Rossendale, we rise up over the Campbell strandline (Figure 20), eroded by Lake Agassiz waves into the distal silts and fine sands of the Assiniboine fan-delta; these sediments were examined at STOP 3 and will again be seen at STOP 9. The Assiniboine River, which has been incised into the fan-delta, changes to an aggrading river downstream from the Campbell strandline. The Rossendale area has been important for Lake Agassiz studies since Elson (1955) discovered buried organic material at the Rossendale gully and a fossiliferous valley fill on terraces in the Assiniboine valley to the south (Figure 20). As we approach the Assiniboine River valley, multiple topographic levels may be observed, including the modern floodplain, the Campbell-level fill terrace, the sharply-defined break between the terrace and the surface of the Assiniboine fan-delta, and the still-higher glacial landscape of the Tiger Hills in the distance.

STOP 9
DISTAL FAN SEDIMENTS

As a prelude to examining the fossiliferous sediments of the valley fill terrace, a brief stop will be made at the crest of the slope which rises from the terrace. In a small pit, the nonfossiliferous silty lower-fan sediments of the Assiniboine fan-delta may be seen. The presence of metres of undisturbed horizontal and rippled lamination, as well as lack of fossils or bioturbation, is compatible with rapid sedimentation in the fan. The section exposes the following sediments:

0.0 - 0.4 m	Ap horizon, silt
0.4 - 0.8 m	Silt, massive, blocky
0.8 - 1.4 m	Silt, laminated and rippled, rodent burrows
1.4 - 2.4 m	Silt, fining upward, faintly defined strata diminishing in thickness from 5 cm upward to 1 cm, isolated ripples
2.4 - 3.4 m	Silty fine sand, rippled

STOP 10
VALLEY FILL SEDIMENTS (LATIMER SITE)

Attention to the Rossendale area was first drawn by Elson (1955). Whereas Upham (1895) did not recognize transgressions of Lake Agassiz, and Johnston (1916) only cited northwestern Ontario stratigraphy to support his belief in a major lake level rise, Elson added a new dimension to Lake Agassiz studies with the recognition of fossiliferous valley

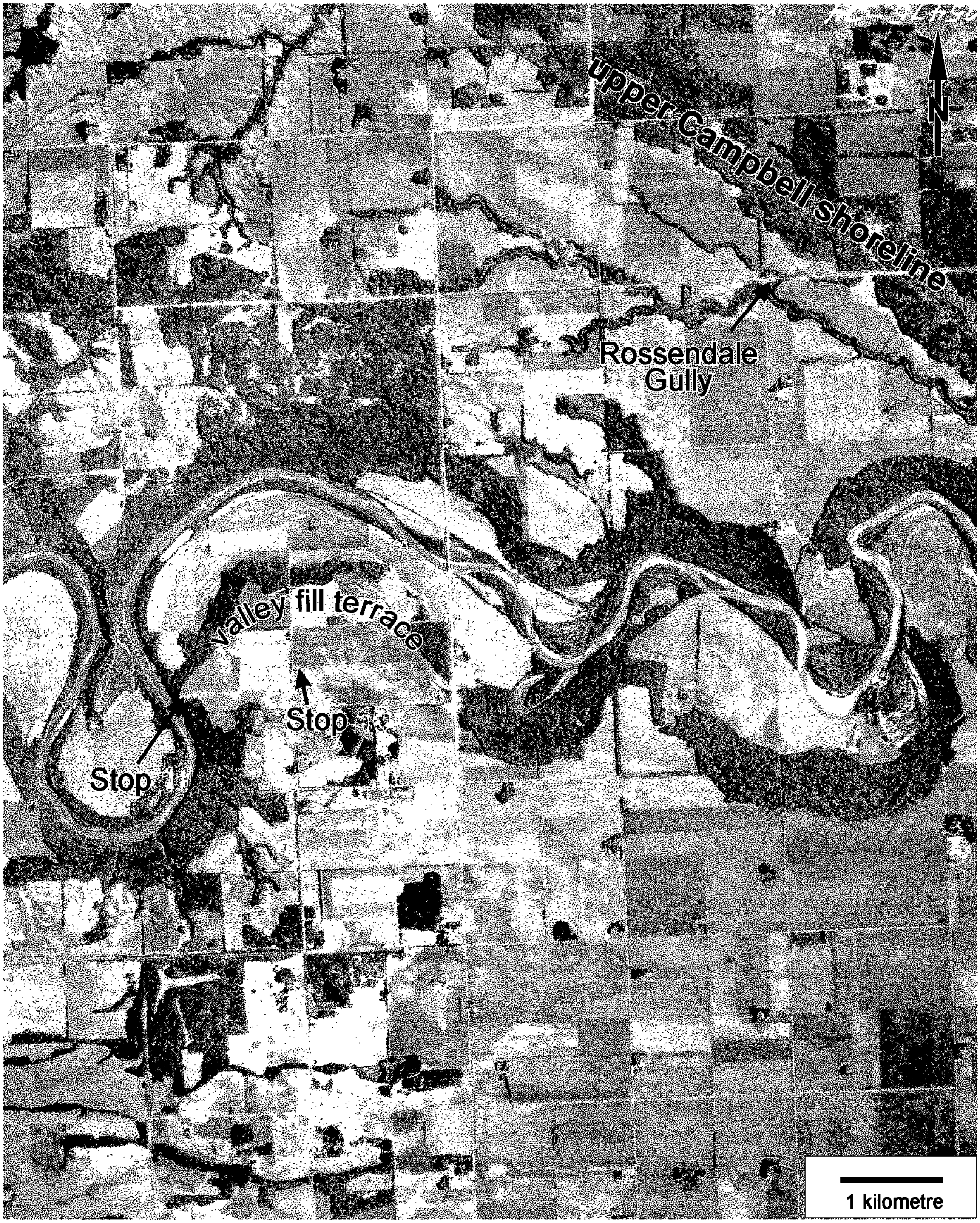

Figure 20. *Aerial photo of valley fill terrace along Assiniboine River, showing location of STOPS 9 and 10, the upper Campbell beach, and the Rossendale gully eroded into paleolagoon sediments dated at 9.5 - 9.6 ka by Teller (1989).*

fills. Best developed are a fill at the Norcross level in the Pembina Spillway, and the Campbell-level fill of the Assiniboine at a lower level. A less clear Norcross-level fill in the Assiniboine valley was confirmed by Elson (1967). Comparable terraces were reported by Brophy (1967) for the Sheyenne valley of North Dakota, another major spillway entering the Lake Agassiz basin from the west.

Elson (1955, 1967) described the Campbell-level fill as a set of paired terraces, rising from 1050' (320 m) at the upper Campbell shoreline to about 1080' (330 m) north of Glenboro, in the Spruce Woods area. The fill sediments were described as fossiliferous sand, silt, and clay which undergoes a transition from a fluvial deposit north of Glenboro to an estuarine-like facies south of Rossendale. Fossiliferous sediments in a flat-bottomed gully near Rossendale were regarded as also having been deposited during a rise to the Campbell shoreline. Additional data, including ostracode and mollusc identifications from the terrace south of Rossendale, were presented by Klassen and Elson (1972). Klassen (1972, 1975, 1983) also contributed additional data regarding the terraces in the Assiniboine valley near Rossendale, referred to by Klassen as the Campbell terrace.

Klassen (1983) summarized radiocarbon dates obtained from the Rossendale area fossiliferous deposits. These dates ranged from 9.7 to 12.4 ka, and were used to support what has been referred to as the 'old chronology' for Des Moines lobe deglaciation. The Rossendale site was reexamined by Teller (1989), who demonstrated that the dates of 12,400 ± 420 yr B.P. (Y-165) and 12,100 ± 160 yr B.P. (GSC-1319) should be rejected due to the tendency of the dated subaquatic moss species, *Scorpidium scorpioides*, to accumulate dissolved old carbon. New dates of 9600 ± 70 yr B.P. (TO-534) and 9510 ± 90 yr B.P. (GSC-4490) were reported for wood from a similar depth at the same site. Paleontological data were cited by Teller (1989) to conclude that the Rossendale deposit was separated from a large lake, hence the gully was infilled as part of a protected lagoonal environment.

A fundamental question regarding the Rossendale area fossiliferous deposits is whether a rise in lake level is indeed required, or whether the fossiliferous valley fill terraces could be explained by a prolonged stable lake level.

If it accepted that the deposits are proof of a major rise in lake level, two scenarios may be considered:

1) the sediments were deposited when Lake Agassiz rose to a Campbell level at the beginning of the Emerson phase, and the dates presented by Teller (1989) represent prolonged occupation of the Campbell level; or
2) the deposits are the result of a rise in lake level that took place in mid- to late-Emerson phase time, perhaps due to the closure of a temporarily-opened eastern outlet. This event could then be called upon to explain the existence of the lower Campbell shoreline, due to differential uplift during the regression. The terraces are, however, 6 to 8 metres higher than the lower Campbell shore.

At this stop, we will examine a gully cut into the upstream margin of the Campbell-

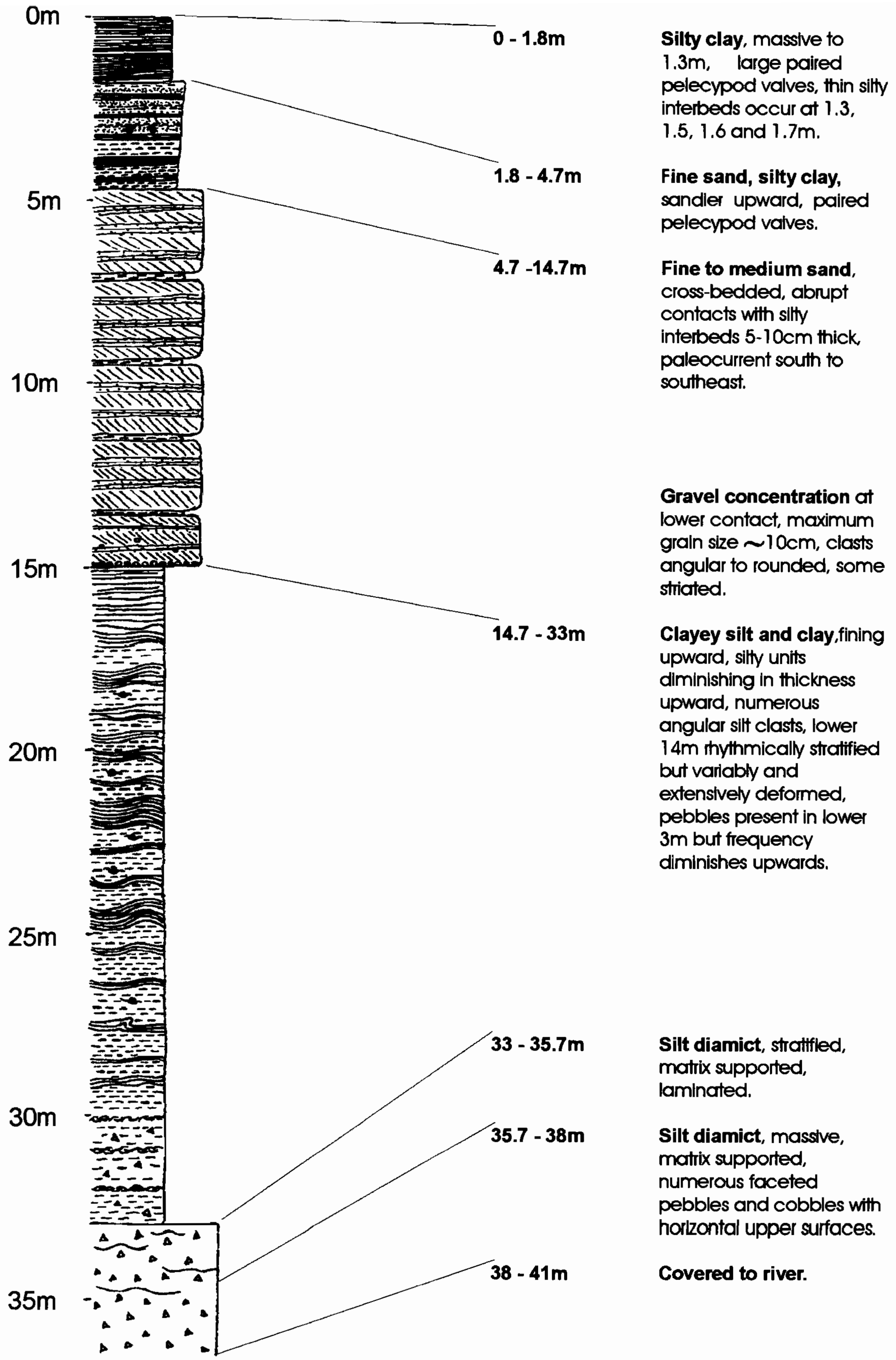

Figure 21. Stratigraphic sequence through the Rossendale valley-fill terrace at the Latimer site.

level terrace in the Assiniboine valley. The site was first described by Elson (1955). This exposure of 38 m of sediments was subdivided from bottom to top into six major units (Figure 21) on the basis of 1995 field work:

1. **Massive diamict:** The lowest exposed unit in the gully is a massive, matrix supported diamict. An origin as basal till is implied by the massive structure of the deposit, as well as by the presence of faceted and striated clasts.

   Analysis of one sample at the GSC indicates that the <2 mm matrix of the till consists of 49% sand, 45% silt, and 7% clay (<2 micron). The <63 micron fraction includes 8% calcite and 39% dolomite, or a total of 47% carbonate. The 8-16 mm fraction consists of 71% carbonate, 6% shale, 21% igneous and metamorphic clasts, and 2% other. The 4-8 mm fraction consists of 70% carbonate, 5% shale, 23% igneous and metamorphic clasts, and 2% other. These analyses imply that this unit likely is derived from the northeast, and may have been deposited by a late-glacial advance from the Interlake, perhaps the advance to the Darlingford and Edinburg Moraines which occurred prior to about 11 ka.

2. **Stratified diamict:** Above a sharp contact with the massive diamict is stratified diamict from 33 to 35.7 m (Figure 21). This deposit may be basal till comparable to subglacial meltout till, but deposition in a proximal glaciolacustrine environment by slumping and rainout seems equally, or perhaps more, likely.

3. **Deformed clay and silt rhythmites:** A unit over 18 m in thickness, from 14,7 to 33 m, consists of clay and silt. The sequence fines upward, with silty units diminishing in thickness upward. Extensive deformation and brecciated silt may be observed. Pebbles are present, with greatest abundance low in the sequence.

   The stratigraphic position of these sediments (*viz.* immediately overlying basal and/or proximal glacial sediments) and the clear indication of rhythmic sedimentation imply that these sediments were derived from the ice margin and were deposited in a glaciolacustrine environment. Deformation may be related to iceberg scour, and pebbles probably are dropstones.

4. **Cross-bedded sand:** A 10-m-thick, non-fossiliferous, cross-bedded sand unit overlies an abrupt contact with the clay and underlies fossiliferous sand. Silty interbeds 5 to 10 cm in thickness are present. The down-dip direction of the crossbeds is to the south and southeast. About five cycles of sedimentation may be observed in this unit. At the lower contact of this unit is a concentration of angular to rounded gravel clasts up to 10 cm in size.

   Three possible origins for this unit were considered: (1) subaqueous outwash derived from the ice margin; (2) turbidity fan sediments of the Assiniboine fan-delta, deposited by catastrophic inflow from the Assiniboine Spillway; and (3) fluvial sediments deposited by the Assiniboine River as part of the valley fill sequence.

   Derivation of this relatively coarse unit from the ice margin seems unlikely, due to the position of the sediments above distal deposits. A resurgence of glaciofluvial sedimentation could be called upon, but an event of this nature has not been reported from other sites in the region and no coarsening upward was observed.

Deposition of this deposit by the Assiniboine River seems unlikely due to its thickness (exceeding typical fluvial deposits in the area), structural and textural simplicity, and lack of bioturbation or fossils.

Rapid deposition from traction currents associated with turbid inflow from the Assiniboine Spillway seems the most likely origin, hence the deposit is interpreted as part of the Assiniboine fan-delta complex, deposited prior to 11 ka during the Lockhart Phase of Lake Agassiz. The lag at the lower contact indicates that some erosion took place, unless the clasts were ice-rafted. Perhaps the clasts are dropstones concentrated from the underlying clay.

5. **Fossiliferous sand:** Near the top of the sequence close to the surface of this terrace is the fossiliferous valley fill discussed by Elson (1955). The lower portion of this fossiliferous sequence, from 1.8 to 4.7 m consists of fossiliferous silty sand. Richly fossiliferous sediments from 3 m in depth were examined by E. Pip (University of Winnipeg). The following molluscan taxa were observed:

    Valvata tricarinata (Say, 1817)
    Probythinella lacustris (Baker, 1928)
    Amnicola limosa (Say, 1817)
    Lymnaea stagnalis (Linne, 1758)
    Stagnicola palustris (Muller, 1774)
    Fossaria exigua (Lea, 1841)
    F. parva (Lea, 1841)
    Physa gyrina (Say, 1821)
    Helisoma anceps (Manke, 1830)
    Gyraulus parvus (Say, 1817)
    G. circumstriatus (Tryon, 1866)
    Anodonta grandis (Say, 1829)
    Lampsilis radiata siliquoidea (Barnes, 1823)
    Sphaerium transversum (Say, 1817)
    Pisidium spp.

    The following macrophyte taxa also were observed:

    Potomageton sp.
    Utricularia vulgaris

    Sediments from this unit also were examined for ostracodes by C.G. Rodrigues (University of Windsor). From 3 m, *Candona obtusa*, *Cyclopris* sp., *Ilyocypris gibba,* and *Limnocythere* sp. were recovered. This ostracode assemblage indicates a relatively shallow water environment. As was noted at STOP 6, *Ilyocypris gibba* is mainly confined to shallow, moving water (Delorme, 1989). Thus, assemblages containing *Ilyocypris* *gibba*, provided there is no reworking, indicate a stream, spring, or margin of a lake. This tremendously diverse biota indicates a shallow, eutrophic, lacustrine environment with high productivity and an abundance of organic matter. Shells from this unit were dated at 11,625 ± 130 (BGS-1819). This date is regarded as being much too old, on the basis of correlation of the terrace to deposits of the Rossendale gully, dated at about 9.5 ka by Teller (1989). Furthermore, a wood date of 9.7 ka was obtained from the terrace by Klassen (1983). The fossiliferous sediments therefore are assigned to the Emerson phase of Lake Agassiz, so an unconformity and a gap of many centuries is associated with the lower contact of this unit.

6. **Fossiliferous clay:** Capping the sequence is a silty clay unit from the terrace surface to 1.8 m depth. Silty interbeds are present, and fossiliferous portions of the unit include large paired pelecypod valves. Sediments from this unit also were examined for ostracodes by C.G. Rodrigues (University of Windsor). A sample from 1 m depth contained *Cytherissa* sp. And *Ilyocypris gibba*. Hence deposition in shallow, moving water can again be inferred on the basis of the ostracode fauna.

In summary, subglacial and proglacial sediments at the Latimer site are overlain by underflow sediments of the Assiniboine fan-delta. Capping the sequence, above an unconformity that represents a long gap in time, are fossiliferous sediments containing a fauna indicative of shallow, flowing water in a biologically-productive environment. The valley fill deposits are correlated to the Rossendale gully deposits, dated at 9.5 ka by Teller (1989), on the basis of association with the upper Campbell shoreline. The deposits imply that erosion of the valley was followed by aggradation in response to an incursion by the lake, as envisaged by Elson (1955), or at least prolonged occupation of the upper Campbell level. The similarity of the elevation of the terraces, about 320 m, to that of the nearby upper Campbell shoreline implies contemporaneity.

STOP 11
DEEP BASIN LANDFORMS OF LAKE AGASSIZ

Approaching Winnipeg from the west, along Highway 3 about 5 km past Brunkild, the flat clay plain begins to take on a gently rolling topography. This ridge and swale topography (linear clay ridges) is very subtle, but is noticeable using the highway and oncoming traffic as a reference. Vehicles will essentially disappear behind the upcoming ridge and reappear on its crest.

Linear clay ridges are one of two prominent landforms found on the former floor of glacial Lake Agassiz, the other being iceberg scours. Both features are difficult to see at ground level, but are ubiquitous from the air. Both features have a long history of being misunderstood.

The linear clay ridges parallel regional ice flow and are up to 3 m high with a spacing of 1-3 km; they are found in areas of thick (commonly > 10 m) clay deposits (Figure 22). Three hypotheses have been put forward as to their mode of formation: 1) the clay surface is mimicking the underlying fluted till surface; 2) the clay surface is mimicking the underlying fluted till surface because of differential compaction due to dewatering of the lake bottom sediments after the draining of Lake Agassiz; and 3) the clay surface itself was fluted by glacial ice advancing into glacial Lake Agassiz.

In the fall of 1995, a series of auger holes were drilled into two of these clay ridges and the adjacent troughs. As can be seen in the cross section (Figure 23), the clay surface is mimicking the underlying till surface. Water content in the sediments ranges from 8% in the compact basal till, to 17% in the soft glaciolacustine diamict, to 37% at the base of the clay, and to 24% in the near surface clay. Water content in the top 10 m of lake bottom sediments in Lake Winnipeg today range from 50% to 75%. Mathematically, re-watering the clay in the clay ridges would add approximately 3.4 m to the clay thickness in the trough and only 1.8 m above the till ridge. This would effectively flatten the clay surface, which is a more likely lake bottom configuration.

The glaciolacustrine diamict at the base of the lacustrine clay (Figure 24), generally referred to by engineers as 'putty till', is a common feature in the stratigraphy of the Winnipeg region. Genetically it can be either iceberg turbate, lacustrine clays with an abundance of ice-rafted detritus, or ice-proximal gravity flow sediment.

Another prominent landform of the clay plain in the Winnipeg region are iceberg scour marks (Figure 1). The iceberg scour marks are up to 150 m wide and can be in excess of 10 km long. Figure I shows a group of scours which are as large as any in Manitoba, but they are also found in clusters which are significantly narrower and shorter. As a rule, iceberg scours which are located in depositional basins are positive silty features (ridges) (Woodworth-Lynas & Guigne, 1990), while those found on the till plain adjacent to the depositional basin are grooves (Figure 25). Where they appear as ridges today, there is still a groove in the clay below that ridge, which was infilled and partially buried by silt.

The iceberg-scour ridges have long puzzled geologists. Hypotheses for the formation of these features have varied from permafrost action to the mimicking of the underlying bedrock fracture pattern (see Mollard, 1983). Clayton *et al.* (1965) was the first to suggest that the marks were the result of ice impinging on the lake floor, with lake ice as the suggested vehicle for pushing up the ridges. Dionne (1977) reported similar features from Lake Ojibway in Quebec and confidently attributed them to icebergs. Nielsen and Matile (1982) stated that by infilling the iceberg scours with silt from density underflows on the floor of Lake Agassiz, subsequent differential compaction due to dewatering after the draining of Lake Agassiz would cause the clay to significantly decrease in volume, while the silt essentially maintained its original volume. In this way, silt-infilled scours become ridges.

Burial of the scours in the Winnipeg region by the silt has had the effect of preserving the ice-scoured clay morphology, internal structures, and the surface outline of the scours. In 1987 and 1988, studies by the Centre for Cold Ocean Resources Engineering (Woodworth-Lynas & Guigne, 1990) delineated a series of low-angle thrust faults and high-angle normal faults, which resulted in the creation of a detailed model for the construction of these infilled grooves by the keel of icebergs impinging on the floor of Lake Agassiz (Figure 26). This model has since been invaluable in the design of ocean bottom pipelines.

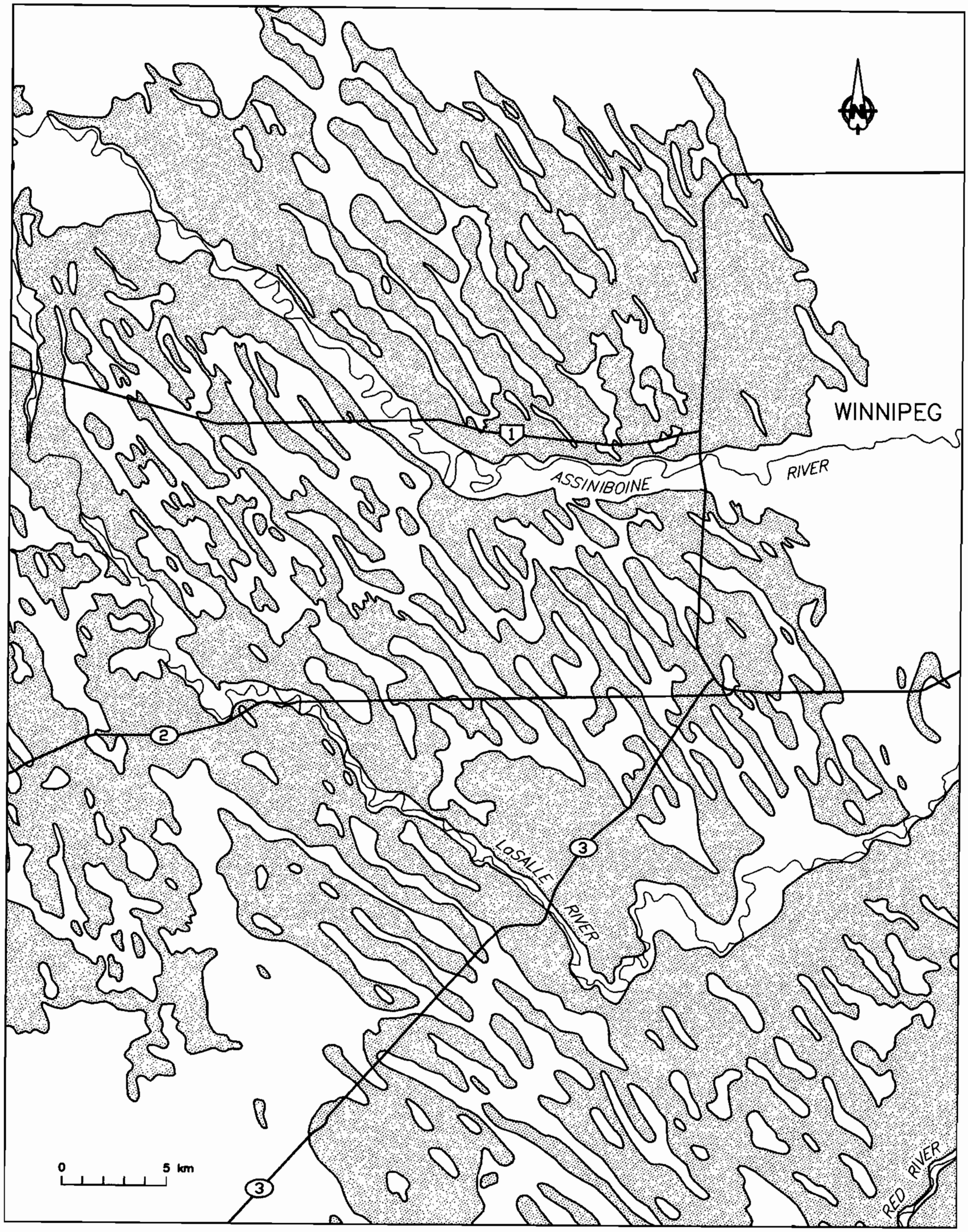

Figure 22. Linear clay ridges west of Winnipeg. Shaded areas are well-drained Red River Clay soils, whereas the intervening areas are generally more poorly drained soil types in slightly lower areas (based on Ehrlich et al., 1953, map). Note that both the Assiniboine and La Salle Rivers have in part been "captured" by one of these troughs.

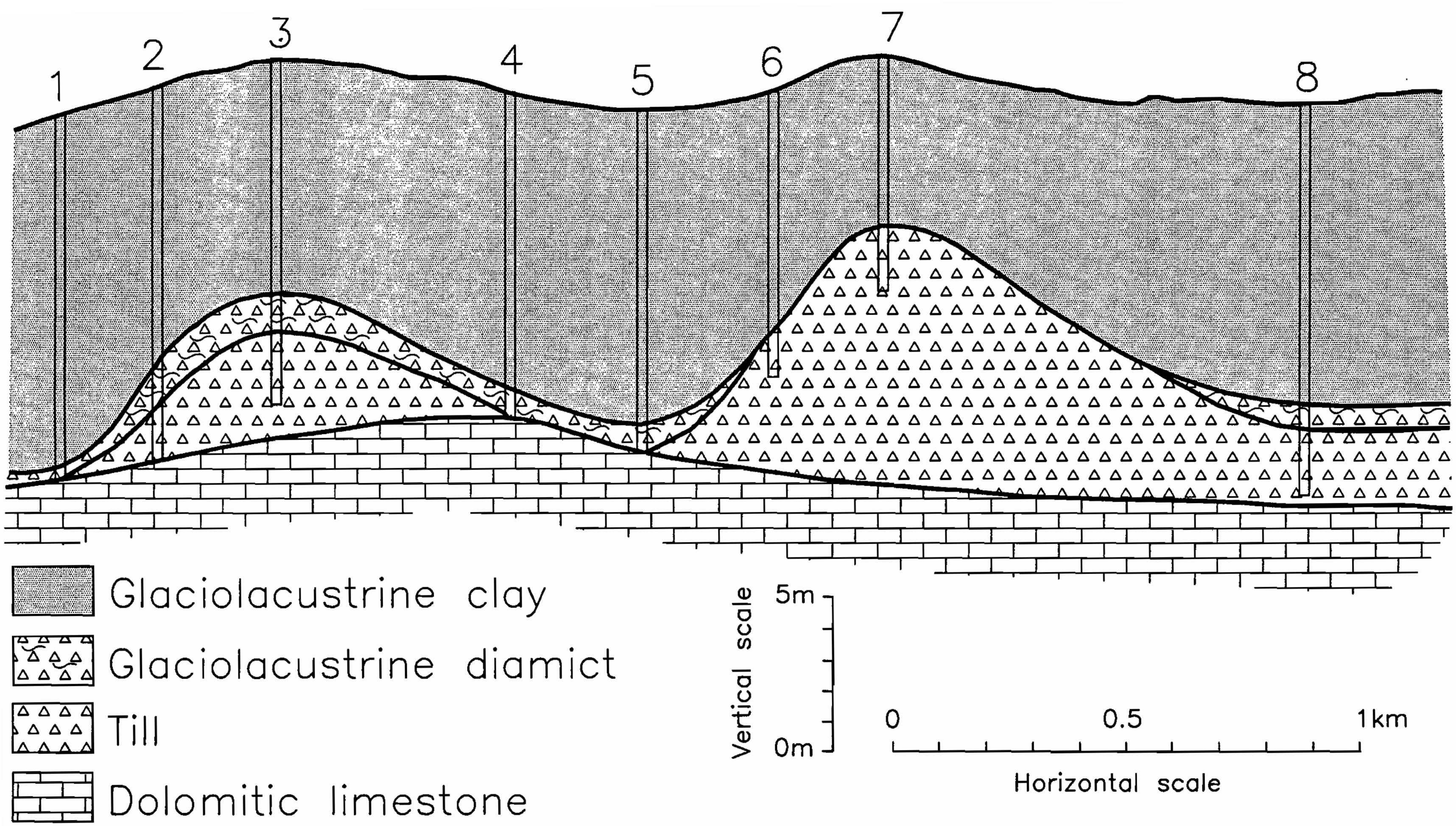

Figure 23. Cross section across two of the parallel linear clay ridges along Highway 3 near Oak Bluff, west of Winnipeg, showing the underlying till surface which they mimick.

Figure 24. Stratified diamict northeast of Winnipeg, interpreted as an ice-proximal sediment gravity flow on the floor of Lake Agassiz.

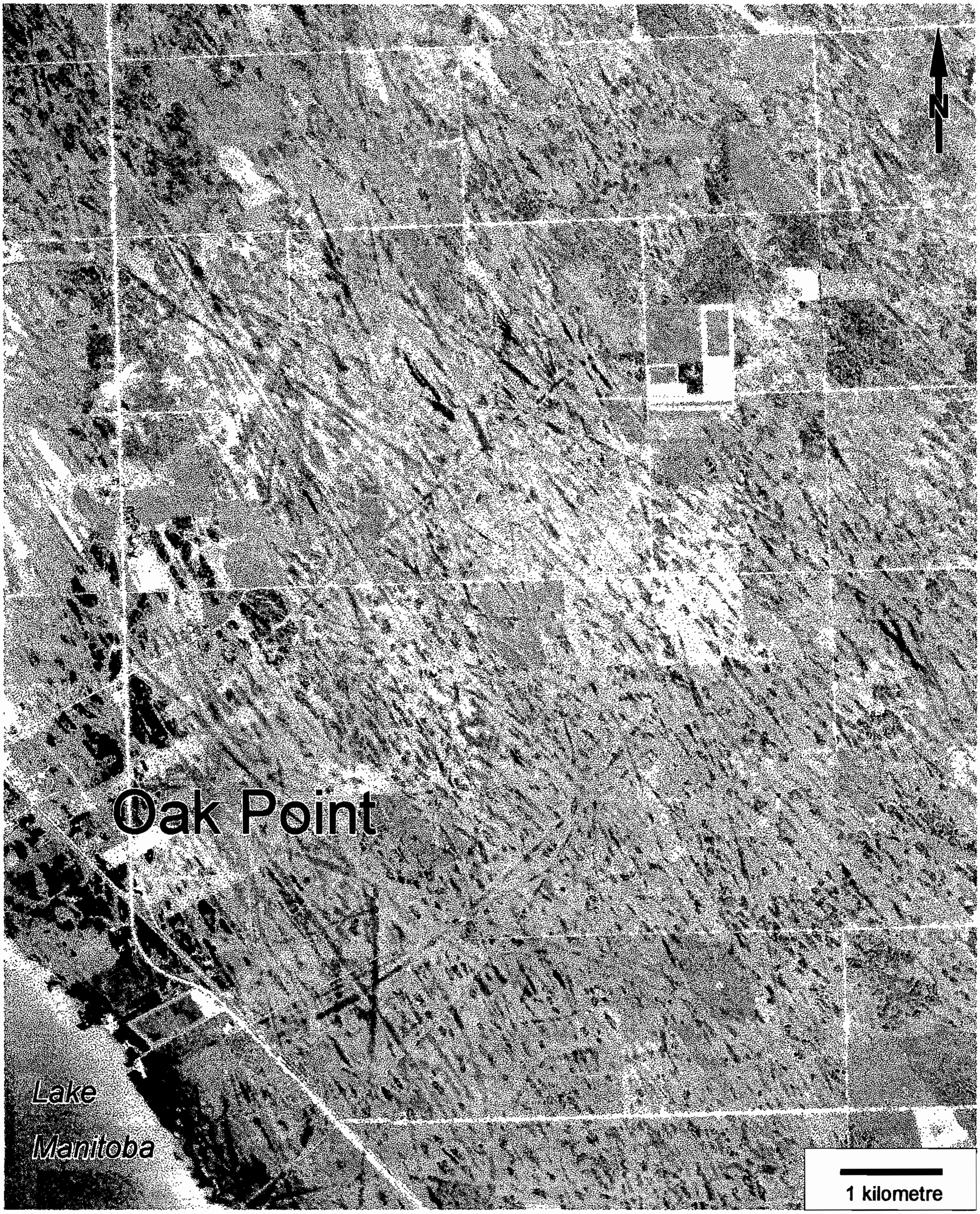

Figure 25. Aerial photo of iceberg scour marks in till in the Interlake region.

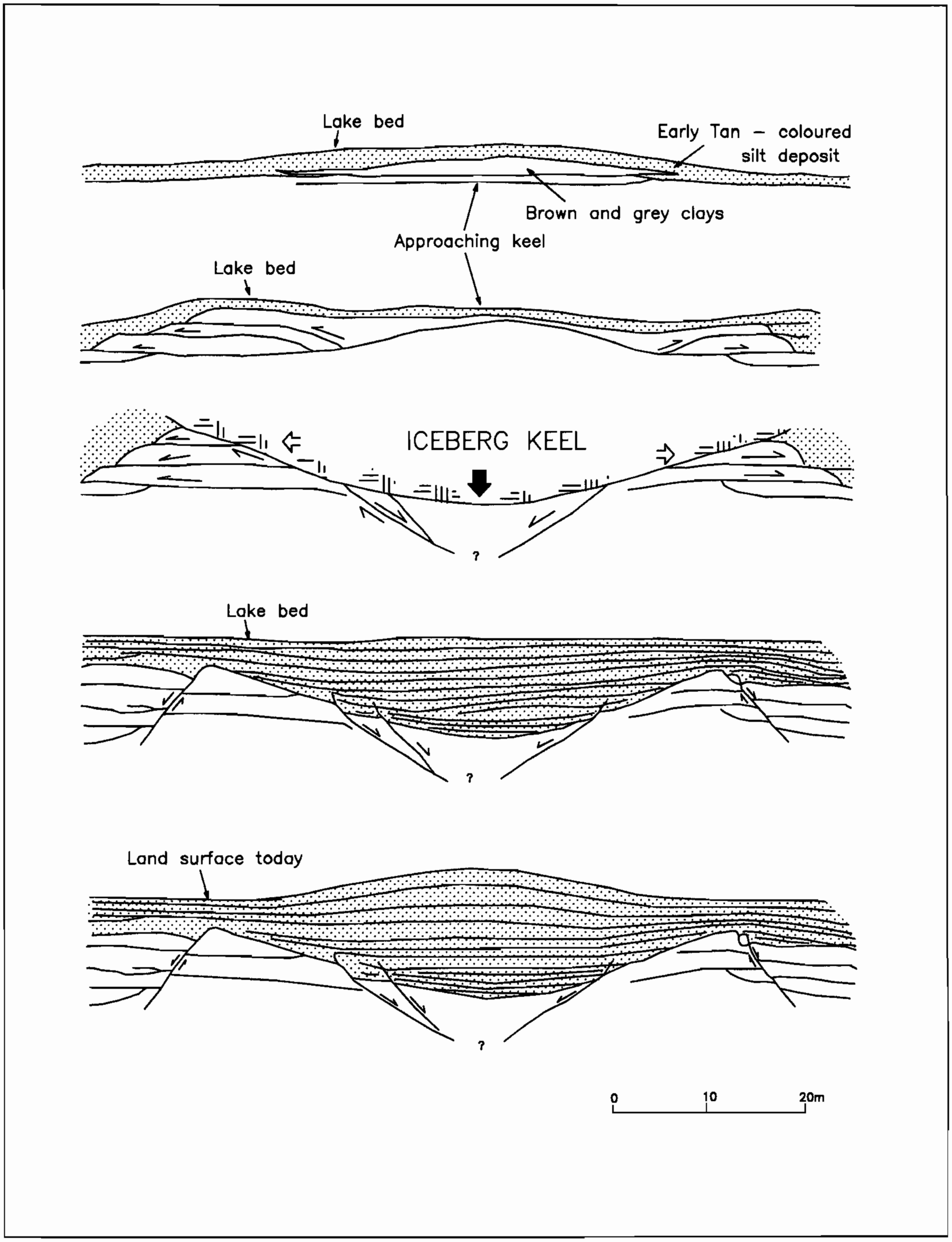

Figure 26. Geological model of the construction of an iceberg scour mark on the floor of Lake Agassiz, modified from Woodworth-Lynas and Guigne (1990). The fifth diagram depicts the ridge as we see it today, following dewatering of the clay.

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APPENDIX

REVIEW OF LAKE AGASSIZ HISTORY

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ABSTRACT

Lake Agassiz research has been influenced over the past century by the consequences of the fact that Warren Upham did not recognize a major transgression of the lake. Johnston established that two high lake level phases had occurred, but efforts to defend Upham by claiming that the second high level, the Emerson Phase, only reached the Campbell level (the lowest level to drain to the Minnesota River) became entrenched. Alluvial fills in valleys to the west, and the elevation of sediments deposited in the Emerson Phase, indicate, however, that the Emerson limit is the Norcross level. Furthermore, it appears likely that drainage switched between the Minnesota River and outlets across the Mesabi Range during the first high-water period, the Lockhart Phase. The Moorhead Phase, a time of low lake level between 11 and 10 ka, was initiated by drainage to Lake Superior, was maintained by drainage to the Clearwater River in Saskatchewan, and was terminated by a major re-advance of the ice margin at about 10 ka. Following the Emerson Phase, the Morris Phase was a time of step-wise drainage to the Great Lakes and Lake Ojibway, culminating in eastward drainage to incipient Hudson Bay.

INTRODUCTION

Lake Agassiz was an immense ice-dammed lake that formed in the Hudson Bay drainage of central North America during the last deglaciation (Upham, 1880, 1895), between about 12 and 8 ka BP (Clayton and Moran, 1982). Meltwater and nonglacial runoff from an area exceeding two million km^2 were channelled through the lake. Morphological features and sediments downstream in the Mississippi, Great Lakes, and Mackenzie basins, as well as oceanic circulation, were influenced as Lake Agassiz discharge was redirected and as the volume of water stored in the basin fluctuated (Teller, 1987, 1990; Broecker et *al.*, 1989; Fisher and Smith, 1994; Lewis *et al.,* 1994).

The lake was initiated as the margin of the Laurentide Ice Sheet retreated from the Hudson Bay - Atlantic drainage divide, and land sloping to the north in the Red River valley was inundated up to the level of the lowest available outlet across the divide. Initial drainage to the Gulf of Mexico, mainly through the southern outlet at the Minnesota River valley, was interrupted whenever lower outlets to the Atlantic or the Arctic Ocean were uncovered by ice recession. The basin was tilted by differential postglacial isostatic rebound, so shorelines now rise in elevation to the northeast, and outlets in the north rose relative to those in the south during evolution of the lake. Lake Agassiz finally drained after the Laurentide Ice Sheet was breached by Hudson Bay.

Fluctuations in lake level are recorded in offshore stratigraphy. In North Dakota and Minnesota (Harris *et al.*, 1974), a major fluctuation in water level is indicated by Sherack Formation clay, which is separated from underlying Brenna Formation clay by wood and other subaerial organic remains, dated about 11 to 10 ka. Along the western shore are large sandy subaqueous turbidity fans, including the Sheyenne and Pembina Deltas in North Dakota, and the extensive Assiniboine Delta, with its apex at Brandon, Manitoba. Partially filling the narrow valleys which were cut into these fans by the rivers at their heads are fossiliferous valley fills which mark transgressions of the lake (Elson, 1956, 1967). Radiocarbon dates and fossils from these deposits indicate the timing and extent of major lake level fluctuations and the paleoecology of these episodes.

Shoreline features record a series of lake levels (Upham, 1895; Leverett, 1932; Johnston; 1946; Elson, 1967; Teller and Thorleifson, 1983). In Minnesota, east of and clearly connected to the southern outlet, four well developed shorelines are present near the towns, in descending order, of Herman, Norcross, Tintah, and Campbell (Upham, 1890). North of the southern outlet are discontinuous higher shorelines and multiple features whose correlation to these type localities is uncertain. Several more levels were drained by lower outlets. The position of these shorelines relative to outlets and ice margins, and the chronological implications of their degree of differential uplift, indicate the paleogeography of the lake.

Lake Agassiz has been examined over the past 170 years by scientific explorers, government geological surveys, and academic researchers. Much work by Warren Upham was summarized in a United States Geological Survey (USGS) Monograph (Upham, 1895). Additional work by the USGS was summarized by Leverett (1932) and work by the Geological Survey of Canada (GSC) culminated in a report by Johnston (1946). A Lake Agassiz conference held at the University of Manitoba in 1966 (Mayer-Oakes, *ed.*, 1967) included a review of progress by Tamplin (1967) and led to a bibliography by Bannatyne et al. (1970). A Lake Agassiz symposium held at the Geological Association of Canada meeting in Winnipeg in 1982 (Teller and Clayton, *eds.*, 1983) included a synthesis of progress (Elson, 1983). These symposia drew attention to the increasing role played by academic research as well as by provincial and state geological surveys.

HISTORY OF LAKE AGASSIZ RESEARCH

The former existence of an extensive body of water in the Red River valley was first documented by scientific explorers. Winchell (1873), who initiated geological surveys for the state of Minnesota, was first to attribute the lake to glacial blockage of northward drainage.

The first major phase of research on Lake Agassiz began when Winchell hired Warren Upham. Following education and geological survey work in New Hampshire, Upham was employed by the Minnesota Geological Survey from 1879 to 1885 (Emmons, 1935). His first field season, in 1879, was directed at the geology of central and western Minnesota. A compelling aspect of his first report (Upham, 1880), is his meticulous presentation of the glacial theory. Due to the suggested close association of the lake which had filled the Red River valley with the retreating ice margin, the name Lake Agassiz was proposed in recognition of the first prominent advocate of the glacial theory. Upham subsequently conducted additional mapping for the Minnesota Survey, was then supported by USGS, and, in 1887, extended mapping into Manitoba as far north as Riding Mountain with the support of GSC. Upham (1895) summarized his work in the Lake Agassiz basin, south of Lakes Manitoba and Winnipeg, in a USGS monograph.

Upham made immense contributions to the advancement of science in general, and studies of Lake Agassiz in particular. It is important, however, to recognize his errors, and to understand that the implications of these errors persist to the present:

1) The Sherack problem resulted from Upham's mistaken interpretation of at least the upper clay in the Red River valley as postglacial alluvium. Upham reasoned that the location of the deposits near the Red River, and not throughout the area of former inundation, and the presence of organic remains such as logs and turf deep within the sequence, indicated an alluvial origin. Upham therefore had no reason to suspect that Lake Agassiz did anything other than form and ebb, while enlarging northward, during a single episode of ice retreat, and it was to this model that his interpretations were fitted. Upham therefore did not recognize the second high phase of Lake Agassiz.

2) The Herman problem was recognized in print when Chamberlin (1895), Upham's supervisor, placed a dissenting chapter in his monograph. Chamberlin noted that the Herman beach is poorly developed and does not vary across moraines, and suggested that this feature formed by the limit of a transgression. These observations conflicted with Upham's view that the Herman formed during ice recession from South Dakota to Riding Mountain. Chamberlin suggested that crustal subsidence was not complete until after the early history of Lake Agassiz, causing a rise to the Herman.

Given Upham's earnest advocacy of the glacial theory, attribution of the filling of Lake Agassiz to undetermined forces to the north would have hindered his cause. He did not relent, but Chamberlin's dissent was followed by the claim by Tyrrell (1896, 1898) that Lake Agassiz had formed by contact in northern Manitoba between a retreating western glacier and an advancing eastern glacier. Tyrrell had dismantled the concept of the Laurentide Ice Sheet, claiming instead that multiple ice sheets had formed in differing space and time. This concept was resisted by Upham (1913, 1914), but the concept of a single Laurentide Ice Sheet, with a synchronous margin from the Prairies to the Atlantic, was lost until its re-establishment by Flint (1943). Upham's legacy suffered additional erosion when Leverett (1912a, 1912b, 1932) concluded that Upham's glacial history required substantial revision.

The Johnston era of Lake Agassiz research established the concept of two periods of high lake level. Johnston (1914, 1915), who mapped the Rainy River district of Ontario for the GSC, solved the Sherack problem by arguing convincingly that thick, calcareous silt and clay well above the limit of fluvial action, and overlying organic material, must indicate the re-filling of Lake Agassiz. Johnston (1916) went on to attempt a solution to the Herman problem, adding to Chamberlin's concerns about the lack of boulders on the Herman shore, which supposedly formed in contact with the ice margin. After indicating that the upper clay reaches elevations slightly above the Campbell level, Johnston suggested that Lake Agassiz had re-filled to the Herman shoreline.

In response, Upham (1917) maintained that Johnston's views were in error, and he appealed for a defence of his monograph. In support of his model, he cited the abundant sediment of the deltas formed in the Herman level, which must largely have been contributed by a nearby ice margin to the west, and he suggested that rapid ablation had prevented the formation of bergs which would have distributed boulders. Leverett (1917), in what seemed to be an attempt to cool tempers, suggested, in defiance of Johnston's reasoning, that it was probable that the southern outlet had been fully eroded by the early high lake phase, so Upham's model only required a supplementary chapter rather than a radical modification. Johnston (1921) appeared to acquiesce to Leverett's compromise, later stating that the Campbell shoreline probably marks the upper limit of the second stage of the lake.

The Elson era was marked by the advent of radiocarbon dating, aerial photographs, and topographic maps. Elson (1956, 1967), who built on work by Zoltai (e.g. 1965) and Prest (1963), added much detail to the Lake Agassiz model, and drew attention to valley fills at the Norcross and Campbell levels in the Pembina and Assiniboine valleys, features later reported for the Sheyenne River by Brophy (1967). A bewildering array of radiocarbon dates on various materials caused Elson (1967) to suggest that southern Manitoba had been deglaciated at about 13 ka, and that multiple oscillations of lake level had occurred. The Clayton and Moran era was a time of simplification. Dates on material other than wood were rejected by Clayton and Moran (1982), and a 'young chronology', with deglaciation of southern Lake Agassiz at about 11.7 ka, resulted. Their lake level reconstruction was more consistent with lithostratigraphy than Elson's model. Valley fills were, however, ignored, and complete erosion of the southern outlet in the first high level phase was maintained. Elson's rise to Norcross was forgotten, apparently due to the lack of radiocarbon confirmation for a rise above Campbell at ~9.9 ka.

In summary, the most contentious issue in Lake Agassiz research has been the limit of the second high phase. That this should be an issue is remarkable, given that there is no evidence, nor has anyone ever claimed there is any evidence, indicating that the southern outlet was fully eroded in the first high-water phase, now known as the Lockhart Phase. At best, it can only be said that Warren Upham was able to fit his observations into a model with no transgressions above Campbell. This is not surprising, given his failure to correctly interpret the stratigraphy of fine-grained sediments in the Red River Valley. It must also be recalled that Upham's supervisor was so unconvinced that he placed a dissenting chapter in Upham's monograph.

HISTORY OF LAKE AGASSIZ

a. Ice margins

A series of ice marginal positions apparently uninterrupted by extensive readvances has been defined for Saskatchewan by Christiansen (1979) and Schreiner (1983). In contrast, major readvances and prolonged pauses occurred in the Red River and Superior basins. Ice margins in southern Saskatchewan were correlated to the Dakotas, southern Manitoba and Minnesota by Clayton and Moran (1982). Attig *et al.* (1985) correlated margins along the southern Superior basin, and the sequence of moraines in northwestern Ontario were summarized by Prest (1970). The late phases of Lake Agassiz were in contact with ice margins discussed by Schreiner (1983), Klassen (1983), and Dredge (1983).

b. Shorelines

The lake level defined on foot by Warren Upham in the 1880's has not been thoroughly reexamined since the advent of aerial photographs and topographic maps. Consequently, the model has been repeatedly challenged (e.g. Kupsch, 1967). Nevertheless, the compilation of data by Johnston (1946), and fragmentary reexamination by various authors, indicates that much of the model is reasonable. The isobases compiled by Teller and Thorleifson (1983; Figure 28) and a shoreline diagram which portrays the levels discussed here (Figure 29) are meant to be schematic portrayals which outline the broad structure of lake level variations. A full re-mapping of the shorelines, with definition of water planes on the basis of appropriately plotted data, is urgently required, but will require much effort.

The Alice level (Figure 28 and 29) mapped by Bluemle (1974) is here associated with northward drainage to Lake Koochiching in northern Minnesota, as discussed by Hobbs (1983), Fenton *et al.* (1983), and Clayton (1983). Correlations of northern Minnesota levels presented by Hobbs (1983) were modified using steeper water plane gradients. One level of an independent Lake Koochiching, the Mizpah level, is attributed to the time before Lake Agassiz merged with Lake Koochiching at the Trail level. A second major modification is the suggestion that many mapped berms, particularly in North Dakota, are subaqueous offshore bars, as proposed by Laird (1944, 1964) and Elson (1971). Consequently, significantly fewer levels are portrayed. Levels in the north represent a correlation to scattered data presented by Dredge (1983) and Klassen (1983), guided by the gradients of late shorelines in the Lake Ojibway basin at correlative isobases, as presented by Vincent and Hardy (1979) and Veillette (1994). Two schematic Moorhead Phase levels, early and late, are presented on the basis of a compilation by Warman (1991). The gradient of marine limit is also plotted in order to constrain speculation regarding the present gradient of late Lake Agassiz levels.

c. Outlets

Several pathways of discharge are known with confidence, are inferred, or may have drained Lake Agassiz. These include the southern outlet, northern Minnesota outlets, Thunder Bay area outlets, eastern outlets to Lake Nipigon and directly to Lake Superior, northwestern outlets to the Arctic Ocean through the Clearwater spillway and possibly Wollaston Lake, and poorly known probable connections to Lake Ojibway and possibly to early Hudson Bay.

The southern outlet is a clearly defined spillway extending from the well developed shorelines of the Red River valley to the Mississippi River (Matsch and Wright, 1967; Matsch, 1983). This channel initially was cut by the downstream, ice marginal extension of the Sheyenne spillway (Baker, 1966).

The Embarrass and Prairie outlets across the Mesabi range in northern Minnesota, and the McIntosh spillway extending from Lake Agassiz to northern Minnesota have been discussed by Hobbs (1983), Clayton (1983), and Fenton *et al.* (1983). These outlets were above the level of Lake Agassiz at the time of final ice retreat from northern Minnesota due to differential uplift.

Thunder Bay area outlets, in the Shebandowan and Savanne area where glaciolacustrine sediments have been observed to cross the drainage divide (Johnston, 1946), have been inferred by Elson (1967) and Teller and Thorleifson (1983). Thorleifson (1983) concluded that eastward drainage at the beginning of the Lockhart Phase took place in the vicinity of the Minnesota/Ontario border, at North Lake, Flatrock Lake, and Matawin River.

Eastern outlets to the Nipigon basin are a complex of channels located between the drainage divide and a former high level of Lake Nipigon confluent with the Superior basin (Antevs, 1931; Elson, 1957; Zoltai, 1967; Teller and Thorleifson, 1983). Additional possible Lake Agassiz outlets directly to Lake Superior are located at White Otter and Nagagami.

A single large channel in northwestern Saskatchewan, the Clearwater spillway, clearly carried an immense discharge from the Lake Agassiz drainage basin, possibly from Lake Agassiz itself, to the Arctic Ocean (Elson, 1967). A lack of clearly defined offshore sediments and shorelines connecting the main body of Lake Agassiz to the channel led Christiansen (1979) and Schreiner (1983) to conclude that the Clearwater channel was the outlet of a lake fed by meltwater derived from the nearby ice margin only. Smith and Fisher (1993) and Fisher and Smith (1994), however, have demonstrated a likely connection to Lake Agassiz. Another possible outlet to the Mackenzie River is located at Wollaston Lake in northern Saskatchewan.

Lake Agassiz probably was confluent with the Kinojévis levels of Lake Ojibway after a connection was permitted by retreat from the Agutua moraine (Teller and Thorleifson, 1983; Dyke and Prest, 1987; Lewis and Anderson, 1989). Drainage would have been controlled by outlets to the Ottawa River (Vincent and Hardy, 1979; Veillette, 1994). The Fidler level of Lake Agassiz reported by Klassen (1983) is lower than the lowest Ojibway outlets, so marine waters must have first entered James Bay, and Lake Agassiz must have been drained eastward to the sea. The Echoing spillway in northern Manitoba, on the Ontario border, is slightly below marine limit, but rapid final drainage of Lake Agassiz could have had sufficient head to scour to this depth. McDonald (1969) identified sediments probably related to this drainage on the upper Fawn River in Ontario. Klassen (1983) and Shoemaker (1992) have speculated that Lake Agassiz may have drained subglacially across Hudson Bay, but no observations call for this to have happened.

d. Stratigraphy

There have been several particularly important observations about the stratigraphy in the Lake Agassiz basin that have contributed to our understanding of the history of the lake, namely (1) lacustrine clay is overlain by till north of the Edinburg moraine (Arndt, 1977), (2) subaerial organic material lies between offshore clay units in the southern Red River valley and Lake of the Woods region, (3) fossiliferous valley fills are graded to Norcross and lower Campbell strandlines in the Assiniboine, Pembina, and Sheyenne valleys, and (4) there are sequences of rhythmites of probable annual origin in northwestern Ontario (Antevs, 1951; Warman, 1991), which constrain dates for deglaciation as well as fluctuations in the rate and provenance of sedimentation.

e. Geochronology

The age of inception of Lake Agassiz has been estimated at 11.7 ka by Clayton and

Moran (1982), on the basis of interpolation between an advance into South Dakota at 12.3 ka and the beginning of the Moorhead Phase at 10.9 ka. The final pre-Lake Agassiz advance in the Red River valley was correlated to the advance that overrode the Two Creeks forest bed in the Lake Michigan basin.

Wood dates from the unconformity of the southern Lake Agassiz basin extend from about 9.9 to 10.9 ka (Moran *et al.*, 1973). The former limit is much better constrained than the latter. The upper alluvial fill of the western valleys is undated. Sediments correlative to the lower fill were dated at 9.5 ka by Teller (1989).

The varve record in Lake Superior (Teller and Mahnic, 1988; Thorleifson and Kristjansson, 1993) indicates enhanced sedimentation correlated to the construction of the Agutua-Nakina moraines at about 8.2 ka. The ensuing cessation of varve sedimentation in Lake Superior at 8.0 ka (Mothershill, 1988) is correlated to the diversion of outwash to Lake Ojibway, probably coinciding with the diversion of Lake Agassiz outflow to the Ottawa River.

f. Proposed history

The western margin of Lake Agassiz indicates three major events: 1) spillway erosion and deposition of nonfossiliferous sediments as large subaqueous fans in the Herman level during deglaciation, 2) a major drop in lake level followed by a rise to about the 350 m or Norcross level in the Pembina and Assiniboine valleys, 3) another drop in lake level followed by a rise to 320 m in these valleys, the Campbell level. The stratigraphy of the southern Red River valley indicates: 1) sedimentation of fine-grained sediments during fluctuating ice retreat, 2) subaerial exposure, and 3) a transgression whose fine-grained deposits can be confidently traced at least up to the Campbell shoreline. The upper limit of Brenna sediments, which were at least in part deposited in a Herman level, indicates that clay sedimentation in Lake Agassiz did not occur in less than tens of metres of water depth. It therefore is far more reasonable to correlate the Sherack Formation with the upper or Norcross alluvial fill, rather than the lower. The lower fill may readily be correlated to a late Emerson Phase temporary opening of an eastern outlet which has been suggested by, for example, Johnston (1946).

An Emerson Phase maximum well above Campbell offers the following benefits: 1) unconvincing claims of clay sedimentation in negligible depth would no longer be required, 2) a satisfactory explanation for the upper alluvial fill in the western valleys, 3) a reasonable scenario for distribution of red clay across northwestern Ontario, 4) an explanation for glaciolacustrine sediments across the drainage divide east of Thunder Bay, and 5) an explanation for the gradient of the uplifted upper Campbell water plane, which is only slightly steeper than lower Campbell and is identical at correlative isobases to that of the post-10-ka Duluth level in Lake Superior.

A solution to the Herman problem is proposed here (Figure 32). Temporary drainage to Lake Superior, followed by differential uplift and re-occupation of the southern outlet, would have produced the highest Herman beach on a landscape which had, slightly earlier, been underlain by much stagnant ice (Bluemle, 1974).

Because the Cass phase, as defined by, for example, Fenton *et al.* (1983), is insignificant relative to Lake Agassiz history as a whole, this very early period of Lake Agassiz history is here combined with the Lockhart Phase (Figure 29). The southern outlet may have been intermittently abandoned late in the Lockhart Phase, so the Moorhead Phase is defined as the first drop in water level which desiccated the Fargo area. The Emerson Phase is marked by the reoccupation of the southern outlet until the abandonment of the lower Campbell shoreline. Drainage may have been routed to the Mackenzie River through Wollaston Lake during a portion of Post-Emerson time, so the term Morris Phase, introduced by Bjork and Keister (1983) and Bjork (1985), is used rather than Nipigon and Ojibway phases, introduced by Teller and Thorleifson (1983).

The Lockhart Phase (Figures 30 to 33), which began after about 12 ka, was a time in which the Minnesota River and Mesabi Range outlets drained the lake, the basin south of Winnipeg was deglaciated, large discharges from the west cut the Sheyenne, Pembina and Assiniboine spillways and deposited extensive subaqueous fans, and the Brenna Formation which makes up the majority of clay in the southern basin was deposited. The southern outlet was eroded 15 m, down to the Herman level, by the ice marginal Sheyenne River which flowed through Lake Agassiz.

The Moorhead Phase (Figures 34 and 35) began at about 10.9 ka as the ice margin retreated from the Steep Rock moraine at the Minnesota - Ontario border, initiating eastward drainage to Lake Superior. The Eagle-Finlayson moraine was deposited with its eastern extension reoriented, possibly by this drainage, to its Brule position (Zoltai, 1961, 1963, 1965, Prest, 1970). Continued ice retreat opened the Shebandowan outlet to Thunder Bay and lake level fell dramatically. Subaerial vegetation colonized the US part of the basin, southeastern Manitoba, and the Rainy River district. During a millennium of lowered lake level, differential uplift would have gradually raised lake level, but opening of the Clearwater outlet in late Moorhead time maintained low levels.

The Emerson Phase (Figures 36 and 37) began at about 9.9 ka as the Superior lobe dammed the Kaministikwia basin at the Marks moraine. Correlative ice advance to the Dog Lake and Lac Seul moraines and the Arran till limit in western Manitoba closed access to Clearwater and Lake Agassiz filled to the Norcross level, inaugurating the Emerson Phase. Lake Agassiz was confluent with Lake Kaministikwia and red clay was deposited in deep water across northwestern Ontario for two to three decades. Western rivers aggraded their lower courses and Sherack Formation silty clay was deposited offshore. The southern outlet was eroded to resistant bedrock and the strongly developed Campbell beach formed. Ice retreat opened the Kaiashk outlet but, after about two centuries, a readvance to the Sioux Lookout moraine raised Lake Agassiz to the lower Campbell level due to differential uplift. Western rivers deposited another valley fill.

The Morris Phase, the final chapter of Lake Agassiz (Figures 38 to 41) was initiated by retreat from the Sioux Lookout moraine. Final deglaciation of the Kaiashk outlet caused abandonment of the southern outlet. The Clearwater channel was now above Lake Agassiz. A succession of outlets to Lake Nipigon and later directly to Lake Superior were opened. Drainage to the Ottawa River through Lake Ojibway may have occurred before and after a readvance at about 8.2 ka to the Agutua moraine. Marine inundation of Hudson

Bay occurred as Lake Ojibway was abruptly drained northward, leaving brecciated clays in its wake (Skinner, 1973). A deep Lake Agassiz, mostly resting north of a nearly dry Lake Winnipeg, was held back in northern Manitoba, until rapid eastward drainage scoured the Echoing spillway, which carried water to early Hudson Bay. Postglacial uplift has reduced the area of Hudson Bay, but has greatly expanded the extent of Lake Winnipeg.

SUMMARY

The principal points emphasized here are: 1) that the earliest levels of Lake Agassiz built shorelines on stagnant ice well above the Herman level, 2) that two outlets across the Mesabi Range controlled Lake Agassiz for a portion of its early history, 3) that the highest Herman shoreline was formed as outlets to the Superior basin were uplifted until the southern outlet to the Minnesota River was re-occupied, 4) that Lake Agassiz was lowered from the high level Lockhart Phase to the low Moorhead Phase by the opening of outlets to Thunder Bay and on to the Atlantic from 10.9 to about 10.3 ka, 5) that drainage to the Arctic Ocean via the Clearwater River in Saskatchewan maintained low levels from about 10.3 to 9.9 ka, 6) that a readvance at the start of the Emerson Phase at 9.9 ka, which blocked western and eastern outlets, returned drainage to the southern outlet and hence the Gulf of Mexico by raising Lake Agassiz to the Norcross level, at which time the upper fossiliferous fill of valleys along the western shore was deposited, 7) that the southern outlet was eroded to the Campbell level for the first time during the Emerson Phase, 8) that a brief episode of eastward outflow in late Emerson time ended with a rise to the lower Campbell shoreline and deposition of another alluvial fill, and 9) that the Morris Phase history of the lake, after final abandonment of the southern outlet, involved drainage to Lake Nipigon and, in the very late history of the lake at about 8 ka, a possible brief episode of northwestward drainage through Wollaston Lake, probable direct drainage to northeastern Lake Superior, confluence with Lake Ojibway, and final drainage eastward to Hudson Bay.

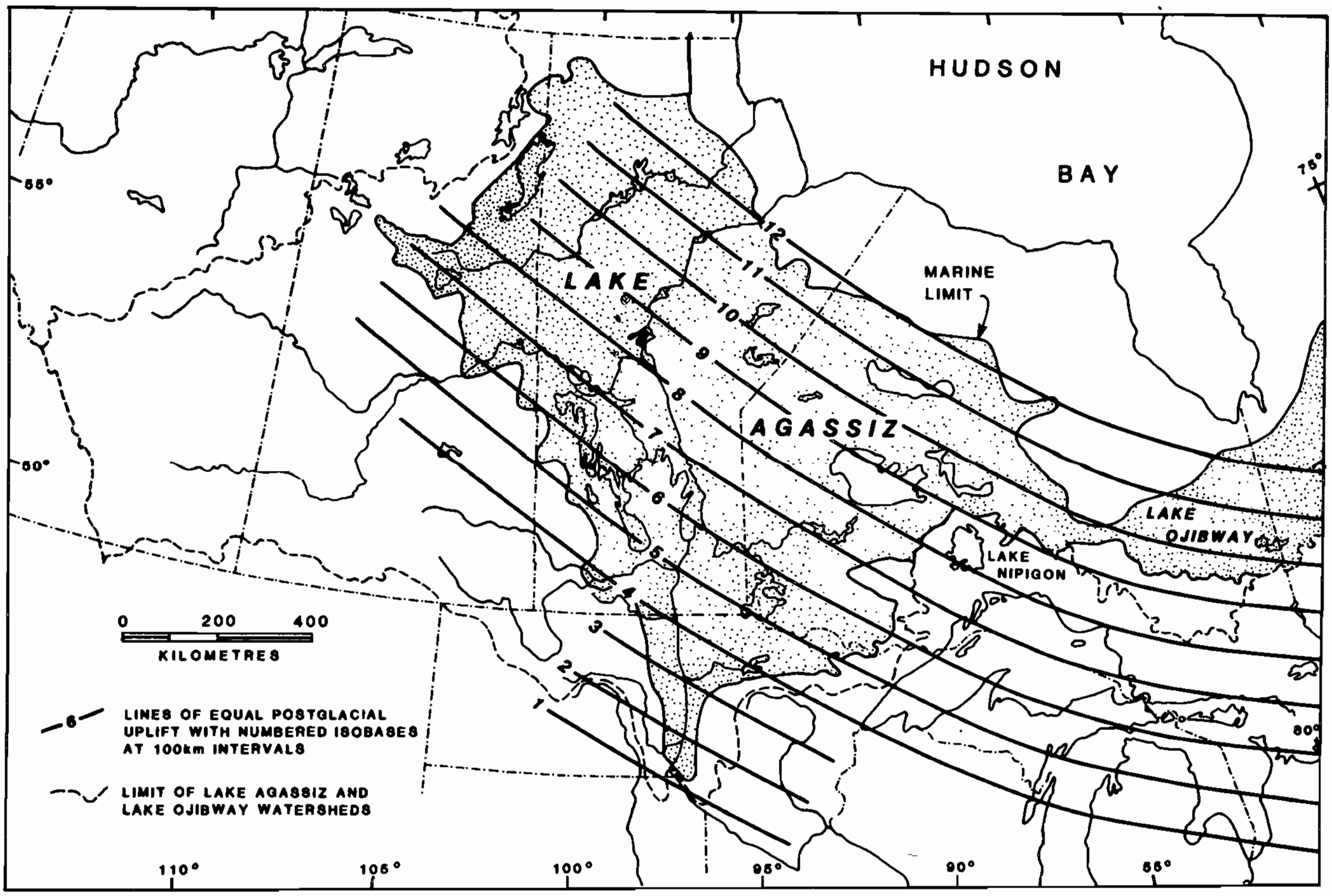

Figure 27. ***Maximum extent of Lake Agassiz (after Teller et al., 1983) and the late stages of Lake Ojibway (Vincent and Hardy, 1979; Veillette, 1994), with isobases showing mean trend of lines of equal postglacial uplift.*** *Data were compiled by Teller and Thorleifson (1983) from the work of Johnston (1946), Walcott (1972), and Vincent and Hardy (1979). Isobases are speculative outside southern Lake Agassiz and Great Lakes region.*

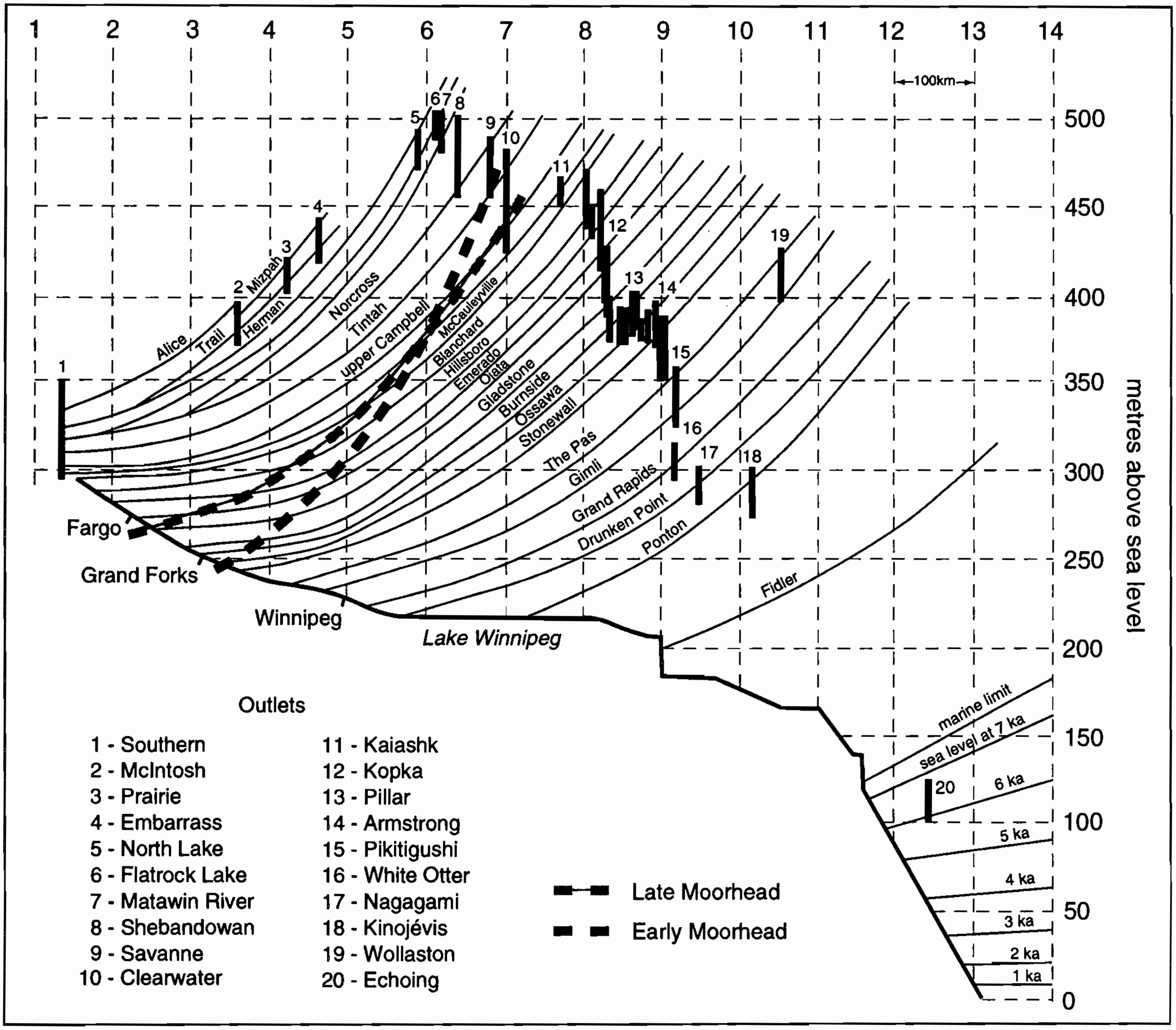

Figure 28. ***Schematic model for general configuration of Lake Agassiz shorelines, showing former water planes and outlets (vertical bars) projected onto a vertical plane orthogonal to the isobases depicted in Figure 27.*** *Water planes are based on data from Johnston (1946), Bluemle (1974), Hobbs (1983), Dredge (1983), Klassen (1983), and Craig (1969). Dashed Moorhead Phase levels (Warman, 1991) are schematic approximations of shorelines which were inundated by the subsequent rise in lake level.*

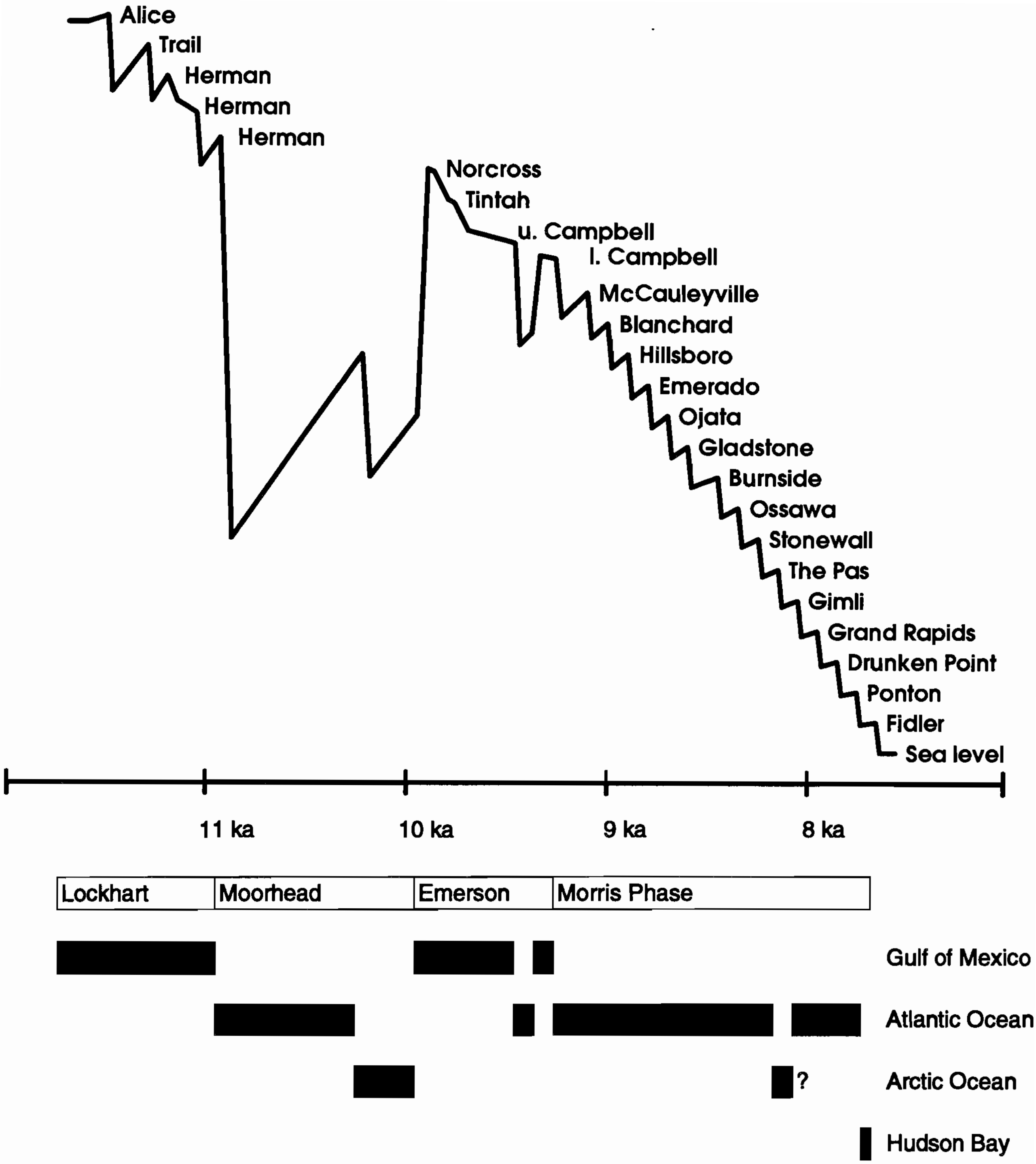

Figure 29. ***Schematic model for the trend and approximate relative magnitude of Lake Agassiz water level fluctuations, and subdivision of Lake Agassiz history as four phases.*** *Suggested vertical range of Moorhead Phase levels are plotted with reference to the elevation of later Morris Phase shorelines in North Dakota. Proposed routing of freshwater discharge from the Lake Agassiz basin also is indicated.*

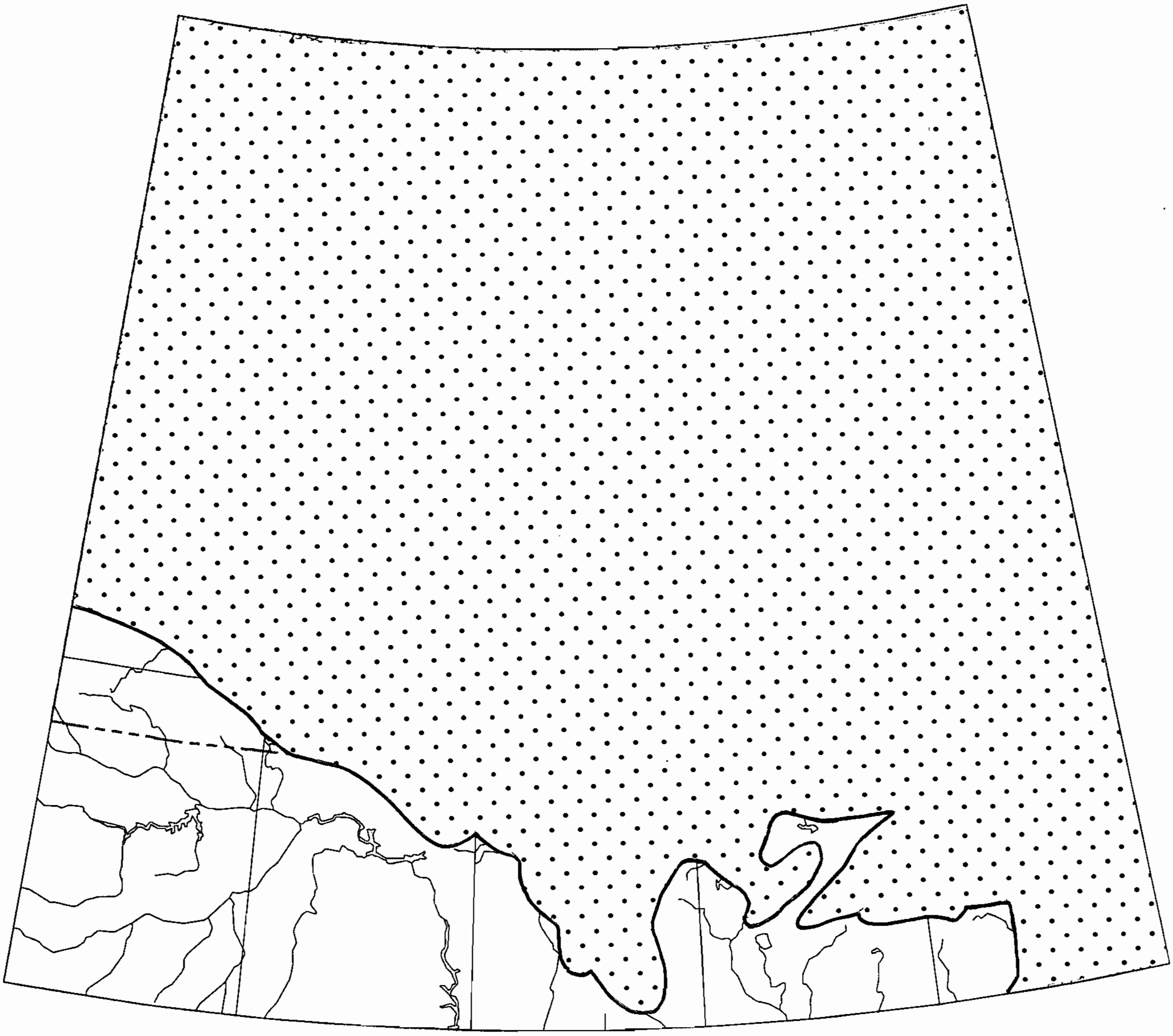

Figure 30. ***Lake Agassiz region paleogeography at ~11.7 ka BP:*** *Clayton and Moran (1982) correlated an advance to the Big Stone moraine, the final pre-Lake Agassiz margin in the Red River valley, to the Maxwelton margin defined for southern Saskatchewan by Parizek (1964), to the Kensal and North Viking phases in North Dakota, and the maximum or Alborn phase, or possibly a retreating margin (Attig et al., 1985), of the St. Louis sublobe in northern Minnesota. Attig et al. (1985) correlated this event to the second occupation of the Nickerson moraine of the southwestern Superior basin, which deposited the Barnum Formation, and to the Marenisco and late Athelstane margins in Wisconsin and adjacent Michigan which correlate to the over-riding of the Two Creeks forest bed. As ice retreated from the Big Stone moraine, flowing water and small bodies of standing water in the Lake Agassiz basin maintained a level of about 355 m, 30 m above the Herman level and graded to the crest of the moraine (Baker, 1966; Fenton et al., 1983; Clayton, 1983). Lake Agassiz formed, surrounded by stagnant ice, as ice retreated sufficiently from the Big Stone moraine for the lake to extend from the North Dakota to the Minnesota shores of the valley. The southern outlet was eroded to the Alice level (Bluemle, 1974), the highest shoreline here attributed to Lake Agassiz. As ice retreated in northern Minnesota, the Mizpah level of Lake Koochiching (Hobbs, 1983) was drained by both the Prairie and Embarrass spillways. A beach at 427 m (1400') at Mizpah reported by Hobbs (1983) is here correlated to the 436 m (1430') level at Togo and to the Norwood level at 443 m (1450') at the Embarrass outlet.*

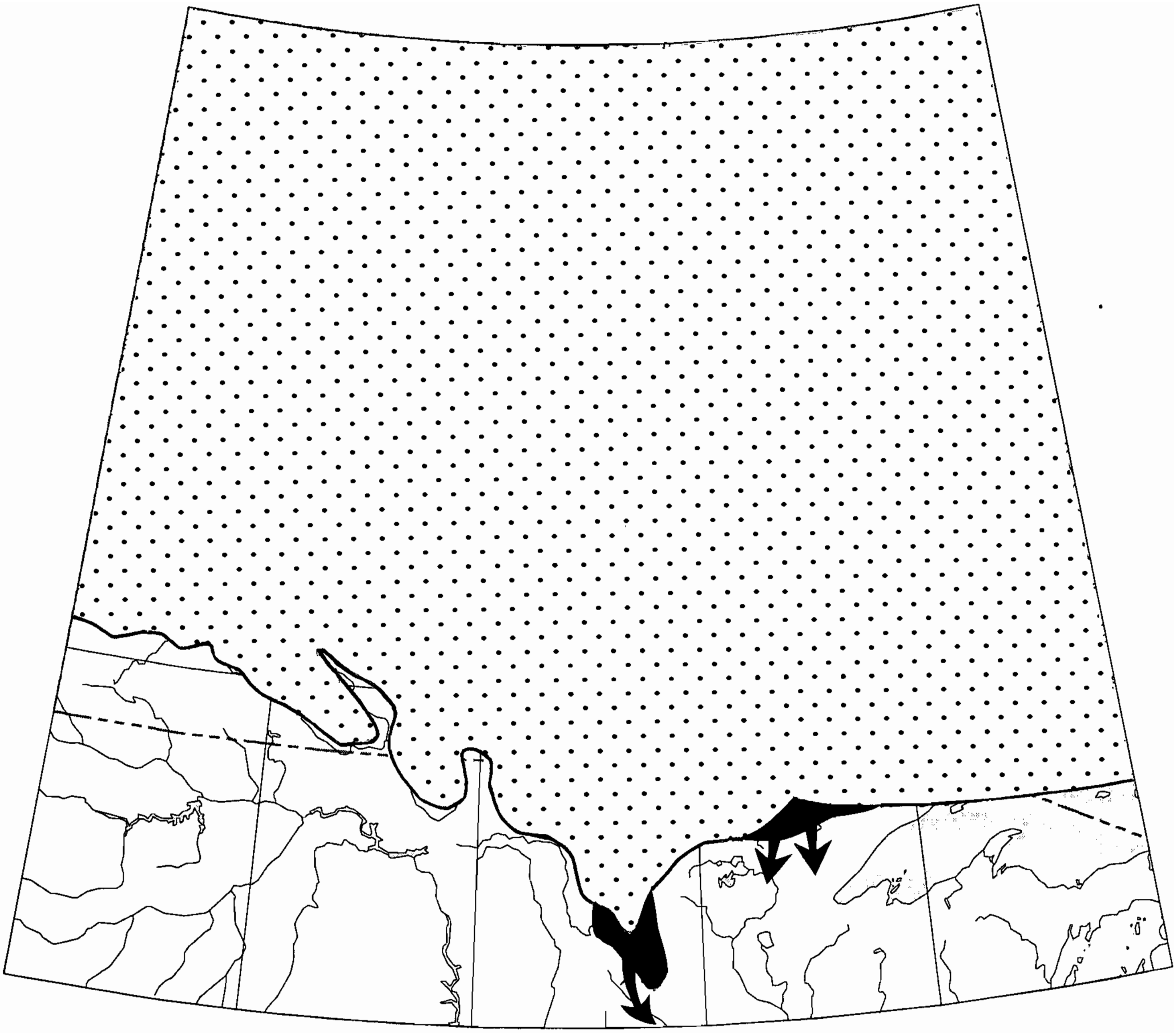

Figure 31. ***Lake Agassiz region paleogeography at ~11.5 ka BP:*** *As the Wahpeton and Erskine moraines (Leverett, 1932) were built in the Alice level of Lake Agassiz and the Mizpah level of Lake Koochiching, the ice margin to the west formed the Weyburn and Souris lobes (Clayton and Moran, 1982). To the east, the St. Louis sublobe margin had retreated north of the Mesabi range and ice in the Superior basin had retreated from the Marenisco and correlative positions (Attig et al., 1985). As ice retreated from this position, floodwater cut the Sheyenne spillway (Brophy, 1967; Brophy and Bluemle, 1983; Kehew and Lord, 1986; Kehew and Teller, 1994) and deposited the Sheyenne fan in the Alice level at 335 m, 12 m above the Herman level (Fenton et al., 1983). A brief rise of Lake Agassiz during this flood may have caused the formation of the McIntosh spillway, which extends from the Red River valley to northern Minnesota and which deposited a fan at 389 m (1275') in the Mizpah level of Lake Koochiching. Downcutting of the southern outlet to a water level of 323 m (1060'), later to govern the Herman level, caused flow to reverse in the McIntosh channel. Downcutting of the Mesabi outlets lowered Lake Koochiching. As ice continued to retreat, the southern outlet was abandoned as Lake Agassiz fell to become confluent with this lower level of Lake Koochiching. Differential uplift of the Mesabi outlets caused Lake Agassiz to transgress and re-occupy the southern outlet, culminating in a three-outlet Trail level, which formed several shorelines mapped by Hobbs (1983), including the 436 m (1430') terrace at Embarrass, a shoreline at 412 m (1350') at Gemmell, and the upper shoreline at Trail (381 m; 1250').*

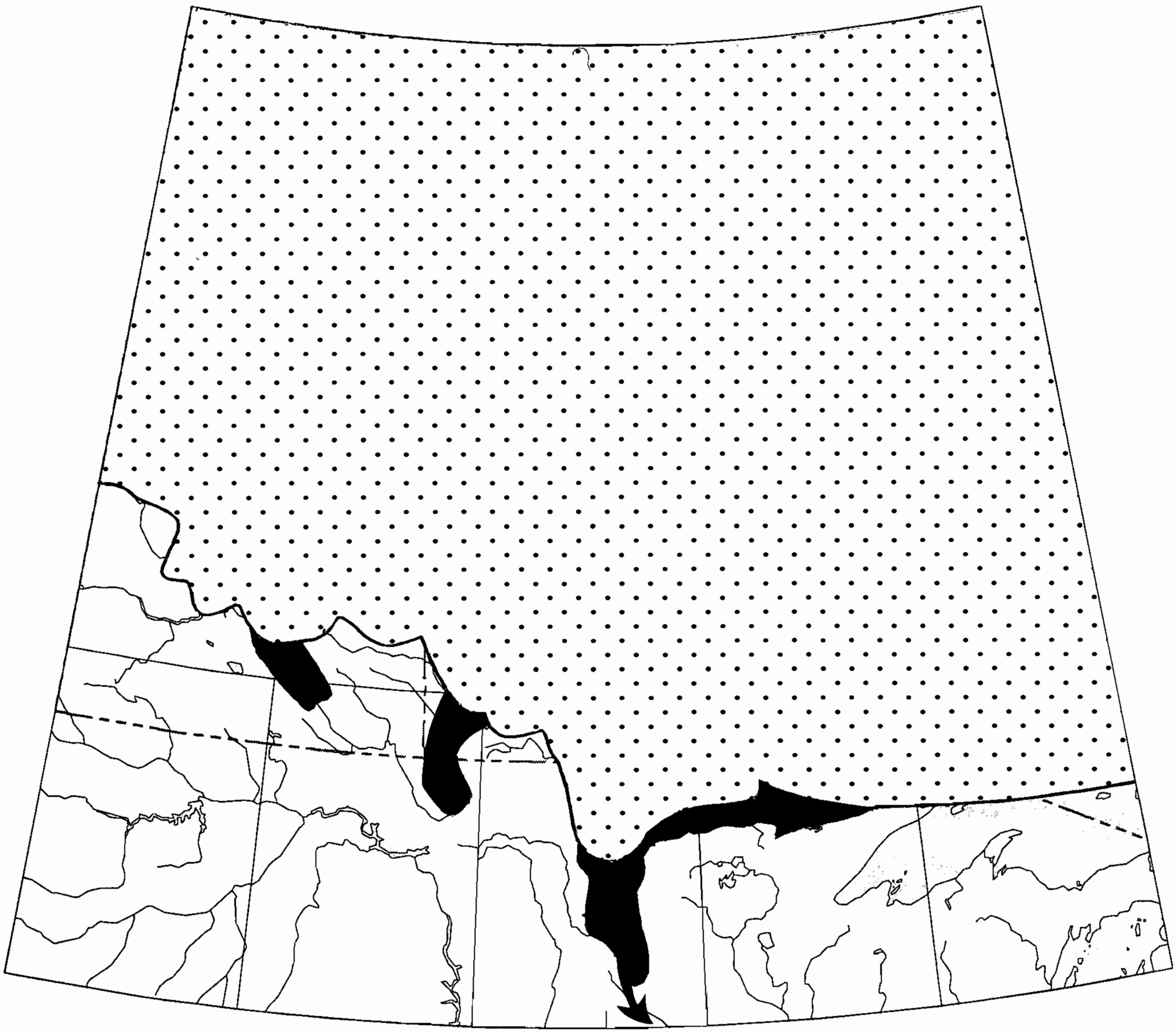

Figure 32. ***Lake Agassiz region paleogeography at ~11.2 ka BP:*** *The Condie moraine at Regina is correlated to the Darlingford moraine in southwestern Manitoba, the Edinburg and Holt moraines of the Red River valley, the Rainy River interlobate moraine (Bajc, 1991) of Ontario, and a later overridden margin to the east in the Superior basin. Glacial Lake Souris had formed and was drained by the Sheyenne spillway. As ice retreated, a portion of Lake Agassiz discharge was diverted from the Mesabi and/or southern outlets to North Lake, located on the Minnesota-Ontario border. Opening of the nearby Flatrock Lake outlet probably caused abandonment of the southern outlet, but differential uplift of these outlets caused a transgression which at least partially returned flow to the southern outlet and formed what has been known as the highest Herman beach (Upham, 1895; Johnston, 1946). The Mesabi outlets were at this point elevated above the southern outlet by differential uplift. By this time, stagnant ice in eastern North Dakota had melted, so the Herman shoreline is the highest undisturbed, readily apparent beach (Bluemle, 1974). This rising level explains the transgressive character of the Herman shoreline, as discussed by Chamberlin (1895). As ice to the west retreated, glacial Lakes Regina and Souris were drained. Floodwater first eroded the Souris/Pembina spillway and subsequently the Qu'Appelle and Assiniboine spillways, depositing the Pembina and Assiniboine fans in the Herman level. Elson (1967) reported an uninterrupted sequence of 500 varves, beginning with proximal units, on Lake of the Woods, at a site exposed during the Moorhead Phase. This would imply deglaciation of Lake of the Woods at about 11.4 ka, earlier than implied here.*

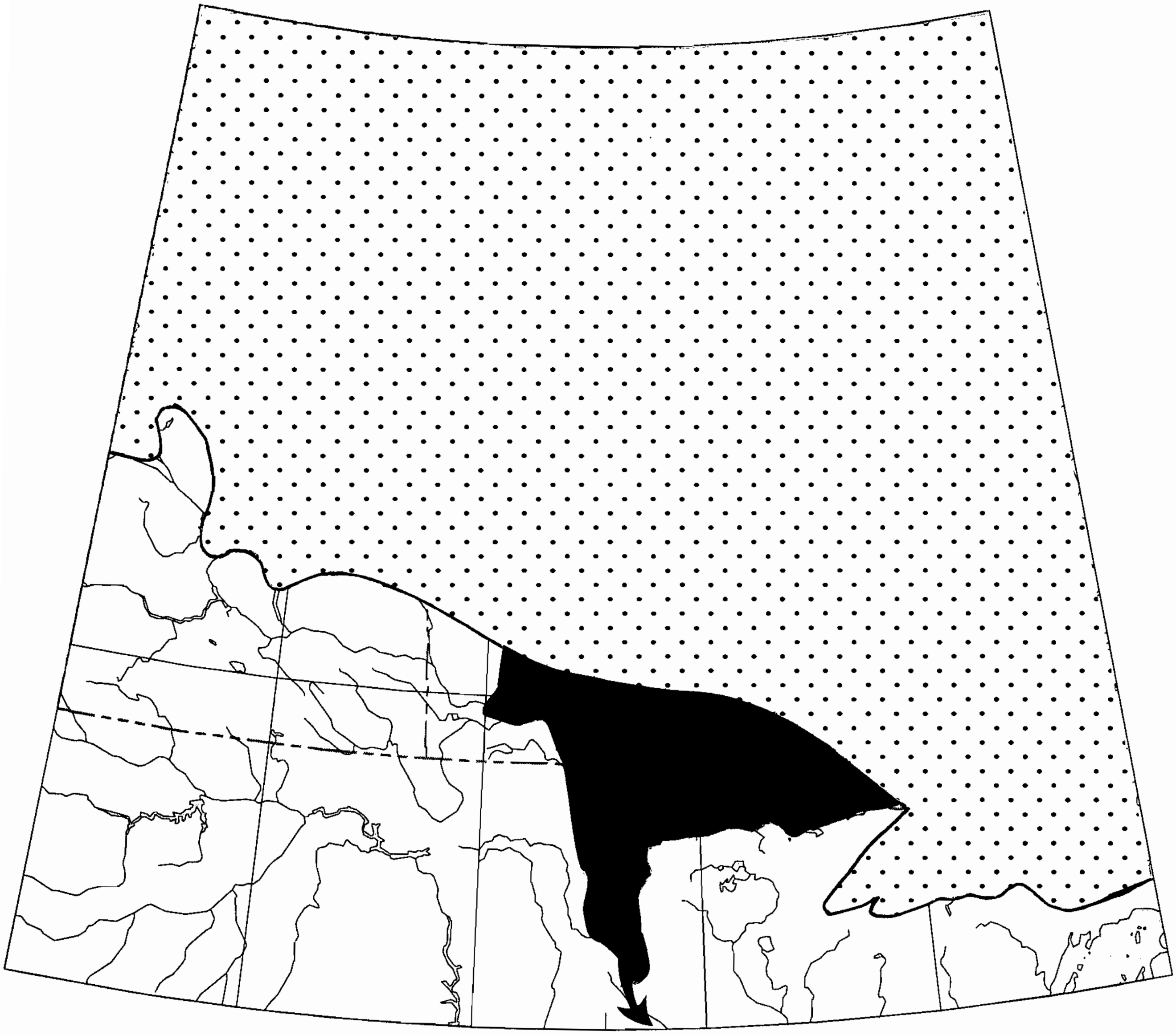

Figure 33. **Lake Agassiz region paleogeography at ~11.1 ka BP:** *An ice margin between Saskatoon and Regina, defined by Christiansen (1979), is here correlated to later overridden margins in Manitoba and much of northwestern Ontario, to formation of the Steep Rock moraine on the Minnesota-Ontario border, and to the Porcupine phase of Lake Superior (Clayton, 1983; Clayton, 1984; Attig et al., 1985). The southern outlet was eroded to a water level at 317 m (1040'), producing a second Herman level.*

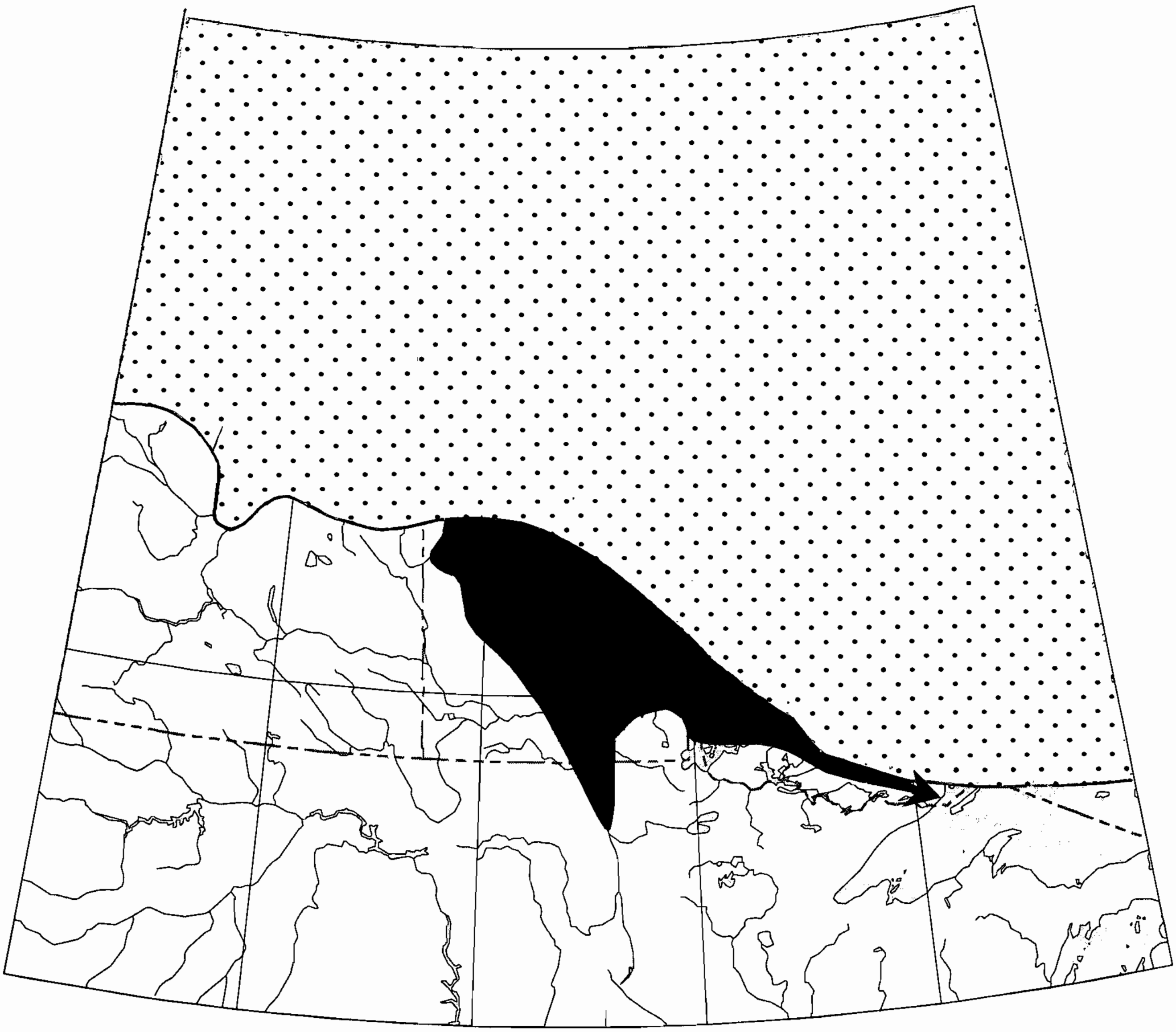

Figure 34. ***Lake Agassiz region paleogeography at ~10.9 ka BP:*** *An ice margin at Saskatoon (Christiansen, 1979) is correlated to a later overridden position in Manitoba and to the construction by enhanced glaciofluvial discharge (Sharpe and Cowan, 1990) of the Eagle-Finlayson and Brule moraines. In the Superior basin, the ice margin was retreating from the Porcupine maximum (Attig et al., 1985). At about this time, opening of drainage to Thunder Bay through the Shebandowan and Savanne area first caused abandonment of the southern outlet as water fell to a level portrayed here as a third Herman level (Figure 28), and subsequently caused a dramatic lowering of Lake Agassiz to initiate the Moorhead Phase. A water plane drawn a constant increment lower than the final Herman shoreline (Figure 28; Warman, 1991) defines a shoreline determined by eastward drainage and which accounts for the lowest Moorhead Phase level, in the Grand Forks area, defined by Arndt (1977). As discussed by Warman (1991) and Barnett (1992), there is no evidence for, and varve evidence against, a major readvance in northwestern Ontario, so the Savanne outlet probably was the northernmost eastern outlet available in the Moorhead Phase. The Clearwater outlet of Saskatchewan appears not to have been deglaciated in early Moorhead time, so drainage of Lake Agassiz was to Lake Superior, and, according to Clayton (1983), to the Atlantic at this time.*

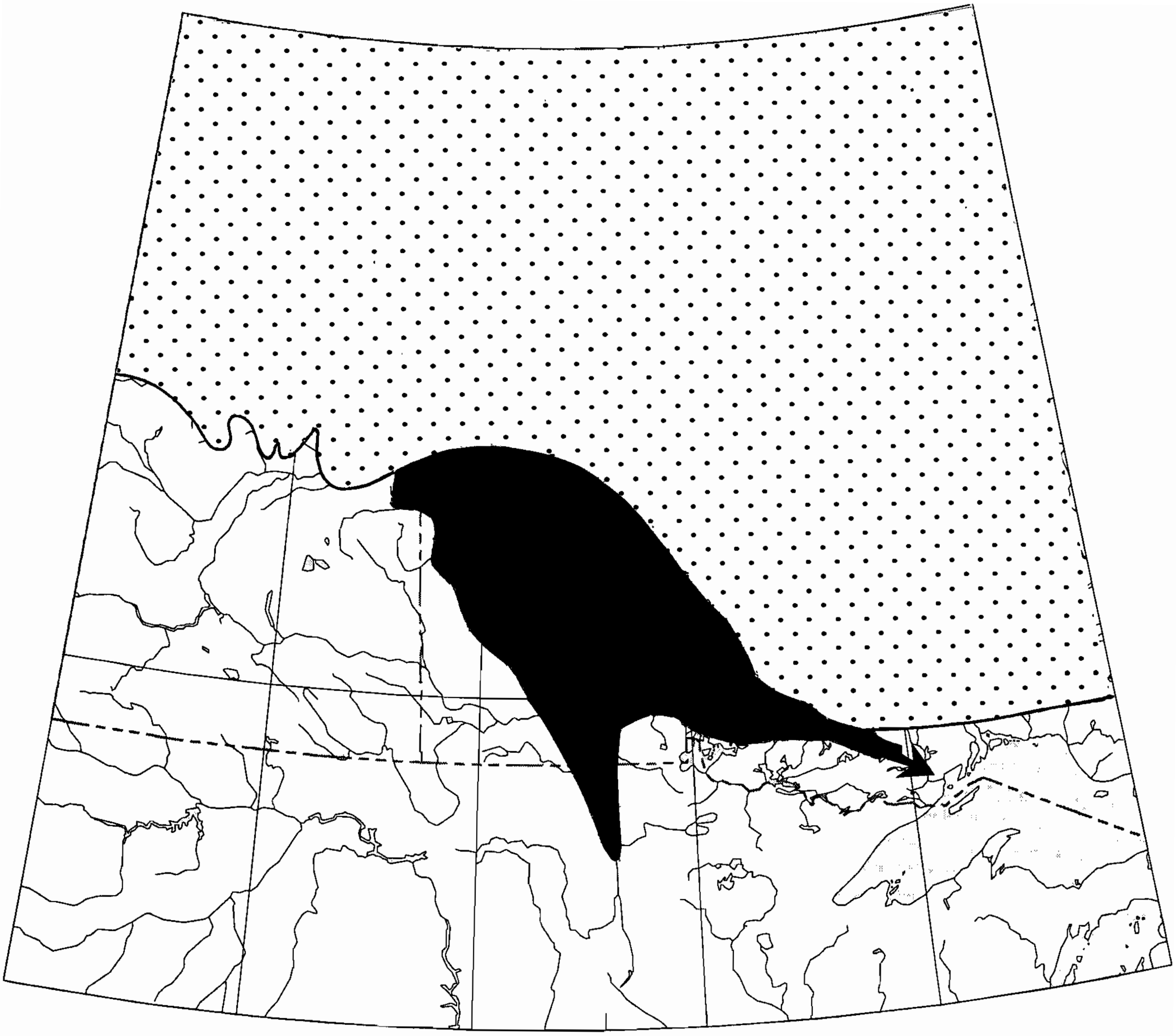

Figure 35. ***Lake Agassiz region paleogeography at ~10.4 ka BP:*** *An ice margin north of Prince Albert, Saskatchewan (Christiansen, 1979) is correlated to a later overridden position in Manitoba, the construction of the Hartman moraine which was dated to 10.4 ka by Warman (1991) on the basis of 440 varves below the red clay marker at Dryden, and to an undetermined position north of Lake Superior. Differential uplift of the Savanne outlet caused Lake Agassiz to transgress from Grand Forks to south of Fargo. By late Moorhead time, differential uplift of the Thunder Bay outlets would have drowned sites in southeastern Manitoba which now yield subaerial organic material dated to late Moorhead time (Matile and Thorleifson, 1996). It therefore is apparent that the Clearwater spillway in western Saskatchewan must have been the late Moorhead outlet of Lake Agassiz, as suggested by Fisher and Smith (1994). Opening of this outlet again lowered Lake Agassiz, and a delta was deposited in the Fargo area at this time (Arndt, 1977; Fenton et al., 1983).*

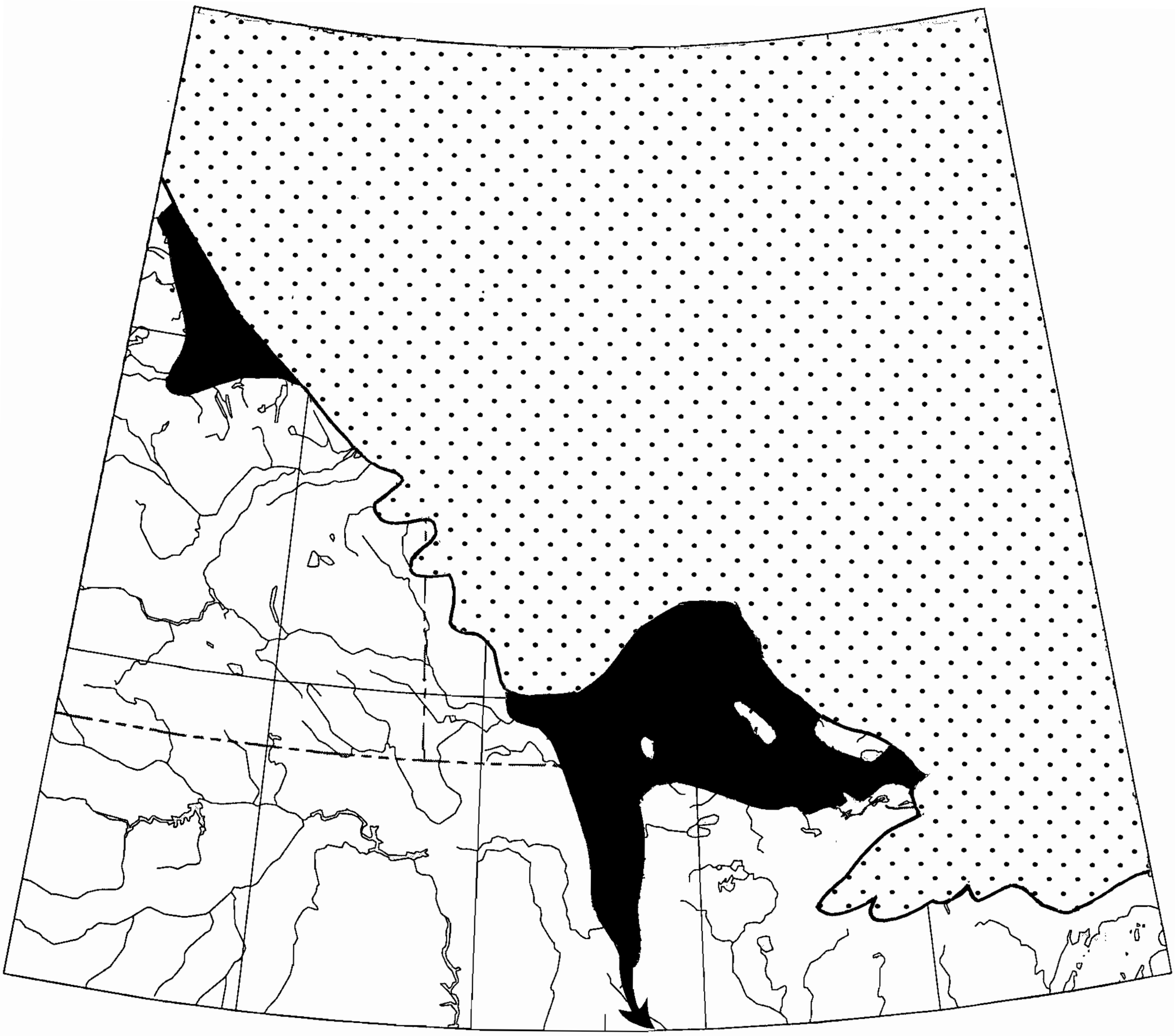

Figure 36. ***Lake Agassiz region paleogeography at ~9.9 ka BP:*** *A major readvance at ~10 ka returned drainage to the southern outlet, causing deposition of Sherack Formation clay. The distribution of lacustrine sediments indicates that the rise reached the Norcross level, depositing fills in the Assiniboine, Pembina, and Sheyenne valleys. To permit a rise to Norcross, ice must have blocked access to the Clearwater spillway, but this feature was not overridden. The ice margin was at the southwestern limit of southwestward ice flow in Saskatchewan (Prest et al., 1968), the Arran limit in eastern Saskatchewan (Moran, 1969) and western Manitoba (Klassen, 1979), a buried moraine at Portage la Prairie, Manitoba (Fenton, 1970; Fenton and Anderson, 1971), the Lac Seul moraine in northwestern Ontario (Warman, 1991), the Marks moraine near Thunder Bay, and the Marquette maximum in Lake Superior (Hughes, 1978; Attig et al., 1985). The final Herman level was not reoccupied due to differential uplift. Lake Agassiz was confluent with Lake Kaministikwia in the Thunder Bay area, and glaciolacustrine sediments were deposited across the drainage divide. Red clay was transferred into Lake Agassiz from the Kaministikwia basin for two to three decades (Warman, 1991). As the southern outlet was eroded, the Norcross and Tintah shorelines were formed. As resistant Precambrian rocks were exposed in the floor of the outlet, Lake Agassiz stabilized at the Campbell level and slowly regressed due to differential uplift. Existence of the lower Campbell shoreline, which diverges from the upper Campbell northward, implies that the Kaiashk outlet near Lake Nipigon temporarily opened late in the Emerson Phase.*

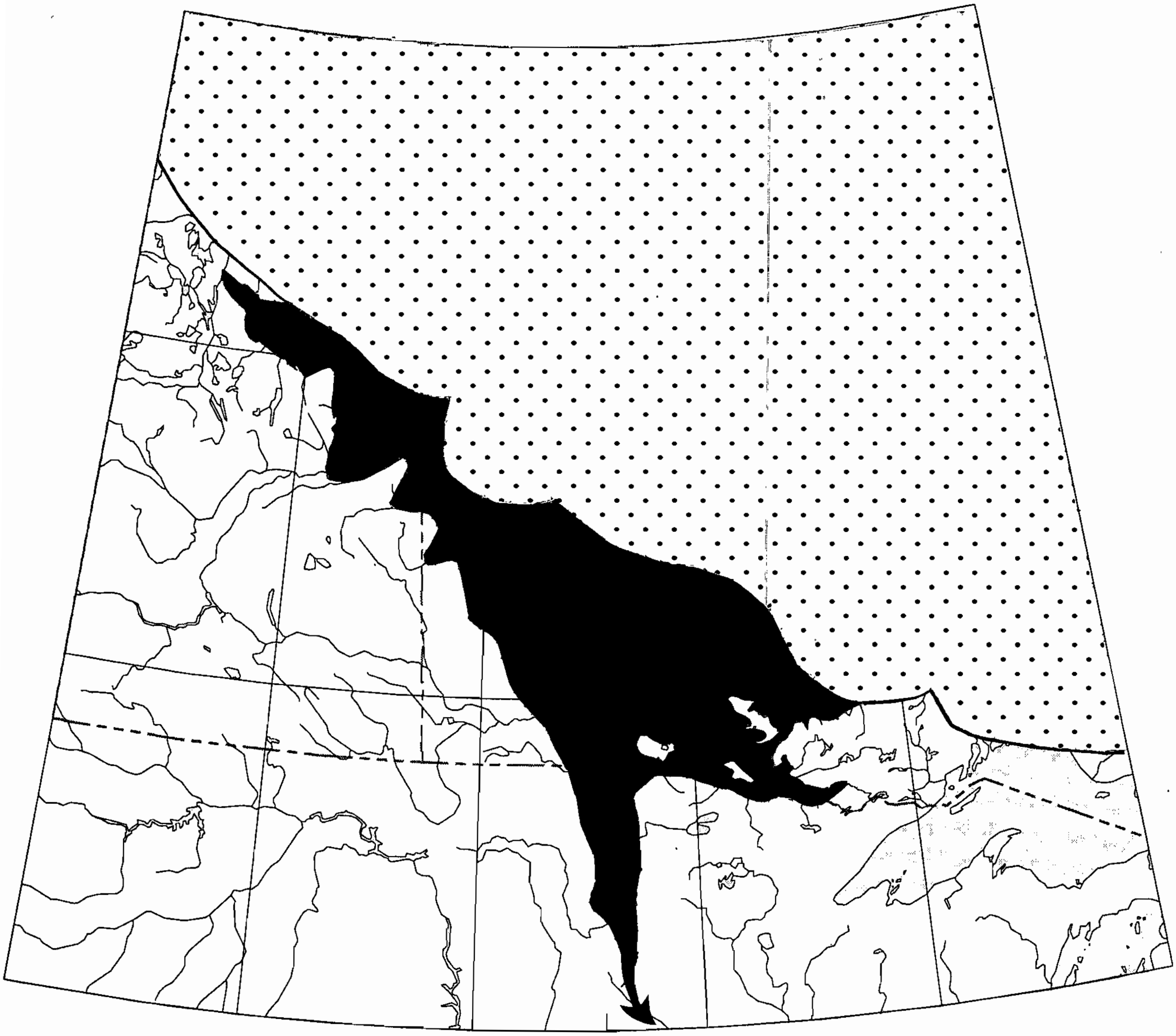

Figure 37. ***Lake Agassiz region paleogeography at ~9.3 ka BP:*** *A retreating position of the ice margin in Saskatchewan is correlated to the Pas and George Island (Todd and Lewis, 1996) moraines in Manitoba, the Sioux Lookout moraine in northwestern Ontario, and the southern portion of the Nipigon moraine, and the North Shore ice margin of Lake Superior (Farrand, 1960). The advance to this margin at Sioux Lookout overrode varved clay (Hurst, 1933) and closed the Kaiashk outlet. Lake Agassiz rose to the southern outlet, which produced the lower Campbell shoreline due to differential uplift since abandonment of the Campbell. A fossiliferous valley fill was deposited by the Assiniboine, Pembina, and Sheyenne Rivers. Correlative fill at Rossendale was dated at 9.5 ka by Teller (1989). As ice retreated from the Sioux Lookout moraine, a succession of outlets that drained into Lake Nipigon, and which were described by Elson (1957), Zoltai (1967), and Teller and Thorleifson (1983), were opened. Lake Agassiz was lowered in step-wise manner, although every lowering was followed by slow transgression everywhere south of the outlet isobase due to differential uplift of the outlet. Consequently, these shorelines converge to the south.*

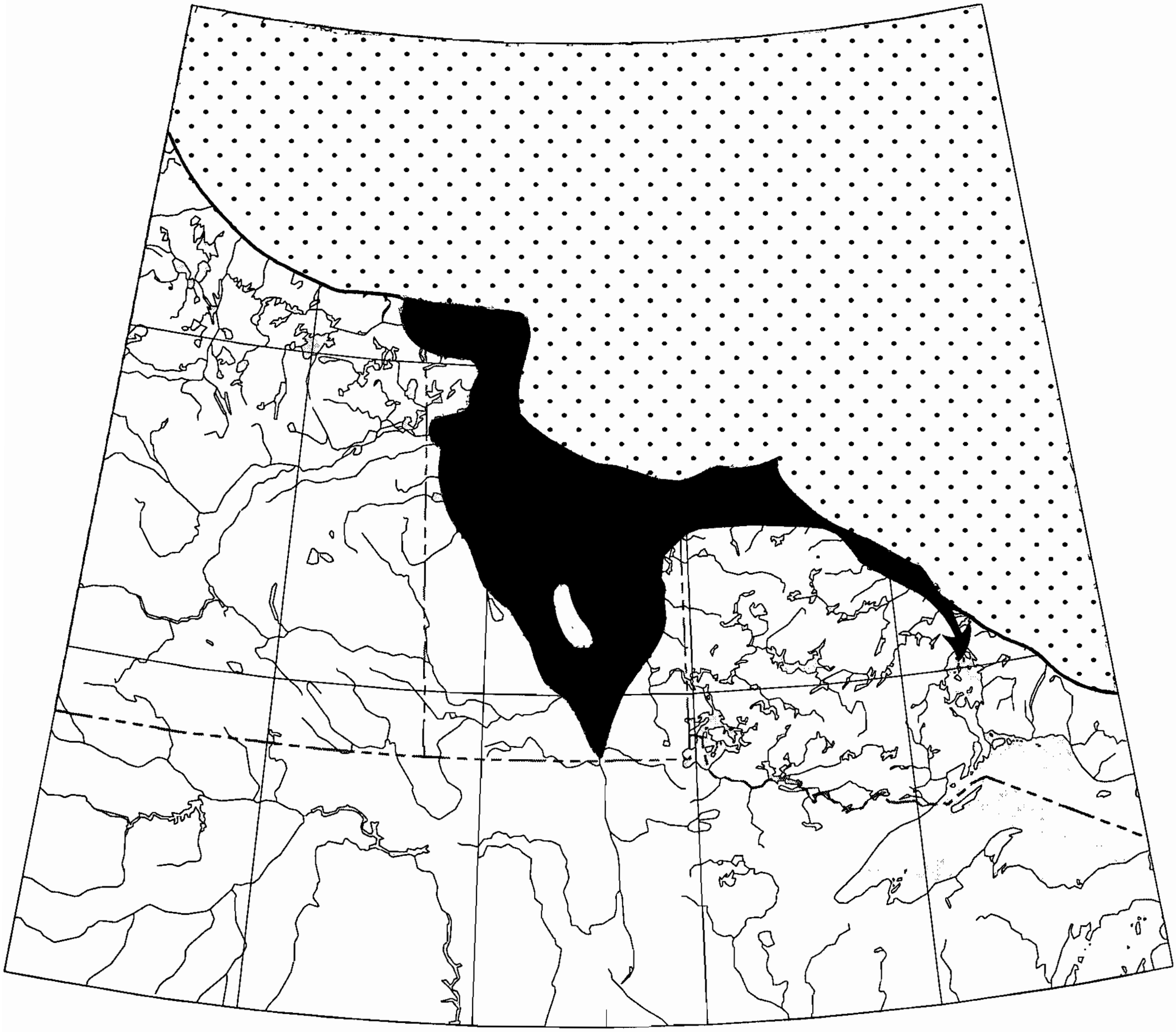

Figure 38. ***Lake Agassiz region paleogeography at ~8.2 ka BP:*** *The final major moraine-building episode occurred as the Cree Lake moraine in Saskatchewan (Schreiner, 1984), the Hargrave moraine in Manitoba (Klassen, 1983), and the Agutua-Nakina moraines in Ontario (Prest, 1963; 1970) were deposited. Deposition of the Agutua-Nakina moraines by enhanced glaciofluvial discharge was dated at about 8.2 ka by Thorleifson and Kristjansson (1993), on the basis of a Lake Superior varve record reported by Teller and Mahnic (1988). Construction of these moraines coincided with the final episode of discharge to Lake Nipigon, from the Pas and possibly Gimli levels of Lake Agassiz.*

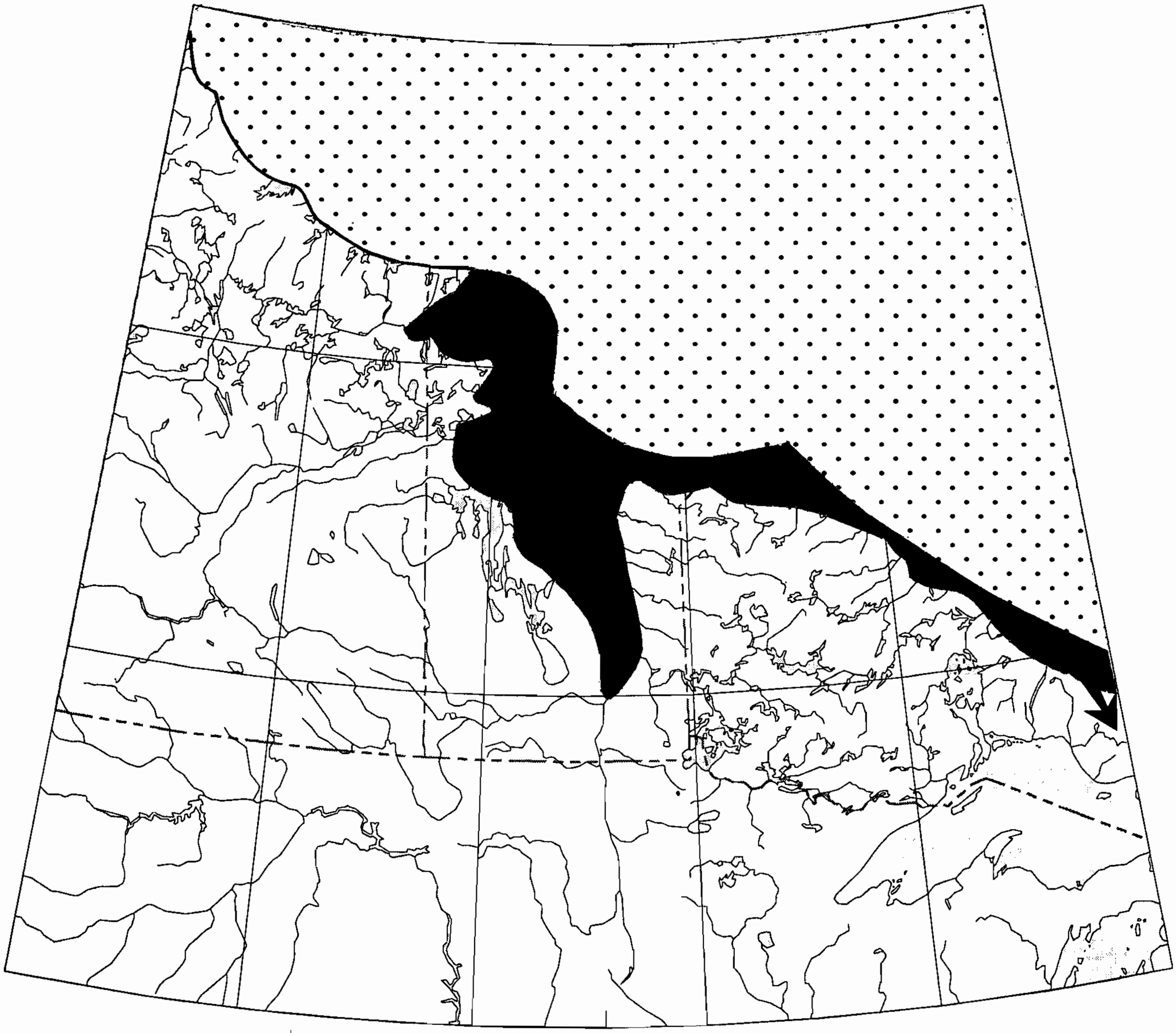

Figure 39. ***Lake Agassiz region paleogeography at ~8.0 ka BP:*** *As the ice margin retreated from the 8.2 ka margin, no major moraines were built. In fact, the ice mass seems generally to have been stagnant, because eskers and lineated till features related to the 8.2 ka margin are virtually undisturbed. Small moraine segments were built at Sipiwesk in Manitoba (Klassen, 1983) and Big Beaverhouse in Ontario (Prest, 1963). As ice retreated in Ontario, discharge may have been diverted to outlets directly entering Lake Superior at White Otter and Nagagami, producing the Grand Rapids and Drunken Point shorelines of Lake Agassiz. If the ice margin in Saskatchewan was farther north than depicted here, or if the margin retreated and readvanced, it is possible that Lake Agassiz drained to the Arctic Ocean via Wollaston Lake, at about this time.*

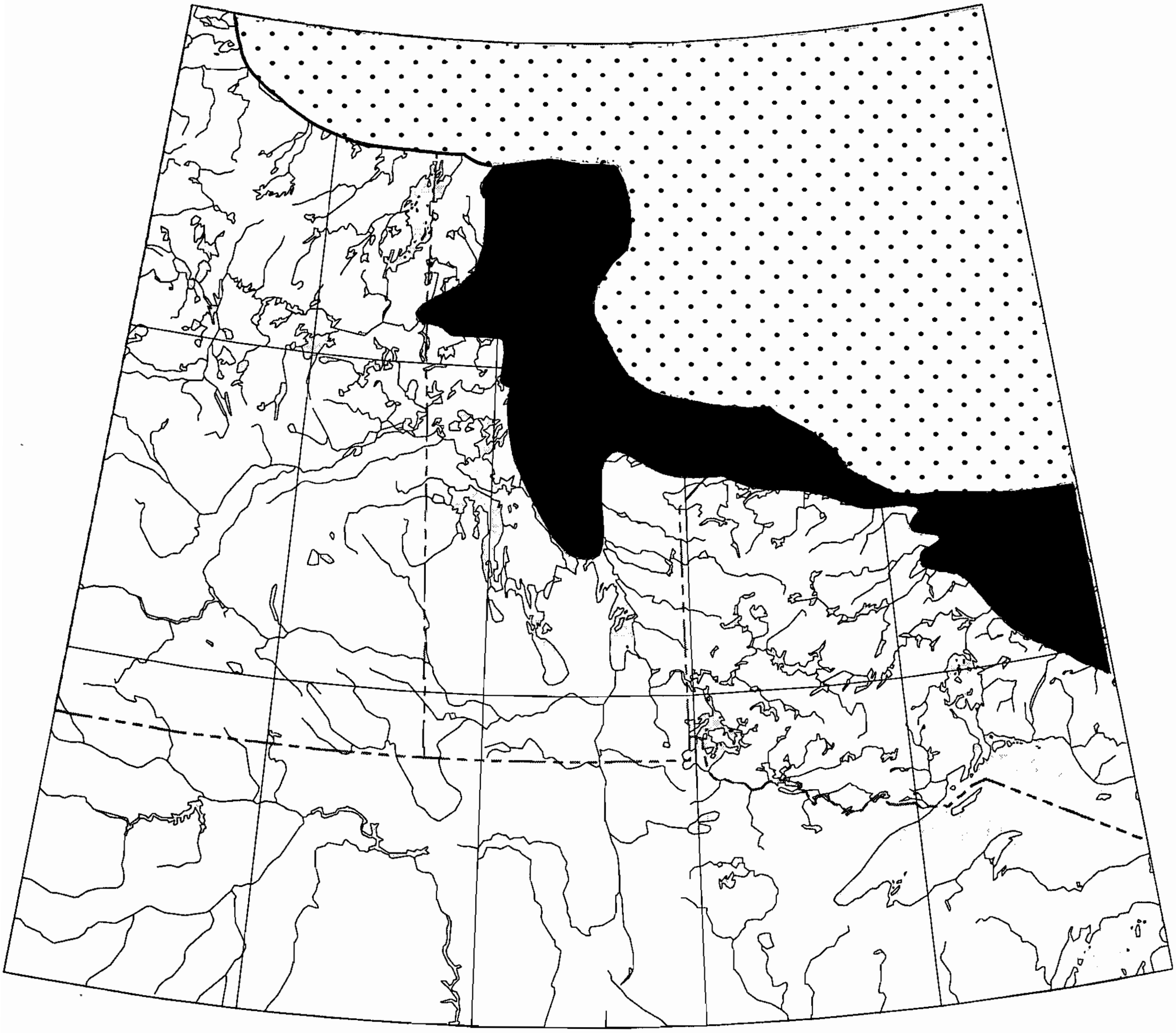

Figure 40. ***Lake Agassiz region paleogeography at ~7.8 ka BP:*** *As the margin of the stagnant ice mass retreated by calving, to an undetermined retreat position, Lake Agassiz fell to the Ponton level, confluent with the late Kinojévis level of Lake Ojibway, which drained to the Ottawa River. Due to tilting of the basins, relative to present, the southern portions of major lakes with outlets to the north, such as Lake Winnipeg, Lake Manitoba, and Lake of the Woods, were dry. Lake Nipigon, with an outlet to the south, was more extensive than present.*

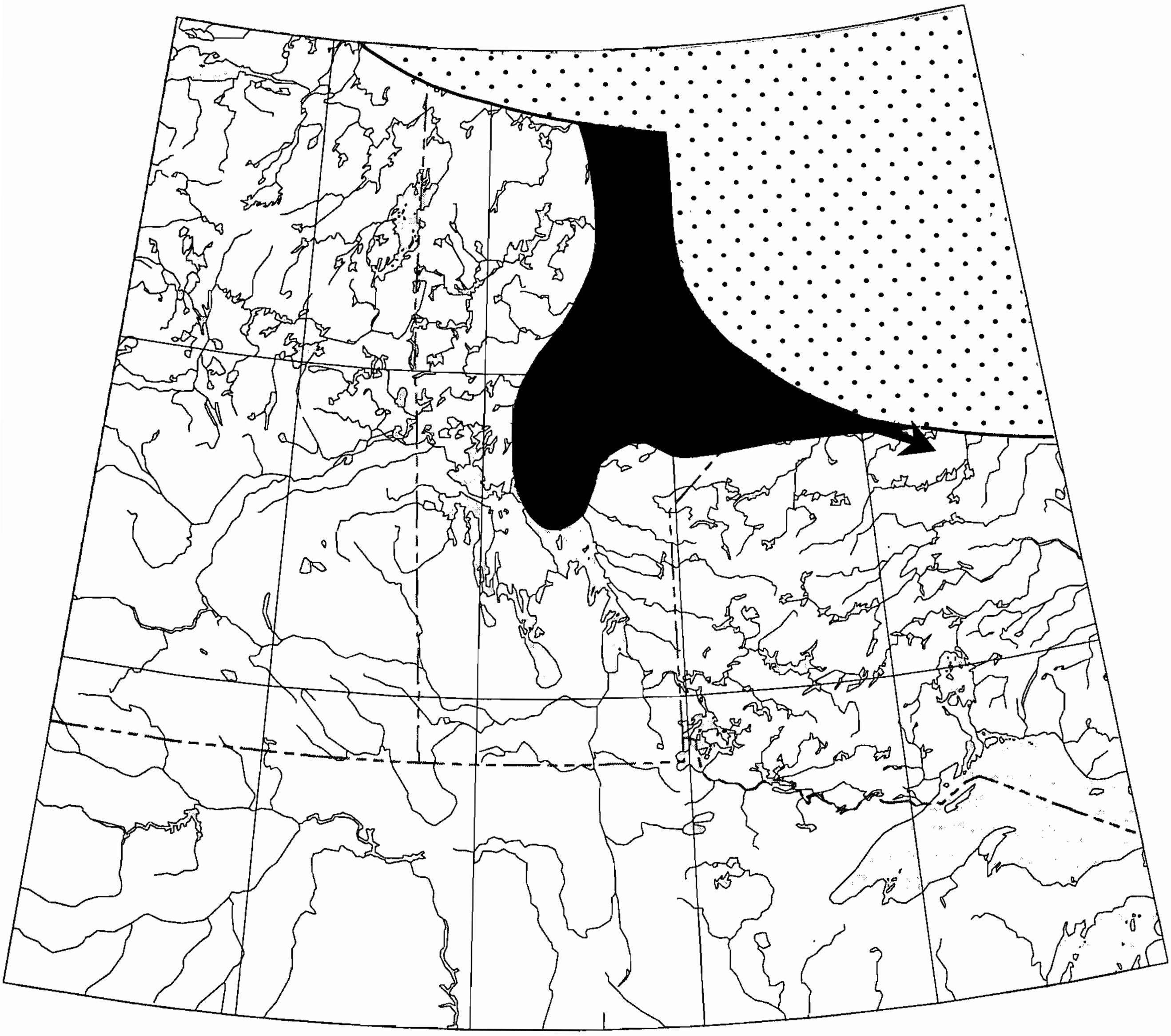

Figure 41. ***Lake Agassiz region paleogeography at ~7.7 ka BP:*** *The existence of the Fidler level of Lake Agassiz (Klassen, 1983), which is lower than the final outlet to the Ottawa River, indicates that Lake Ojibway must have drained to Hudson Bay, and marine water entered the James Bay Lowland, while Lake Agassiz was still in existence. Hence there must have been an outlet from the Fidler level to James Bay. As the ice retreated, Lake Agassiz was completely drained to the east. Due to the head available, the Echoing River channels were scoured to a depth below sea level. Sediments carried eastward and deposited along the upper Fawn River in Ontario by this discharge were dated at 7400 BP (GSC-877) by McDonald (1969).*

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HISTORY OF LATE GLACIAL RUNOFF ALONG THE SOUTHWESTERN MARGIN OF THE LAURENTIDE ICE SHEET

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Abstract — The routing of runoff from the retreating Laurentide Ice Sheet was controlled by a combination of ice-marginal position and topography; isostatic depression also played an important role in the evolution of the ice-marginal landscape. A major redirection of drainage occurred after ice retreated north of the Missouri Escarpment about 12 ka BP. Runoff that had previously flowed south to the Missouri River could then be routed eastward to Lake Agassiz, which in turn drained to the upper Mississippi River valley.

Changes in drainage routing in the region west of glacial Lake Agassiz typically were abrupt, as new and lower paths along the retreating ice margin became available. Each new basin was larger than its predecessor and its headwaters extended farther west. Four distinct, sequential phases of ice-marginal drainage occurred between 12 and 10.7 ka BP: the James, Sheyenne, Souris–Pembina and Qu'Appelle–Assiniboine. Starting with the Qu'Appelle–Assiniboine phase about 11.2 ka BP, Lake Agassiz collected runoff from as far west as the Rocky Mountains. By about 10.7 ka BP, ice had retreated north of the lower reaches of the Saskatchewan River, allowing meltwater to by-pass the Qu'Appelle–Assiniboine basin. At about the same time, the eastern outlets of Lake Agassiz opened, initiating drainage to the North Atlantic Ocean from this vast region.

Drainage routes evolved by a common sequence of events, beginning with the impoundment of proglacial lakes along the ice margin. Most of these lakes drained catastrophically, resulting in the formation of spillways with distinctive geomorphic features that include broad, scoured and streamlined subupland channels along with subsequently incised narrow, deep inner channels. Characteristic deposits of these outbursts include boulder mantles on subupland channel floors, boulder-gravel bars within the inner channels, and large, primarily subaqueous fans in lake basins that received the flood bursts. These outbursts progressed from basin to basin, causing the drainage of other proglacial lakes, until they reached Lake Agassiz. Spillways associated with the next younger drainage phase were commonly incised across the abandoned lake floors.

INTRODUCTION

As the Laurentide Ice Sheet wasted northward through the Interior Plains of the U.S. and Canada, meltwater from the ice flowed in rapidly shifting, ice-marginal drainage basins composed of deeply incised spillways connecting proglacial-lake basins. Landforms and deposits associated with these drainageways reflect the storage and rapid release of meltwater from proglacial lakes.

In the Interior Plains, widespread stagnation and downwasting followed numerous, rapid advances of lobes of the Laurentide Ice Sheet (Clayton *et al.*, 1980; Clayton and Moran, 1982; Klassen, 1989; Christiansen, 1979). Ablation of the stagnant ice was delayed by supraglacial debris that formed a discontinuous cover over the ice. As the ice melted, isolated water-filled depressions coalesced into ice-marginal lakes. Because of the regional northward slope, accentuated by differential isostatic depression, lakes were ponded against the active ice margins to the north at successively lower elevations during deglaciation.

Drainage of these proglacial lakes occurred periodically and abruptly as marginal barriers of ice or sediment failed, often because of large influxes of meltwater, allowing drainage in a matter of weeks or months. These outbursts commonly modified existing drainageways or formed new spillways with distinct characteristics. Discharge from the lakes flowed in spillways that led to lower proglacial lakes to the east and south. From the inception of glacial Lake Agassiz at about 11.7 ka BP until its eastern outlets opened about 10.7 ka BP, proglacial lakes east of the Missouri Escarpment drained in a west to east sequence that ended at Lake Agassiz. During this time interval, Lake Agassiz overflowed through its southern outlet into the Mississippi River valley and ultimately to the Gulf of Mexico before it abruptly shifted to overflow through its eastern outlets to the Great Lakes (e.g. Fenton *et al.*, 1983). The purpose of this paper is to describe the major meltwater drainageways that formed along the ice margin and to summarize their chronological development, as it is currently understood, during this period of time.

ICE-MARGINAL LAKE DRAINAGE AND GENERAL OVERVIEW OF SPILLWAY GEOMORPHOLOGY

The geomorphology of proglacial lake spillways in this and adjacent regions has been described by Kehew and Lord (1986, 1987, 1989). The history of deglaciation in this region has been summarized by Klassen (1972, 1975, 1989), Christiansen (1979) and Clayton and Moran (1982). Many spillways in the Interior Plains were originally shallow valleys containing braided meltwater streams that headed near an ice margin prior to their entrenchment by ice-marginal lake drainage (Fig. 1A). The entrenchment phase of many spillways occurred rapidly due to the release of erosive outbursts from the lakes. During the initial phase of an outburst from an ice-marginal lake, the flow covered the broad, shallow valley floor and at times spilled out of its channel and over drainage divides, unless the existing valley was deep enough to contain the flow. Where no valley was present, the initial outflow from the lake scoured broad shallow tracts as much as 10 km wide.

As the lake continued to drain, deep incision ensued, forming a relatively straight inner channel 1–2 km wide and 30–100 m deep (Fig. 1B). Remnants of the initial phase of lake drainage were left as scoured uplands flanking the deeply incised inner channel. Remnants of cross-bedded outwash deposits also may remain at high levels along one or both sides of the newly formed channel if the spillway had an early braided-stream phase. Landforms and sediments characteristic of the scoured upland surfaces formed during the early part of the outburst (Fig. 2) and include erosionally streamlined longitudinal hills, longitudinal grooves, potholes, erosional and depositional megaripples and boulder mantles (Lord and Kehew, 1986, 1987; Lord *et al.*, 1992; Sun, 1993). With the exception of Holocene alluvium, deposits within the deep inner channels are scarce, but where they do occur, consist of poorly sorted, very coarse-grained bars or bedrock terraces veneered by coarse gravel suggestive of high discharge, hyperconcentrated flow (Lord and Kehew, 1987). Paleohydraulic estimates for these outburst discharges, based on slope-area and competence techniques, range between 1×10^4 and 1×10^6 m^3/sec (Lord and Kehew, 1987; Komar, 1989; Kehew and Teller, 1994).

The specific causes for the drainage of proglacial lakes by the abrupt release of large slugs of water will be discussed for individual spillway networks, but in most cases, appear to relate to the failure or breaching and rapid erosion of their confining margins. The morphology of these valleys differs from normal glacial outwash streams, which are aggradational systems, and from other

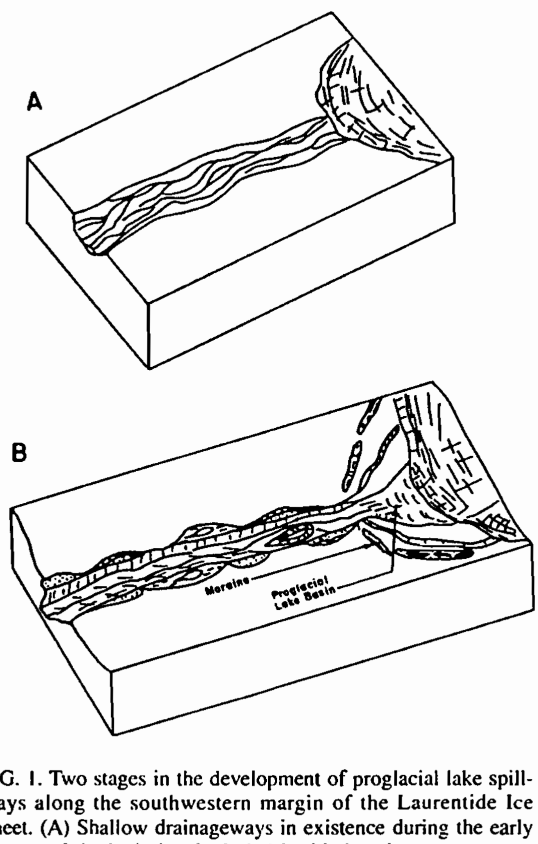

FIG. 1. Two stages in the development of proglacial lake spillways along the southwestern margin of the Laurentide Ice Sheet. (A) Shallow drainageways in existence during the early phases of deglaciation included braided meltwater streams, subglacial channels and older valleys that were not completely buried by the subsequent ice advance. (B) Proglacial lakes formed along the ice margin due to isostatic depression of the crust and topographic barriers of stagnant ice or morainal deposits. As the lakes grew during wasting of the ice sheet, the outlets were abruptly deepened during outburst floods and spillways downstream were deeply incised. Remnants of the former valleys were left as high-level terraces.

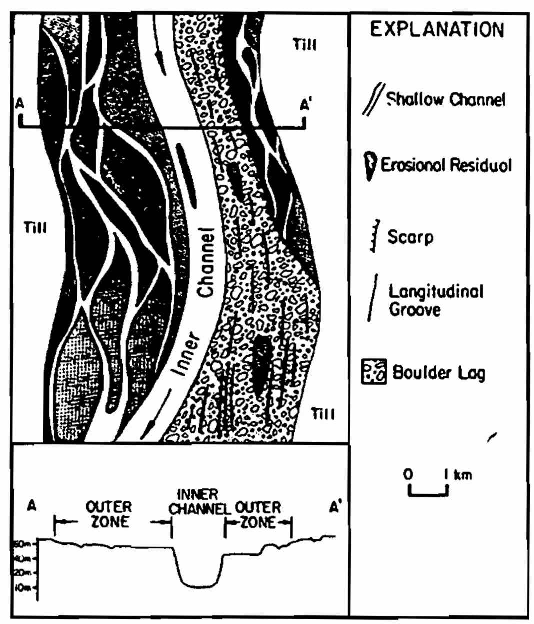

FIG. 2. Model of spillway geomorphology. When existing valleys were too small to contain outburst discharges, these floods scoured broad subupland areas (outer zone), forming streamlined erosional residuals, longitudinal grooves and boulder mantles. Incision of the spillway formed the inner channel, which was cut below the subupland surface. (From Kehew and Lord, 1986.)

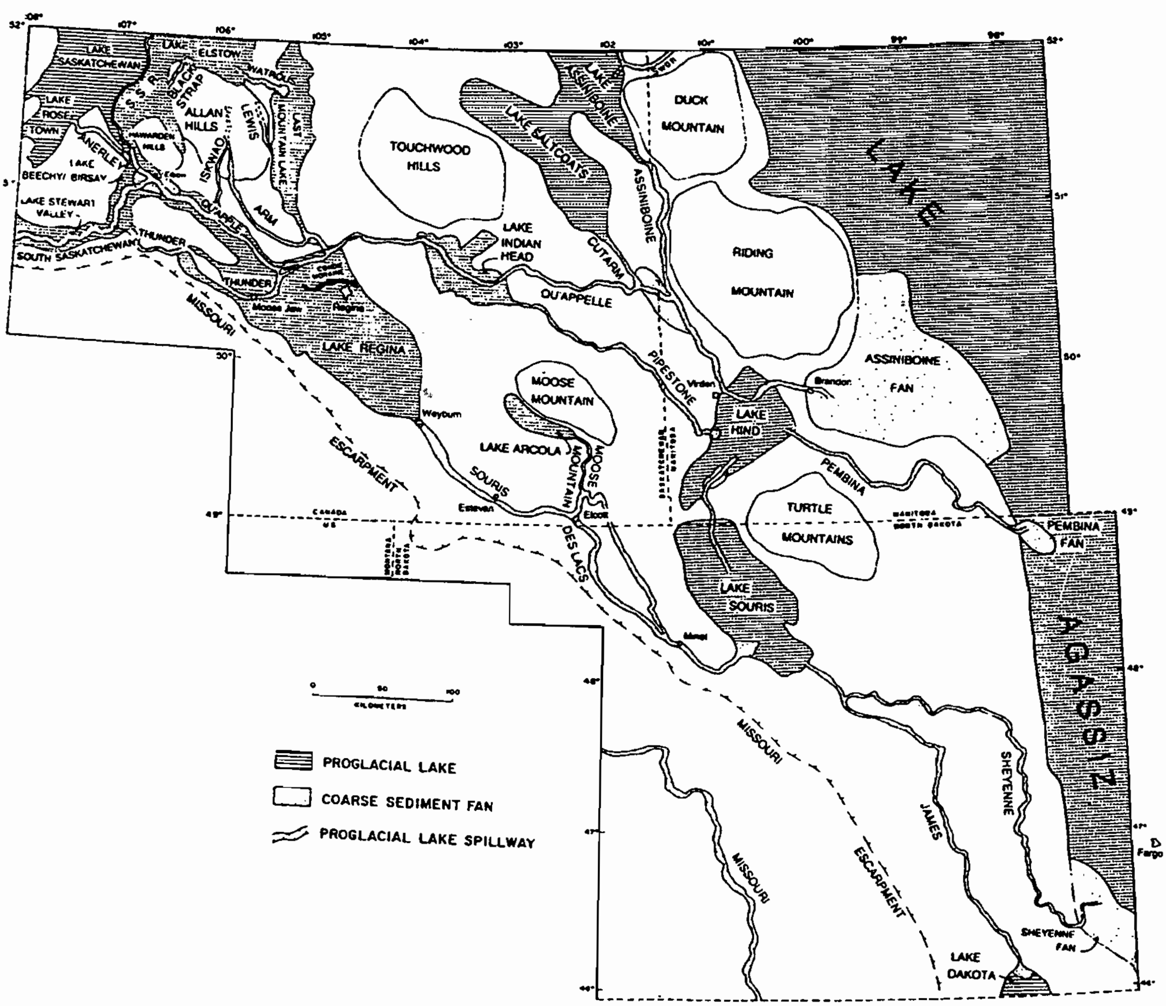

FIG. 3. Map of area discussed in this paper, showing prominent features of drainage evolution. Area of coarse-grained fans shown in proglacial lake basins west of Lake Agassiz is limited to gravelly facies at head of fan.

fluvial systems in which downcutting occurs by headward erosion due to base-level lowering. In the latter case, the valley becomes progressively smaller upstream and the cross-valley profiles are much less regular and commonly asymmetric (Kehew and Lord, 1986). Glacial-lake spillways in this region, by contrast, maintain a relatively constant size and shape from their head at the lake outlet to their mouth.

Proglacial lakes commonly drained sequentially from upstream to downstream, in a domino-like reaction to an initial lake outburst (Kehew and Clayton, 1983; Kehew and Lord, 1986, 1987). In fact, most proglacial lake basins have large coarse-grained sediment fans on their western or northern (upstream) sides at the mouth of the main channel entering the basin (Fig. 3); the bulk of the sediment in these fans is stratigraphically related to the last lacustrine event in that lake basin. Thus, a cause and effect relationship seems likely: a flood burst into the proglacial lake basin deposited the fan and simultaneously increased the discharge from the lake outlet, which rapidly scoured and incised the outlet and the spillway downstream from the lake (Kehew and Clayton, 1983; Lord, 1991). The morphological similarity of spillways throughout the region suggests that the events described above shaped most large valleys and were common and widespread along the southwestern margin of the Laurentide Ice Sheet.

The occurrence of terrace remnants in some spillways provides evidence for multiple outburst floods. The fact that some trunk spillways such as the Qu'Appelle (Figs 3 and 4) have several flood-scoured tributaries that head in different proglacial lake basins supports the likelihood that multiple flood events passed through the spillway to Lake Agassiz. For spillways that have a more complicated history, it is likely that only the first outburst was involved in the formation of the scoured uplands. Later floods probably were confined to the entrenched inner channel and resulted in a further deepening and enlargement. Slope-area discharge estimates for spillways of this compound type, therefore, would overestimate the discharge achieved during any single flood.

Proglacial lake basins along the southwestern margin

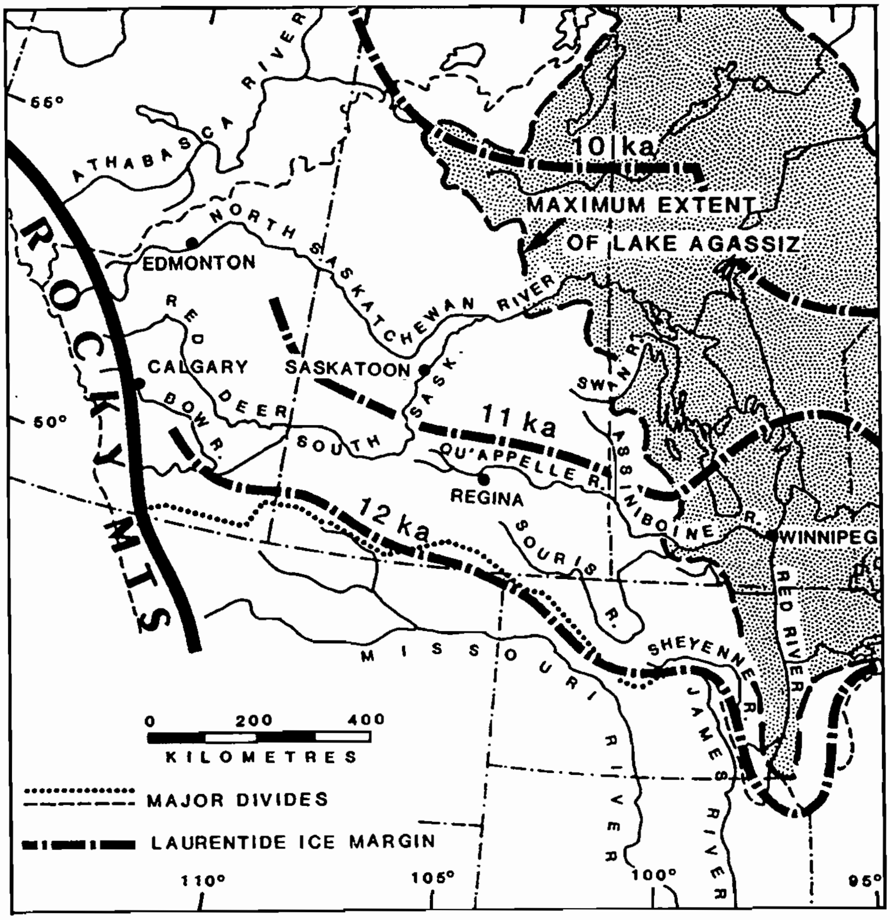

FIG. 4. Generalized ice-marginal positions of the southwestern margin of the Laurentide Ice Sheet between Lake Agassiz and the Rocky Mountains showing relationships with major drainage routes formed during deglaciation. Missouri Escarpment (dashed) roughly coincides with the 12 ka ice margin along much of its eastern extent.

of the Laurentide Ice Sheet contain a diverse suite of landforms and sediments. Coarse-grained fans, which occur at the mouths of spillway inlets, include subaerial deposits as well as subaqueous underflow fan deposits. Although fine-grained offshore lake sediment occurs at the surface near the center of some basins, the floors of many are covered by areally extensive deposits of thick, homogenous sand that resulted from flow expansion of incoming outburst floods containing large volumes of coarse sediment eroded from the spillway leading to the lake. Glacial Lake Souris (Fig. 3), which contains the best known fan of this type (Lord, 1991), illustrates the effect of such a flood on a proglacial lake basin. The Lake Souris fan, which covers nearly half the basin, contains only the sand and coarser fraction of the flood sediment (approximately 8 km^3), even though fine sediment eroded from till and Tertiary sediment in the spillways must have constituted the bulk of the total load. The lack of the silt and clay component indicates that the incoming flood initiated downcutting of the lake's outlet and led to drainage of the entire lake before the fine sediment could settle out of suspension (Lord, 1991). Where there is a total absence of fan or lacustrine deposits younger than the main fan in these basins, it indicates that the lake was drained at the time of spillway incision. Subsequent runoff into the basin eroded channels across the basin floor, some of which have characteristics that are typical of flood-eroded channels. In many basins, these channels do not lead across the lake floor to the former outlet through which drainage occurred, but are eroded instead across the northern side of the basin toward lower terrain that must have been blocked by ice at the time of lake drainage. The Lake Regina, Lake Indian Head, Lake Assiniboine and Lake Hind basins are good examples of this relationship (Fig. 3).

The elevational relationships between lake outlets are consistent from basin to basin; southern outlets are higher in elevation than are northern outlets. Greater isostatic depression of the crust at the time of glaciation in the northern ends of these basins would have produced an even greater relative difference in outlet elevations. Thus, when these lakes overflowed through their southern out-

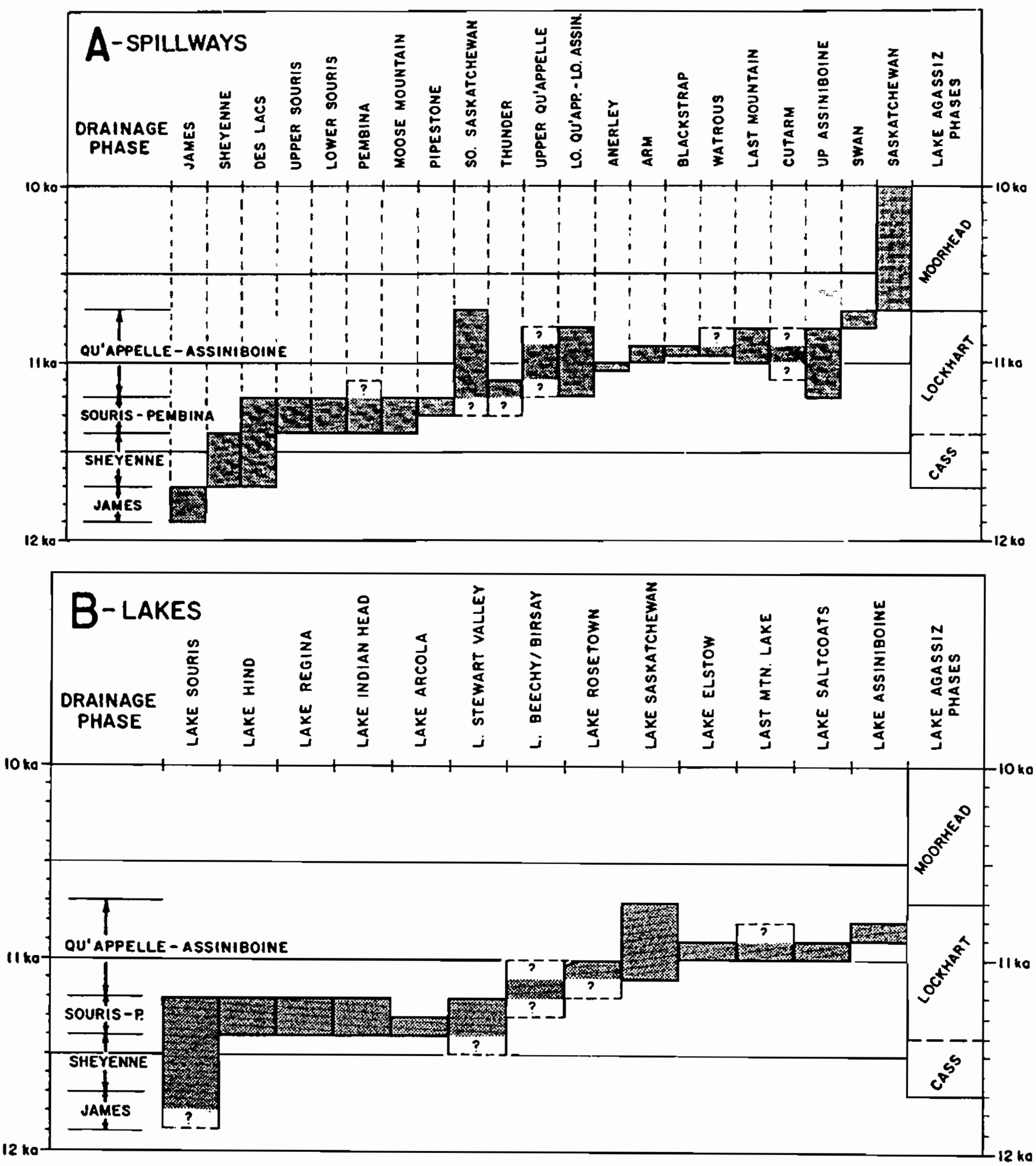

FIG. 5. Chronology of meltwater flow through various proglacial spillways (A) and the life span of related proglacial lakes (B) in the region west of Lake Agassiz. The scale is subdivided into named drainage phases; Lake Agassiz phases are also shown.

lets, they must have been impounded by ice on their northern margins. Only when the ice to the north had thinned or receded could northern outlets have developed.

Large sediment fans were also deposited in Lake Agassiz at the mouths of drainage systems that carried outflow discharges from the lakes to the west. Three major fans, the Sheyenne, Pembina and Assiniboine (Fig. 3), accumulated during the very short periods (200–500 years; Fig. 5) when their drainage basins carried runoff from the Laurentide ice margin to Lake Agassiz during its Cass and Lockhart phases. As concluded by Kehew and Clayton (1983), Fenton *et al.* (1983) and others, we believe that the bulk of these fans were deposited as a result of flood bursts from upstream proglacial lakes during these active periods of runoff.

The effects of meltwater outbursts on Lake Agassiz itself from proglacial lakes to the west are not well understood. Landforms in its southern overflow spillway, the Minnesota River valley, are characteristic of high-discharge erosive flows (Matsch, 1983; Kehew and Lord, 1986) and it is possible that they formed as a result of incoming floods from abruptly drained proglacial lakes along the southwestern margin of the Laurentide Ice Sheet which, in turn, led to an abrupt increase in outflow from Lake Agassiz. However, the volume of inflow from these relatively small lakes to the west may have had little impact on the outflow from Lake Agassiz because these inflows would not have raised the level of this giant lake significantly, and therefore, would not have changed the discharge at the upstream end of the Minnesota River spillway. This as well as the fact that the floor of the southern outlet is on resistant igneous and metamorphic rocks explain why Lake Agassiz did not drain completely during these high-discharge events as did smaller lakes to the west.

HISTORY OF DRAINAGE DEVELOPMENT

Proglacial drainage phases defined in this paper are intervals during which meltwater was routed away from the ice sheet through spillways and proglacial lakes north of the Missouri Escarpment. In all but the initial phase, these waters flowed eastward to glacial Lake Agassiz, which overflowed into the headwaters of the Mississippi River. Delineation of the northern boundary of the drainage network that was active during each phase is based on topographic and geomorphic criteria. Chief among these is the blockage of lower, more northerly drainage routes by the Laurentide Ice Sheet. The northern boundary of each system mapped thus serves as a proxy ice-margin indicator, which is not necessarily based on moraines and other traditional ice-margin indicators.

A steep, isostatically-induced regional slope toward the north was very important throughout the evolution of the drainage networks, and old spillways to the south were episodically abandoned for new routes along the retreating ice margin. After one system was abandoned in favor of a lower, more northerly one, old spillways were occupied by non-glacial streams that evolved into the modern drainage system.

The history of ice-marginal drainage basin evolution discussed in this paper begins just after 12 ka BP, when ice retreated north of the Missouri Escarpment, and ends after 11 ka BP, when ice retreated north of the lower Saskatchewan River valley (Figs 4 and 5).

During the zenith of the Wisconsinan glaciation, ice had pushed southward across the Missouri Escarpment to and beyond the Missouri River (Fig. 4). Meltwater drainage was down the Missouri to the Mississippi River at this time. As ice wasted back north of the Missouri Escarpment, a fundamental change in drainage took place. The Escarpment was established as a drainage divide that ultimately routed meltwater east toward the Lake Agassiz basin and to the Mississippi River rather than through the Missouri River basin.

A series of ice-marginal drainage basins formed north of the Missouri Escarpment as Laurentide ice retreated downslope. The relationship of ice-margin configuration and land-surface morphology determined the location and extent of proglacial lakes and spillways in the newly deglaciated or stagnant-ice terrain. Only when this water was able to shift to lower routes — made available by the downslope retreat of the ice margin — were drainage systems abandoned by meltwater.

Drainage north of the Missouri Escarpment can be divided into four phases (Fig. 5), each represented by an ice-marginal drainage network of lake basins and connecting spillways. Lower parts of older drainage systems were beheaded and abandoned as the younger systems integrated new drainageways to the north. The active periods of the proglacial-lake spillways and associated lake basins are chronologically tabulated in Fig. 5, which also shows the phases in the history of Lake Agassiz. Figures 6–8 show the northward shifting drainage systems that carried (or impounded) meltwater as the Laurentide Ice Sheet retreated downslope. All but the first of these systems discharged into Lake Agassiz.

Lake Agassiz came into existence in the lowlands of the Red River valley about 11.7 ka BP (Fenton *et al.*, 1983), as ice retreated north of the divide between the Missouri–Mississippi River drainage basin and the northward-sloping Hudson Bay basin. To the west, ice retreat allowed runoff from the ice to be routed between the Missouri Escarpment and the ice margin and then to overflow into the Missouri River basin. Within the area discussed in this paper, the James spillway (Fig. 6), which is (and was) tributary to the Missouri River (Clayton and Moran, 1982), served as the overflow route for about 200 years (Fig. 5). This interval represents the last overflow from the southwestern Laurentide Ice Sheet to the Missouri River basin (cf. Dyke and Prest, 1987; Clayton and Moran, 1982; Klassen, 1989).

As ice continued to retreat, Lake Agassiz began to develop and expand northward during the Cass phase. A lower route for runoff from the southwestern margin of the ice sheet was opened through the Sheyenne spillway (Fig. 7), the next spillway to the north. Most of the Sheyenne fan was deposited in Lake Agassiz at this time (Brophy and Bluemle, 1983). After a brief re-advance of the Red River Lobe into the Lake Agassiz basin at about 11.4 ka BP (Clayton and Moran, 1982; Fenton *et al.*, 1983), drainage expanded downslope again; this time overflow shifted from the Sheyenne to the Souris–Pembina drainage system (Figs 5 and 8). This basin drained to the Pembina fan in Lake Agassiz during the Lockhart phase. The next drainage change initiated flow through the Qu'Appelle–Assiniboine ice-marginal basin (Fig. 3), which extended west to the Rocky Mountains (Fig. 4). This routing continued until ice retreated far enough north to allow the headwaters of the Qu'Appelle River west of about 107° at Elbow, Saskatchewan (Fig. 3), to be routed north, establishing the modern South Saskatchewan River from the Rocky Mountains to the Lake Agassiz basin (Fig. 4). Deposition of the Assiniboine fan in Lake Agassiz occurred during the Qu'Appelle–Assiniboine proglacial drainage phase, which functioned until about the time that the eastern Agassiz outlets opened at 10.7 ka BP, ending the Lockhart phase of Lake Agassiz and beginning the Moorhead phase.

JAMES DRAINAGE PHASE

Although the emergence of the Missouri Escarpment

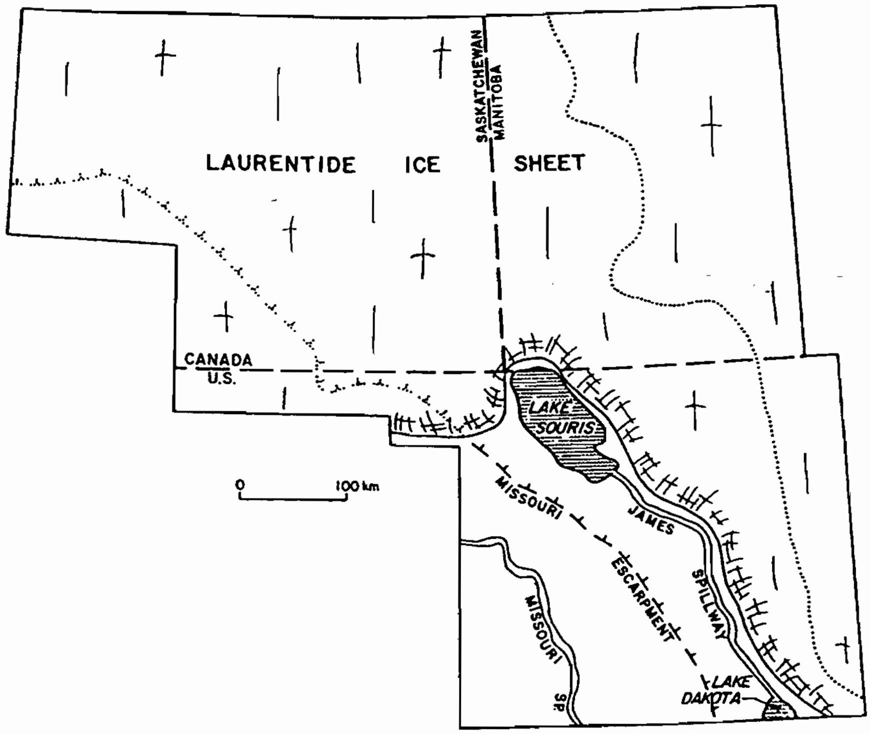

FIG. 6. Meltwater routes and lakes during the James drainage phase.

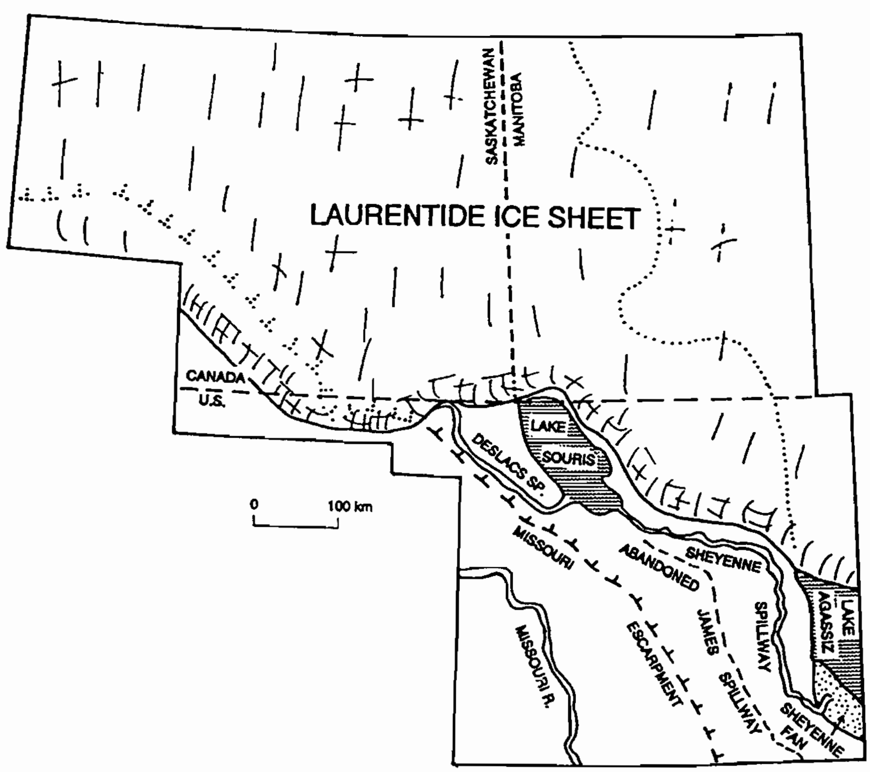

FIG. 7. Meltwater routes and lakes during the Sheyenne drainage phase.

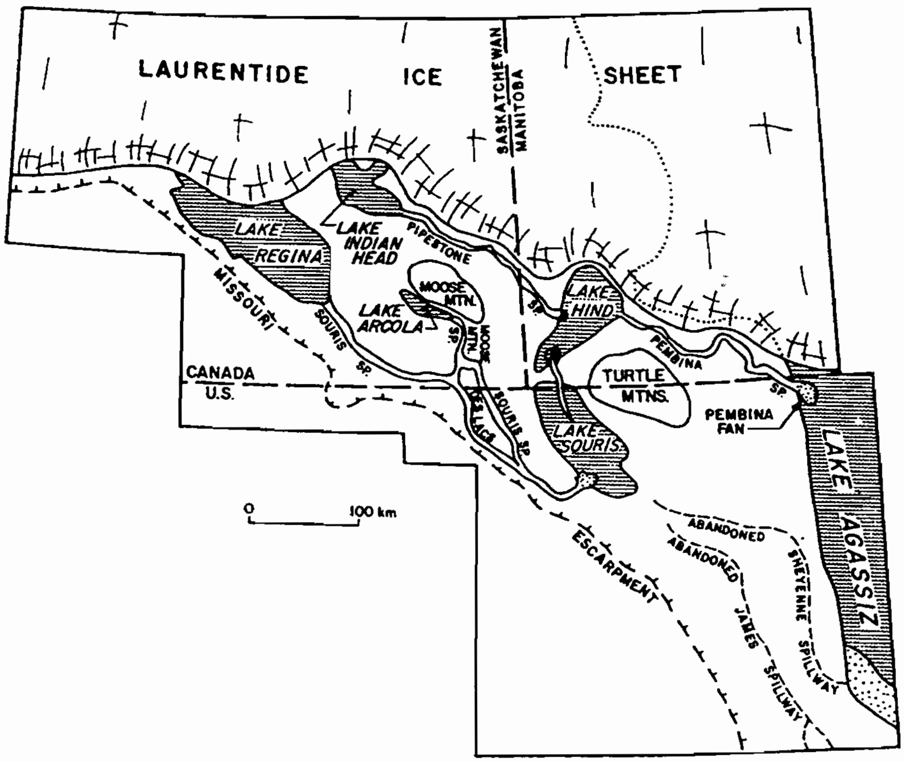

FIG. 8. Meltwater routes and lakes during the Souris–Pembina drainage phase.

from beneath the ice sheet caused a major shift in drainage paths, the first drainageways northeast of the Escarpment (Fig. 6) still were connected to the Missouri River. This configuration, only a small part of which is shown on Fig. 6, includes the James spillway and Lake Dakota. Meltwater that cut the James spillway largely originated from the lake that first occupied the Lake Souris basin (Fig. 6). Because of subsequent events in this lake basin, which probably included a glacial advance across the basin (Schnacke, 1982; Kehew and Clayton, 1983; Lord, 1988), the extent of the lake during the James phase is not known. Ice of the Red River Lobe, which occupied the Lake Agassiz basin at this time (Clayton *et al.*, 1980), appears to have prevented eastward drainage into the Sheyenne drainage system. Thus, runoff from most of the southwestern Laurentide Ice Sheet (west of the Agassiz basin) continued to be through the Missouri River basin until about 11.7 ka BP (Clayton and Moran, 1982; cf. Christiansen, 1979, Klassen, 1989, Fig. 2.25c).

SHEYENNE DRAINAGE PHASE

As the Red River Lobe retreated, Lake Agassiz formed in the Red River lowland south of the ice, expanding northward through time. A succession of proglacial drainage systems evolved on the northward-sloping surface to the west of the Agassiz basin and large, coarse-grained fans were deposited at the mouths of the rivers that carried meltwater from each of them. The first period of proglacial drainage into Lake Agassiz is designated as the Sheyenne phase (Fig. 7). Brophy and Bluemle (1983) summarized the history of ice-marginal fluctuations and drainage during this stage of Lake Agassiz.

The Sheyenne proglacial drainage basin extended west from the Sheyenne fan in Lake Agassiz to glacial Lake Souris, which supplied the water for subsequent entrenchment of the Sheyenne spillway. The basin may also have included surface runoff from as far west as glacial Lake Regina, although we speculate that the region immediately west of Lake Souris was still ice covered at this time. A northwest–southeast oriented esker near Weyburn in the southern outlet of Lake Regina (Fig. 3) (Christiansen, 1956) indicates that early meltwater flow from this region, perhaps including that routed through the Sheyenne and James basins was through glacial ice. Another possible route for this runoff was the Des Lacs spillway (Fig. 3), which probably formed before it carried water from the Lake Regina outburst during the Souris–Pembina drainage phase (Lord, 1984, 1988).

Meltwater drainage development in the Sheyenne

ABRIDGED FROM
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Glacial-lake outbursts along the mid-continent margins of the Laurentide ice-sheet

Alan E. Kehew and Mark L. Lord

As water levels rose in the proglacial lakes, ice or debris dams failed causing the incision of outlets and the release of tremendous volumes of water. Torrential discharges from the glacial-lake outbursts were diverted along the ice margin or around ridges of ice-marginal sediment in the direction of decreasing elevation. Commonly, these flow paths led to lower glacial lakes. The voluminous influx of meltwater overloaded the capacity of smaller lakes, causing their outlets to be widened and deepened. If the outlet substrate was non-resistant, complete drainage of the lake took place. The outburst then migrated to the next lower lake basin augmented by the flow from the lake just inundated (Kehew and Clayton, 1983).

The first outburst recognized in the Great Plains originated from Lake Regina in Saskatchewan (Kehew, 1982) and proceeded to flow through Lakes Souris and Hind on its path to Lake Agassiz. Recent investigations have shown that the Lake Regina outburst was not a rare, isolated event. In fact, it appears that the catastrophic partial or total drainage of glacial lakes was probably the most typical meltwater process along many segments of the Laurentide Ice Sheet. Other examples in the Lake Agassiz region were described by Clayton (1983), Matsch (1983), and Teller and Thorleifson (1983). In the eastern Great Lakes region, outbursts from Lake Maumee were reported by Vaughn and Ash (1983) and Fraser and Bleuer (1985). An event of this type in northern Illinois and Indiana was recognized as early as 1925 by Ekblaw and Athy.

The purpose of this paper is to review the erosional and depositional effects of Laurentide glacial-lake outbursts and to present new examples of these phenomena. Reconnaissance investigations suggest that glacial-lake outbursts are so common in the area of the Laurentide Ice Sheet that they should be incorporated into all models of deglaciation. The effects of these events must figure prominently in the history of the Mississippi Valley, the evolution of the Mississippi fan and Louisiana slope in the Gulf of Mexico, and in the development of the drainage system of most of northern North America.

EROSIONAL EFFECTS

The erosional capacity of glacial-lake outbursts was truly enormous. The spillways connecting glacial-lake basins on the prairies are easily visible on landsat imagery and rank among the most pronounced topographic features of the midcontinent region (Fig. 1). The volume of sediment eroded by outburst floods is equivalent to the volume of the spillway channels. Present spillway dimensions underestimate erosion because of thick sections of Holocene fill deposited in the spillways. The sediment eroded from spillway channels leading to the Mississippi Valley was deposited in the lower Mississippi Valley and/or the Gulf of Mexico. In the western plains, material eroded from spillways was deposited in glacial-lake basins.

Perhaps even more remarkable than the amount of erosion was the intensity of erosion. Outburst floods quickly trenched through the glacial drift and continued downcutting into bedrock if the bedrock was poorly indurated. Cretaceous and Tertiary sedimentary rocks on the Great Plains were easily cut through by the floods. Studies of spillway fills in North Dakota (Boettger, 1986) indicate that there was no continuous depositional phase in the spillway channels--Holocene sediment directly overlies bedrock on the spillway bottoms. Discharges therefore were short-lived and abruptly terminated.

Investigation of spillway morphology in the Dakotas, Minnesota, Manitoba, and Saskatchewan (Fig. 2) (Kehew and Lord, 1986) led to the development of a generalized geomorphic model of spillway erosional features (Fig. 3). The most diagnostic criteria are listed in Table 1; these features are described in detail in Kehew and Lord (1986). In general, spillways are wide, deep channels that begin abruptly at glacial-lake outlets and maintain their size and shape until they end at other lake basins. A flood origin for the valleys is indicated by the lack of tributary valleys of comparable size with the exception of other spillways. Spillways commonly terminate where they meet other glacial-lake basins (Kehew and Clayton, 1983), thus implying that glacial lakes were suddenly subjected to huge inflows of water. The depositional phenomena accompanying these cataclysmic inflows will be discussed in the following section.

Many spillways consist of two prominent components--a broad, gently sloping upper level (the outer zone), and a centrally positioned trench-like inner channel (Figs. 3 and 4). Shallow flow initially covered the outer zone if no existing drainageway lay in the path of the outburst flood. The common lack of distinct outer-zone margins suggests that stagnant ice formed the original channel bottom. As incision of the outer zone progressed, boulders from the underlying till began to accumulate as a coarse channel-bottom lag. This boulder concentration increased the resistance to flow in the broad, shallow outer-zone channel. In response, the flow began to carve out a narrow, deep inner channel at the center of the outer zone which provided less resistance to flow. In addition to the lag-covered surface, other characteristic outer-zone features include longitudinal grooves and streamlined residual hills (Kehew and Lord, 1986). Streamlined erosional residuals are protected

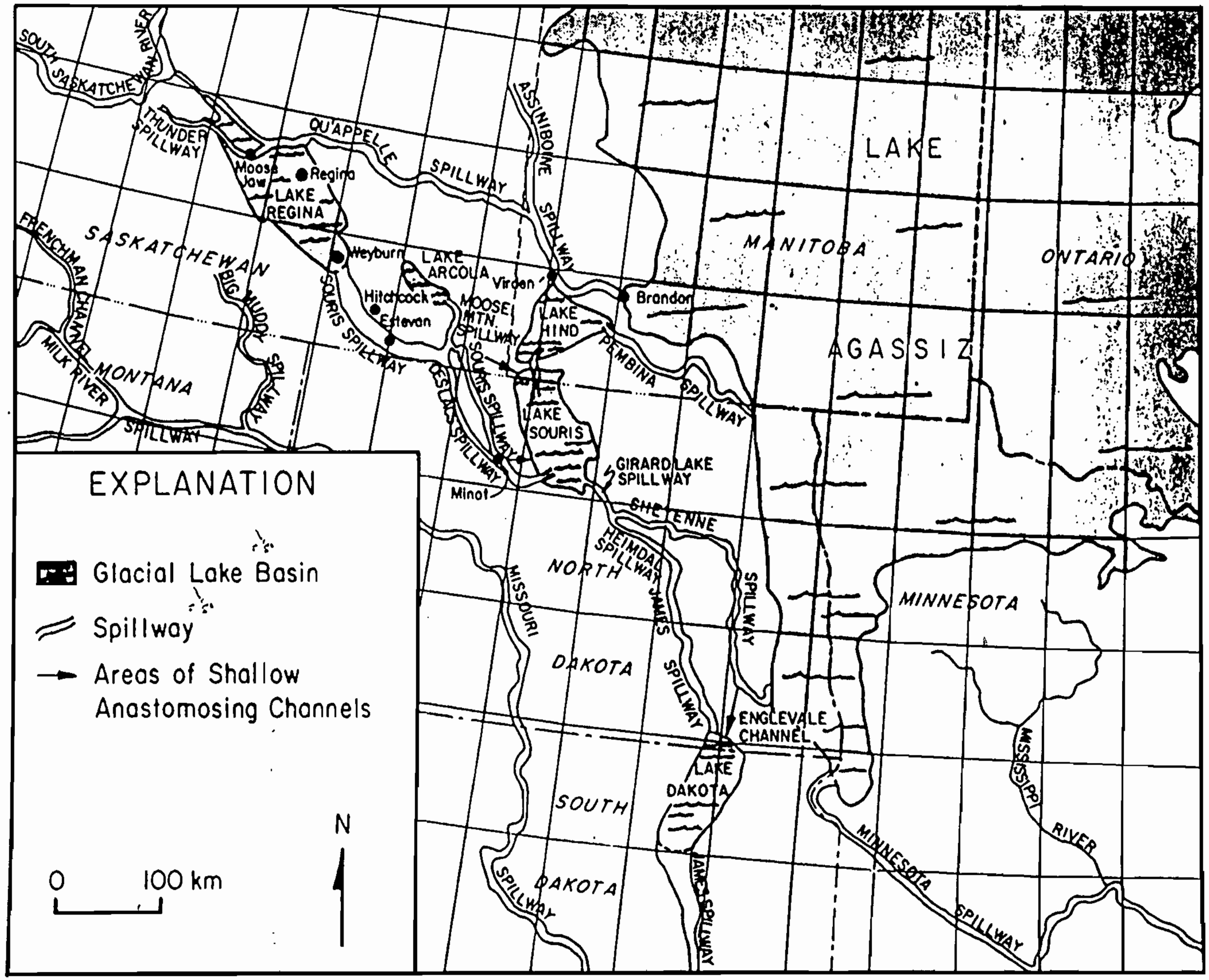

Figure 2. Location map of glacial lakes and spillways west of Lake Agassiz.

Table 1. Geomorphic features of spillways produced by glacial outbursts.

General

Lack of tributaries other than small Holocene valleys or other spillways.

Usually contain underfit Holocene streams.

Deeply entrenched.

Trend at an angle to regional slope.

Often parallel to ridges of ice-marginal deposits.

Constant size from lake outlet to termination at topographically lower lake basin or junction with other spillway.

Terminate at large, coarse-grained sediment fans in glacial-lake basins.

Channels eroded by flow that exceeded channel capacity may lead away from spillway across drainage divides.

May contain terraces representing multiple outbursts.

Inner channel

Trench-like shape.

Uniform width and side slopes.

Regular meander bends.

Occasional bifurcation to form parallel or anastomosing channels separated by linear ridges or streamlined erosional residuals.

May contain isolated erosional residual hills, usually streamlined.

1-3 km in width.

25-100 m in depth.

Lack of slip-off slopes.

May contain bars of very coarse sediment along channel sides at infrequent intervals.

Outer zone

Broad, scoured surface.

May contain shallow anastomosing channels.

Boulder lag may be developed by incision into till.

Longitudinal grooves.

Streamlined erosional hills.

from erosion during flow as they attain the streamlined shape of an airfoil. This shape, which is approximated in plan view by the lemniscate loop (Komar, 1984), minimizes flow separation and drag on the landform (Fig. 5).

The inner channel of the spillways is the most diagnostic indicator of outburst flooding. The size and shape of these channels are not consistent with an origin by gradual downcutting by small streams. The width and depth of the inner channels correlate with discharges of 10^5 m^3s^{-1} or more if bankfull flow is assumed (Kehew and Clayton, 1983; Lord, 1984). The uniform channel shape, constant valley-side slope, and lack of slipoff slopes rule out gradual erosion of the channels. A compound origin, however, including multiple episodes of outburst flooding separated by other types of fluvial activity, is very likely for some of the spillways. Multiple events are indicated by terraces underlain by outwash sediment and/or outburst flood sediment. In addition, some spillways were occupied by glacial meltwater streams after their incision.

Inner channels commonly contain erosional remnants produced by the bifurcation of the flow around obstacles or the simultaneous incision of two or more branches of the inner channel. Long, parallel inner-channel segments separated by narrow bedrock or till highs were presumably produced by concurrent erosion of two longitudinal grooves within the outer zone (Kehew and Lord, 1986). Shorter erosional remnants in the inner channels include streamlined hills which were preserved in the highly erosional flow because of their shape.

DEPOSITIONAL EFFECTS

As the capacity for erosion by glacial-lake outbursts was enormous, so was the propensity for formation of deposits reflective of the huge volumes of sediment-charged water that formed them. Eroded sediment was deposited as gravel bars within the inner channels and large, coarse-grained fans in glacial lakes recipient to the outburst discharges. Deposition of the fine-grained fraction was limited to large basins that did not drain completely during outburst inflows. In the glaciated plains west of Lake Agassiz (Fig. 2), the source sediment for the outburst deposits consists of poorly indurated till and bedrock. The till typically is composed of approximately equal parts sand, silt, and clay, with only a few percent gravel. Subjacent to the till is poorly indurated Paleocene and Cretaceous bedrock composed of silt and sand with minor amounts of fine sand and lignite. Most of the descriptions of deposits in this section are based on detailed work in the Souris and Des Lacs spillways (Fig. 2). Similar deposits are present in the Qu'Appelle, Minnesota, Sheyenne, Assiniboine, and Thunder spillways.

The discharge hydrographs of glacial-lake outbursts were probably similar to those of historical jökulhlaups, showing a steadily increasing discharge to a peak followed by a sudden cessation of flow--a pattern opposite to that of storm hydrographs (Marcus, 1960). As a consequence of the outburst discharge characteristics, flow competence within the spillways seldom decreased enough to permit deposi-

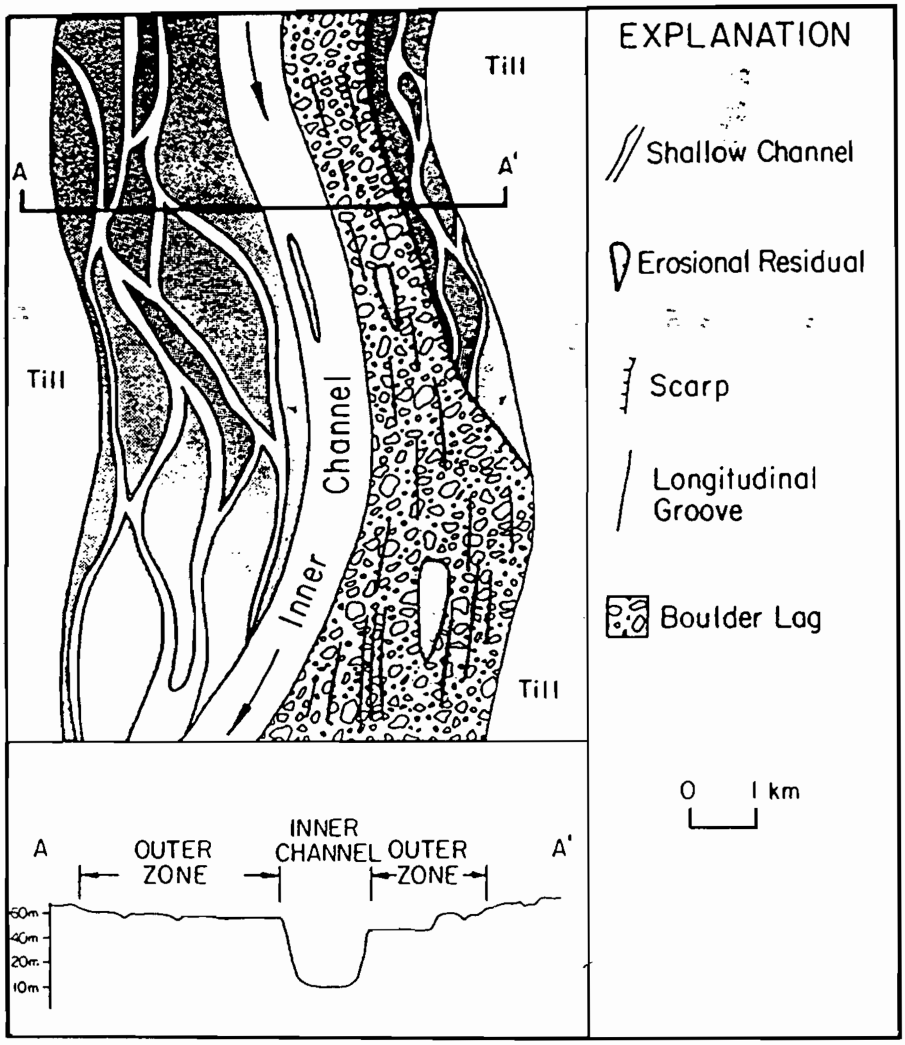

Figure 3. Generalized geomorphic model of spillway morphology. (From Kehew and Lord, 1986)

tion of any of the sediment load. Most sediment eroded by glacial-lake outbursts in the region west of lake Agassiz was dumped in glacial lakes. Though outburst deposits are relatively rare in the spillways, they are morphologically and texturally distinct from other glaciofluvial deposits.

Spillway Deposits

Glacial-lake outburst deposits in spillways consist of large scale bars. The bars occur in two depositional settings within the spillways: point bar positions and in alcoves formed by landslides during incision of the spillways (Fig. 6). The average dimensions of the bars are approximately 2 km in length, 0. 5 km in width, and 20 m in thickness (Fig. 7). These bars are primary bedforms with little or no subsequent alteration since their deposition. Bars of similar dimensions have been

Figure 4. Landsat image of Souris spillway in southeastern Saskatchewan. Outer zone is visible as the broad, non-cultivated area (5-10 km wide) flanking the incised inner channel. Cultivation is prevented by surficial boulder lag in outer zone. North at top. Width of image approximately 200 km. Arrow indicates direction of flow in spillway.

described in association with other catastrophic floods, most notably those from the Lake Bonneville and Glacial Lake Missoula floods (Malde, 1968; Baker, 1973).

Internally, the bars consist of massive, matrix-supported, very poorly sorted cobble gravel (Fig. 8) commonly containing boulders 1 m in diameter with exceptional boulders up to 3 m in diameter. The maximum grain size in the bars decreases in the downstream direction and toward the side of the spillway. Texturally, in sharp contrast to the material eroded to form the spillways, the average composition of the bars is 2 percent clay plus silt, 17 percent sand, and 81 percent gravel (Lord, 1984). The erosive power of the outbursts can be shown by comparing the amount eroded by the discharges with the amount deposited within the spillways. For example, the Lake Regina outburst eroded an estimated 26 km^3 of sediment

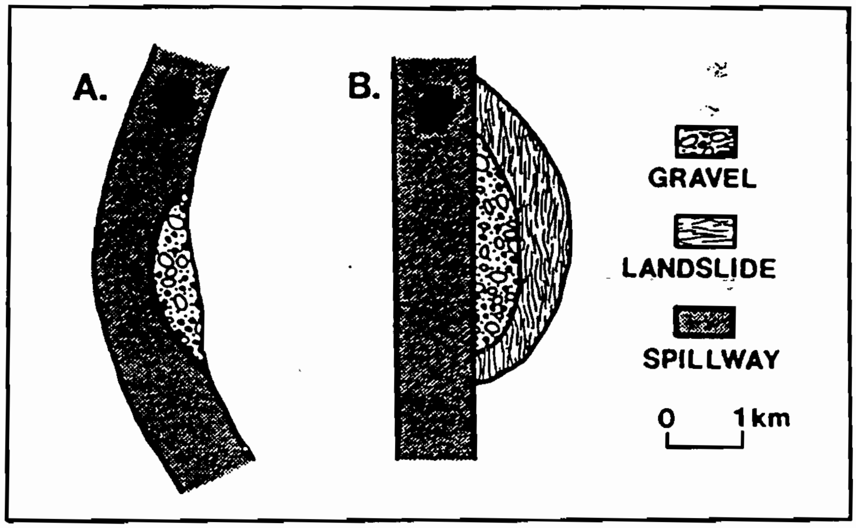

Figure 6. Schematic diagram showing two positions in which large-scale bars formed within spillways: A. Point bar position deposit, B. Alcove deposit on the spillway side of a landslide.

Table 2. Comparison of textures and volumes of material eroded versus redeposited (within spillways) by the Glacial Lake Regina outburst.

	Sediment eroded		Sediment deposited		Percent material redeposited in spillway
	Percent*	Volume (km^3)	Percent**	Volume (km^3)	
CLAY+SILT	67	17.4	2	0.01	0.1
SAND	28	7.3	17	0.13	1.8
GRAVEL	5	1.3	81	0.61	46.1
TOTAL	100	26.0	100	0.75	2.9

* Data from Kehew (1983)

** Data from Lord (1984)

(mostly till) from the Souris and Des Lacs spillways (Fig. 2). By comparison, only an estimated 0. 75 km^3 of sediment, or 2. 9 percent, was redeposited within the spillways (Table 2) (Lord and Kehew, in review). The vast majority of sand and almost all of the silt and clay were conveyed through the spillway system into downstream glacial lakes.

Flow during the outburst discharges was probably hyperconcentrated. This interpretation is consistent with estimated sediment-water concentrations, the high competence required to transport the sediment load, and the intermediate character of the bars between debris-flow and clear-water deposits. It is also likely that, once material was entrained into flow, most clast sizes, including gravel, were transported in suspension. Deposition within the spillways was limited to areas of substantial flow expansion such as at bends in the spillways or alcoves created by landslides, and did not occur along the spillway bottoms. When deposition was triggered by flow expansion, it occurred indiscriminate of clast size, resulting in deposits of massive, matrix-supported gravel.

Glacial Lake Deposits

The glacial lakes that received torrential discharges from the outbursts were the major sediment sinks for the eroded material. The form taken by the deposits in the lakes was largely dependent on three factors: (1) the density of the inflow, (2) the lake basin morphology, and (3) the volume of the lake basin. The density of the sediment-laden outburst discharges was significantly greater than that of the stilled lake water. As a result, inflows formed density currents that flowed along the lake bottom and deposited underflow fans (Kehew and Clayton, 1983). Underflows, because of their high density and low turbulence, may transport coarse sediment many kilometers past the inlet (Church and Gilbert, 1975). Underflow fans, because they do not prograde into lakes at the water level by fluvial action, tend to be very well sorted and gradually fine away from the inlet (Fenton and others, 1983; Kehew and Clayton, 1983). Underflow fans deposited by glacial-lake outbursts can be identified by their large areal extent and position at the mouth of a spillway eroded by a glacial-lake outburst. Outwash processes are ruled out for these deposits because they are not associated with valley trains, outwash plains, or moraines.

Lakes with irregular bottom topography developed local flow conditions that resulted in complex sediment deposition patterns. The capacity of the lake to contain the added volume of catastrophic inflows determined whether the recipient lake itself drained in a domino fashion. Where the influx triggered drainage of the recipient lakes, most of the fine-grained sediment--silt and clay, did not have sufficient time to settle out and was transported through the lake. In these cases, the lake acted similar to a very wide reach of a river with continuous throughflow rather than a lake in which the entire sediment load is dumped. Morphologic evidence of a domino-triggered lake drainage is incision into the fan by the spillway when the lake started to drain. Additional evidence for simultaneous outburst inflow and outflow from lake basins is summarized in Kehew and Clayton (1983).

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