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Sedimentological and stratigraphic investigations of a sequence of 106 varves from glacial Lake Assiniboine, Saskatchewan

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Abstract

Sediments retrieved from a long core on the floor of glacial Lake Assiniboine, Saskatchewan, expose 106 couplets, consisting of thick, light coloured, silt-rich beds and thin, dark, clay-rich beds. The couplets contain sharp lower and upper contacts of the silt bed, silty and clayey laminations within both the silt and clay beds, and ice-rafted debris in the silt beds, which are features characteristic of glacial varves.

Seasonal variations in runoff are reflected in grain size profiles of individual silt beds in the varves. Mean grain size maxima in the lower portion of the silt bed suggest that snow accumulation during the previous winter had been substantial and that a warm spring combined with a rapid melting rate generated significant volumes of nival meltwater runoff. Coarse laminae higher in the silty part of the couplet imply that substantial meltwater inflow was produced by summer melting of glacier ice.

Vertical trends in clay bed thicknesses, silt bed thicknesses, and total couplet thicknesses were strongly influenced by the proximity of meltwater inflow channels and lake depth. These interpretations, and correlation of the core to varve exposures at the surface, formed the framework for a paleohydrological reconstruction. Close to 11,000 BP, ice dammed the outlet of glacial Lake Assiniboine and the water depth rose about 2 m yr^{-1}. Eventually the lake became deep enough for couplets to form. Varve years 1-40 contain thick clay beds, silt beds, and couplets as a result of the proximal inflow of meltwater. A decline in silt bed and couplet thicknesses from varve years 41-85 occurred in response to ice retreat and more distal inflow. Varve deposition ceased in the shallow part of the basin probably because underflow currents from the distal source were redirected. Varve years 86-106 are distinguished by an increase in silt bed and couplet thicknesses and a decrease in clay bed thickness caused by a reduction in water depth and a return to proximal inflow. Varved sedimentation terminated when Lake Assiniboine drained through the Assiniboine valley to Lake Agassiz.

Introduction

Tyrrell (1892), during his frontier explorations of northwestern Manitoba and parts of Saskatchewan, was the first to recognize an "ancient" lake basin from Little Boggy Creek to Fort Pelly, Saskatchewan. He suggested that terraces along the base of the Duck Mountain Upland were relict shorelines of "glacial Lake Assiniboine". Located at the upstream end of the Assiniboine spillway (Fig. 1), this ice-dammed lake covered nearly

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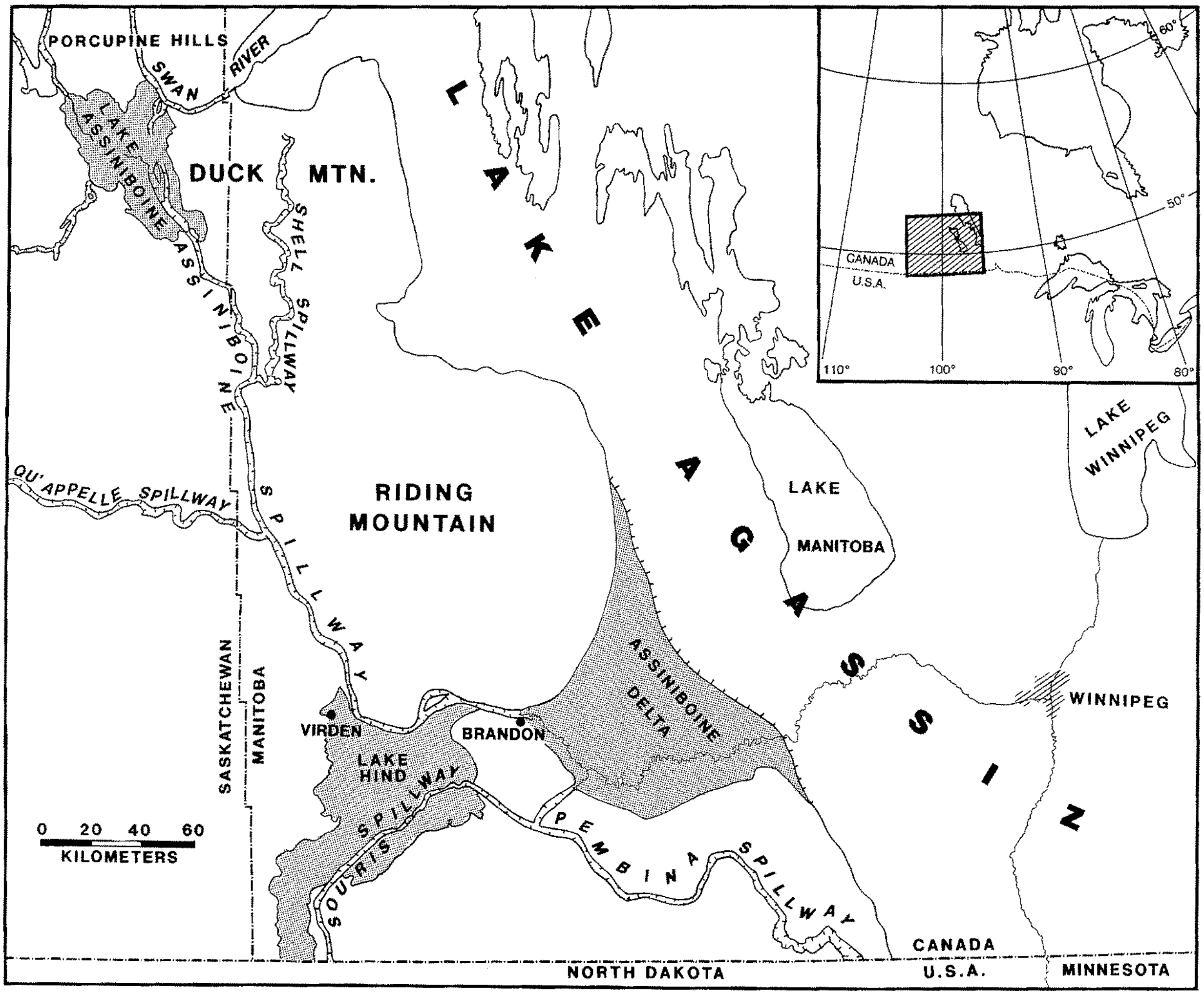


Fig. 1. Regional location of glacial Lake Assiniboine, showing major spillway systems eroded during the last deglaciation. Glacial Lake Hind basin is also stippled, as is the main part of the Assiniboine fan delta which was deposited into Lake Agassiz by meltwater floods through the Assiniboine-Qu'Appelle system.

1500 km^2 and, similar to many glacial lakes of the Prairie region, formed along the margin of the retreating Laurentide Ice Sheet.

Early attempts at resolving the histories of ice-dammed lakes in the Prairies have been accomplished principally through the interpretation of strandline and spillway elevations (e.g. St. Onge, 1972; Klassen, 1972, 1975; Christiansen, 1979). A multifaceted approach that centres largely on the glaciolacustrine sedimentary record, in addition to geomorphology, was used to decipher the geological history of glacial Lake Assiniboine (Wolfe, 1993).

Rhythmically bedded sediments discovered in outcrop and analysis of an 11 m, 106 rhythmite sequence collected from the basin during the summer of 1990, proved critical to the interpretation of the history of glacial Lake Assiniboine. Unlike many studies that have not taken the opportunity to explore the nature of glacial rhythmite deposition in detail, sedimentological and stratigraphic investigations on these several *centimetre* thick couplets, revealed a highly sensitive record of seasonal events to decadal phases in the Lake Assiniboine basin.

Local geologic setting and surface features

Modern drainage and physiography

Much of the Lake Assiniboine Plain (Fig. 2) is relatively flat, except for Holocene gully erosion that has occurred along river channels and valleys. Presently, the Assiniboine River heads in the Porcupine Hills area and flows eastward across the lake basin. It then abruptly elbows to the north before turning south into the Assiniboine valley at the former outlet of glacial Lake Assiniboine, 10 km southeast of Kamsack, Saskatchewan (Fig. 2). The Whitesand River flows east across the basin where it joins with the Assiniboine River near Kamsack. Little Boggy Creek drains the southern region of the Duck Mountain Upland and meets the Assiniboine River south of Kamsack. The modern Swan River occupies an 80 m deep, 1 km wide spillway that today bypasses the Lake Assiniboine Plain and flows to the northeast.

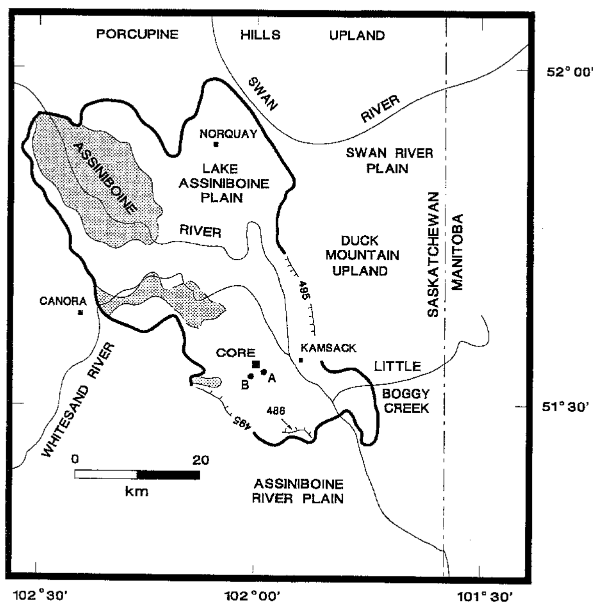


Fig. 2. Outline of surface features in the Lake Assiniboine region including the modern drainage system, physiography, strandlines (hatched lines) with elevations (m), and coarse grained glaciolacustrine sediments (stippled zones) deposited by former inflowing channels. Also shown are thick rhythmite exposures (A, B), and the coring site in the southern part of the basin.

Predecessors of the Assiniboine and Whitesand Rivers supplied most of the sediment to glacial Lake Assiniboine, depositing sand and gravel where they entered the lake (Fig. 2).

Strandlines

The few strandlines that were identified in the field and traced on aerial photographs were fragmentary and poorly developed, suggesting that the lake was at least partly bounded by ice for much of its existence. The most prominent strandline occurs in the southwestern corner of the basin (Fig. 2) and has a distinct gravelly surface indicative of former wave reworking and deposition. This strandline occurs at 495 m, measured near the top of the slope, and is at the same elevation as eastern strandline segments found along the Duck Mountain Upland (Fig. 2). An isolated, though fairly well developed lower strandline at 488 m was recognized in the southern part of the basin (Fig. 2). These observations indicate that Lake Assiniboine established at least 2 lake levels: 495 m and 488 m.

Outcrops of rhythmic sediment

In places, glaciolacustrine sediment in the basin is rhythmically bedded and characterized by repeated thin, dark coloured clay-rich beds overlying relatively thick, light coloured, silty clay to clayey silt beds. In outcrop, couplets range in thickness from 1 to 40 cm, although the silty bed comprises most of the couplet. Basal and upper contacts of the silty bed are sharp. Occasionally, current structures such as ripples were found in the silt beds but most often they contain horizontal laminations. Sedimentary structures were difficult to assess in the clay beds but they appeared to be horizontally laminated. Rhythmites are restricted to the southern part of the basin and at two thick sections, sites A and B (Fig. 2), 69 and 52 couplets were counted, respectively. Toward the north, exposures reveal apparently massive clay to silty clay.

Subsurface investigations

Materials and methods

To acquire a complete record of the rhythmic interval, a core was obtained from the deepest part of the basin (Fig. 2). In total, 23 m were retrieved using a truck mounted, mechanical hollow stem auger drilling system equipped with 0.6-m-long, 7-cm-diameter Shelby tubes. In the lab, sediment was extruded using a hydraulic piston, placed in split 0.6-m-long plastic tubes, covered in heavy plastic bags to decelerate drying, and labelled. Core segments were scraped down to a flat surface and then left uncovered for approximately 48 hours which, because of differential drying, brought out subtle sedimentary features. All core segments were then photographed.

Each 0.6 m section was described in detail and graphic logs were prepared following recommendations given by Anderton (1985), at a scale of 1:2 (see Wolfe, 1993). Sedimentary structures were recorded and estimates of grain size were made. Clay bed thickness, silt bed thickness, and the total couplet thickness were measured to the nearest millimetre.

A two-tiered approach was used for grain size analysis to examine the internal nature of the rhythmites. First, all completely recovered couplets were analyzed to obtain general grain size curves. Most were then classified into three categories based on curve shape, and several couplets were chosen for high resolution analysis.

Initially, a minimum of 3 samples were collected from the silty part of the couplet: one from the bottom third, one from the middle, and another from the top third, even if no clear grain size difference could be seen. Then, as required, more samples were taken as visible grain size fluctuations dictated, to a maximum of 8 samples per silt unit. For couplets that were not completely preserved in the core, a single representative sample was taken. At least one sample was also taken from the clayey portion in most couplets. Then for selected rhythmites, subsamples were taken at 1 cm intervals.

Samples collected for grain size analysis were examined using the laser-operated Brinkmann PSA 2010 Particle Size Analyzer at the University of Manitoba. Milligram-size samples were collected and stored in small pill bottles. Samples were immersed in 2% Calgon solution, shaken and covered overnight to allow for sediment dispersal. Immediately prior to grain size analysis, each sample was placed in an ultrasonic bath for 2 minutes to permit further disaggregation. Several drops of the slurry were transferred to a cuvette, again diluted with 2% Calgon solution, and placed in the Brinkmann electro-optical unit for analysis.

Results

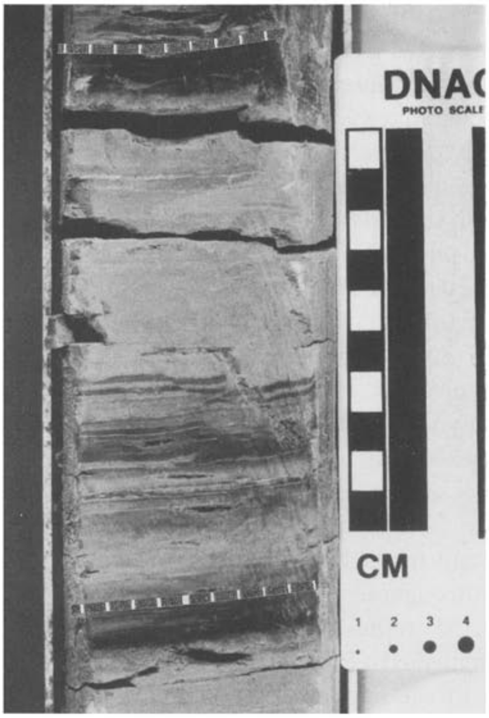
Sedimentological analysis

Between a depth of 4.0 and 14.9 m, 106 couplets were counted, which consisted of alternating dark coloured, clayey beds and light coloured, silty beds similar to the surface exposures. The sequence of couplets is abruptly overlain by 1.3 m of medium to coarse grained silt that contains trough cross laminations at the base and horizontal laminations near the top.

Contained within many of the thicker silty beds are thin, slightly coarser, light-coloured, horizontal, silty laminations and darker, clayey laminations. These laminations occur variably throughout most silt beds (Fig. 3a). Similarly, the clay component is frequently interrupted by horizontal silty to silty clay laminations (Fig. 3b). Ice-rafted debris is only present in the lower part of the rhythmite sequence (up to couplet 25) and commonly occurs as thin laminations, consisting of poorly sorted silt and clay with occasional

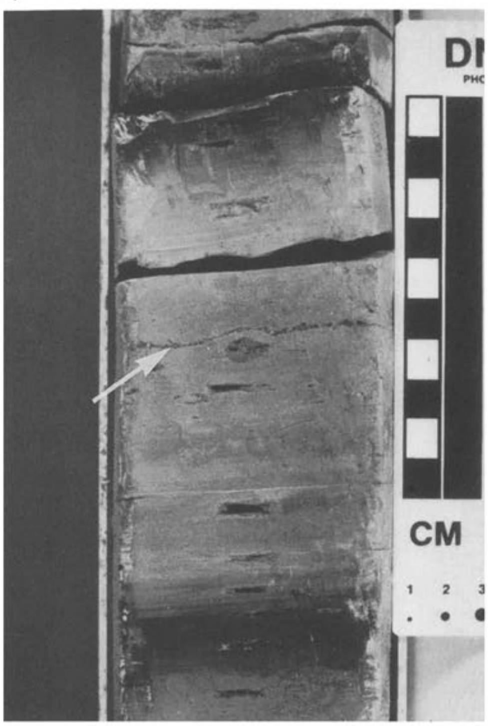
Fig. 3. Sedimentology of cored rhythmic interval. a) Couplet 47. Lower and upper contacts are highlighted. Notice several dark, clayey laminations in the lower portion of the silt bed. b) Couplet 52. Frequent silty to silty clay laminations are present within the lower clay bed. c) Couplet 4. Notice the thin lamination of diamicton near the centre of the silt bed is bent around the silty clay to clayey silt clast. d) Couplet 17 Disturbed laminations and a recumbent fold can be seen near the centre of the photo.

a)



b)

c)



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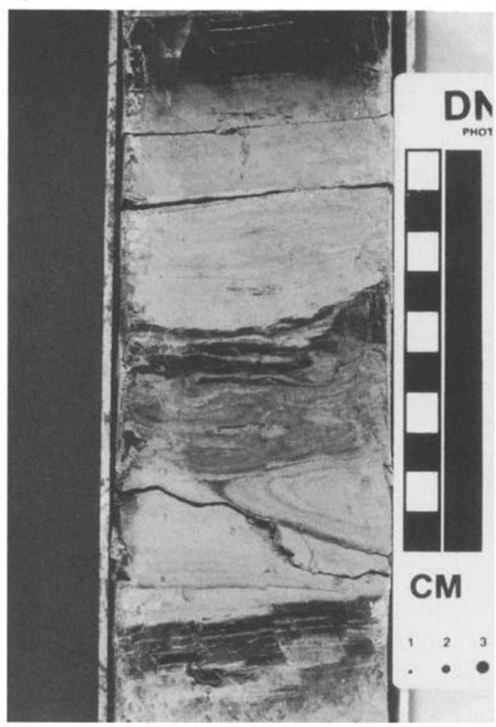

granules, or scattered clasts often near the middle of the silt bed (Fig. 3c).

Horizontal bedding is pervasive throughout the unit. Other types of bedding are relatively rare but cross lamination occasionally occurs in the silty part of the couplets. Several of the couplets have their laminations disturbed and many show displacement along vertically restricted faults. As well, convoluted bedding was found in the form of recumbent folding (Fig. 3d). Disturbance due to coring is likely responsible for the deformation at the top of several Shelby tube sections.

Contacts at both the base and top of the silt beds are nearly always sharp and regular. Rarely is there evidence of scouring at the top of the clay bed. Periodically, both upper and lower contacts appear gradational over less than 1-2 cm. Gradational contacts, in places, coincide with underlying clay beds that contain internal laminations of silty sediment (Fig. 3b).

Stratigraphic analysis

In general, total and silt bed thicknesses are more variable and thickest in the lower half of the sequence, with an upward tendency towards thinner beds (Fig. 4). Significant reversals in this trend occur between couplets 15 and 21, 31 and 40, and 86 and 98, which are revealed by the smoothed silt bed and total couplet thickness curves (Fig. 4). Clearly, the variation in the total couplet thickness is almost entirely dependent on the fluctuations in silt bed thickness (silt:couplet, $R^2 = 0.945$ for 11 point running mean data). Conversely, the clay bed thickness is relatively constant compared to the total thickness (clay:couplet, $R^2 = 0.270$ for 11 point running mean data), although it too thins upward. However, the clay bed thickness roughly varies with the silt bed thickness. A notable exception to this occurs in the upper portion of the sequence where silt bed thicknesses increase from couplet 86 to 98 and then remain constant through to couplet 106, while clay beds thin throughout this interval.

Based on these observations, 3 subunits can be distinguished (Fig. 4). Subunit 1 contains couplets 1 to 40 and, although highly variable, the silt bed thickness and total couplet thicknesses decrease upward. The clay beds also become thin-

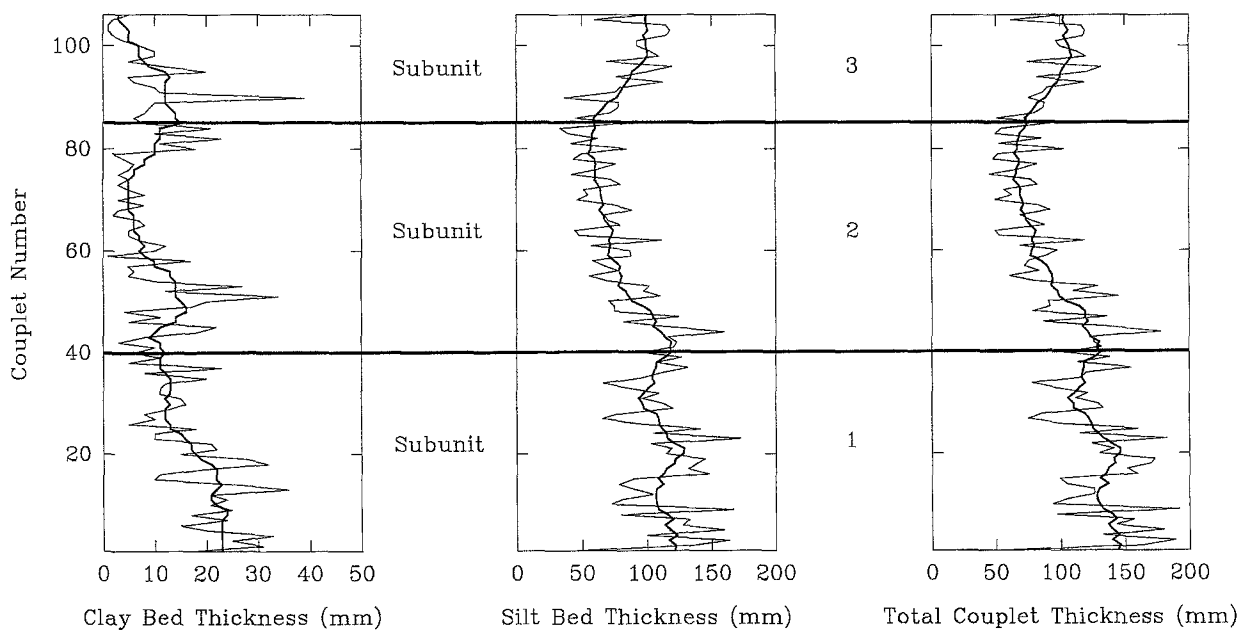

Fig. 4. Clay, silt, and total couplet thickness versus couplet number. Raw data and 11 point running mean curves are shown. Note horizontal scale is different for clay bed thickness. Sequence is divided into three subunits.

ner upward through this interval. Subunit 1 contains the thickest silt beds, clay beds, and couplets.

Subunit 2 is defined by couplet interval 41 to 85 and is characterized by a decrease in silt bed and total couplet thicknesses. This unit includes the thinnest silt and couplet thicknesses, with the lowest values near the top of the interval. Clay beds are thinner in subunit 2 compared to 1.

Subunit 3 occurs at the top of the sequence from couplets 86-106 and is demarcated by an increase in silt bed and couplet thickness in the lower part of this interval. Throughout this subunit, the clay bed thickness decreases. Silt and couplet thicknesses are intermediate with respect to subunits 1 and 2.

Grain Size Analysis

In general, the mean grain size of the silt bed narrowly ranges from very fine silt to fine silt whereas the mean grain size of the clay bed is usually less than 4.0 μm (Fig. 5). If the maximum mean particle size is selected from each completely preserved couplet, nearly all of the data points plot between 7.8 and 15.6 μm (i.e. fine silt; Fig. 6). The maximum mean grain size is nearly constant for much of the series except for three relatively coarse intervals: couplets 2-7, 48-53,

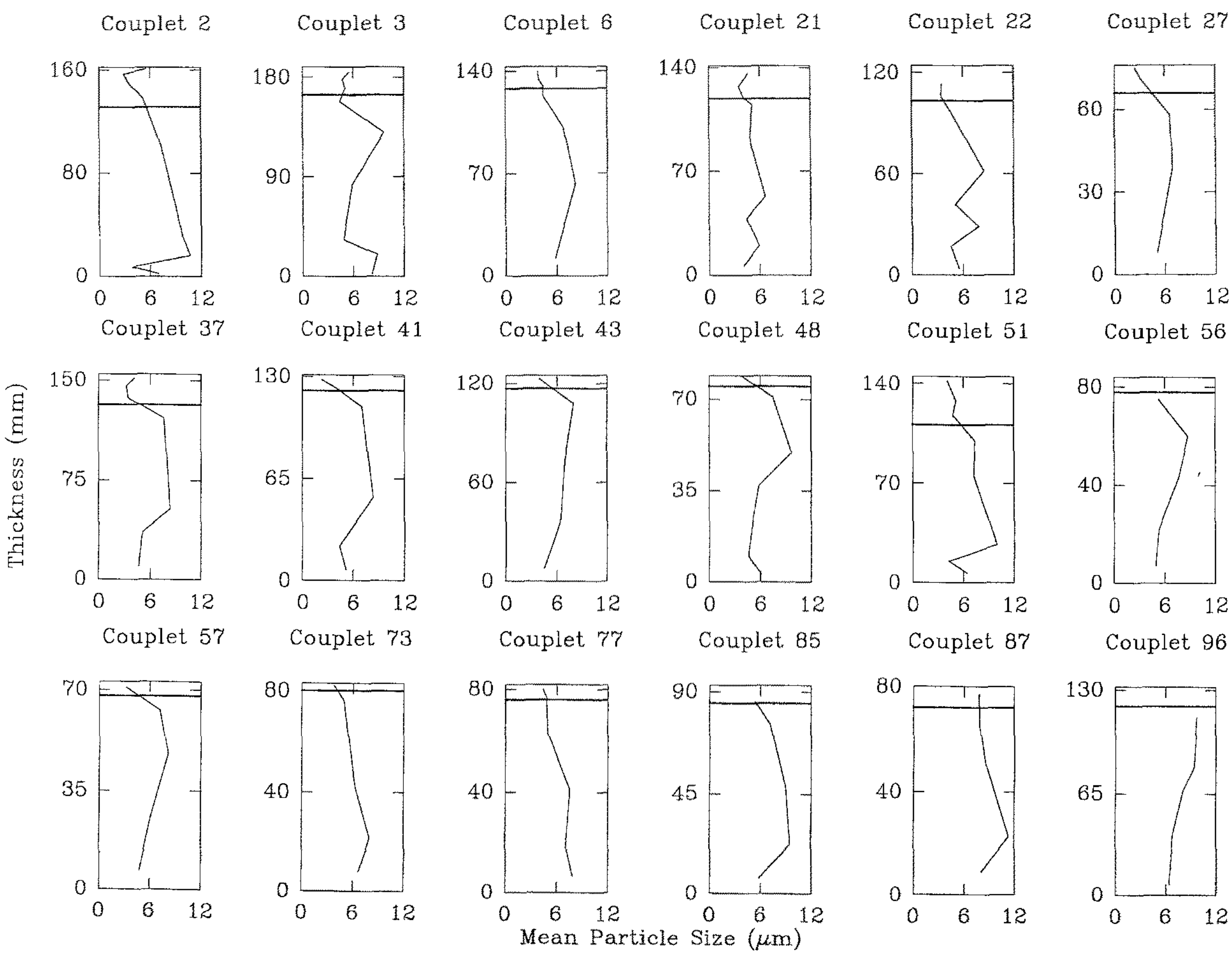

Fig. 5. Selected mean grain size profiles of completely preserved couplets. Note vertical scaies are variable. Horizontal line near top of profiles represents stratigraphic contact between the silt and clay bed.

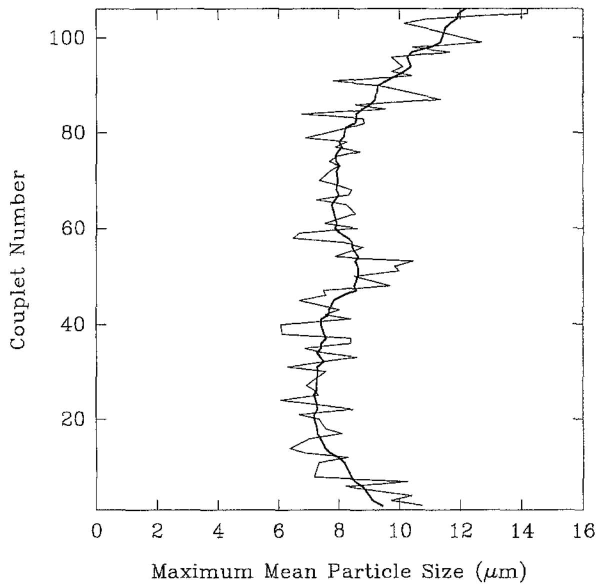

Fig. 6. Maximum mean particle size selected from the multiple analyses performed on each completely preserved silt bed versus couplet number. Both raw data and 11 point running mean curves are shown.

and 87-106. Couplets 2-7 are coarsest at the base and then fine upward. Maximum mean grain size generally increases from couplet 87 to the top of the sequence.

Most rhythmites were visually classified into one of three general types (Fig. 7) using the curve shapes of the silty part of the couplets. Type 1 is coarsest at the base and fines upward. Type 2 is characterized by a coarsening upward trend. Type 3 exhibits two peaks, one near the base and the other near the top of the silt bed. Type 3 is the most common type of trend, accounting for over

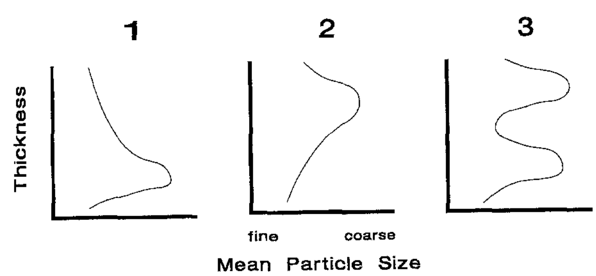

Fig. 7. Diagrammatic mean particle size distribution curves for silt components of couplets: (1) fining-upward, (2) coarsening-upward, and (3) bimodal.

a third of the analyzed rhythmites, followed by type 2 and type 1.

Six rhythmites, including couplets 2, 3, 6, 56, 77, and 96, were analyzed in greater detail to 1) confirm the trends indicated by the initial analyses and 2) to use as models to examine seasonal variations meltwater inflow.

In couplets 2 and 77 (Fig. 8), "type 1" rhythmites, the fining-upward trends are consistent with that predicted by the initial, more widely spaced size analysis (compare with Fig. 5). Similarly, couplet 56, a "type 2" rhythmite, displays a coarsening upward trend through most of the silty portion of the couplet, which also was indicated by the initial analysis (Fig. 5). The grain size trend revealed by a one centimetre sampling interval suggested that couplet 6 was most similar to "type 2". However in another "type 2" rhythmite, (couplet 96, Fig. 8) the high resolution analysis revealed at least two grain size peaks; one near the base and another in the mid to upper third of the silt bed. A third peak may also be present near the upper silt bed contact. This couplet should, therefore, be re-classified as "type 3". In couplet 3, a "type 3" rhythmite, two increases in particle size are present in the silty portion, consistent with the more general results (see Fig. 5). In contrast to couplet 96 (Fig. 8), the larger peak occurs at the base of the silt bed.

Interpretation

Sedimentology

Several mechanisms may be responsible for rhythmically bedded sediments in glaciolacustrine environments. The cause most often recognized for rhythmic sedimentation in glacial regimes (or areas with annual snow cover) is the seasonally controlled alternation between the warmer melt season and the ice-covered months. Here, the rhythmically bedded lacustrine sediments represent accumulation over a one year period, (i.e. a varve), partly deposited during the summer (coarser), partly during the winter (finer).

Other forms of rhythmic sedimentation, which may represent significantly different time intervals, include slump-induced surge current depos-

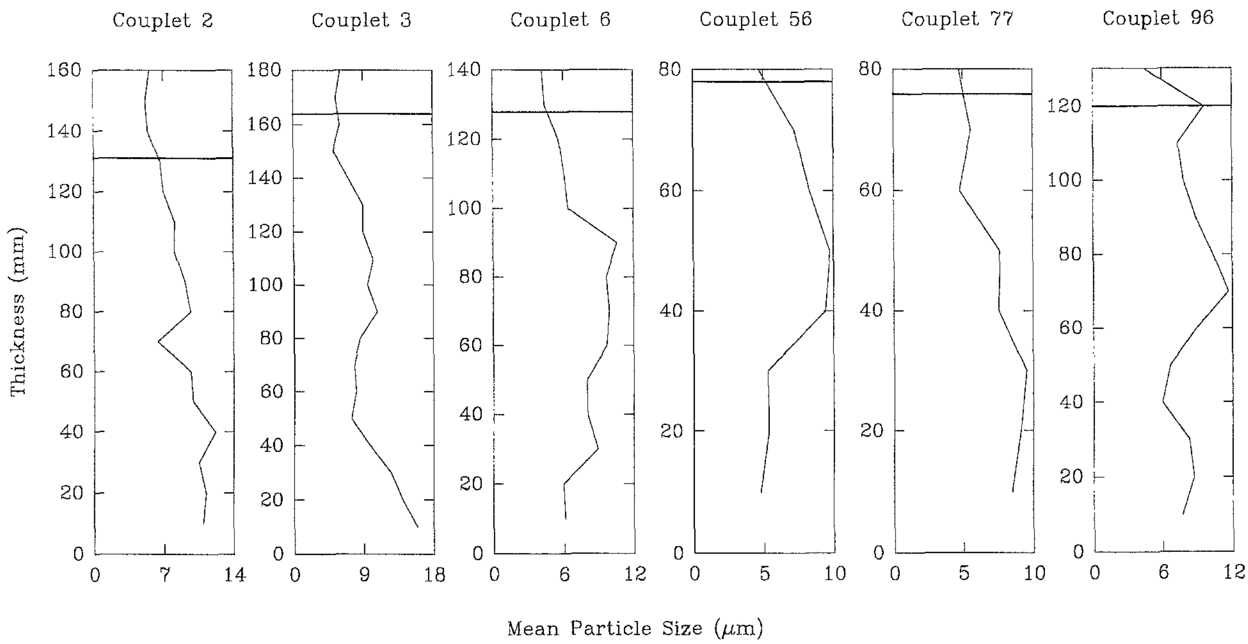

Fig. 8. Mean particle size profiles for couplets 2, 3, 6, 56, 77, and 96. Horizontal line near top of profiles represents stratigraphic contact between the silt and clay bed. Note variable horizontal scales.

its, sediments generated by fluctuations in runoff or river flow to the lake, or changes in wave energy; all of which are possible over the course of minutes, hours, days, or weeks (Smith & Ashley, 1985). Surge currents are provoked by movement on unstable basin slopes (Ashley, 1988), continue no more than several minutes, and produce waning flow deposits similar to turbidites (Smith & Ashley, 1985). These deposits may form at any time within the year and, therefore, repetitive couplets can be generated through the irregular recurrence of this process (Smith & Ashley, 1985) .

During the melting season, sediment-laden river water commonly produces an underflow current (Ashley, 1988) that can last for several days (Smith & Ashley, 1985). Thus, each river surge can generate a fining upward (e.g. silt to clay) bed that may be similar in appearance to an annual deposit. Commonly, underflow currents cease to exist during the winter and finer sediment previously held buoyant settles from suspension over top of the single or multiple couplets of sediment (Smith & Ashley, 1985). Clearly, sedimentological distinction of varves versus other non-annual coarse-fine couplets is critical in establishing a lake chronology, as well as in understanding the nature of the lake's history.

Both non-varve and varve couplets are composed of a relatively coarse (typically sandy or silty) unit overlain by a clay bed, possess sharp contacts at the base of the coarse bed, and are variable in thickness (Smith & Ashley, 1985). However several key sedimentological characteristics of the rhythmites described from glacial Lake Assiniboine indicate that they are annual deposits.

Although both non-varve and varve deposits commonly have sharp contacts at the base of the coarse bed, the upper silt contact can be used as a distinctive feature. A continuous, surge current-produced bed normally would contain a gradational upper silt contact (Smith & Ashley, 1985). However, the rhythmites of Lake Assiniboine normally display sharp upper silt contacts, suggesting a temporal break in the formation of the two beds. This is attributed to an abrupt break in deposition at the time of the change in energy conditions at the surface when the lake alters from ice-free to ice-covered status.

Surge deposits are generated over very short time intervals and a single continuous graded bed would be expected to form from each event. In contrast, varves are produced over the entire year, and irregular variations in sediment influx are likely to occur during each season. In the Lake Assiniboine rhythmites, laminations in the silty portion of the couplet are quite variable (e.g. Fig. 3a), and may be related to fluctuating supply of sediment from inflowing rivers, melting of glacial ice or snow, rainfall-generated runoff, or periodic fluctuations in wave energy impinging on the lake floor.

Silty laminations that occasionally appear in the otherwise massive clayey bed (e.g. Fig. 3b) are indicative of occasional high energy events in the winter. Factors contributing to the generation of these structures include low sediment concentrations in winter lake water, permitting relatively dilute underflow currents to flow across the lake floor (Gilbert, 1975; Shaw *et al.*, 1978). These currents may be created by winter storms if the lake is ice free, slope failure on delta fronts (Shaw *et al.*, 1978), or slumping of sediment near the ice margin (Gravenor & Coyle, 1985). Regardless of their origin, they indicate that the lake bottom was occasionally supplied with silt-sized sediment during the winter and deposition did not exclusively consist of clay settling from suspension.

Discrete laminations of diamicton (poorly sorted sandy, silty clay) occur exclusively in the silty part of some couplets mainly in the lower part of the rhythmic interval (e.g. Fig. 3c). This sediment could potentially represent dumping from winter ice, river ice, or icebergs, or may be due to mass movement. Sharp upper contacts and absence of associated waning flow structures such as horizontal laminations or ripples suggests that deposition by surge currents generated by mass movements is unlikely. Because the thin diamicton units are present only in the high energy portion of the couplets, and often near the middle of the silt bed, they are probably related to occasional dumping of icebergs that may have been produced during peak summer melting in this ice-marginal lake. Although the diamicton laminae occur strictly in the lower part of the rhythmically bedded interval, their absence in the younger lake sediments does not preclude the existence of floating icebergs (e.g. Gilbert & Desloges, 1987).

In summary, frequent sharp basal and upper contacts of the silt bed, occasional silt and clay laminations indicative of discrete sedimentation events, and diamicton near the centre of the silt bed deposited by melting icebergs, all indicate that the couplets are classic glacial varves (e.g. DeGeer, 1912; Ashley, 1975; O'Sullivan, 1983) and justifies their utilization as a relative timescale. Furthermore, these sediments indicate that Lake Assiniboine was capable of producing varves for about 106 years.

Varve thickness stratigraphy

In general, thicknesses of the clay beds, silt beds, and total varves in the cored sequence decline irregularly during the first 40 years (subunit 1, Fig. 4). Then, a more steady upward decline in clay, silt, and couplet thickness occurs. Exception to this is near the top, where the silt and total varve thicknesses increase while the clay beds decrease. These stratigraphic intervals, subunits 1, 2, and 3, are interpreted to represent *deep-proximal, deep-distal,* and *shallow varve* deposits, respectively, which are elaborated on below.

Thinning of couplets, often recognized in vertical sequences over several years, is usually interpreted to represent a retreating active ice terminus and, therefore, a more distal meltwater source (e.g. Agterberg & Bannerjee, 1969; Bannerjee, 1973; Ashley, 1975; Shaw & Archer, 1978 and many others). At the base of this thinning-upward sequence (subunit 1: varve years 1-40), are the thickest silt beds, clay beds, and overall thickest couplets produced when the ice margin was nearest to the coring site (i.e. "deep-proximal" varves). Varves become thinner upsection, indicating ice margin retreat. As well, there is a decrease in the maximum mean particle size (Fig. 6) during this period.

Subunit 2 (41-85) contains the thinnest silt and couplet thicknesses. Throughout this interval, the silt and total couplet thicknesses also decrease upwardly as they do in subunit 1, while clay bed thicknesses are variable (Fig. 4). These "deep-distal" varves were deposited when the ice margin was and continued to become comparatively remote. The overall low values for the maximum particle sizes are interrupted by a five year rise (Fig. 6; varves 48-53), which may reflect an increase in summer storm activity.

Subunit 3 (86-106) is distinguished by an increase in silt bed and couplet thicknesses in the lower part of the subunit, while clay bed thicknesses decrease. An increase in total varve thicknesses near the top of a varve sequence has, in other situations, been interpreted as (1) an ice advance (e.g. Bannerjee, 1973), (2) an increase in runoff by overland streams to the lake (e.g. Ringberg, 1991), or (3) a decrease in lake level which permitted more proximal influx (e.g. Catto, 1987). The first two mechanisms would tend to produce thicker silt and clay beds because there is an increase in sediment supplied to the basin. In contrast, a decrease in water depth would increase energy on the lake floor and result in the resuspension of already-deposited silts and clays in the shallowest areas of the basin. In turn, there would be an increase in sedimentation of more silty sediment (and, potentially, more total sediment) at the core site. The increase in silt bed thickness (and total couplet thickness) and the decrease in clay bed thickness in subunit 3 (Fig. 4) are compatible with this interpretation, as is the increase in maximum mean particle size (Fig. 6). After varve year 98, the silt bed becomes constant in thickness, so lake level may have stabilized.

Variations in *individual* silt bed and varve thickness (versus sequence trends) also occur throughout the rhythmic interval. In studies of other ice-dammed lakes, these variations have been related to annual climatic fluctuations which affected the melting of ice (Perkins & Sims, 1983; Leonard, 1986; Ringberg, 1991). For example, Perkins & Sims (1983), in their studies of varve deposits at Skilak Lake, Alaska, indicate that warm years produce greater than normal melting of the Skilak Glacier and an increase in meltwater and sediment supply to the lake; all of which is reflected in a thicker varve. Conversely, thin varves are generated during relatively cool years.

Irregular individual peaks and short-term thickness trend reversals, such as between varve years 15 and 21, and 31 and 40 (Fig. 4) may be due to seasonal variations, similar to those described at Skilak Lake. For example, higher mean annual temperatures allowing more summer melting of the glacier may be responsible for these anomalies. Alternatively, they may reflect periods when multiple river systems were contributing sediment, thus increasing the supply to the coring site. Fluctuations in silt and varve thicknesses are much less pronounced in the deep-distal interval (subunit 2, Fig. 4) compared to the deep-proximal interval (subunit 1), because the distal site is less sensitive to irregularities in annual meltwater-sediment supply.

There are several mechanisms that have been responsible for annual variations in the clay bed thickness. Slight changes in the length of the ice-covered season (viz. the time when wind energy is kept from the lake waters) can modify the available length of settling time critical to deposition of the easily suspended clay-sized particles. Intermittent sedimentation of silt during the winter has caused an increase in the thickness of some clay beds. Also, as in Lillooet Lake (Gilbert, 1975), large sediment-water influxes in the fall may have substantially increased the concentration of suspended particles in the lake water, which then settled onto the lake floor during the winter beneath the ice.

Annual grain size variations

Variations in seasonal and interseasonal fluxes in runoff to the basin can be seen in Figure 5. Grain size trends within the summer portion of a varve have been recognized for their paleoclimatic significance (e.g. Peach & Perrie, 1975; Smith, 1978). In his discussion of the varved sediments of Hector Lake, Alberta, Smith (1978) indicates that approximately half of the summer beds in the intermediate-proximal facies contain two silty laminations separated by a clayey to clayey silt lamination. These strata were interpreted to represent two distinct seasonal inflow maxima; spring nival melt and summer glacial melt. Other varves that do not contain the clayey sublaminae, and are normally or inversely graded, are indicative of seasons in which significant nival or glacial meltwater influxes did not occur (Smith, 1978).

At Hector Lake, significant nival melting is dependent on sufficient snow accumulation and a rapid melting rate while glacier ice melting is enhanced by periods of warm, clear weather (Smith, 1978). Water derived from glacier melting is normally greater than that produced from the melting of snow, and a phase of reduced meltwater flow often occurs between spring snow melt maxima and summer glacial melt maxima (Smith, 1978). Distinction of the two seasonal meltwater flow intervals may be enhanced by a delay in glacial melt runoff (Stenborg, 1970; Smith, 1978) and may be responsible for intervening clayey to clayey silt lamination in the silty, summer part of a varve. Causes for this delay include temporary meltwater storage in deep firn packs and the closure of major drainage conduits by freezing during the winter (Stenborg, 1970). In all cases, the influence of melting ice (and its snow cover) may be directly into the ice marginal lake or via overland flow beyond the margin of the ice.

To assess the annual climatic or hydrological implications of grain size variations in individual varves, six couplets were studied in detail (Fig. 8). On the basis of mean particle size distribution in the silty part of these varves, and in combination with the less-well-controlled size distribution of other complete couplets (Fig. 5), three major patterns of summer (ice-free season) grain size changes were identified, and are shown in Figure 7.

Spring nival melt varves (Fig. 7 – type 1) are represented by the fining upward trends displayed in varve years 2 and 77 (Fig. 8). Absence of an upper stratigraphic maxima (type 1 vs. type 3) may be indicative of cool summers during these varve years or glacial melting that is evenly distributed, yet progressively less intense, throughout the summer months.

Summer glacial melt varves (Fig. 7 – type 2) are illustrated by varve years 56 and 6 (Fig. 8). Differences in the shape of these profiles may be indicative of the conditions of summer melting. In varve year 56 (Fig. 8) summer melting is concentrated in the last part of the summer. In contrast, the trend in varve year 6 is more gradual suggesting a steady increase in summer temperature until a maximum was achieved. The absence of nival peaks in these varves indicates a lack of significant, previous winter snowfall accumulation or a cool spring.

Finally, some varves display both spring nival melt and summer glacial melt peaks, separated by an interval of decreased meltwater input (i.e. *Dual-melt varves;* Fig. 7 – Type 3). Interestingly, in varve 96, the spring nival peak is less than the summer glacial peak (Fig. 8), while the opposite occurs in varve 3 (Fig. 8). The very strong nival peak in varve 3 may be indicative of a warm spring. The minor peak near the upper winter contact in varve 96 (Fig. 8) is likely reflective of a late summer or fall storm event.

Synthesis

Stratigraphic model

Correlation of nearby extensive varved outcrops (A, B; Fig. 2) to the deep sediment core provide key additional data and lateral support for the stratigraphic changes observed in the thickness of the varves (Fig. 9). The lower part of A section fits between varve years 17 and 62, and the upper interval between varve years 76 and 98. At B, varve deposition appears to have been initiated at varve year 8 and terminated at varve year 59. Strikingly similar profiles indicate that the varved sediments are laterally continuous and, more importantly, suggests that varve thickness variations are reflective of basin-wide changes and are not site-specific. Therefore, intervals based on varve thickness trends described and interpreted from the deep sediment core can be used to designate phases in the Lake Assiniboine history. In addition, A and B provide additional information concerning (1) changing influx patterns, sources, and distribution, (2) the effects of basin topography on sedimentation, and (3) lake level fluctuations. All or some of these factors may be responsible for the incomplete records seen in the surface

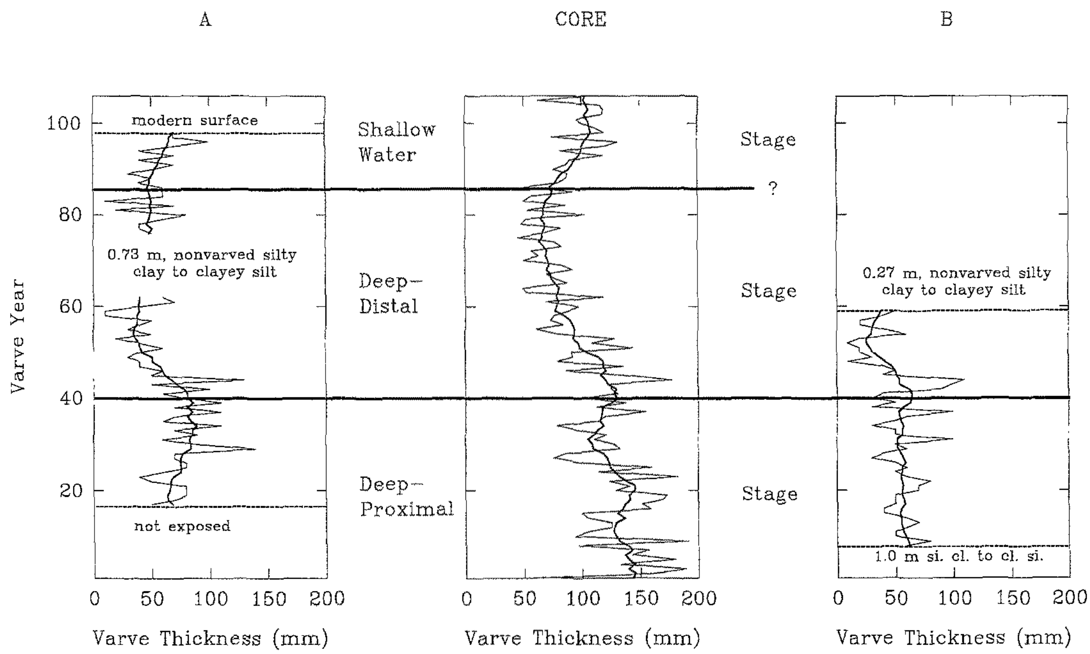

Fig. 9. Correlation of varve thickness profiles for A (left), the deep sediment core (middle), and B (right). A and B are located 1 and 3 km from the core, respectively (see Fig. 2). Both raw data and 11 point running mean curves are shown. Gap in A record is due to disturbed interval where rhythmic sediments could not be identified. Varve year scales for A and B have been adopted from the core on the basis of correlated thickness peaks and troughs.

exposures. A brief synopsis of the history of glacial Lake Assiniboine, represented by the varved interval, is presented below.

Deep-proximal stage (varve years 1-40)

An ice dam formed at the southern outlet of glacial Lake Assiniboine (Fig. 10a), which caused the lake level to rise and initiate varve sediment deposition in at least the southern part of the basin. Lacustrine sediments prior to this were coarser with no dark, clayey beds, suggesting the lake was more shallow. Strandlines at 495 m in the southwestern part of the lake and along the Duck Mountain Upland (Fig. 2) developed where the lake was not in direct contact with the ice (Fig. 10a), and indicate that the lake stabilized at this elevation.

The 40 varves at the base of the sediment core are the thickest in the sequence and characterize this phase (Fig. 9). Because lake level was rising, varve formation in more elevated sections of the lake floor, such as B, did not begin until later; thus only varve years 8-40 appear to be represented at this site (Fig. 9). By using the elevation difference (16 m) at the base of the varves in the sediment core (450 m) and nearby B (466 m), the delay in the beginning of the varve sequence at B (8 varve years) would suggest that the lake level rose at least 16 m (466 – 450) in 8 years (i.e. 2 m yr^{-1}) in order to allow varves to form and be deep enough below wave base at both the core site and section B. Prior to initiation of varve deposition, Gilbert-type delta foreset beds indicate that lake level stood at 472 m (Wolfe, 1993) so based on a delay in varve sedimentation noted above, a rise of 16 m would have brought lake level up to 488 m. At this rate it probably took an additional 3.5 years (11.5 years total) for the lake to rise another 7 m (23 m total) to the maximum strandline level (495 m). Therefore, lake level likely reached 495 m by varve year 12.

Thick couplets in the lower part of the varved interval in the sediment core (1-40), and in sections B (8-40) and A (17-40; Fig. 9) are due to a relatively close source of influx on the western side of the lake (Fig. 10a) . Sand was deposited

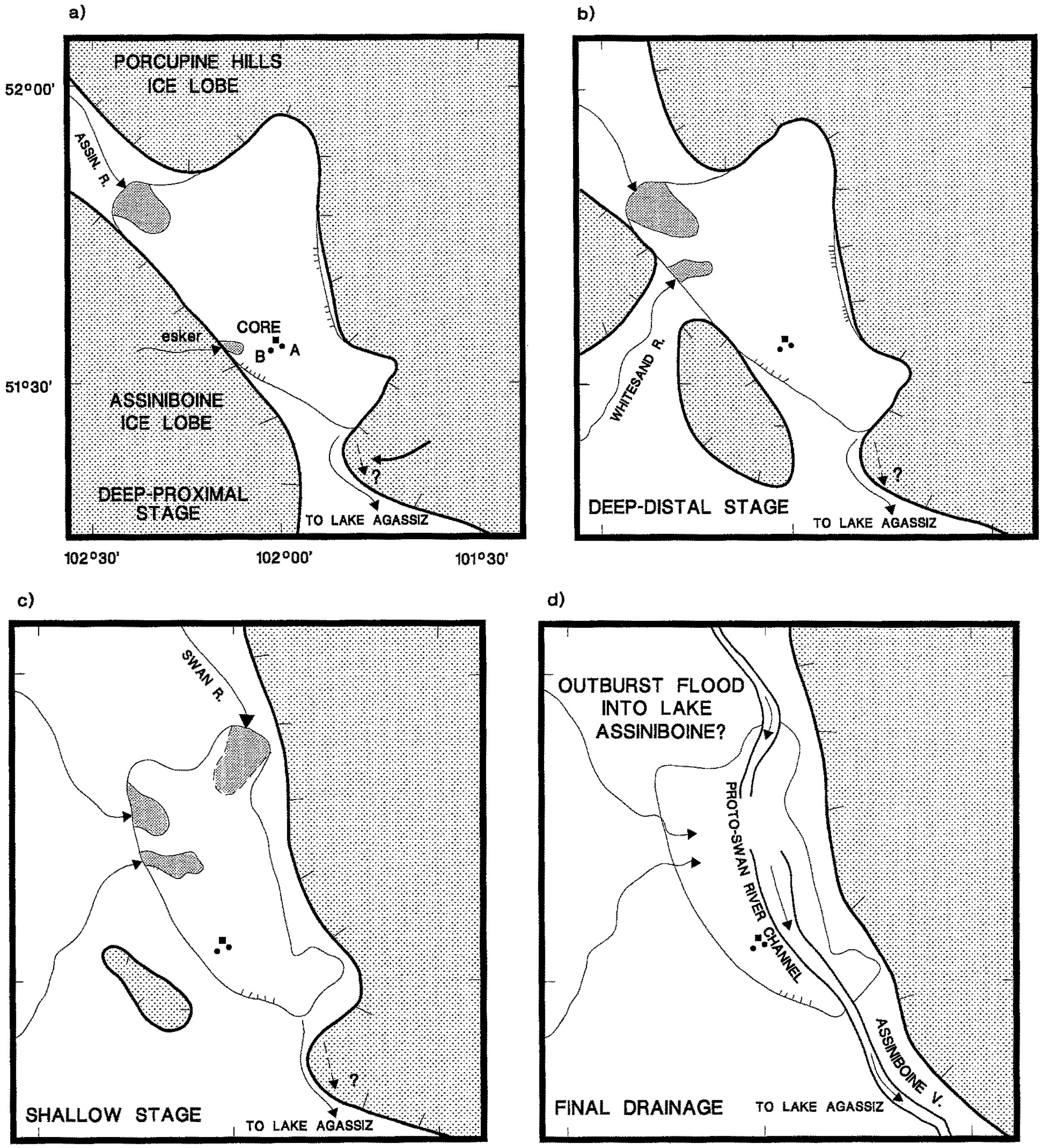

Fig. 10. Stages in the history of glacial Lake Assiniboine represented by the 106 varves. a) Deep-Proximal Stage (varve years 1-40), b) Deep-Distal Stage (varve years 41-85), c) Shallow Stage (varve years 86-106), and d) Final Drainage.

near the end of what is now a series of eskers at a maximum elevation of 495 m, where meltwater from the ice entered Lake Assiniboine. The upper Assiniboine River, which carried meltwater from the Porcupine Hills region, also provided inflow to the northern part of Lake Assiniboine, which may have contributed to the deep-proximal sediment package.

Deep-distal stage (varve years 41-85)
A decrease in silt and couplet thickness in the sequence overlying the deep-proximal varves, in addition to a similar but more irregular decline in clay bed thickness, indicates that the influx of sediment became increasingly more distal to the coring site in varve years 41-85 (Fig. 4). This change in varve character marks the beginning of the deep-distal stage (Figs. 9,10b). With westward disintegration of the Assiniboine Ice Lobe, the Whitesand River began to carry meltwater and deposit sand and gravel, with minor silt and clay, where it entered Lake Assiniboine from the southwest (Fig. 10b). Abundant kettle lakes in the area south of where the Whitesand River entered Lake Assiniboine suggests that stagnant ice remained after active ice of the Assiniboine Lobe wasted away, laterally restricting southwestern expansion of the lake. Ice to the east probably remained active, resupplied from higher elevations of Duck Mountain. Lake level probably remained at 495 m, as ice appears to have controlled overflow from the lake into the Assiniboine River valley during this stage.

A similar decrease in couplet thickness is seen in sections A and B during this period (Fig. 9). However, varved deposition ceased at varve year 59 at B (Fig. 9) at an elevation of 469 m. Considering B's elevated position in the basin relative to A and the core site, it is probable that new underflow current distribution patterns that emanated from the more distal Whitesand River avoided this site and, as a result, poorly laminated clayey silt to silty clay was deposited over top of the varved sequence (Fig. 9); alternatively, a later erosional stage may have removed additional varves deposited after varve year 59, before depositing the overlying sediment.

Shallow stage (varve years 86-106)
Near varve year 86, a major influx of water from the Swan River entered Lake Assiniboine and was responsible for erosion at the lake outlet (Wolfe, 1993). These events caused lake level to drop (Fig. 10c) and resulted in an increase in both silt bed thicknesses and couplet thicknesses while the clayey part of the couplets became thinner. A similar increase in couplet thickness occurs at section A from varve year 86 to 98 (Fig. 9). Constricting lake margins caused the mouths of rivers to encroach toward the coring site (Fig. 10c) and this resulted in greater sediment accumulation during the ice-free season. Higher energy associated with shallow water transported more of the clay-sized material out of the basin. From varve years 98 to 106, silt and couplet thicknesses are relatively constant, suggesting that the lake stabilized, likely at the 488 m level.

Final drainage
The varved sediments in the core abruptly end and are overlain by 1.3 m of silt that represents material deposited by a floodburst into the northern part of the lake (Kehew and Lord, 1987; Fig. 10d) . This influx spilled over the outlet, ultimately causing Lake Assiniboine to breach its southern margin and erode deep lake outlet channels. This may have occurred subglacially although it seems just as likely that the ice dam at the southern end of the lake was unstable and failed as a new flood of water entered the basin. Eventually a channel near the center of the broad scoured area, the present-day Assiniboine River valley, was chosen as the preferred drainage route, and it became entrenched to 434 m, 54 m below the previous level of Lake Assiniboine. Presumably Lake Assiniboine catastrophically drained at this time (Kehew and Lord, 1987). The date of the final event is estimated to have been about 11,000 BP, based on regional correlations and their relationship to glacial Lake Agassiz (*cf.* Fenton *et al,* 1983; Nielsen, 1988).

Conclusions

For 106 years, glacial Lake Assiniboine deposited annually bedded, silt-clay couplets. Both summer variations in sediment influx and long-term trends in this influx can be identified by grain size analyses of the summer, silt-dominated part of each couplet. Spring runoff maxima are featured in some varves and summer meltwater maxima characterize others, whereas the most common varves contain both snow melt and glacier melt peaks of sedimentation separated by periods of reduced meltwater inflow.

Phases in the lake's history, are revealed by changes in clay bed thickness, silt bed thickness, and couplet thickness. Initially, ice advanced, probably from the Duck Mountain Upland, dammed the outlet from Lake Assiniboine, and lake level rose to a minimum of 495 m by about varve year 12. For at least 40 varve years, sediment influx to the core site was dominated by flow through the old Assiniboine valley that drained Porcupine Hills and by runoff from ice that lay west of the lake basin. Wastage of the ice margin along the western side of the basin allowed the Whitesand River to enter Lake Assiniboine for at least 45 varve years. Following 85 years of varve deposition, a major flow into the northern part of Lake Assiniboine through the Swan River valley caused the outlet of Lake Assiniboine to be incised and lake level fell 7 m to the 488 m level. At least 20 varve years later, a second major influx caused Lake Assiniboine to breach its southern shore and entrench a deep valley. As a result, Lake Assiniboine drained through the Assiniboine River valley to Lake Agassiz around 11,000 BP.

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