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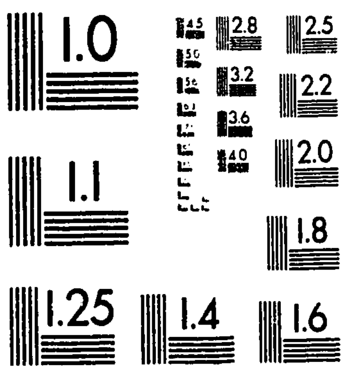

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UNIVERSITY OF ALBERTA

Glacial Environments of the Edmonton Region, Alberta

by

Kent M. Holden

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH

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ABSTRACT

The Duffield Sand Complex west of Stony Plain is interpreted to have formed subglacially by erosional and depositional meltwater processes operating in a large cavity. The cavity was formed as meltwater pressure lifted the ice sheet from its bed northwest of Stony Plain. Sedimentation in the cavity during its deepest phase created extensive and thick cross-stratified and cross-laminated bedding units. Progressive sedimentation in the cavity, erosion of scallops in the cavity roof, and late-stage drainage produced hummocks and eskers.

The Big Lake Sand Complex, west of the city of Edmonton, formed in the eastern side of the cavity by meltwater draining through the Sturgeon Valley. Cross-cutting of the western side of this complex by a channel containing an esker demonstrates its subglacial origin. Gradients of large anabranching channels, south of Stony Plain, are both down-flow and reversed and are measured at about 1 m per km over 10 kilometers. These gradients demonstrate that channels formed subglacially.

An esker superimposed on the Bon Accord Sand Complex, above the northeastern wall of the Sturgeon Valley near Bon Accord, indicates a subglacial genesis for the sand complex. Descent of the esker from the surface of these deposits into the Sturgeon Valley indicates that the valley existed before both deposition of the Bon Accord Sand Complex and formation of the esker.

Drainage through the Sturgeon Valley prior to deposition of the Bon Accord Sand Complex was by conduit flow which supplied meltwater to the regional-scale cavity to the southwest. A local subglacial meltwater event, prior to deposition of the complex, flowed over a large area in and southeast of the valley, creating a scour field over 50 km^2 in area. This flow event also eroded numerous flutes along the southern rim of the upper Sturgeon Valley; some beginning near the valley floor extend kilometers across the adjacent plain.

Retreat and stagnation of ice over the region produced regressive, arcuate ridges over the Bon Accord Sand Complex. These ridges are interpreted to have formed by the raising and lowering of buoyant ice in a retreating ice-marginal lake, which caused subglacial sediment to be squeezed into the lake at the grounding line.

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Thanks to Dr. J. Ronald Eyton who created TERRA FIRMA: the software package used in this thesis to transform mere numbers into meaningful maps and statistics.

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The provision of Digital Elevation Moldels by Mr. Marvin Weiss at the Land Information Services Division of Forestry, Lands and Wildlife, Governement of Alberta, is greatly appreciated.

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Most of all, thanks to my wife Sylvie, who endured me waking up in the middle of the night to jot down ideas, and for her support these past two years.

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CHAPTER 1

GENERAL INTRODUCTION

This thesis focuses on the late-Wisconsinan sedimentary environments and landforms of a region north and west of Edmonton, Alberta. The work is presented in two main chapters, written in the form of papers, each discussing portions of an integrated glacial system. Both papers include Thematic Mapper (TM) data, and Digital Elevation Models (DEMs) as tools by which sequences of events are interpreted from visual and derived numerical elements. Field surveying and detailed sedimentological logging provided the main basis for my environmental interpretations.

The observations of previous work (Hughes 1958; Bayrock and Hughes 1962; Bayrock and Berg 1966; Rains 1969; Westgate 1969; Bayrock 1972; Kathol and McPherson 1975; Shaw 1975, 1982, 1987; Andriashek 1989; Shetsen 1990) are largely sound. However, there are some fundamental differences between the conclusions in this thesis and past environmental interpretations. In places, reinterpretations of sediment and landform genesis from ice-marginal to subglacial environments of formation are warranted by new evidence.

Brodzikowski and Van Loon's (1991) systematic review of literature on glacigenic sediments revealed only one type of subglacial fluvial facies where tunnel-flow controled the sedimentation regime. Recent work on subglacial fluvial sedimentation, however, by Shaw and Gorrell (1991) extended the conduit environment to include dunes formed in a tunnel-channel. Subsequent work by Fisher and Shaw (1992), concluded that Rogen moraine on the Avalon Peninsula, Newfoundland, was deposited by subglacial meltwater flowing as a broad sheet rather than in a conduit. The production of vast tracts of scoured topography, in which drumlin swarms and eskers are formed, have also been attributed to such meltwater sheet-flows (Shaw 1983; Shaw and Kvill 1984; Shaw et al. 1989; Rains et al. 1993). Hummocks formed by subglacial fluvial accretion are now included in this group of landforms. The process is summarized in Ashley et al. (1985).

Subglacial deltaic and lacustrine facies described by Brodzikowski and Van Loon (1991) are up to 15 m thick, were deposited in subglacial lakes generally less than a few hundred meters in diameter, and resemble the textural and structural characteristics of 'normal' deltaic and lacustrine sediments. Some European glaciolacustrine sediments are also interpreted to have been deposited subglacially - though there has been much opposition to this interpretation. These range in type and thickness from thin, varved clay units (Carruthers 1953) to nearly 65 m of stratified, sandy sediments (Hoppe 1963). A subglacial lake presently existing beneath the Antarctic ice sheet (Oswald and Robin 1973), over 100 km long, indicates that there is no reason why large-scale subglacial lacustrine systems should not have existed beneath the Laurentide ice sheet.

The second chapter of this thesis concludes that hummocks, eskers, and meltwater channels in the Edmonton region were formed in subglacial lacustrine and fluvial systems rather than proglacially, as was previously suggested. The third chapter investigates the genesis of certain landforms and the progression from subglacial to ice-marginal environments of formation.

The study area is divided into two parts. Each is composed of a set of landforms and sediments that illustrate, mainly by superposition and cross-cutting relationships, their progressive formation in a changing environment. The two main chapters deal mainly with late-Wisconsinan glacial sediments ranging from those deposited in subglacial fluvial and lacustrine environments to those deposited or formed ice marginally. Though the second chapter deals almost exclusively with subglacial aqueous processes, the third integrates both subglacial and ice-marginal environments by examining the sequential development of landforms as the environment changed during deglaciation.

Landforms and their assemblages in the study area were first evaluated by inspection of a LANDSAT image. A number of image-enhanceing techniques were used to produce enlarged views of the terrain. These enlarged views then served as a regional reference guide to coordinate the study of site-specific geomorphology and sedimentology. DEMs provided a further means of studying landforms and landform assemblages. These models supply detailed information about landform curvature, height, width, elevation differences and topographic gradient, all of which are used to characterize each landform type. Satellite imagery and DEMs provided regional landform trends and simple tonal differences, which were the basis of preliminary interpretations. Mapping from aerial photographs and field reconnaissance refined these interpretations based on large-scale geomorphic and sedimentological characteristics.

CHAPTER 2

Subglacial Lacustrine and Fluvial Environments: The Stony Plain Area, Alberta

Introduction

The effect of subglacial meltwater on landform development has been the focus of much attention over the last decade (Shaw 1983; Shaw and Kvill 1984; Shaw and Sharpe 1987; Mooers 1989; Sharpe and Shaw 1989; Shaw et al. 1989; Kor et al. 1991). Both erosional and depositional subglacial fluvial processes have been inferred at scales much larger than normally accepted. Landforms such as eskers show, however, that subglacial water was indeed present under continental ice, regardless of its distribution, throughout the last glaciation.

Literature on subglacial fluvial sedimentation has been, until recently, largely restricted to the description and interpretation of eskers. The recognition of subglacial meltwater sediments of this origin is hindered by the limited number of depositional landforms considered diagnostic of the subglacial fluvial environment. It has been suggested that hummocks of various sizes and shapes may be formed by flowing water beneath ice, and largely composed of glaciofluvial and/or glacigenic sediments (Shaw and Gorrell 1991; Fisher and Shaw 1992; Holden and Shaw, in prep.). Rains et al. (1993) outline the possible importance of subglacial meltwater sheet-erosion in the development of hummocky terrain in Alberta. The potential for the discovery of other landforms produced by meltwater is considerable.

Debate on the formation of drumlins (Shaw 1983; Boulton 1987; Menzies 1989; and Boyce and Eyles 1991) has fueled much controversy but has also prompted increased research on subglacial fluvial environments, landforms and sediments. Subglacially formed gravel dunes, for instance, within tunnel channels commonly associated with eskers were recently discovered (Shaw and Gorrell 1991). As well Fisher and Shaw (1992) interpreted Rogen moraine composed of interbedded diamicton and stratified water-lain sediment as subglacially-deposited fluvial landforms.

Hummocks of an extensive tract of hummocky terrain west of Edmonton are proposed to have formed in subglacial cavities created by the flow of turbulent subglacial meltwater. These hummocks, which are composed of stratified silt and sand, with secondary clay and gravel, were previously interpreted as part of a pitted delta (Hughes 1958; Bayrock and Hughes 1962; Andriashek 1989; and Shetsen 1990). They are reinterpreted here as products of subglacial meltwater sedimentation — based on field evidence, interpretation of aerial photographs, enhanced satellite images, computer assisted three-dimensional models, and their association with other subglacial landforms. Such features as eskers superimposed on the Duffield Sand Complex, eskers feeding into meltwater channel systems, and anabranching channels with significant convex-up profiles, provide the evidence for the subglacial formation of the hummocky terrain.

Regional setting

Study location

The study area extends over 2500 km^2 from Big Lake in the east to Wabamun Lake in the west, and from north of Onoway to the North Saskatchewan River in the south (Fig. 2.1).

Bedrock geology and topography

Most of the study area is underlain by weakly consolidated bentonitic sandstone, shale and coal of the Late Cretaceous Horseshoe Canyon Formation which dips gently toward the southwest (Carlson 1967). Sandstone of the Tertiary Paskapoo Formation is found only toward the southwest of the study area (Kathol and McPherson 1975). Late Tertiary erosion by rivers draining toward the northeast excavated preglacial valleys, for example the Beverly and Onoway valleys (Kathol and McPherson 1975), which extend through the central and northern parts of the study area.

Sediments of the Empress Formation, locally termed the 'Saskatchewan sands and gravels', were deposited in these preglacial valleys (Bayrock and Hughes 1962; Stalker 1968; Kathol and McPherson 1975). The nearest bedrock source for the quartzitic clasts in this formation is to the west in the Cordillera. This led Tyrrell (1886) and Antoniuk (1954) to suggest that the 'Saskatchewan sands and gravels' are outwash sediments transported eastward from the mountains by large meltwater rivers.

Surficial geology and geomorphology

Recent publications and reports on the surficial geology of the Edmonton region are numerous (Hughes 1958; Bayrock and Hughes 1962; Bayrock and Berg 1966; Westgate 1969; Bayrock 1972; Kathol and McPherson 1975; Andriashek 1989; Shetsen 1990; Godfrey 1993). However, only a few of these cover the entire study area.

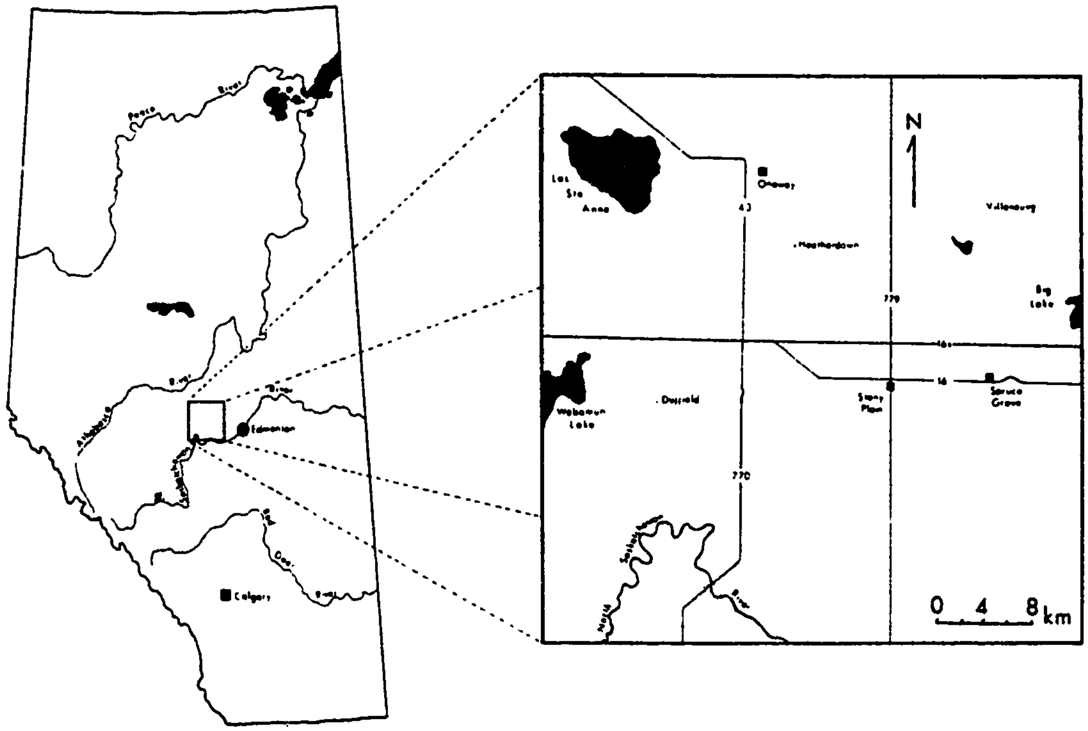

Fig. 2.1. Location of the study area showing major lakes, rivers, cities, and highways.

Coarse silt to medium sand predominate to the east where they are mapped as ice-contact lacustrine, deltaic and fluvial deposits associated with recessional phases of glacial Lake Edmonton (Bayrock 1972; St-Onge 1972; Shaw 1975, Andriashek 1989; Shetsen 1990). A large area of hummocky topography, composed of stratified gravel, sand and silt, west of Stony Plain and a smaller tract south of Big Lake were previously interpreted as pitted deltas. The origin of these deposits is addressed later as a major question regarding the evolution of glacial environments in the Edmonton region.

Draped moraine and ground moraine occur in the northeast (Bayrock 1972; Shetsen 1990), but the thickest and most extensive till deposits in the study area are found to the west around Duffield and northwest around Lac Ste. Anne (Shetsen 1990). Topographic characteristics of the area vary from the extremely hummocky topography of the so-called pitted deltas to nearly flat-lying northeastern areas capped by glaciolacustrine silts and clays. Several isolated channels in the region include a major anabranching paleochannel system (Bayrock and Hughes 1962; Shetsen (1990) which dissects a gently undulating plain north of the North Saskatchewan River, leaving upstanding erosional remnants. Several nested channels are also mapped near the western margin of the study area. These channels begin north of Heatherdown and may be traced southward along a rectilinear boundary between the Duffield moraine to the west and the large hummocky sand complex to the east.

Satellite Imagery

Satellite imagery is especially useful in the analysis of terrain under snow cover where surface features are small and landforms normally obscured by the various reflectances of vegetation are more easily observed. Using snow-cover imagery, Skoye (1990) detected 'mega-flutes' between the Spondin channeled scablands and the Sheerness high in south-

Fig. 2.2 Digital Elevation Model (DEM) of the study area showing glaciotectonism at (1), (2) and (3). Also shown are (A) a large meltwater channel; (B) a series of inset meltwater channels associated with eskers; (C) a diamicton mass superimposed on a meltwater channel; (D) a small sinuous ridge; (E) a large sinuous ridge; (F) an anabranching channel system; and part of the North Saskatchewan River at lower left.


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central Alberta that had not been previously documented because of their extremely low relief. The appropriately named 'Snow Scenes' are most effective at relatively low sun elevations, when even small features cast a shadow. When the land surface and its vegetation is covered by snow, shadows accentuate relief and enable detection of otherwise unobservable features.

The principal geomorphological use of satellite imagery is to observe spatial relationships between individual geomorphic features and to discuss them as landform assemblages. A Landsat 5 image used in this study was acquired December 12, 1985 when much of the region was covered by a thin blanket of snow. Snow cover and low sun angle, about 11.2° above the horizon, reduced potentially obscuring reflectances from vegetation and increased the detectability of subtle landforms.

Modelling Techniques

Three-dimensional modelling of topography also aids the interpretation of regional landscapes. Topographic maps depict general three-dimensional shapes via contour lines, but 3-D models of landforms allow ready visualization of forms and their spatial relationships. By observing landforms in this manner, we gain insight into their formation. A technique termed 'hillshading' makes use of various lighting effects on digital elevation models (DEMs) which aid the detection of low-lying features. A DEM of the study area provides a hillshaded model of the topography (Fig. 2.2). Of particular interest is the identification of sinuous landforms with associated marginal fans and branching ridges superimposed on a hummocky sand complex, nested meltwater channels, and an anabranching system of channels.

Local Topography

Sinuous Landforms, Marginal Fans, and Branching Ridges

The Duffield Sand Complex is discernible on satellite images and DEMs as a large tract of hummocks, approximately 25 km across from east to west and 50 to 60 km across from north to south. Hundreds of lakes are distributed throughout the complex, occupying inter-hummock lows and isolated, oriented depressions at the base of numerous small channels. Concordant surfaces in the hummock zone are few except at the eastern margin where the relief is subdued and oriented sand flutes and hummocks are common. Hummock amplitude generally decreases from north to south, except in the northeastern part of the area where smaller, low-centered hummocks occur in diamicton.

A regularly sinuous ridge oriented parallel to the western margin of the hummocks is differentiated from individual hummock forms near Mink Lake (E on Figs. 2.2 and 2.3). This ridge is approximately 12 km long, 500 m wide and has smaller ridges and fans branching from it. In places the marginal fans overlap. A second, more subdued, sinuous ridge to the south, measuring 8 km in length and about 500 m in width, displays similar textural characteristics on satellite images and photographs as the first ridge. It also has marginal fans but its sinuosity is less regular than that of the first ridge.

Cross-laminated fine sand in the upper 3 m of the main ridge indicates paleo-flow from northeast to southwest, parallel to the ridge. Unfortunately, there are no extensive exposures for detailed sedimentological investigation. However, a well-log of ridge sediment records approximately 10 m of sand and gravel beneath 2 m of surficial silt. Three additional borehole records from neighboring farms indicate occasional lenses of gravel beginning at about 5 m from the surface.

A map of landform curvature derived from a DEM of the main ridge and surrounding area shows a number of overlapping ridges extending from it (Fig. 2.4). Where dark tones

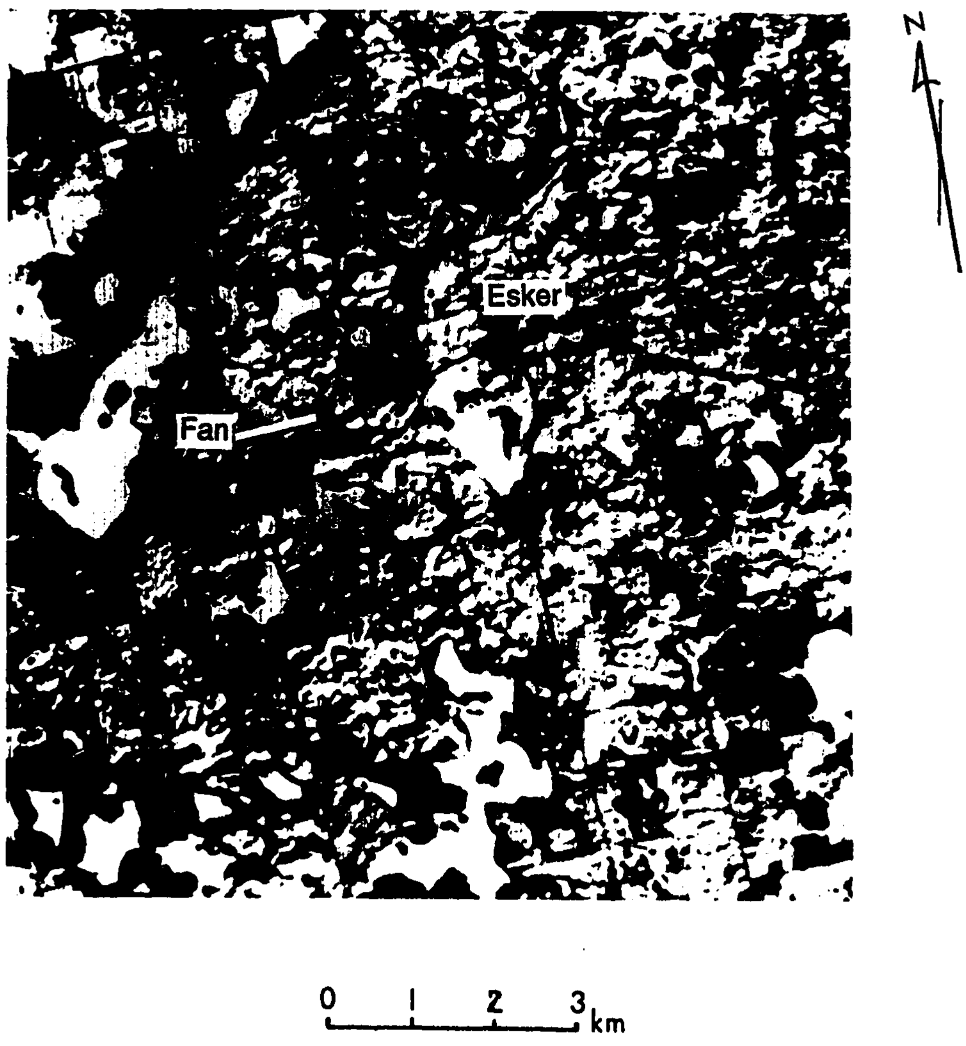

Fig. 2.3 Landsat 5 satellite image showing the main part of a large esker on the Duffield Sand Complex. Note a major fan at the outside of the second left hand bend, and a smaller ridge joining the main esker through the lake at the second right hand bend. The surface texture of the esker is smooth whereas the surrounding topography is hummocky.

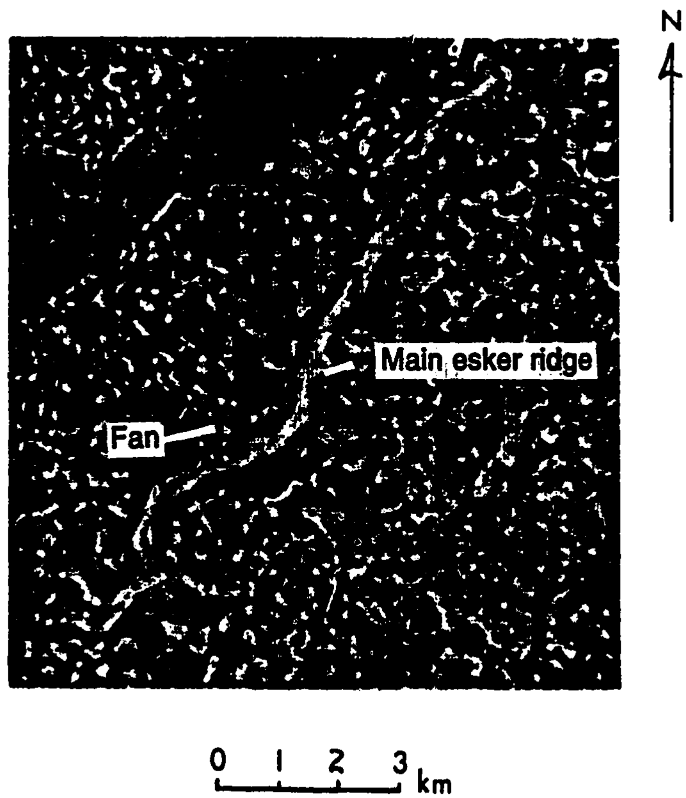

Fig. 2.4. Landform map of down-slope curvature generated from a DEM of the main esker on the Duffield Sand Complex. Note the northwestward trending ridge appearing to the right of the 'Fan' label. The area shown on this landform curvature map corresponds with the area shown on the satellite image in Fig. 2.3.

on the image represent areas of concave-up topography, lighter tones represent areas of convexity. Linear landform crests are illustrated by continuous thin, dark lines between high curvature (lightly shaded) areas. The main ridge, which trends from northeast to southwest, is easily recognized by a thin dark line representing its crest. Dark lines branching off to the east and south mark the crests of smaller features. One ridge (Fig. 2.5 at center) extends southward from the main ridge over another ridge trending eastwards.

A landform map (Fig. 2.6) produced from aerial photographs, DEMs and satellite images shows distinct differences between the main ridge, which does not contain isolated depressions, and the surrounding topography in which depressions are common. Marginal fans occur along the length of the main ridge and vary in shape from lobate to triangular, and tend to taper away from the ridge (Fig. 2.6). Fan sediments are thickest closer to the ridge and thin toward the edges where individual fans may overlap.

Fans usually join the main ridge at the outside of bends, but sometimes branch from both sides at the same point on the ridge. Where two fans join it in this manner, the ridge surface elevation drops considerably (A on Fig. 2.6) and undulating anabranching ridges are superimposed on the fans perpendicular to the main ridge. The southern set of anabranching ridges (Fig. 2.6) is well defined in outline and occupies an intermediate position between the main ridge and a more subdued one to the south.

At its northern end, the main ridge ends abruptly, the topography changing from the gently undulating ridge surface to a series of subparallel ridges and swales (B on Fig. 2.6). The ridges and swales trend perpendicularly to the main ridge and are located on top of the lowered surface and continue only a short distance over a small marginal fan.

Minor parallel to subparallel ridges occur at numerous locations along the main ridge and on some fans. At D and E (Fig. 2.6), smaller ridges on marginal fans trend at 90° to the main ridge and parallel to the long axis of the fans. At F, G, and H (Fig. 2.6), larger ridges trend parallel to the main ridge and in places form an anabranching pattern.

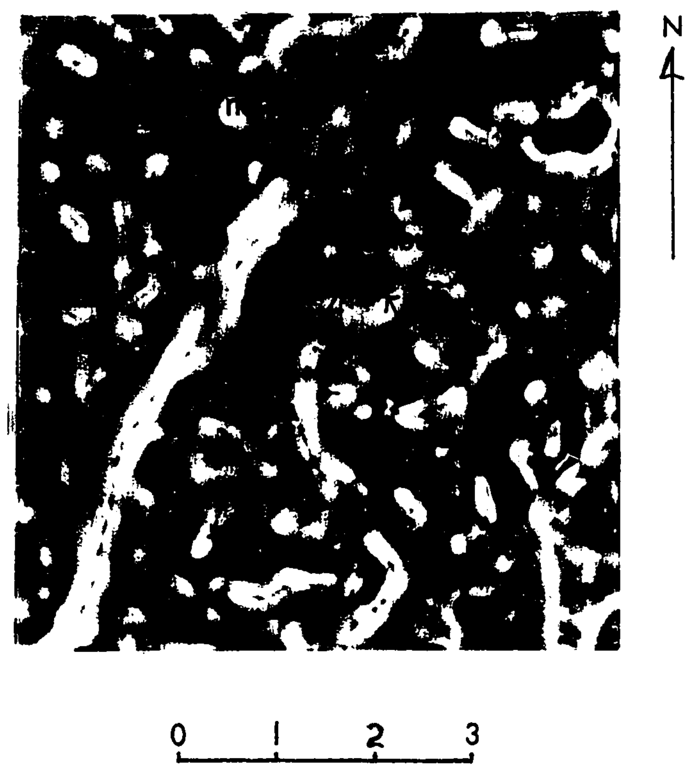

Fig. 2.5 Enlarged view of the down-slope curvature map of the northern part of the esker. Note the main ridge has the highest curvature of the surrounding landforms (lightest). Note also the flatter marginal fan area at bottom left of the esker, and overlapping distributary ridges at center. Such ridges cover the entire central portion of the Duffield Sand Complex.

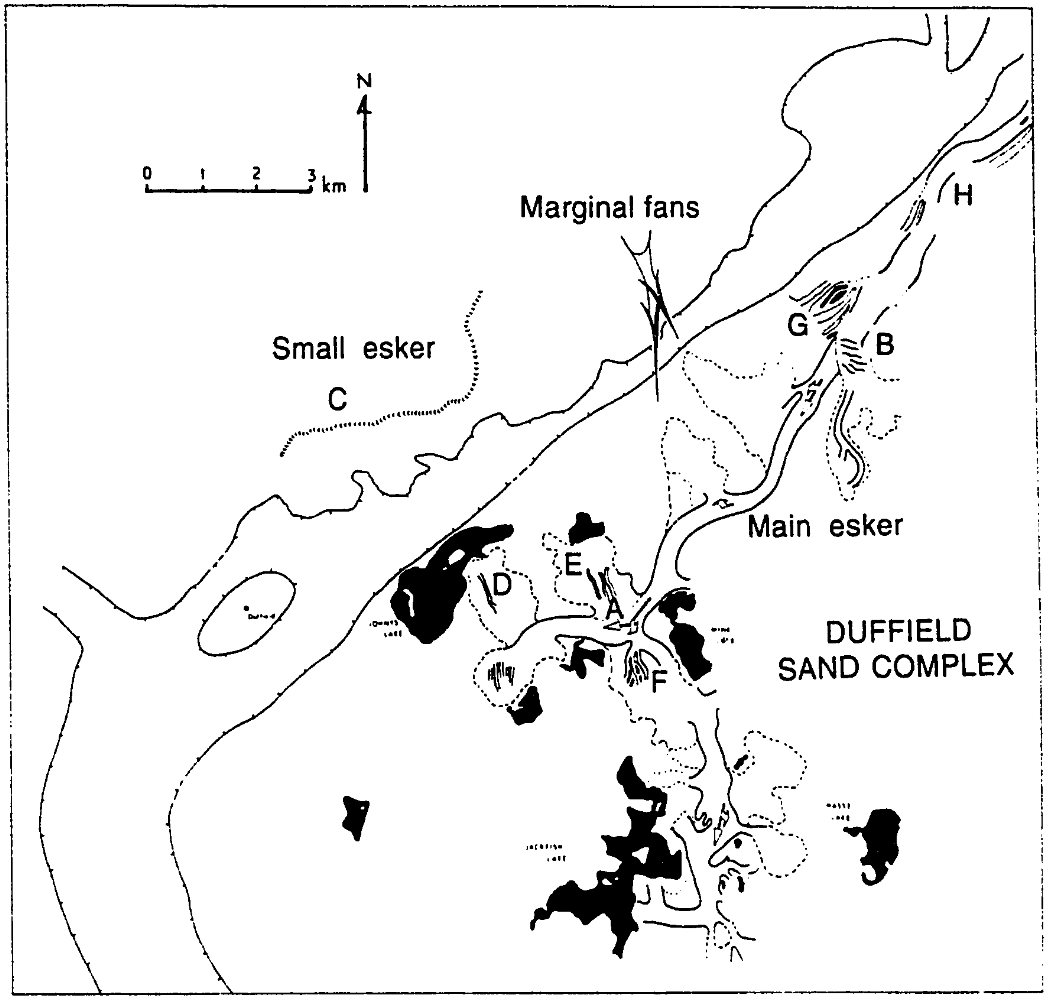

Fig. 2.6. Landform map mainly compiled from aerial photographs showing principal lakes around the main and secondary eskers, all marginal and terminal fans, a large meltwater channel and smaller esker at left. Also shown is a portion of the Duffield meltwater channel to the left and bottom left.
A: Depressed esker surface at junction of two fans. B: Sub-parallel ridges and swales on a marginal fan. C: Esker over diamicton mass. D and E: Ridges on marginal fans. F, G, and H: Larger ridges parallel to main esker ridge and in some places forming anabranching patterns.

Sinuous ridges and marginal fans - Interpretation

Röthlisberger (1968), Shreve (1972), Weertman (1972) and Nye (1976), indicate that subglacial meltwater flow is generally perpendicular to the ice margin and is driven by the pressure gradient controlled by ice-surface gradient. Hughes (1958), Bayrock and Hughes (1962), Westgate (1969) and Shaw (1982) estimated the regional direction of ice flow in the study area to have been toward the southwest. The orientation of the main ridge west of Mink Lake is, therefore, as expected for subglacial meltwater flow in a conduit given the regional pattern of ice flow. The ridge is interpreted to be an esker.

Smaller ridges seen on DEM images extending from the main esker resemble the distributary ridges associated with the Lanark and Tweed eskers, Ontario (Gorrell and Shaw 1991; Brennand, in press) which are said to be of subglacial origin. Stacking of linear ridges (Fig. 2.5) without disturbance of the surrounding terrain, can only be explained by infilling of sequentially-formed subglacial conduits.

Increased resolution provided by large-scale air photos affords details that also supports the interpretation of the main ridge as a subglacial landform. Features mapped from air photos as marginal fans closely correspond to large overlapping ridges seen on the maps of down-slope curvature derived from the DEMs. The smaller anabranching ridges to the south (F on Fig. 2.6), and the sub-parallel ridges and swales to the north (B on Fig. 2.6) are superimposed on these larger ridges.

The shape of the anabranching system between the northern and southern esker ridges resembles anabranching eskers at Livingstone Lake, northern Saskatchewan (Shaw et al. 1989). This resemblance may reflect similarities in processes of formation, though at a very different scale. The Livingstone Lake eskers are on average about 15 m in height and more than 200 m wide whereas the anabranching eskers described here attain a maximum height of 3 m and are 15 to 20 m wide. The location of these small ridges on a fan between two larger eskers suggests that their formation is related to late-stage deposition dominated by conduit flow.

Sub-parallel ridges and swales occurring on the northern lower esker surface, and above part of a marginal fan, are transverse bedforms. These large dune-type forms document an outburst flow from the main esker conduit toward the east during late-stage sedimentation within it. The event truncated the upper part of the esker and probably involved piracy, causing complete abandonment of flow through the southern portion of the main conduit in favor of the new passage.

Hummocky terrain

Theories on the formation of hummocks vary from topographic subsidence caused by ice-block melt-out (Gravenor and Kupsch 1959; Parizek 1969) to the filling of subglacial cavities either by squeezing (Hoppe 1952; Stalker 1960) or by glaciofluvial infilling (Shaw and Kvill 1984; Sharpe and Cowan 1990). The unique sedimentary characteristics of the landforms west of Edmonton suggest that neither subsidence nor squeezing were significant processes related to their genesis. Furthermore, an esker located within the hummock zone clearly postdates, or was formed synchronously with, the deposition of the hummocks and the sand complex.

Topography of the western "pitted delta" varies from gently undulating to extremely hummocky. The formation of these landforms by melting of underlying ice, as described in the theory of pitted delta genesis, is expected to produce characteristic deformation features such as large-scale faulting (McDonald and Shilts 1975; Brodzikowski and Van Loon 1991). Sand pit exposures in a hummock, however, do not substantiate hummock formation by subsidence of supraglacial sediments. Furthermore, exposures do not support deposition in a broad delta. Detailed observations leading to these conclusions were made at selected sites.

Hummock sediments

Important sedimentary evidence was obtained from the C and H Sand pit (NTS 83 G/9, UTM 985430) located 5 km north of Stony Plain. The pit is excavated near the northeastern margin of the Duffield Sand Complex (Fig. 2.2). The surrounding topography is hummocky with relief of individual landforms up to 20 m, and up to 300 m or so across. Larger features are commonly superimposed by smaller hummocks. Sedimentary sequences exposed in the C & H pit show a distinct proximal to distal transition with the proximal deposits being much coarser and including evidence of vigorous erosion. Proximal deposits are described before distal sediment to demonstrate this trend.

Proximal deposits - Description

Proximal deposits are exposed beneath the steep northern slope of the hummock (Fig. 2.7 Section 1). The proximal position is established from paleocurrent measurements which indicate flow from north-northwest. These proximal sediments are predominantly plane bedded, planar cross-stratified, and trough cross-stratified pea gravel and coarse sands. One bed contains large, undeformed soft-sediment clasts within a matrix of clast-supported gravel (Figs. 2.8 and 2.9). The soft-sediment clasts range from a few centimeters to more than 2.5 m in diameter. Their pronounced angularity suggest short transport distances; they were eroded from a position very close to the point of deposition. Paleocurrents from this section show evidence for transport under variable flow directions ranging from between northeast and northwest. The clasts were probably derived from the depression immediately to the north of the hummock, suggesting that the terrain results from both erosion of the depressions and deposition of the hummocks

The rip-up clast and gravel bed lies in an erosional trough within which the size of soft-sediment rip-up clasts increases from the margins to the centre. The size of the largest sediment clast at any vertical section of the trough usually corresponds with the maximum thickness of the sedimentary unit at that location (Figs. 2.8 and 2.9). Stratification within

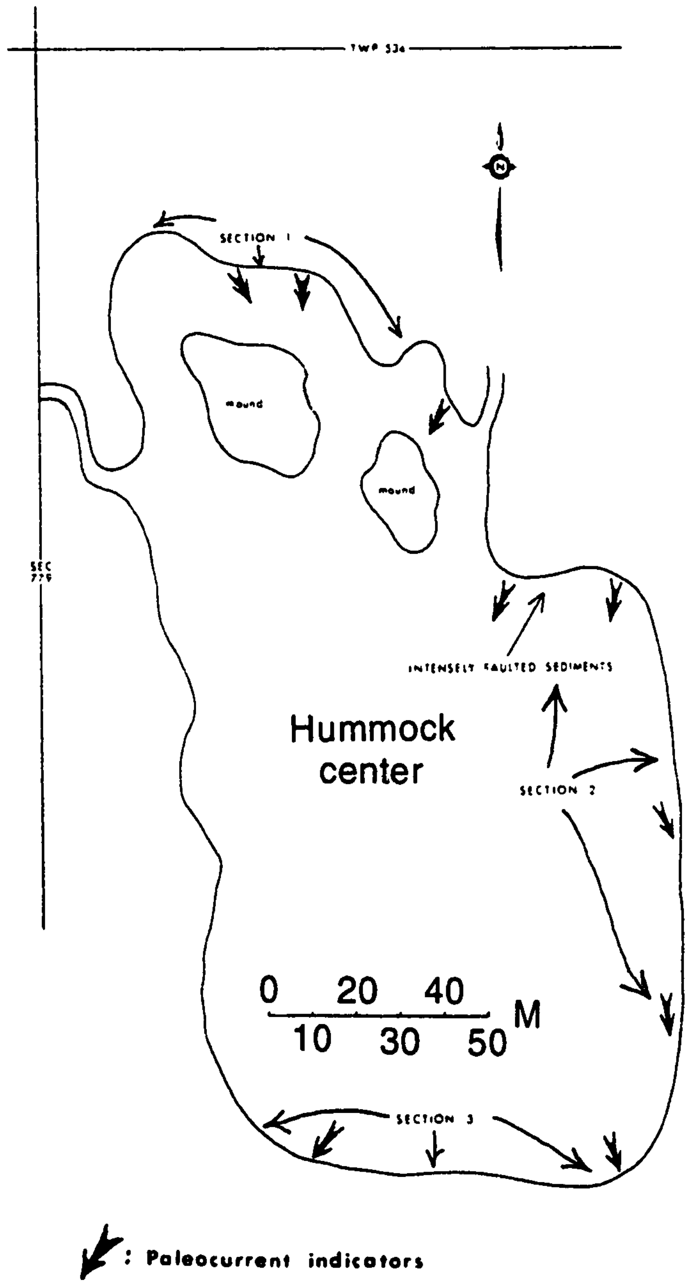

Fig. 2.7. Location of the C & H Sand pit with predominant paleocurrent directions and section locations.

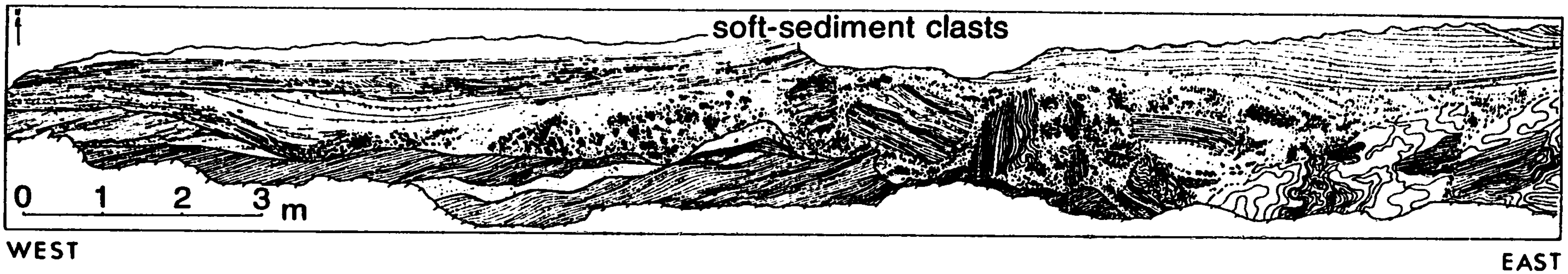

Fig. 2.8. Proximal deposits in Section 1 of the C & H Sand pit illustrating large soft-sediment rip-up clasts ranging in size from 1 cm to 3 m in length. The matrix is composed of sand (small dots) to cobble-sized clasts (large Black dots). The rip-up clast unit is capped by high flow-regime plane-beds which dip northward but are composed of southward climbing laminae.

Fig. 2.9. Photograph shows sandy cobble gravel as matrix between rip-up clasts. The unit is capped by high flow regime plane beds. Note the range of soft sediment clast sizes.

the unit is only observable on the upper west side of the exposure in trough cross-beds. They are limited to about 1 m laterally and grade into overlying coarse pebbly plane-beds.

Proximal deposition - Interpretation

The architecture of the sedimentary units in Section 1 shows that the soft-sediment clasts were deposited in a trough of little more than 4 m depth that was probably scoured beneath a broader, deeper flow. Given the internal characteristics of the hummock, the inter-hummock swale immediately up-flow of this section, and the broad hummock down-flow, any one or combination of three environmental conditions might have prevailed;

(i) the flow transporting the soft-sediment clasts and gravel had enough hydraulic head to flow over the 15 m high hummock;

(ii) the swale is an erosional scour and the hummock a complementary depositional ridge;

(iii) the original depositional surface was faulted post-depositionally into a ridge and swale topography, of which the swales, according to previous interpretations of this terrain, are kettle holes.

A subglacial hydraulic head would most easily explain up-slope transport over the proximal side of the hummock. This may also be accomplished by an unconfined aqueous flow, provided that the hummock was submerged in the flow or there was sufficient kinetic energy to carry the flow over the hummock bedform. Sediment gravity flows may generate sufficient velocity and turbulence, depending on slope angle, to scour and rip up underlying sediments while at the same time overriding minor opposing slopes (Middleton and Hampton 1976). The incorporation of large soft-sediment clasts in a turbulent, hyperconcentrated flow may be accomplished by channel wall failure or through lift forces explained by the Bernoulli principle. This principle shows that an increase in fluid velocity at a specific location causes a decrease in local pressure. Pore fluid and weakly-consolidated sediment under higher pressure may then be drawn toward a zone of lower fluid pressure caused by high velocities at flow constrictions (Fig. 2.10). Consequently, blocks of sediment may become detached from the bed.

Differences in density between the turbid flow and the soft-sediment clasts determine whether these 'float' within the flow matrix, or saltate along the bed. Excellent preservation of clast angularity either suggests that the rip-up clasts were buoyant within a hyperconcentrated fluid (Postma et al. 1988) or they were frozen, thus clast edges were protected.

Depositional mechanisms responsible for the emplacement of soft-sediment rip-up clasts in hyperconcentrated flows relate to the process of fluidization (Lowe 1976), whereby an increase in strength in the bottom part of the flow is created as water is forced upwards. Once this strength exceeds shearing stress within part of the flow that part "freezes", preserving the solid phase of the flow with approximately the same structure as it had when it was mobile. Lack of grading in the sequence and pronounced angularity of all sediment clast sizes suggest such a "freezing" process. Deposition of the unit must, therefore, have resulted from rapid and substantial deceleration of the entire sediment package shortly after rip-up.

Liquified flow initiation in a subaqueous environment is dependent on local topographic slope which must be greater than 4° (Lowe 1976). Sustenance of such flows is maintained by gravity and varies with the characteristics of the surface slope over which it travels. Minor obstacles may be overcome but larger ones, which cannot be overtopped, cause rapid deposition from hyperconcentrated suspension.

The proximal deposits (Fig. 2.7 Section 1 and Fig. 2.8) are situated on high ground a minimum of 3 km from the nearest down-flow slope, up-flow from the hummock. A subaqueous turbidity current would have been required to travel over a highly irregular topography with a gradient of less than 2.3°, which is significantly less than Lowe's (1976) 4°, without substantial deceleration for more than 3 km before coming to rest on the northern hummock slope. It is unlikely that such a normal

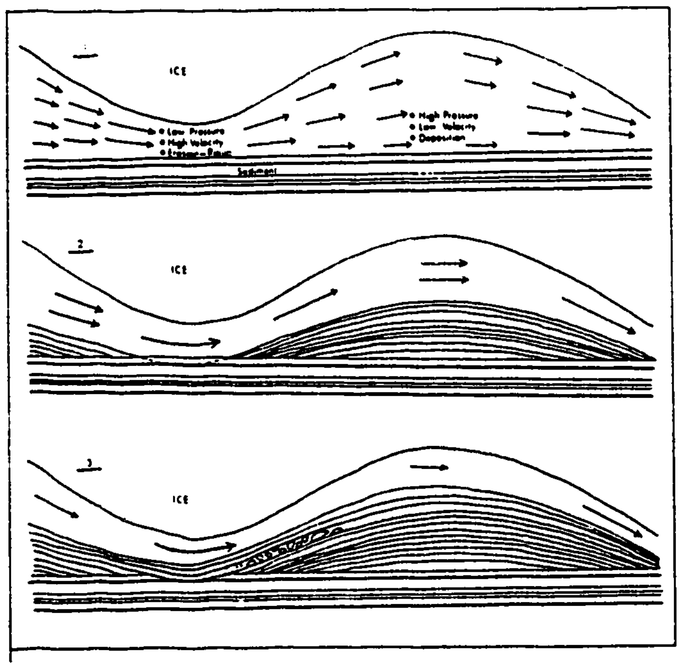

Fig. 2.10. Illustration shows the Bernoulli effect as a possible mechanism for the subglacial erosion and emplacement of soft sediment rip-up clasts.

current could exist at all, let alone generate shear stresses capable of eroding meter sized sand blocks.

Scouring may produce a swale in the form of a stoss-side trough around hummocks (Allen 1982) or sichelwannen (Kor et al. 1991). However, there must be a pre-existing obstacle for this to occur. In both cases, bedding in the hummocks should be truncated by the scour surface formed by the erosion that created the local topography. On the contrary, in the C & H pit, plane beds and cross strata in beds with reverse gradient conformably overlie and grade into the gravel and soft-sediment clast bed and dip toward the swale, closely following the surface slope of the hummock. The Section 1 sediments, therefore, do not show evidence of post-depositional scouring on a scale necessary to produce a 7 m swale; they indicate hummock formation largely by accretion.

The production of ridge and swale topography by faulting of a continuous aggradational surface, be it of deltaic or other origin, necessitates the presence of faults or other evidence of subsidence in the sediments. Yet, Section 1 is not disturbed over its 25 m length by faulting, nor is there any folding of the beds. Thus the topography cannot be related to differential subsidence or glaciotectonics.

Strata of Section 1 reflect the surface slope of the landform, and paleocurrents indicate flow up and over the landform. Because there is no folding or faulting of the proximal sediments, the swale and hummock are considered complementary primary forms, erosional and depositional respectively. They are not products of post-depositional faulting and subsidence. The environment of deposition for the sediments at Section 1, based on the need for a hydraulic head and the conditions for hyperconcentrated suspension sedimentation, must have been subglacial. The primary nature of the hummock and swale indicate that the formative flow submerged both. The energy necessary to drive the flow over the hummock could, therefore, only have been derived from the potential energy in a confined subglacial system. Consequently, the hummock is believed to have been formed by infilling of a subglacial cavity.

Distal deposits - Description

Distal deposits (Fig. 2.7, Sections 2 and 3) are well exposed approximately 50 m south of the proximal deposits of Section 1 (Fig. 2.8). The exposed section extends for 60 m from north to south. The distal sediments mark a major change in grain size and structure from the proximal deposits. Unfortunately, sediments between the two are not exposed, and actual transitions were not traced.

The lower 4 m of the distal sediments are composed of sequences of alternating cross-laminated fine sand and thin beds of parallel-laminated and cross-laminated silt. Two clay beds, each about 2 cm thick and approximately 1 m apart, extend continuously over several tens of meters at approximately the same elevation. Fining upwards below the clay beds is confined to only about 3 cm. Below that, the alternating beds do not show any obvious vertical trends in structure or grain size.

Faulting is quite extensive in Section 2 (Figs 2.7 and 2.11), but cannot account for the relief of the landform. Macdonald and Shilts (1975) propose that differential compaction in thick sedimentary sequences may produce faulting. Detailed analysis of both inter- and intra-formational normal faults in the distal units reveals that displacement occurred mainly below the highest point of the hummock, not at its margins, and accounts for only 1.5 m of accumulated vertical displacement. Maximum fault displacement in the zone of maximum sediment thickness is as expected for faulting caused by sediment consolidation.

A lateral facies change within the distal sediment (Fig. 2.7, Sections 2 and 3) demonstrates that the relief is predominantly a result of continuous primary sedimentation. These sections contain sedimentary units which fine and thin distally. Climbing ripple drift in medium sand is the most common facies in the lower sequence of Section 2, but it rarely occurs in Section 3 where climbing ripples in fine sand and silt dominate. Moreover, the thickness of individual ripple cosets also decreases southwards. The decrease in total thickness of the lower sequence between Sections 2 and 3 is between 0.6 and 0.7 m, which

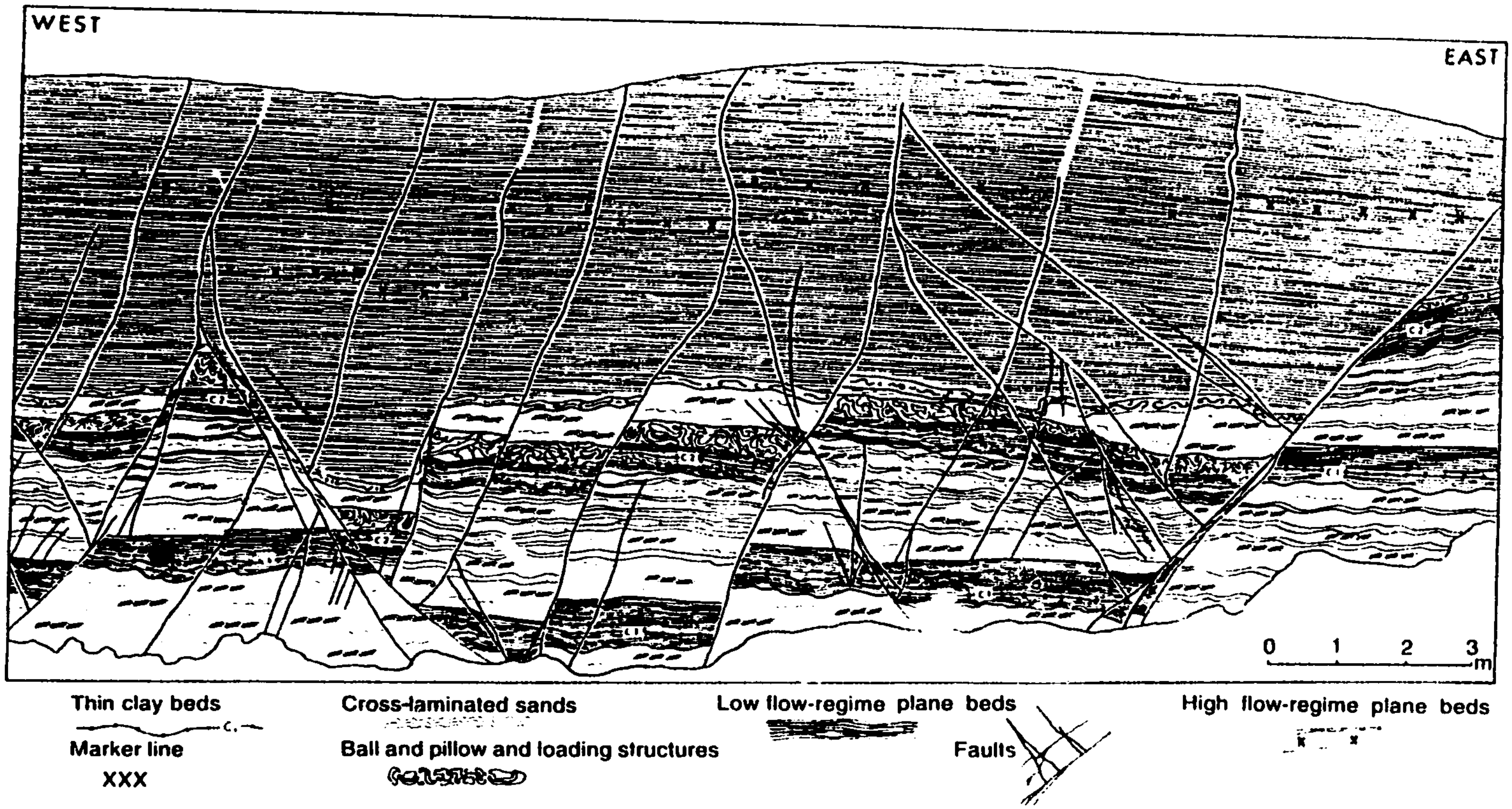

Fig. 2.11. Distal deposits in Section 2 of the C & H Sand pit. The right of the diagram is closest to the hummock margin whereas the left is nearest the hummock center. Note that normal faulting is away from the right hummock margin and is greatest at center. C1 and C2 are clay bands, and X's are merely used as a marker bed to illustrate faulting of the upper sediments.

only partly accounts for the total drop in elevation along the section of more than 1.5 m. This is a minimum estimate for the total drop since faulting below the hummock crest actually decreased the primary relief of the landform.

The upper sequence between Sections 2 and 3 further explains the primary relief. This sequence is composed of upper-flow-regime, plane bedded, coarse sand of fairly constant grain size, although there is a minor decrease in grain size distally. The plane bed laminae, or beds, vary in thickness from about 3 grains to more than 2.5 cm and may have either erosional or conformable lower contacts, the former of which predominate at Section 2 where thick cut and fill sequences are common (Fig. 2.12). The total thickness of plane beds in the upper part of Section 2 is approximately 4 to 4.5 m, whereas they are only 3 to 3.5 m thick in the equivalent part of Section 3. This minimum difference in plane bed thickness accounts for an additional 1 m of surface relief between the section locations. Thus, the observed relief can be accounted for entirely by primary sedimentation.

A pronounced erosional contact between the upper and lower sequences of Section 2 is marked in places by deep troughs filled by massive or crudely bedded medium sands which grade upward into plane beds in coarse sand. These cut-and-fill sequences may relate to the same flow that eroded and redeposited the more proximal soft-sediment rip-up clasts. Though the lack of continuous exposure prevents verification of this, the magnitude of the erosional flow is exceptional in both sets of deposits.

Faulting of the most distal sediment (Fig. 2.7, Section 3) is minimal and shows a maximum total displacement of only 30 cm. However, these sediments are disturbed by intense diapirism caused by loading of the upper clay layer. Diapirs occur as dikes which range in height from 0.25 to 1.0 m, and 0.25 to 2.0 m in width (Fig. 2.13). Small-scale faulting of the surrounding sediments is associated with the diapirism. Fault traces longer than 1 m on vertical sections are rare, although one fault extends from the base of the exposure to the top of the largest diapir, but is truncated by plane beds above. The faulting and diapirism are clearly intra-formational.

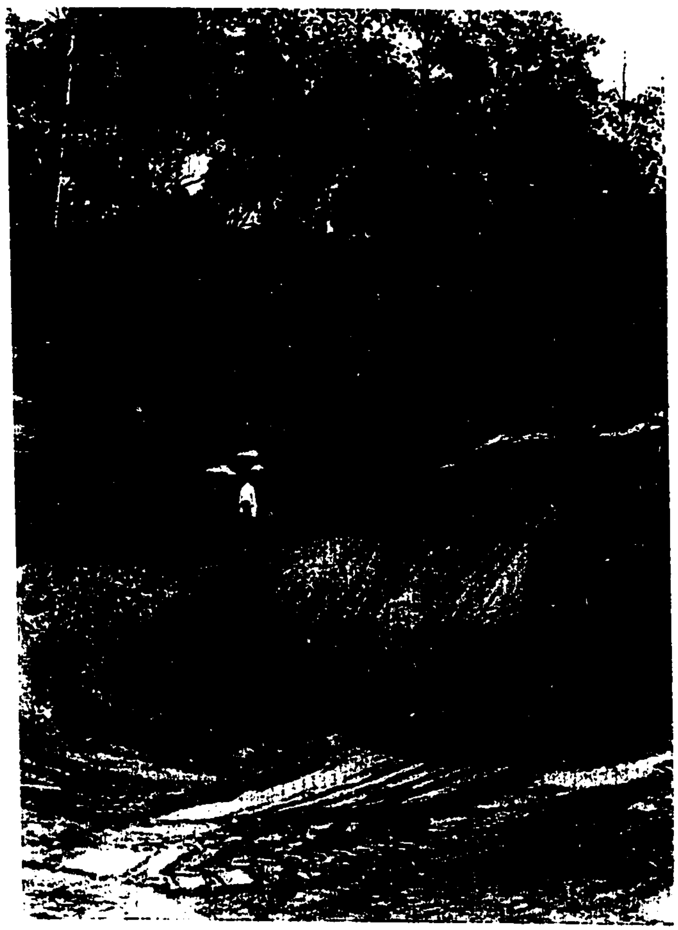

Fig. 2.12 Scour and fill structures near the top of the ripple bedded sand sequence of Section 2 in distal sediments, immediately below high flow regime plane beds. The scours may be synchronous with rip-up and deposition of soft-sediment megaclasts on Section 1.

Fig. 2.13. Diapir in Section 3 with central injection column. Micro-faulting occurs in sediments to the left and right of the diapir. Faulting does not occur above, and indicates that the finer silt beds below were loaded by only 1 m of sand, and the diapiric material was therefore extruded into open water. This is confirmed by down-flow entrainment of the extruded sediment.

Distal deposition - Interpretation

The sediment exposed in the C and H Sand pit shows a progression in depositional conditions from north to south along a 200 m transect. Erosive flows are inferred at the proximal hummock margin producing the rip-up clasts at Section 1. Scour surfaces are also present at the northern most part of Section 2. They are associated with coarsest sediments exposed at the hummock medial and distal sites, and are represented by the southern portion of Section 2, and Section 3 where fine sands and silts were deposited.

The drop in elevation of approximately 2 m between Section 2 and Section 3 is accounted for by differences in primary accretion. Erosion may have also affected the relief in addition to the contribution from accretion. Deposition on a reverse slope (backset beds, Johansson 1973) conformable with the hummock surface at Section 1 supports the conclusion that the hummock may have resulted entirely by deposition. Any contribution to the relief of the landform by erosion is difficult to estimate. Thus, the hummock resembles the megaforms and complex primary bedforms in eskers described by Brennand (in press).

Thin, horizontal clay bands traced from the base of Section 2 to Section 3 represent temporary but total cessation of flow. The very thin fining-upwards sequence immediately beneath the lowest clay band records a relatively sudden decrease in flow. Such flow cessations may indicate that the subglacial hydraulic system was linked to, or controlled by, supraglacial meltwater sources with annual discharge cycles, or that input conduits were subject to rapid conduit closure (Gorrell and Shaw 1991).

The coarsening-upward sequence in these sections, from ripple bedded sands and silts to plane beds in coarse sand, indicates a change in flow conditions over time from a relatively low to high flow regime. This change in regime may reflect an increase in flow discharge for the upper deposits or it may be a result of local constrictions between the ice cover and the bed. For the case of constriction, as the cross-sectional area of a flow decreases at a faster rate than discharge, fluid velocity and bed shear stress increase causing erosion and transport of coarse sediment in high-flow-regime bedforms.

As deposition progressed subglacially, probably over a surface equivalent to the area of the Duffield Sand Complex, and the ice was lowered towards its bed, it is reasonable to assume that cross-sectional area decreased and progressively coarser sediments were deposited under increasing flow velocities. At this stage, thermal and mechanical erosion of the ice bed produced inverted scallops (Shaw 1983). Rip-up of the proximal, soft-sediment clasts and scouring of the swale resulted from increased shear stresses and high pressure gradients. Continued sediment deposition within the scallops accentuated erosion of the basal ice and deepened the swales to form the swale and hummock landscape. This process continued until the hummock-sized cavities through which the meltwater flowed were filled with sediment or until they were sealed by settling of the ice. It may be that the scour and plane beds of the upper sequence represent a single and final drainage event immediately preceding let-down and slow in situ melting of the ice.

Two sizes of subglacial cavities are proposed to have existed within the area occupied by the Duffield Sand Complex. The larger one encompassed the complex and may have extended further east. It is bounded by considerable glaciotectonic landforms to the north, west and south (discussed later). Though the Duffield Sand Complex is inferred to have been deposited in a large cavity well behind the glacier front, it may equally represent progressive glaciofluvial and lacustrine deposition under an ice shelf. Sediments exposed at the C & H sand pit would then illustrate erosion and deposition dominated by a hydrostatic flow regime closest to the grounding line. Decreasing amplitude of the hummocks from north to south within the complex is therefore explained by the decreasing effects of the hydraulic head away from the source. These smaller subglacial cavities are believed to have been numerous and covered the entire basal surface of the ice above the sand complex. Once the hummock-sized cavities beneath the ice were closed off and completely filled with sediment, meltwater drained through a system of conduits, the largest of which is illustrated by the Mink Lake esker.

Nested meltwater channels, sinuous ridges and diamicton

A low-angled artificial light source on DEMs reveals a wide channel south and north of Duffield (A on Fig. 2.2) and a series of inset channels immediately north of Heatherdown (B on Fig. 2.2). The Duffield channel forms a broad boundary between the Duffield Moraine to the west and the Duffield Sand Complex to the east. A 12 km wide area of hummocky diamicton crosses this channel (Andriashek 1989; and C on Fig. 2.2). A 7 km long esker (Shetsen 1990) winds across hummocky terrain which also overlaps the meltwater channel (D on Fig. 2.2 and C on Fig. 2.6). However, the sediments upon which the sinuous ridge is superimposed do not have the tone and textural signature of the diamicton, as seen on aerial photographs, and are probably of a different composition.

The contacts between the Duffield channel, the diamicton, and the sediments associated with the esker provide additional information on their genesis. Throughout its length, the floor of the meltwater channel is remarkably flat and its walls are well defined. Where there are hummocky sediments within the channel margins, the floor rises abruptly into ridge and swale topography. In plan view, the channel tapers northeastward as the hummocky terrain encroaches from the northwest.

The entire surface of the diamicton is comprised of low-centered hummocks, otherwise known as prairie mounds, doughnuts, humpies, or rim-ridges (Hoppe 1952; Gravenor and Kupsch 1959; Stalker 1960). Theories on the genesis of rim-ridges include formation by dewatering processes, ice-pressing, topographic inversion (Parizek 1969), or the formation and collapse of open system pingos (Bik 1967). None of these has been unconditionally accepted. Such hummocks within the study area are not limited to particular sediments, but are most abundant over diamicton.

The series of meltwater channels to the north is composed of four channels, two of which are inset, one is a hanging tributary, and another cross-cuts the first two inset channels. The channels are perched on a high, arcuate, preglacial ridge that extends from west of Gladu Lake to Wabamun Lake (B on Figs. 2.2 and 2.14). Their widths increase from 300 m to about 2 km, and channel banks are generally well defined. Integration of the channel system through cross-cutting and inset relationships permits the establishment of a relative chronology.

The largest channel of the series (A on Fig. 2.14) was presumably incised during a period of relatively high discharge. The down-flow end of its main tributary (B on Fig. 2.14) hangs above the main channel base and was, therefore, only in use during the highest phase of meltwater discharge. An esker (Shetsen 1990; and C on Fig. 2.14) at the mouth of a tributary of the hanging channel may indicate that these channels functioned as subglacial meltwater conduits. At the base of the largest channel is a smaller inset channel (D on Fig. 2.14) which presently carries an underfit, southward-draining stream. This smallest channel was evidently formed during the latest stage of meltwater drainage through the largest channel. The last of these integrated channels (E on Fig. 2.14) truncates the northern ends of the two inset channels and, according to morphometry of its tributaries, drained towards the northeast. It was the last channel formed during meltwater evacuation of that area.

Nested meltwater channels, sinuous ridges and diamicton - Interpretation

The nested channels were created by the flow of subglacial meltwater through conduits under grounded ice on the preglacial ridge near Heatherdown. The esker-channel pair, tributary to the largest channel, could only be produced in subglacial meltwater conduits. The esker, therefore, was formed by infilling of a Röthlisberger conduit, and the channel into which it feeds may be a Nye channel (Paterson 1981).

Interpretation of the Duffield channel is made difficult because of its partial burial, but this may be a key to understanding its genesis. Both subglacial and supraglacial meltwater sources may produce discontinuous meltwater channels. If the meltwater channel were of subglacial origin, such a sharp discontinuity may have been easily caused by masking of the channel by superimposition of diamicton from melting of overlying ice. If of proglacial

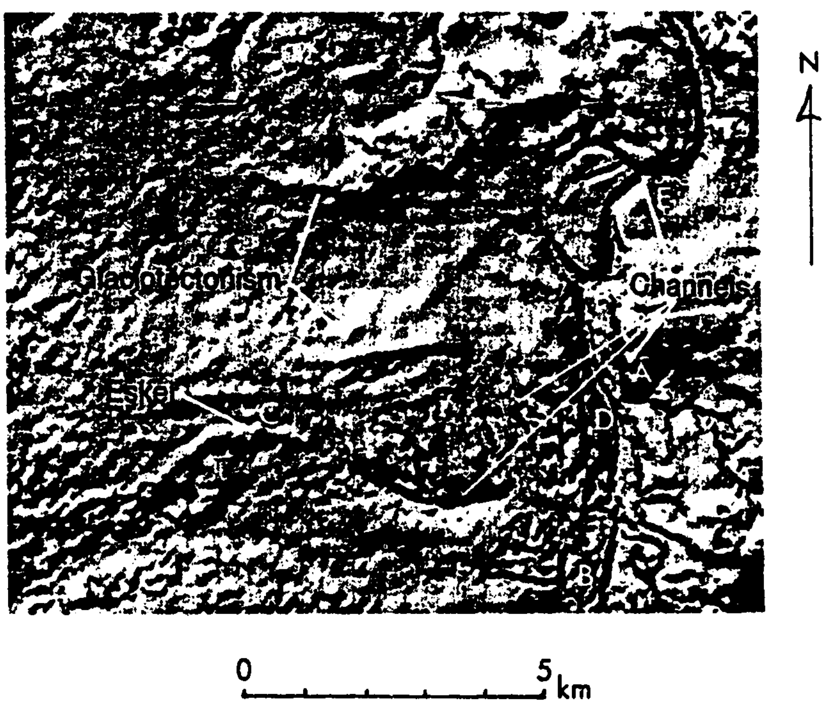

Fig. 2.14. Hillshade map of inset channels near Heatherdown. Inset, cross-cutting and tributary relationships are used to establish a relative chronology. A: the largest channel in the series. B: the down-flow end of the main tributary to the largest channel. C: an esker associated with a tributary of the largest channel. D: the smallest inset channel. E: a channel that cross-cuts the largest channel of the series.

origin, its abrupt northern termination may simply reflect the location at which a supraglacial river flowed from an ice margin onto the land surface. Upstream capture of the supraglacial river might then explain the disappearance of the channel to the north. However, the peculiar position of the diamicton at the precise location of channel disappearance seems too coincidental. Air photo analysis and field reconnaissance reveals that the diamicton is superimposed by or on a 7 km long, sinuous esker (Shetsen 1990) which, in turn, rests on hummocky glaciofluvial sediments within the meltwater channel.

The channels and associated landforms may, therefore, document an evolution of the subglacial landscape through time, beginning with a large southward-flowing subglacial river postdating deposition of the Duffield Sand Complex. The powerful flow eroded into bedrock forming a Nye channel with pronounced margins. With a decrease in meltwater discharge and lowering of basal ice, the esker formed in a relatively small Röthlisberger conduit. Shaw (1983) suggested a similar succession for tunnel valleys and eskers in northern Saskatchewan.

The diamicton (C on Fig. 2.2) over the central portion of the channel may have been deposited from down-wasting ice of a lobe which advanced across the channel, blocking drainage and causing ponding to the north. A small lake about 50 km^2 in size formed (F on Fig. 2.2), and later drained southwards across the hummocky diamicton creating a 6 km long perched channel, much smaller than the Duffield channel. This small channel is confined to the eastern margin of the diamicton, suggesting a probable origin related to diamicton emplacement.

Glaciotectonism

Ice-thrust features are commonly arranged in arcuate belts of linear to curvilinear, parallel to sub-parallel, sharply-crested ridges whose concave sides face away from the ice (Tsui et al. 1989). The latter characteristic presupposes that the physical characteristics of the ice and processes forming the tectonism are constant, and does not allow for variability in such environments. Tectonism at the frontal margin of an ice lobe, along a sinuous margin, or at a protrusion at the base of the ice, may produce parallel, curvilinear ridges with concave sides facing into the direction of ice flow. The characteristic preferred here for determining ice-flow direction from glaciotectonic landforms is their tendency toward gentle stoss slopes and steeper lee slopes.

A series of five, consecutive, arcuate ridges are identified on the DEM, northeast of the Duffield Sand Complex, immediately north of Gladu Lake (1 on Figs. 2.2 and 2.15). The ridges span an area about 10 by 3 km and are up to 40 m high. They rise abruptly at the margin of a relatively flat lacustrine plain adjacent to the hummocky sand complex. These landforms trend from northwest to southeast, perpendicular to the estimated direction of ice flow in the region (Rains 1969; Westgate 1969; Shaw 1982) and have steep, south-facing slopes. Fractured quartzite clasts and thrust bedrock intercalated with till immediately north of Gladu Lake attest to extremely local high glacial stresses. Other ice-thrust sediments and bedrock occur: (i) directly south of Wabamun Lake (Tsui et al. 1989), to the southeast and southwest of the lake where large-scale landforms occur; (ii) northwest of Heatherdown (Babcock et al. 1978); (iii) at the city of Edmonton, and (iv) north of Pigeon Lake.

A single arcuate glaciotectonic ridge southeast of Wabamun Lake (2 on Fig. 2.2) is approximately 30 km long by 8 km wide and up to 80 m high. It is sharply crested, and has a steep down-glacier slope and a more gently dipping up-glacier slope. Another, northwest of Heatherdown (3 on Fig. 2.2) is substantially smaller, rising to approximately 25 m and may be up to 1 km long (Babcock et al. 1978). Analysis of DEMs of this region, however, shows that the ridge studied by Babcock et al. (1978) is, in fact, part of a discontinuous ridge that extends an additional 3 km to the west.

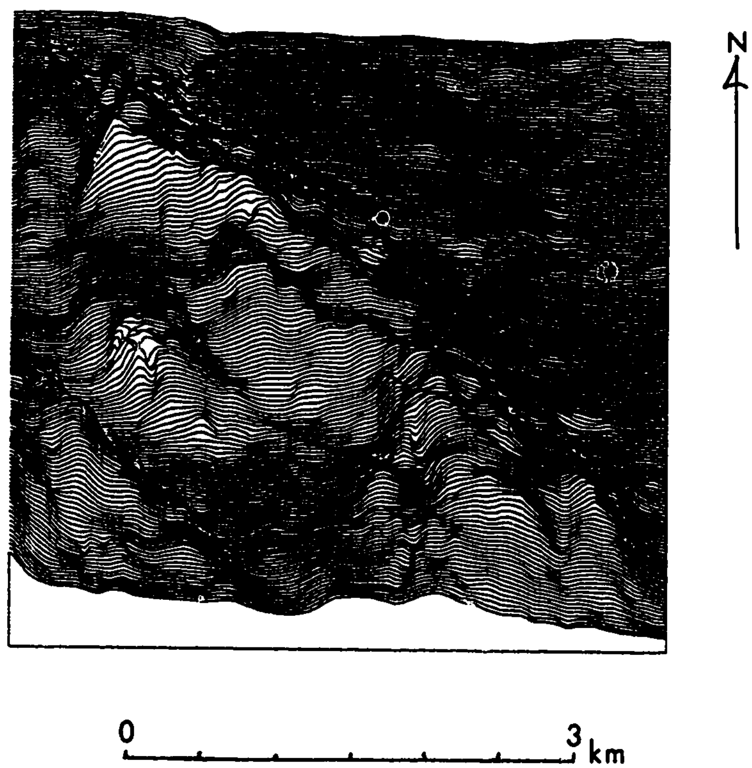

Fig. 2.15. Perspective plot of the glaciotectonic ridges north of Gladu Lake. Five consecutive ridges are differentiated in the right-hand set. A minimum of two additional ridges occur at bottom left, only partially shown. They are the result of high, localized, subglacial stresses.

Glaciotectonism - Interpretation

Though the outline of the Duffield Sand Complex closely approximates the boundaries of the larger subglacial cavity discussed earlier, the position and characteristics of the surrounding glaciotectonic landforms illustrate ice-grounding positions around the cavity. The northern edge of this cavity is outlined by the sharp-crested ridges at Gladu Lake and near Heatherdown. As is shown by their limited extent, the ridges were formed by extremely localized, high basal shear stresses, probably caused by extending ice flow over the central part of the cavity. Down-flow convergence of landform trend poles to planes onto the central cavity axis may suggest that the extending flow over the cavity was significant enough to effectively draw-in ice from either side of the main flow. The resultant glaciotectonized ridges at the up-flow end of the cavity are, therefore, not perpendicular to the central axis of the cavity, but are perpendicular to the direction of ice flow toward the axis. No evidence of glaciotectonism exists between the two sites because meltwater flowed through that area, separating the ice and bed.

The lateral and down-flow cavity margins are also marked by glaciotectonic landforms reflecting the effects of extending glacier flow over the cavity. The southernmost and largest of the glaciotectonic landforms best illustrates this process. This landform is believed to be the eastern continuation of tectonized sediments south of Wabamun Lake studied by Tsui et al. (1989). The 80 m high ridge, at its westernmost extremity, trends roughly perpendicular to the axis of the proposed cavity. Toward the axis of the cavity, however, where ice flow would be at a maximum, the ridge bends sharply and is gradually attenuated southwards.

Anabranching Channels and Residuals

A network of channels east of the Duffield Sand Complex and west of Edmonton was mapped by Shetsen (1990) as a dendritic system, though Bayrock and Hughes (1962) described them as braided or anastomosing. Analysis of DEMs, satellite imagery and air photos (F on Fig. 2.2 and Fig. 2.16) confirms the anabranching pattern, covering a minimum area of 175 km^2. Bayrock and Hughes (1962) thought the channel system to be an earlier course of the postglacial North Saskatchewan River draining toward the northeast, but they did not justify this interpretation.

The term 'anastomosis' has been traditionally used to describe a depositional environment in which levéed channels divide and rejoin around broad, lower interchannel areas. Anastomosing channels described by Bretz (1969) and Kehew and Lord (1986) include those formed by unconfined flows that swept around and eroded obstacles within the flow. Their use of the genetic term 'anastomosis' is inappropriate for a flow whose formative processes and physical characteristics differ from those for which the term was coined. The channels west of Edmonton are anabranching and are strikingly similar to the tunnel valleys of Woodland (1970), Grube (1983), Boyd et al. (1988) and Barnett (1990).

Large-scale anabranching channels on the Scotian Shelf, interpreted by Boyd et al. (1988) as tunnel valleys, have "steep valley walls, variable depth along channel, hanging tributary valleys... and commonly bifurcate and rejoin". The Scotian Shelf tunnel valleys are two to three times wider than those described here, yet the interchannel residuals are quite similar; compare for example the channels in Figure 16 with those in Figure 2 of Boyd et al. (1988). Unfortunately, the Scotian shelf anabranching system is as yet unmapped in its entirety and it is difficult to make a full comparison with the anabranching channels west of Edmonton.

Anabranching channel characteristics

Several channel and residual characteristics indicate that the present morphology cannot be explained by factors controlling modern fluvial anastomosis.

Anastomosis is observed in rapidly aggrading, low gradient, low energy fluvial systems (Smith 1983). Interfluve residual surfaces in the study area are predominantly flat to convex up. Most of these large residuals have nearly flat surfaces, and none display the

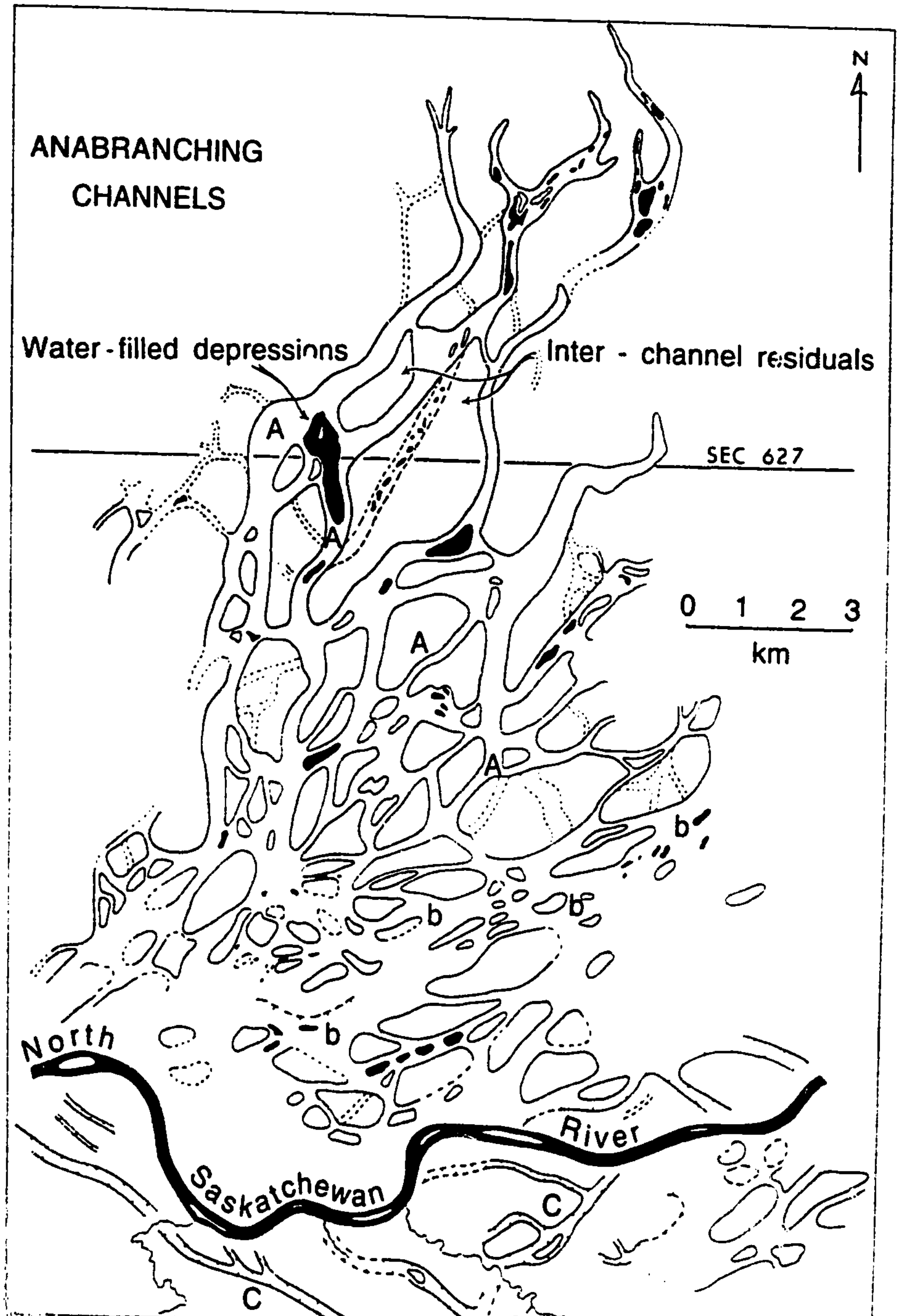

Fig. 2.16. Landform map of anabranching channel system south of Stony Plain and Spruce Grove, compiled from aerial photographs. Note crest-line points (A) that form a linear boundary between up-slope and down-slope channel gradients (approx. 1 m/km) over 20 km. Also note superimposition of residuals by smaller channels, low lying area composed of smaller residuals (B), and proto - North Saskatchewan River channels (C).

concave-up surfaces or levées commonly observed in modern interchannel areas of anastomosing rivers (Smith 1983).

Average gradients observed for anastomosing reaches of the Saskatchewan and Columbia rivers range from 0.000096 to 0.00012 (Smith 1983). For comparison, elevations along the floor of a 20 km long, north to south oriented channel of the anabranching system reveal an average up-flow (reverse) gradient of 0.0009 for the northern-most 10 km, and an average down-flow (normal) gradient of 0.001 for the southern-most portion. These measurements were made on a 2 m contour interval topographic map of the anabranching area constructed from a DEM. All channels within the system have similar configurations of gradient. Up-flow channel gradients of this magnitude are explained only by water flowing under pressure. Depth of the westernmost anabranching channel increases southward from less than 0.5 m south of Stony Plain to more than 18 m just north of the North Saskatchewan River.

Postglacial rebound has not been locally assessed because there are no suitable shorelines or other reference planes. The northeast-southwest orientation of anabranching channels, however, dictates that postglacial rebound would only have decreased an originally steeper reverse gradient.

Points placed on each major channel at the crest separating uphill and downhill slopes align northwest to southeast, perpendicular to the estimated direction of ice advance over the region (A on Fig. 2.16). These points form a crest line that also marks the boundary between flat-bottomed channels to the north and V-shaped channels to the south. Isolated depressions deeper than about 2 m occur exclusively within flat bottomed channels north of the crest line. North-south differences such as these may be caused by syn- or post-formational deposition of fluvial sediment, deglacial lacustrine sedimentation, or by subglacial fluvial erosion. Steepened residual margins occurring only north of the crest line, however, mark a difference in the amount of lateral channel erosion between northern and southern reaches that is not accounted for.

Satellite imagery reveals a relatively flat area in the southeastern part of the anabranching channel zone. It contains multiple, low-relief landforms trending east northeast to west southwest that do not appear to be separated by the distinct channels (B on Fig. 2.16) found elsewhere in this zone. This group of smaller, oriented landforms separates the major anabranching zone to the north from its southern continuation to and across the North Saskatchewan River. Though a dominant flow direction is difficult to estimate from the satellite images of the smaller forms, DEM analysis of landform morphometry and slope consistently indicates an east northeasterly flow at a gradient of 0.012. DEM analysis has also aided the detection of a low-relief trough at the center of the smaller oriented forms sloping eastward with a moderate gradient of 0.06.

Larger anabranching channels and residuals to the south occur mainly north of the North Saskatchewan River but also extend to the southeast. To the south and southwest, distinct proto-North Saskatchewan River channel scars dissect the landscape, trending west to east (C on Fig. 2.16).

Anabranching channel genesis

The anabranching channels in the study area drained a confined basin to the north, which extended to the pre-existing Sturgeon Channel. Shaw and Kvill (1984) and Rains et al. (1993) postulated that during the last glaciation, vast amounts of subglacial water swept over the region from north to south. During the late stages of this so called "Livingstone Lake event", rapidly flowing water is thought to have become channelized either in pre-existing valleys or to have eroded new channel systems. Ponded water in local subglacial basins was subjected to high overburden pressures imposed by the overlying ice mass. The resultant hydraulic head forced water towards the ice margin through small conduits at the ice-bed interface (Röthlisberger 1968), creating steep-walled channels. Flow through these channels may have been prolonged by continued subglacial drainage into the basin from such conduits as the Sturgeon Channel.

With down-wasting and recession of the ice, the anabranching channels and residuals would have become progressively exposed to subaerial conditions and normal fluvial activity. Flowing water in the channels with normal gradients south of the crest line incised typical V-shaped channel floors. Free of subglacial confinement, multiple channels then ceased to operate and a single channel formed and simply carried flows under gravity. This single, large channel is perched above the present North Saskatchewan River.

Alternatively, both north and south channel segments may have been originally V-shaped, those to the north of the crest line being infilled later by lacustrine sediments of Glacial Lake Edmonton, or a phase therein (cf. St-Onge 1972). This alternative requires that Glacial Lake Edmonton did not extend south of the crest line. Draped lacustrine sediments over the entire area of the channels (Shetsen 1990) contradict this alternative. Nevertheless, the channel gradients of the anabranching system could not have formed in any other than a subglacial environment.

Free-flowing meltwater then drained northeastward over a low-lying area southeast of the anabranching channels, forming a braid plain. Continued northeastward drainage over the region produced a single dominant channel, now part of the North Saskatchewan River.

Discussion and Reconstruction of the Paleoenvironment

Subglacial meltwater sheet floods during the Late Wisconsinan glaciation, first suggested by Shaw (1983) have not been widely accepted. The meltwater hypothesis began as a form analogy between erosional marks and drumlins (Shaw 1983) and evolved into a much broader view of subglacial discharge (Shaw et al. 1989). Kor et al. (1991) and Rains et al. (1993) then proposed subglacial meltwater flows about one hundred kilometers wide. Regional interpretations such as these are primarily based on landforms and landform assemblages, and on swaths of extensive scouring that are difficult to explain by direct glacial processes.

Reconstruction

The preglacial landscape of the region west of Edmonton was formed in Cretaceous bedrock and weakly-consolidated, Tertiary sands and gravels. The central and eastern portions of the area are dissected by a wide and deep valley which Kathol and Mcpherson (1975) refer to as the ancestral North Saskatchewan River Valley. This is equivalent to the preglacial Beverly Valley (Carlson 1967). The ancestral Sturgeon Valley, in which part of the Sturgeon River presently flows, extends from northeast of Gibbons southwestwards to Big Lake, where it joins a remnant of the preglacial Beverly Valley (Andriashek 1989). Its southern valley wall becomes subdued near Big Lake whereas the northern wall extends westward and ends abruptly at the Duffield Sand Complex.

The creation of a large subglacial cavity east of the Duffield moraine was caused by blockage of the Sturgeon subglacial meltwater channel. As the hydraulic head increased within the cavity, ice was lifted from its base and the volume of water in the cavity increased. Ice remained grounded on the preglacial ridge to the north, near Heatherdown, where several conduits gave rise to esker-channel pairs. Ice also remained grounded to the northeast, south and southeast of Wabamun Lake where large glaciotectonic ridges were formed. Glaciotectonized sediments south of Wabamun Lake are also discussed by Tsui et al. (1989). Similar glaciotectonic ridges immediately north of Gladu Lake also indicate grounded ice, as may thrusting of bedrock at Pigeon Lake to the south. Glacial thrusting in areas of grounded ice around the cavity was caused by the transfer of basal shear stresses from hydrostatically-uplifted areas of the ice sheet, which offered no resistance and carried no basal stress, to cavity margins where the substrate was deformed.

Subglacial sedimentation

Various eskers identified on DEMs and on satellite imagery are closely associated with the hummocky topography west of Stony Plain. Detailed air-photo interpretation reveals several marginal fans at bends on either side of the largest esker (termed the Mink Lake esker), located in the Duffield hummocky sand complex. Three-dimensional computer assisted modelling of the esker and surrounding area illustrates a stacked network of ridges coincident with the largest marginal fans. Stacking of marginal ridges in this way is explained only by sequential development of subglacial distributary conduits.

Detailed sedimentary analysis of the C and H Sand pit hummock at Stony Plain indicates a primary subglacial genesis for the hummock. Three major sections studied show a north to south, proximal to distal, fining of sediments and a coarsening-upward sequence along a 200 m exposure. Angular, soft-sediment rip-up clasts in the northern part of the hummock, and the accretional sedimentary architecture indicate hummock formation by subglacial meltwater with sufficient hydrostatic head to overtop the landform.

Relief exhibited by the hummock is accounted for by variation in the thickness of sediment which was controlled by flow regime variations in a hummock-sized subglacial cavity. Coarsening upward, with the coarsest beds in the upper 4 m of the hummock sediments, is directly related to the formation of the landform. Sedimentary filling beneath the floating part of the glacier, or lowering of the ice towards its bed, decreased the cross-sectional area through which meltwater flowed. Consequently, for constant discharge, the flow velocity and shear stress on the bed increased, permitting the erosion of scallops at the scale of hummocks on the underside of the ice. Infilling progressed causing upward erosion into the base of the ice and possible accentuation of existing scallop forms. Increasing flow velocity also permitted transportation and deposition of coarser material in high-regime plane-beds which appear near the surface. Intense erosion upstream from the hummock mobilized large soft-sediment clasts and formed a deep swale. This process continued until sudden cessation of flow in the cavity.

The lower 4 m of the sedimentary sequence in the C and H pit documents environmental conditions during sedimentation within a regional-scale, subglacial cavity, or below an ice shelf, which probably extended from Duffield to St. Albert and from Heatherdown to beyond the North Saskatchewan River in the south. Parallel-bedded fine sand intercalated with thick, cross-laminated cosets in medium sand, illustrate sequences of deposition related to flow events that may either be related to seasonal fluctuations, suggesting a supraglacial water source (Brennand, in press), or to opening and closing of meltwater input channels (Gorrell and Shaw 1991).

During the last stages of cavity sedimentation in the western zone, large-scale scours eroded upwards into the ice were infilled by a relatively broad, sediment-rich meltwater flow from the north. As accretion within individual scours in the ice formed hummocks, the glacier came to rest on most of the bed, and meltwater was confined to single conduits, forming eskers and marginal fans superimposed on the Duffield Sand Complex.

Drainage of the southern part of the regional cavity was directed through a system of anabranching channels south of Stony Plain, just north of the present North Saskatchewan River, following settling of the ice onto its bed. Undulating channel long profiles, with reverse and normal gradients of about 1 m per kilometer over 10 km for north and south segments respectively, leave little doubt as to their subglacial origin.

Final drainage of the northeastern, lower part of the subglacial cavity, which may still have been fed by the Sturgeon Valley following abandonment of the anabranching channels, was through a wide, shallow trough trending southeastwards from Spruce Grove (Holden and Shaw in prep.). The trough cross-cuts hummocky sands south of Big Lake as well as southwestward-trending residuals and isolated depressions extending from the Sturgeon Valley west of Big Lake. A 10 km long esker west of Yekaw Lake, trending southeastwards within the trough, also supports this interpretation of subglacial events.

The lowest in a series of four channels northwest of the Duffield Sand Complex near Heatherdown may have been formed by proglacial meltwater during deglaciation. An esker (Shetsen 1990) feeding into a tributary of the highest channel in the series indicates that it may have formed subglacially, or adjacent to the meltwater conduit output during deglaciation. Furthermore, the large meltwater channel, forming the main boundary between the Duffield Moraine and the Duffield Sand Complex (A on Fig. 2.2), is locally superimposed by diamicton. The diamicton mass either covers an esker or the esker was deposited over the diamicton near the southern edge of the mass (C and D respectively on Fig. 2.2). This suggests one of two alternatives:

(i) the meltwater channel operated subglacially and diamicton and a small esker were subsequently deposited in the channel.

(ii) meltwater flowed supraglacially to an ice margin adjacent to the esker where it eroded only the eastern side of the north to south-trending channel, leaving untouched the esker, glaciofluvial sediments, and diamicton in the channel to the immediate west.

The inferred genesis of the large channel and associated landforms between the Duffield Moraine and the Duffield Sand Complex implies that a highly erosive subglacial river which first cut the main channel. Deposition and preservation of debris melted out from overlying ice indicate that flow in the channel ceased. Deposition of an esker, whether it be of subglacial or englacial origin, marks ice stagnation and melting. Finally, an increasing volume of ponded meltwater, either proglacially or subglacially, behind the diamicton 'high' breached the diamicton obstruction, cutting a smaller, short-lived channel.

Ice then receded northwards impounding a succession of small lakes, referred to as the phases of Glacial Lake Edmonton (St-Onge 1972). A large number of drainage channels were formed subaerially by meltwater from the retreating ice and ice-marginal lakes across parts of the Duffield Sand Complex and to the northeast (Holden and Shaw in prep.), and from isolated bodies of stagnant ice throughout the region.

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CHAPTER 3

A Transgressive Model of Subglacial and Ice-Marginal Environments: Landforms North of Edmonton, Alberta

Introduction

In this paper I describe and interpret landforms and sediment in an area north and west of Edmonton to gain insight into the processes of their formation and the environmental conditions when they were formed. These landforms were formed by a combination of subglacial and ice-marginal processes and mark events during the deglaciation of the area. Though some of the landforms are unquestionably of direct glacial origin, others were formed by meltwater flowing subglacially and at the ice margin. Some landforms such as the Yekaw Lake and New Lunnon eskers are confidently interpreted to demonstrate the importance of meltwater activity in the formation of this terrain.

Glaciofluvial deposits such as the Duffield and Big Lake Sand Complexes west of Stony Plain and south of Big Lake attest to the volume of water once occupying the region. Eskers on the Duffield sands, and resting on other sediments previously described as glaciolacustrine and/or deltaic, suggest a more complex depositional history involving progressive subglacial fluvial sedimentation (Holden and Shaw in prep). The latter hypothesis is supported by the sedimentary structure of hummocks which indicate sedimentation by flows under hydrostatic pressure, and by landform associations and cross-cutting relationships.

The origin of fluted surfaces south of the upper and lower Sturgeon Valley segments, northeast of the Canadian Forces Base at Edmonton, and southwest of Big Lake, is problematic if only direct glacial processes are inferred. The role of subglacial meltwater as an erosive agent is preferred for the formation of these flutes because it has the capacity to erode and transport debris from the substrate and from basal ice while producing a sculpted terrain (Shaw et al. 1989). Major meltwater channels within the study area are interpreted to have formed subglacially, ice marginally or proglacially, based on landform associations or on the superposition of glacial and/or glaciofluvial sediments.

Regional Setting

Study location

The study area extends from west of Manawan Lake to the junction of the Sturgeon and North Saskatchewan rivers in the east, and from north of the hamlet of Waugh to the city of Devon in the south (Fig. 3.1). This rectangular area of approximately 40 by 70 km forms a north to south transect through a succession of sediment and landform types.

Bedrock morphology and geology

The bedrock of the Edmonton region forms a northeasterly-sloping surface of Cretaceous age, dissected by several preglacial channels. The extreme northeastern part of the study area is underlain by marine sandstone of the Bearpaw Formation. The remainder of the study area is underlain by brackish-water bentonitic sandstone, shale and coal of the Horseshoe Canyon Formation.

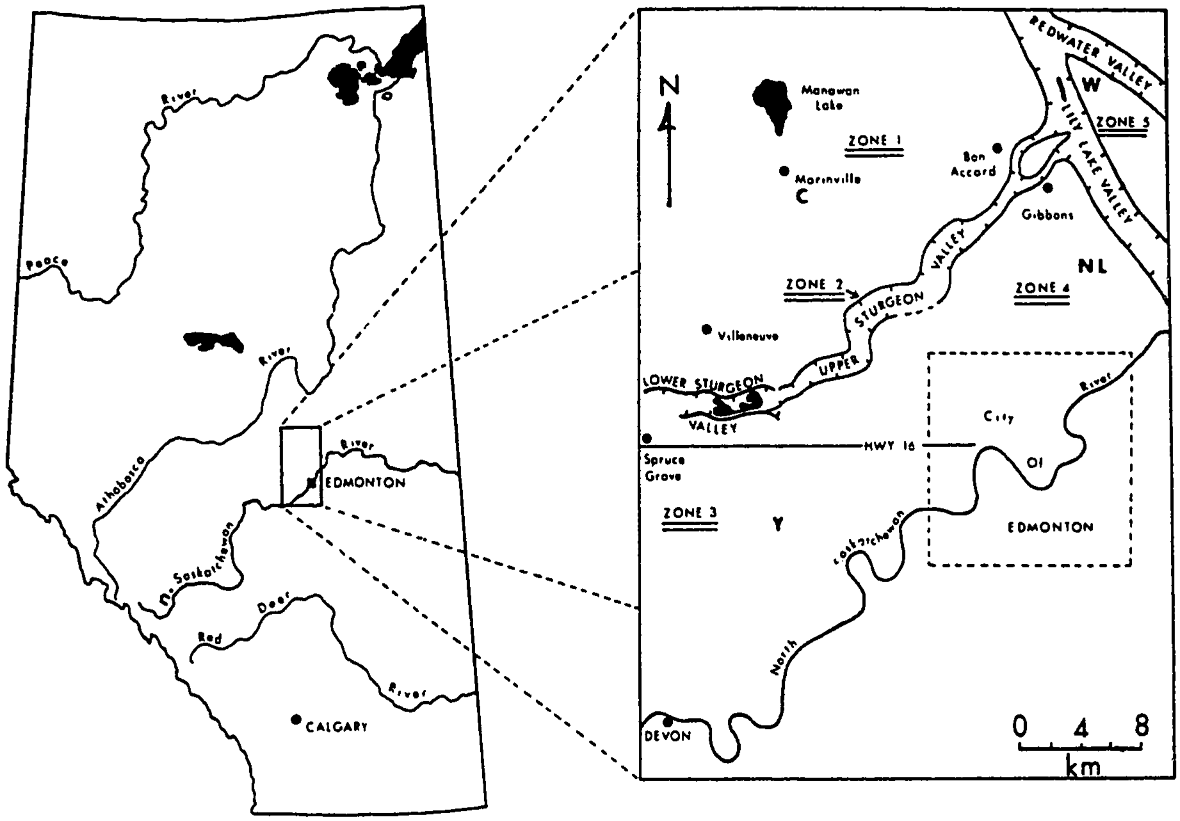

Fig. 3.1. Location map of the study area showing 5 zonal divisions based on heterogeneous landform assemblages and major valley systems.
C: Cardiff NL: New Lunnon W: Waugh Y: Yekaw Lake

Topographic setting

The surface morphology of the study area varies considerably from north to south, indicating its formation in a variety of environments. Generally, proglacial lacustrine sediment covers the area. The lakes, within which the sediment was deposited, were short-lived and their silt and clay rhythmites are too thin to appreciably mask the morphology of underlying landforms. Hummocks and eskers are common in the area along the Sturgeon Valley, south of Big Lake and east of Morinville (Fig. 3.1).

The flatter topography to the north has several landforms, including flutes and arcuate ridges with relief up to 15 m. Groups of oriented depressions are most common amongst the ridges, but also occur on relatively flat terrain to the northeast. Although the New Lunnon esker to the northeast of Edmonton is morphologically unimpressive, it is very important to the environmental reconstructions and deglacial sequences presented here.

A variety of landforms to the south reflect the meltwater processes by which they were formed. Hummocks south of Big Lake are part of a large sand and gravel complex where individual forms attain over 20 m in height. The Yekaw Lake esker located south of this complex, however, was formed in a glacial meltwater conduit.

Landforms of the Study Area

Three valleys traverse the study area; two of which neatly sub-divide the region into zones with distinctive landforms. The Lily Lake Valley and the Sturgeon River Valley are both much broader than the valley containing the present North Saskatchewan River (Figs. 3.1 and 3.2). The latter, however, is cut into bedrock throughout most of the region and, in the study area, is entirely postglacial in origin (Rains 1990).

The Lily Lake Valley (Fig. 3.1) is the most prominent valley in the northeastern study area. It begins north of the hamlet of Waugh at the junction of the Redwater River and Fairydell Creek and trends southeastwards through Lily Lake to the North

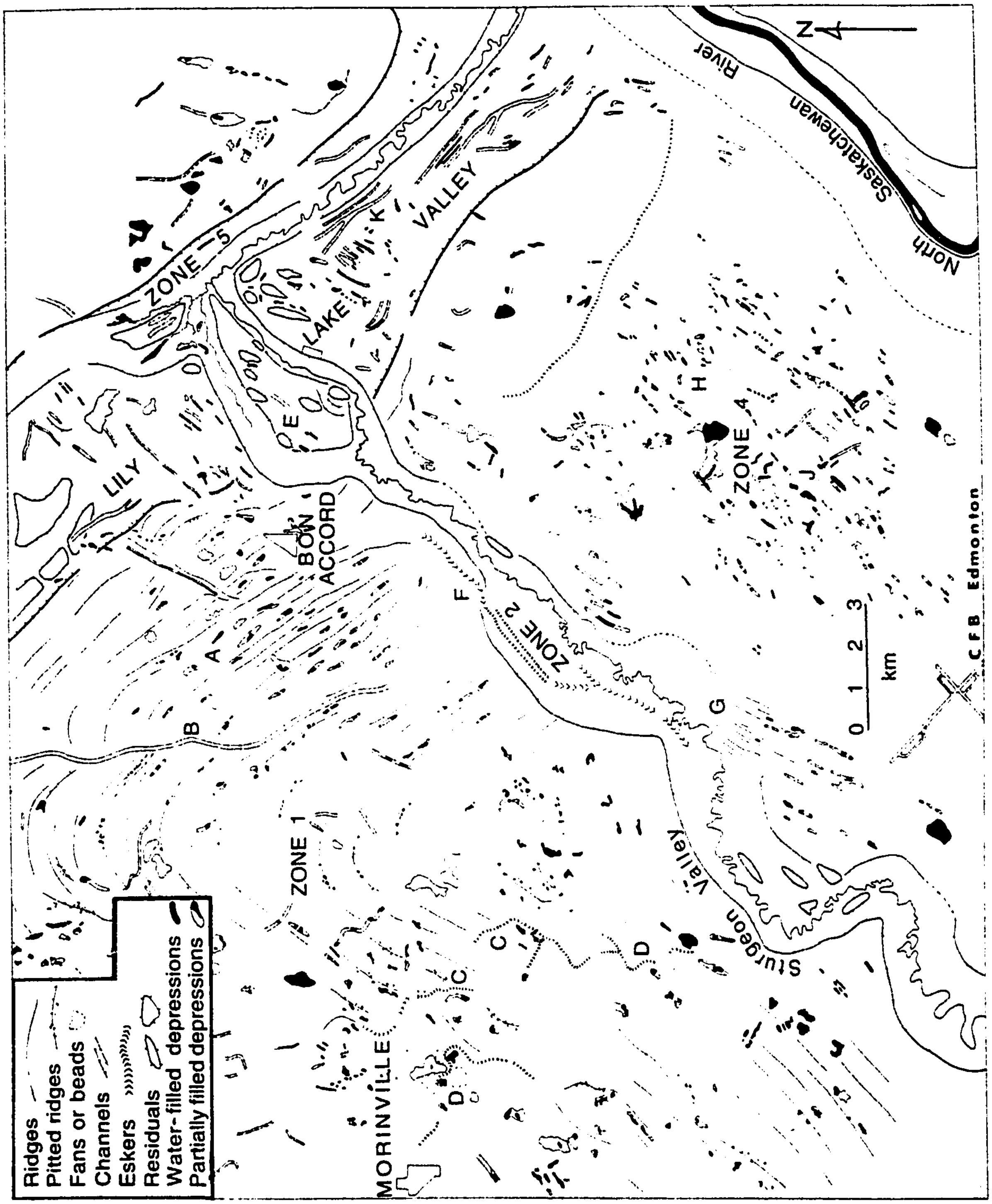

Fig. 3.2. Shows complex relationships between the landform assemblages of Zones 1, 2, 4 & 5. Zone 1: Parallel arcuate ridges and inter-ridge depressions, and eskers. Zone 2: residuals, flutes and eskers within the Sturgeon Valley. Zone 4: short, consistently oriented depressions, small meltwater channels and an esker. Zone 5: Perched channels above the south side of the Lily Lake Valley.

Saskatchewan River. The Sturgeon Valley joins the Lily Lake Valley north of Gibbons. Landforms observed around this junction and within the Sturgeon Valley itself suggest a complex valley genesis involving both subglacial and ice-marginal processes.

The region is mapped in detail and is subdivided into five distinct morphological zones (Fig. 3.1). The first is composed of arcuate ridges and swales which mantle, to varying degrees, the entire northwestern topography. Zones 2 and 3 cover the upper and lower segments of the Sturgeon Valley and the distinctive landforms contained within them. Zone 4 covers the area east of the Sturgeon Valley and south of the Lily Lake Valley and is composed of an intricate assemblage of parallel-oriented depressions. The fifth zone contains the Lily Lake Valley within which there is a series of perched channels with associated sediments.

Zone 1. Arcuate ridges and troughs

Landform morphology

The hummocky topography northwest of the Sturgeon Valley is seen on aerial photographs to form a series of moderately- to well-defined, symmetrically convex-up ridges, approximately 10 m in height. The ridges form a regional, parallel pattern including a distinct, approximately right-angled bend in the center of the ridged area. Northwest of Bon Accord, the ridges are regularly spaced at about 300 to 400 m and are aligned roughly perpendicular to the Sturgeon Valley (A on Fig. 3.2). Most of these ridges extend more than 10 km to the northwest before bending southwestward and running parallel to the valley. The southwestward-trending limbs of most ridges are sinuous and pitted where they descend from the topographic high to the east. Their southwestern limit is located at a drop in elevation and is coincident with Shetsen's (1990) western contact of glaciofluvial sand and silt, of which the ridges are composed, and lacustrine silt and clay. These ridges sometimes occur on a more or less hummocky surface without defined intervening swales. Arcuate ridges and troughs of this type also occur in three other areas: north of Manawan

Lake, where they appear prominently on satellite imagery; south of Morinville and west of Cardiff; and in a remarkable series of regressive, arcuate, troughs and ridges between the Pembina River and Westlock (see also, Fraser 1991).

The majority of the oriented lakes and isolated depressions in Zone 1 occupy inter-ridge lows immediately west of Bon Accord (A on Fig. 3.2). Water-filled depressions between the ridges are seldom more than 0.5 km long whereas the swales that contain them may be as long the confining ridges. To the southwest, oriented lakes and depressions are uncommon except where they occur along the Sturgeon Valley. Other oriented lakes in Zone 1 are found at the base of small meltwater channels. In some cases, detection of these channels on air photos was facilitated by the alignment of numerous, small, water-filled depressions.

There are few well-defined channels in Zone 1. They cross ridges at high angles, generally greater than 60° (B on Fig. 3.2). The longest of these channels extends at least 12 km, is up to 200 m wide, and trends north to south just east of the prominent bend in the ridges. Near its northern end, adjacent northwesterly-trending ridges bend westward toward the channel. Several smaller and poorly-developed channels west of the bend are also perpendicular to local ridges.

Two sinuous ridges southeast of Cardiff are mapped by Shetsen (1990) as eskers (C on Fig. 3.2). Three additional eskers have since been detected (D on Fig. 3.2). The eskers are approximately 100 to 150 m wide, up to 7 km long and may be separated into two or more segments. One of these eskers is located less than two kilometers east of Morinville and has small fans emanating from sharp bends near the center of the ridge. The fans are less than 500 m in diameter and taper away from the main ridge.

Sedimentology

Exposures in the local sediments are rare except along the Sturgeon River southeast of Bon Accord where lateral erosion has kept high river banks relatively clear of debris.

Sections in a wall of the Sturgeon River valley expose over 26 m of stratified sand and silt (Shaw 1975), locally capped by 2 to 3 m of deformed massive silt, and silt-clay rhythmites including dropstones. The exposures are near the southern end of a narrow, 25 km long sediment complex, between Vimy and Bon Accord, previously interpreted as ice-contact lacustrine and fluvial deposits (Shetsen 1990).

The complex facies successions studied by Shaw (1975) at Bon Accord Bridge, Bon Accord West Bridge and the Ridge Top section, illustrate both fining and coarsening-upward trends that suggest changing environments of deposition within the Bon Accord Sand Complex. The deposits there range from parallel-laminated silts and clays through thick units of cross-laminated and cross-bedded sands to horizontally-bedded, coarse sand. Diamicton beneath the ridges west of Bon Accord occurs between 10 and 15 m below the surface (Andriashek 1988), and between 30 and 32 m within the valley itself.

Paleocurrents observed at the Bon Accord Bridge exposures range between southwesterly directions at the base of the section to increasingly easterly and east-northeasterly directions at the central and upper parts (Log 1: Shaw 1975). Although the base of the Bon Accord West Bridge section is not extensively exposed (Logs 3 and 4: Shaw 1975), a southwesterly paleocurrent measurement was made on a cross-stratified coarse sand unit. This log shows an upward succession of paleocurrent measurements to the top of the exposure generally directed to the south-southsoutheast. The Ridge Top sections expose little more than the upper 3 m of the local sediment and indicate upper sequence flow directions toward the southwest. The nearby River Cliff section, however, exposes over 32 m of medium to fine sand cross-lamination and cross-bedding, and illustrates dominant paleocurrents toward the southwest throughout the sequence. Exposures in the eskers located east of Morinville and Cardiff are generally very poor and consist of minor sections at the surface of abandoned sand pits.

Discussion of ridges and swales near Bon Accord

Ridge and swale topography is not uncommon. Gwynne (1942) examined swells and swales of a Wisconsinan ground moraine plain in Iowa and attributed the parallel nature of the swells to the changing outline of the receding Mankato ice lobe. Most of the ridges examined were only broadly arcuate and only a few groups of ridges occurring at the lateral margins of the ridge complex, displayed wavy characteristics with sharp bends. These wavy ridges were explained by Gwynne (1942) as products of the same environment in which the other ridges were formed, the difference resulting from an extremely slow ice retreat in the case of the wavy ridges. This, however, does not actually explain the wavy pattern. Gwynne (1942) suggested that genesis of the ridges was controlled by aggradation of melt-out debris at the ice margin during the summer, followed by minor ice readvance, pushing and overriding the diamicton, during winter. In comparison, the diamicton surface beneath the ridges west of the Sturgeon Valley is between 10 and 15 m below the surface. Therefore, the ridges must have formed in a substantially different environment from those described by Gwynne (1942).

Linear disintegration ridges described by Gravenor and Kupsch (1958) are mainly composed of till but may have pockets of stratified sediments within them. They vary in height up to about 10 m and may be up to 20 km or so in length. Though the ridges are said to be predominantly straight or slightly arcuate, those referred to by Gravenor and Kupsch (1958) are quite discontinuous and are distinctly arcuate. Some (Gravenor and Kupsch 1958, Plate 6), resemble the ridges described by Gwynne (1942), yet not one resembles those west of Bon Accord. Gravenor and Kupsch (1958) considered the disintegration ridges of western Canada to have formed subglacially, at the base of thrust planes in the ice. They followed Elson's (1957) model of ridge genesis, derived from observations on ridges in Manitoba, in which preservation of the ridges resulted from stagnation of the overlying ice mass.

Other studies on parallel, arcuate ridges include research on cross-valley moraines on Baffin Island by Goldthwait (1951), who first referred to them as "sublacustrine moraines," and Andrews (1963). These 10 to 14 m high, irregularly sinuous ridges were believed to be closely associated with former glacial lakes, formed as valley glaciers receded. Fine-grained, sorted sediments in the cross-valley moraines, though, appear only as discrete deposits transverse to, and superimposed on, the moraines and are referred to as "central kames". The ridges themselves are largely composed of diamicton covered by large, angular boulders. Some ridges near the valley margins, however, are composed of poorly-bedded sands and gravels and form a branching system. Andrews (1963) postulated that cross-valley moraine ridges formed in Glacial Lake Lewis were produced, "in a frontal or sub-glacial position under approximately 400 feet of water", based on the positions of lake outlets.

Of the five possible specific interpretations of ridge genesis offered by Andrews (1963), he discredited most on the evidence of ridge morphology and sedimentology. His preferred hypothesis, though not completely accepted, involved the squeezing of sediments (cf. Hoppe 1952) into a system of basal crevasses created by increased ice-flow velocities where drawdown caused extensional flow near the ice margin. The crevasses resulted from tensional fracturing. This theory may apply to the ridges west of Bon Accord where Glacial Lake Edmonton was impounded (St-Onge 1972), but does not explain their extremely arcuate nature.

The surface position of the Bon Accord ridges clearly indicates that deposition of the sand within the ancestral Sturgeon Valley preceded, perhaps immediately, formation of the ridges and re-excavation of the valley. Deposition of the sand and silt beneath the ridges may, therefore, have either occurred subglacially, or in a glacial lake fronting an advancing glacier. Lacking a post-glacial process for the formation of arcuate ridges and troughs of the type west of the Sturgeon Valley, they can only have formed in one of four ice-contact lacustrine environments:

(i) Channelized meltwater flowing along the ice margin after Glacial Lake Edmonton had drained, may have eroded deep troughs. However, the southward slope of the local topography dictates that channelized meltwater would have flowed away from, rather than along, the ice margin.

(ii) Seasonal retreat by melting and calving of the ice front would have caused a northward succession of ice-marginal positions. Oscillating impounded lake levels may have raised and lowered buoyant ice, producing ridges of squeezed sediment at the grounding line. Two to three meters of deformed massive silt exposed at the top of the sedimentary sequence along the Sturgeon Valley at Bon Accord Bridge, beneath 1 m or so of lacustrine rhythmites, supports this hypothesis. Smith (1990) documents 10 m high pressure ridges forming up to 60 m in front of a glacier in a proglacial, lacustrine environment. The ridges are formed in this hypothesis by squeezing and deforming sediment from beneath the glacier to the ice margin.

(iii) Oscillating water levels in a subglacial cavity may have produced the ridges along the interior cavity perimeter through a similar process. These, however, could only be preserved if the cavity was in a continual period of expansion, otherwise, outer ridges would be subjected to deformation by descending ice as water levels dropped and the cavity closed.

(iv) Using a model similar to that of Hoppe (1952) and Andrews (1963), basal sediment may have been squeezed into a crevasse system at the base of the ice formed by tensional fracturing produced during successive buoyant and non-buoyant phases in a deep impounded lake.

Discussion of sediments

Shaw (1975, fig. 2) provides three rose diagrams of the paleocurrents at the Bon Accord Bridge illustrating flows varying widely between northeastwards and southeastwards. These diagrams, however, illustrate flow directions based solely on the number of observations per sedimentary unit. The resulting rose diagrams are, therefore, biased by excessive measurements taken on well-exposed, single-bedding units, and, potentially important units at the base of the sequence are underrepresented. Examination of the paleocurrents at successive elevations within the Bon Accord sequences show that there was a more orderly change in flow direction through their depositional history than Shaw's (1975) rose diagrams suggest.

Upward-changing paleocurrent directions in the Bon Accord Bridge, West Bridge (Log 1 and Log 3: Shaw 1975) and River Cliff sediments, beneath the arcuate ridges, likely reflect the changing influence of the Sturgeon Valley in a broader meltwater flow as it filled with sediment. Paleocurrent measurements taken from coarse sand cross- and plane-bedded units at the base of these sections, document deposition by a relatively high velocity south to southwesterly flow confined by the Sturgeon Valley walls. Deposition of these basal sediments was either accomplished by redirection of meltwater as suggested, or, the sediments were deposited by an earlier flow event. The upper paleocurrents in these three sections, taken together, illustrate a fan-like distribution of flow at the southeastern edge of the Bon Accord Sand Complex. Sedimentation in the upper part of the sequence was, therefore, independent of any influence by the Sturgeon Valley.

The Bon Accord West Bridge section is closest to the central portion of the regional meltwater flow that deposited the Bon Accord Sand Complex, and is composed of the coarsest sediments described in sections along the Sturgeon Valley. Multiple coarsening and fining-upward sequences in the West Bridge sediments, and the scours which they infill (Fig. 3.12), are accounted for by increased susceptibility to variations in meltwater velocity at the center of the flow. Unscoured sediments at the Bon Accord Bridge and River Cliff sections were affected by only moderated flow velocities because of their marginal location relative to the central axis of the meltwater flow.

The ancestral Sturgeon Valley was largely infilled with sediment by a meltwater flow directed toward the southeast. Meltwater, whether in a separate event or at the onset of the larger southeasterly flow, was directed along the valley and produced cross-bedded coarse sand units with paleocurrents toward the southwest. Further sedimentation within the valley was marked by aggradation of thick cross-laminated and plane-bedded medium sands which progressively diminished the valley's channelling effects. As the valley became completely filled, meltwater flowed in a splayed or fan-like manner with diverging flow west and east of Bon Accord. Rhythmic silts southeast of the Sturgeon Valley, described in Zone 4, may be distal sediments correlative to the Bon Accord Sand Complex.

Accepted mechanisms for forming the arcuate ridges superimposed on the Bon Accord Sand Complex are: squeezing of sediment beneath the ice toward the grounding line (cf. Smith 1990) or, squeezing of sediment beneath the ice into an arcuate system of crevasses (cf. Hoppe 1952; and Andrews 1963). These mechanisms suggest that deposition of the Bon Accord Sand Complex, must have pre-dated formation of the ridges, and must, therefore, have been subglacial. Though proglacial deposition of the complex followed by an active glacial advance is possible, the paucity of diamicton and any form of glaciotectonic fracturing in the sediments permits the simpler model of subglacial fluvial accretion.

The preceding section dealt with the events prior to and including the filling of the ancestral Sturgeon Valley and the subsequent formation of arcuate ridges and troughs west of Bon Accord. The next section addresses the cause and timing of the valley reincision and the formation of various landforms within it.

Zone 2. The Upper Sturgeon Valley

The Upper Sturgeon Valley extends from St. Albert to Gibbons where it joins the Lily Lake Valley (Figs. 3.1 and 3.2). Its physical characteristics and the landforms within it make it one of the most spectacular valleys in the study area. Its width varies considerably from 0.75 km to over 3 km. Though the modern,underfit Sturgeon River is less than 10 m wide and has a low northeasterly gradient (0.0004), inclined benches above the base of the valley have a southwesterly gradient ranging from 0.0015 to 0.0020. At other locations along the long profile of the valley, gradients obtained from DEMs are up to 0.02.

Extensive "preglacial" gravels of the Empress Formation nearly 30 m above the base of the Sturgeon Valley at Villeneuve (Fig. 3.1), located approximately 5 km to the north, are dated at 40 ka - 22 ka B.P. (Young et al. *submitted*). The valley is incised into these gravels and was, therefore, formed after about 22 ka B.P. The considerable thickness of sand and silt that completely infilled part of the upper Sturgeon Valley south of Bon Accord may have resulted from deposition by meltwater events contemporaneous with those which produced the Duffield Sand Complex west of Spruce Grove (Holden and Shaw in prep.).

Landform morphology

Analysis of a Landsat 5 image reveals a number of landforms recognized for the first time in the study area. These landforms include: oriented residuals, sharp-crested, sinuous and straight, sandy ridges, and flutes, and are found separately in different parts of the valley.

Oriented, tear-drop shaped residuals on a broad plain below 660 m within the broadest part of the valley between Bon Accord and Gibbons (E on Fig. 3.2) are preferentially aligned south-southwest. This alignment corresponds to the orientation of a number of other, similarly-shaped residuals lying outside the valley margins, northeast of Gibbons. Though the majority of the forms display blunt stoss (NE) and tapered lee (SW) sides indicating flow from the northeast, some display a remarkably conical shape with the apex at the stoss end.

The Sturgeon Valley shrinks to its narrowest width, 2 km southwest of Gibbons in the zone of ice-contact fluvial sediment (Shaw 1975; Shetsen 1990). Less than 3 km to the southwest, a pronounced sinuous ridge lies above, and runs parallel to, the northwestern rim of the narrow valley (F on Fig. 3.2). Where the valley widens southwest of the ice-contact deposits, the sandy ridge descends toward the valley floor and separates into two, parallel ridges. These, in turn, feed into a larger, single ridge that continues for another 4 km before terminating. The larger ridge into which the twin ridges combine is much wider than the ridge from which the twin ridges emanate. It is flanked by four, smaller, fan-like ridges which become progressively lower away from the central axis of the main ridge. The twin ridges are each about 2 km long, 100 m wide and 10 m high. These relatively sharp-crested features (Fig. 3.3), are perched between the valley base and rim, taper toward the southwest and are, at the surface, composed of fine to medium cross-laminated sand with paleocurrents parallel to the valley at 210°. Exposures of deeper sediment were not available.

Several flutes in the Sturgeon Valley just south of the twin ridges, range in length from 300 m to well over 2 km, and are up to 100 m wide (G on Fig. 3.2). Beginning near the valley bottom, they climb more than 30 m southward out of the valley and onto a higher, relatively flat plain. This set of flutes spans nearly 2 km from east to west on the south side of the valley, at the outside of a sharp westward bend. The valley rim is subdued at this location and is almost imperceptible. It was undoubtedly eroded by the same agent that created the flutings. Similar, but more isolated flutes parallel the valley along most of the southern rim.

Discussion of Zone 2 landforms and sediments

Large differences in slope direction and gradient between the river level, the valley floor at wide sections of the valley, and benches, suggest changing flow directions during formation of the Sturgeon Valley. Normal fluvial processes are unlikely to have formed benches in the valley with average estimated southwesterly gradients of about 0.0016 (1.6 m per km) when, less than a kilometer up-valley, the gradient of the same surface increases to nearly 0.0200 (20 m per km). Explanation of this extreme change in gradient by isostatic rebound can be immediately eliminated because of its magnitude. Though the relative uniformity of the basement sediments in the valley at that location (Andriashek 1988) may

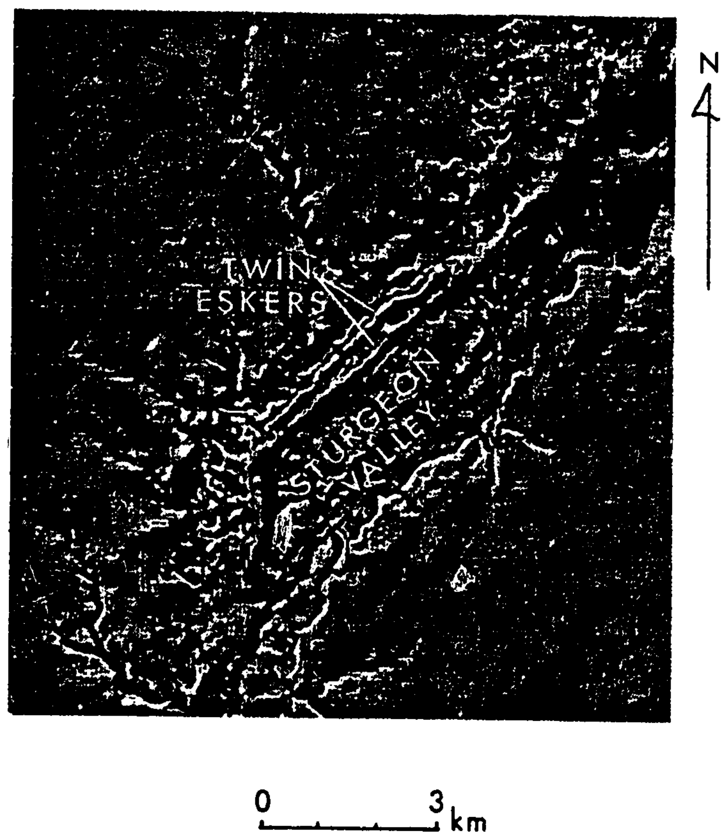

Fig. 3.3. Hillshade with curvature overlay map. Twin ridges are associated with a larger esker above the Sturgeon Valley and are about 2 km long, 100 m wide and 10 m high. These relatively sharp-crested features are perched between the valley base and rim, taper toward the southwest and are, at the surface, composed of fine to medium cross-laminated sand with paleocurrents parallel to the valley at 210°.

eliminate the effects of structural control on incision as explanations for the change in gradient, the positions and origins of the sediments and landforms within the valley are key to understanding its genesis.

The oriented residuals within the wide upper part of the upper Sturgeon Valley (Fig. 3.2) indicate the following:

(i) Based on residual stoss and lee positions, the last significant flow through the valley at elevations above 660 m was from northeast to southwest. This flow was opposite to the present drainage direction. At least a portion of the valley floor was formed by water flowing with the present stream gradient, toward the northeast.

(ii) There was at least one episode of high discharge which overtopped the southern wall of the upper valley and created numerous flutes above 670 m (Fig. 3.2).

A large, moderately sinuous, sandy ridge above a narrow portion of the valley, and fan-like landforms emanating from the lower end of the ridge within the valley, provide strong evidence for subglacial deposition: the ridge is an esker. The location of smaller, yet sharp-crested, parallel ridges in the valley, associated with the larger sinuous ridges in and above the Sturgeon Valley, indicates that their origin was unquestionably related to deposition by flowing water in tunnel conduits, and that, at least, the part of the Sturgeon Valley containing the twinned ridges existed at the time of their formation.

The esker system located above and within the upper Sturgeon Valley south of Bon Accord was, therefore, formed by subglacially-confined water flowing parallel to the valley towards approximately 210°. The formation of an esker in a conduit above the northern valley wall, capping over 25 m of sediment, suggests that the adjacent narrow portion of the valley did not exist when the esker formed and that that part of the subglacial Sturgeon Valley was infilled by sediment, the Bon Accord Sand Complex described earlier, and was only later re-excavated. Though the esker sediments may inter-finger with the upper portion of the Bon Accord Sand Complex, a lack of exposures in the landform precludes interpretation of this capping relationship. It is assumed, however, that 25 m of silt and sand in the Bon Accord Sand Complex could not have been deposited on the esker without completely masking it. Descent of the esker into the valley south of the proposed sedimentary constriction indicates that the larger valley did indeed exist when the esker formed. Flow in the valley at this time was through the meltwater conduit in which the esker formed. Once it had crossed the sedimentary fill, meltwater in the relatively small conduit flowed towards the valley floor. Shreve (1972) explains this in terms of potential flow where flows under high pressure tend to migrate toward low points in the landscape. Infilling of the conduits with fine-grained sediment formed the esker network.

The reason for the splitting of the esker into two, straight, parallel ridges within the valley remains unknown, but the broadening of the esker with distributary fans near its terminus is expected for deposition near a grounding line (Shaw and Gorrell 1991).

Genesis of the flutes less than one kilometer to the southwest of the esker was related to subglacial meltwater flowing over the valley rim during a high discharge event. This event probably preceded deposition of the esker and the Bon Accord Sand Complex, and was probably related to the flow event(s) which formed the residuals to the northeast.

By overlaying a satellite image on a DEM of the Sturgeon Valley, large elevation differences between the southern and northern valley walls are evident. The southern rim is consistently lower than the northern rim by a minimum of 10 m in the vicinity of the flutes. This difference increases to more than 15 m to the northeast in the vicinity of an esker in Zone 4 and indicates that any overflow from the valley would first spill southward. Flutes, mentioned earlier, parallel to and above only the southern valley wall (Fig. 3.2) support this hypothesis. The flutes were, therefore, most likely formed during a major flow event that filled the valley and overtopped its southern wall. This is most likely to have occurred before the valley infill event, probably at the time that the residuals to the northeast were formed.

Re-excavation of the partly-infilled valley, producing the narrow passage adjacent to the esker (Fig. 3.2), postdated esker formation. Though the more elevated part of the valley on which the twin eskers rest is much wider, the valley below is incised to the same depth and width as through the narrow passage. Therefore, it is inferred that the flow that partly re-excavated the infilled part of the valley also cut down into the remainder of the basal valley sediments.

Zone 3. The lower Sturgeon Valley

Big Lake is located within the most easterly part of the lower Sturgeon River Valley (Fig. 3.1). The valley bends westward south of St. Albert and widens to between 3 and 4 km before being buried at the Duffield Sand Complex (Holden and Shaw in prep.). The constant width of the valley west of St. Albert contrasts with its irregularity to the north. The most peculiar characteristic of the lower Sturgeon Valley, however, is its eroded southern valley wall with numerous curved ridges south and west of Big Lake (Fig. 3.4). Other distinctive landforms in this zone illustrate the effects of meltwater.

Landform morphology

Flutes located immediately west and southwest of Big Lake are up to several kilometers in length (Fig. 3.4). Ridge shapes vary from short tear-drop forms with blunt stoss and tapered lee sides within the valley to long, curved features extending as far as 10 km to the south. A number of ridges and sand bars west of the Sturgeon Valley, and closest to the Duffield Sand Complex (A on Fig. 3.4), are oriented more than 70° from the established flow directions of late-Wisconsinan ice in this area (Rains 1969; Westgate 1969; Ramsden and Westgate 1971; Shaw 1982). These linear and tear-drop ridges also diverge where they extend from the Sturgeon Valley towards the west.

Diamicton is not present in the sediments within 20 m of the surface at either of these fluted locations (Andriashek 1988). In all cases, the fluted ridges are formed in glaciofluvial sand and silt and are capped in places by lacustrine silts and clays.

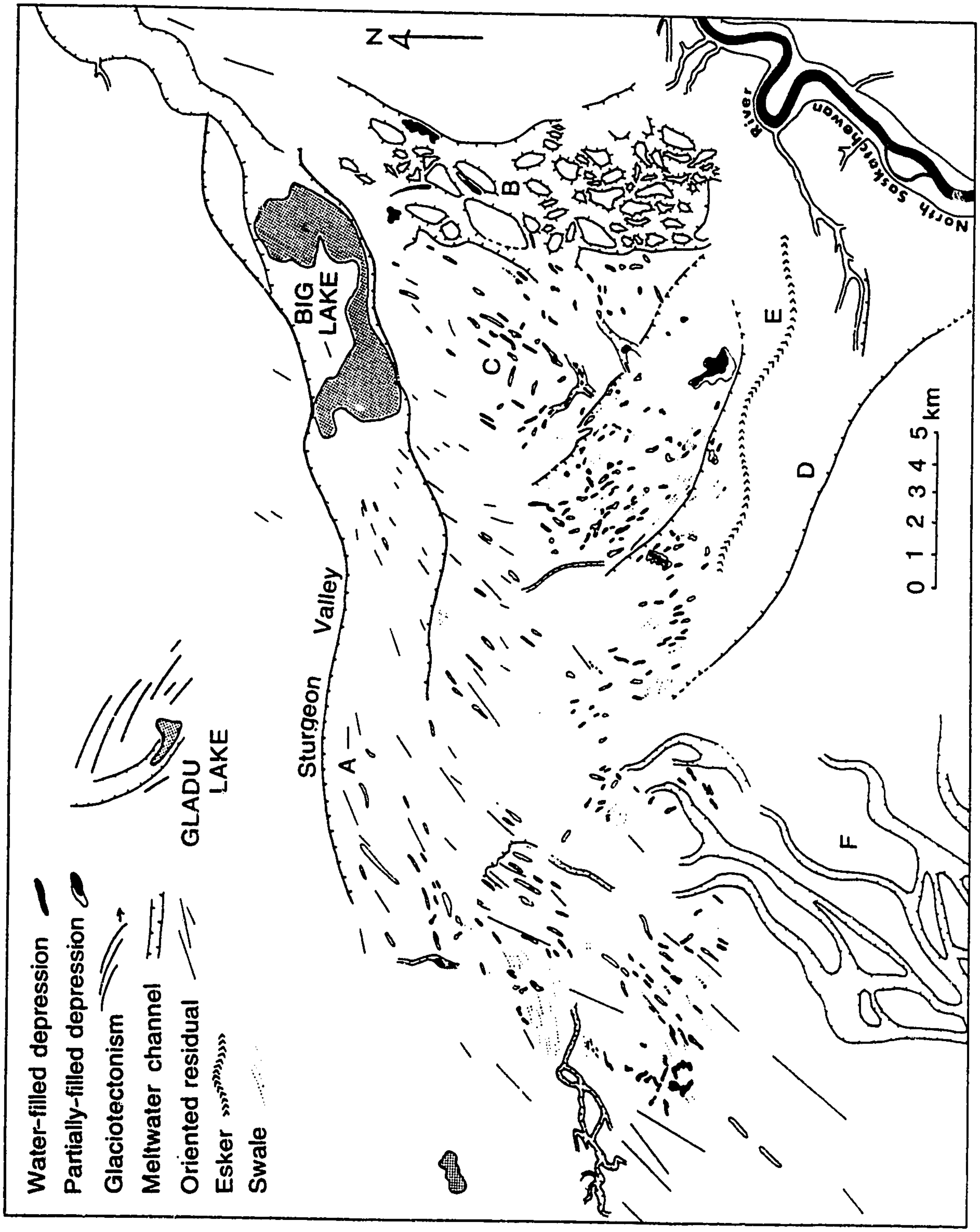

Fig. 3.4. Landform groups of Zone 3 consist of: (A) oriented, splaying ridges issuing from the Sturgeon Valley. (B) anabranching channels and residuals eroded into the eastern side of the Big Lake Sand Complex. (C) oriented depressions and meltwater channels above the Big Lake Sand Complex. (D) a northwest to southeast-trending channel cross-cutting the sand complex, containing oriented depressions, swales and an esker at (E). (F) a large network of anabranching channels south of Stony Plain (Holden and Shaw in prep.).

Detailed airphoto mapping of the area south and southwest of Big Lake reveals numerous channels, swales, oriented depressions and smoothed residual surfaces (Fig 3.5). Large erosional troughs to the east of, and in the sand complex south of Big Lake, illustrate the magnitude of postdepositional meltwater erosion in and around the complex. The most prominent of these landforms are channels and residuals beginning south of St. Albert and decreasing southwards in size and definition (B on Figs. 3.4 and 3.5).

South of St. Albert the channels form an anabranching pattern around large, oriented residual surfaces which are dissected by even smaller channels. At the narrowest reach of the anabranching system, three large channels occupy two thirds of the total width. The largest residual is over 1.5 km wide and 2.5 km long, and is located just beyond the narrowest part of the anabranching channels. Several oriented depressions on the residual surface extend 0.5 to 0.75 km from its up-flow end, paralleling its axis. The three dominant channels then break down into 5 or 6 smaller channels fanning out from northeast to southwest around smaller residuals up to 1 km in length. The most dominant of these channels was formed by meltwater flowing directly southward.

Southwesterly-trending channels branch into smaller, oriented depressions as the elevation increases locally over the Big Lake Sand Complex (C on Fig. 3.4 and Fig. 3.5). Similarly, the southeasterly channels are few, possibly because of an easterly increase in elevation toward the city limits of Edmonton. The remaining southerly-trending anabranching channels then break up into a smaller, less defined multi-channel and residual network. The westerly margin of this net-like channel system is well defined and has only a few small residuals and oriented depressions above it. The easterly margin, however, merely fades to extremely low-lying features that are nearly undetectable. Residuals within the southerly network of channels are small near the edge of the network but increase in size toward the center where the largest and most continuous of the southern channels are found (Fig. 3.5).

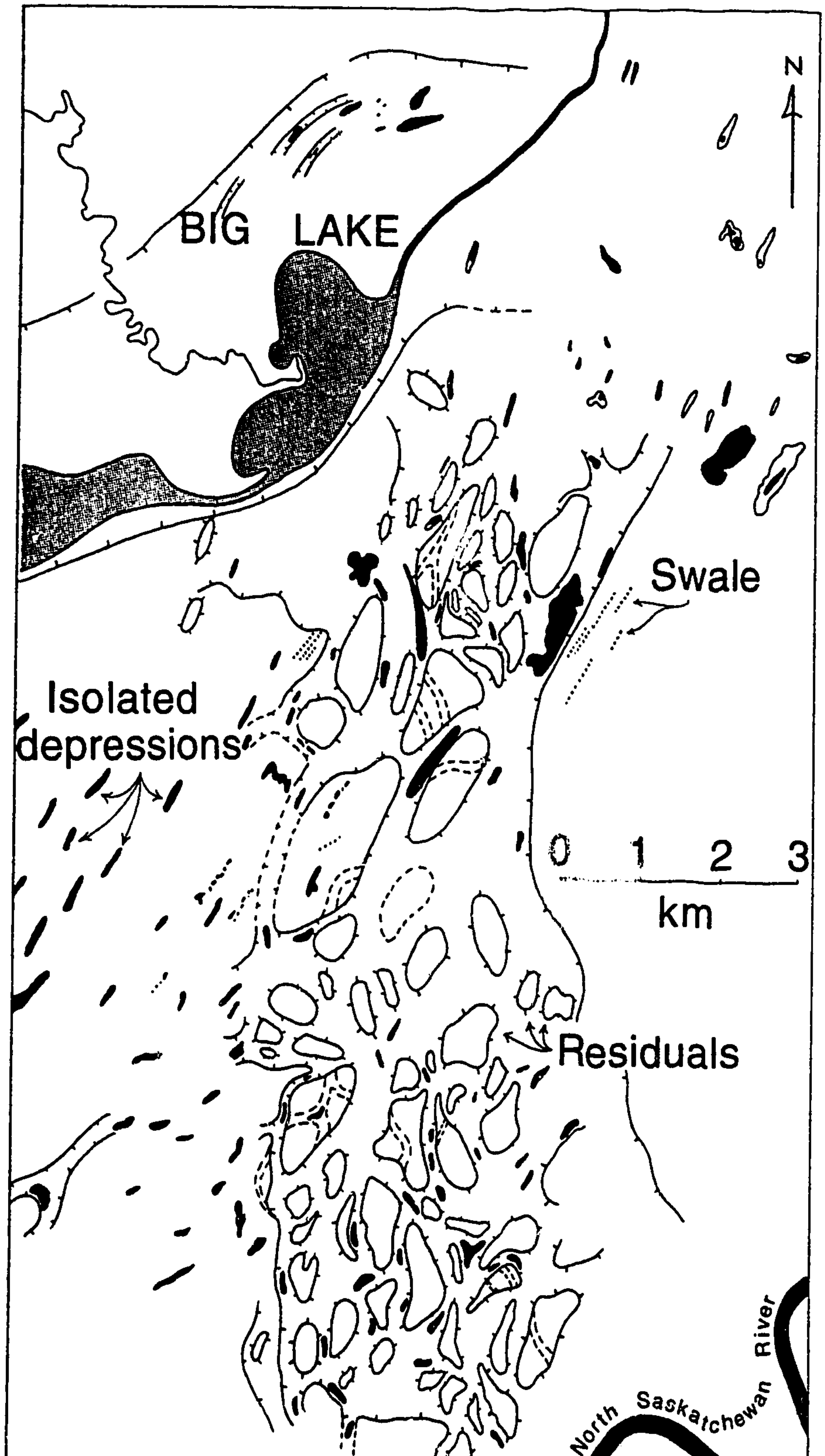

Fig. 3.5. Detailed map of the anabranching channels and residuals above and east of the Big Lake Sand Complex. The landforms begin at the downflow end of the Upper Sturgeon Valley, and mark the position at which the southern wall of the Lower Sturgeon Valley becomes less distinct westwards.

Deep exposures in the local sediments are extremely rare and road-cuts only uncover the uppermost 2 meters of the stratigraphy. One sand pit, however, just south of Hwy 16, west of Edmonton, shows an extremely large cut-and-fill sequence dominated by massive fine sand and thick beds of climbing-ripple lamination in medium sand. Paleocurrents in this section range from 210° to 250° and are similar to paleocurrent measurements taken by Shaw (1975, fig. 2) just 3 km to the east, near Winterburn.

The western margin of the Big Lake Sand Complex is cross-cut by a shallow, 4 to 8 km wide, and 20 km long channel trending from northwest to southeast. The best defined portion of the channel is easily identified on aerial photographs and satellite imagery as a low relief swath of land extending southeastwards from Spruce Grove to the North Saskatchewan River between Devon and Edmonton (D on Fig. 3.4). Two preserved eastern channel margins suggest lateral shifting from east to west during channel formation. The western margin of the sand complex is marked by an abrupt drop in elevation into the channel where its hummocky characteristics are quickly subdued. The topography of the western side of the Big Lake Sand Complex between the two eastern channel margins was not completely subdued though, and is marked by numerous oriented depressions and swales trending along the channel. These were undoubtedly scoured by flowing meltwater. Formation of the cross-cutting channel clearly postdated deposition of the Big Lake Sand Complex.

Large-scale landforms within the channel are few, but one ridge is quite evident on satellite imagery. The ridge is located just west of Yekaw Lake and is sinuous, approximately 10 km long, consistently over 250 m wide and up to 7 m high (E on Fig. 3.4). It is first seen about 5 km southeast of Spruce Grove and winds between the western and eastern channel margins.

Discussion of Zone 3 landforms and sediments

The extensive tract of fluted ridges south and west of the lower Sturgeon Valley reflects formation by an expanding westward meltwater flow, issuing from the Sturgeon Valley and curving southward. The ridges and elongate sand bars are interpreted as meltwater forms, recording high meltwater discharges through this part of the valley. Large tear-drop-shaped forms oriented parallel to and within the valley also record a southwesterly flow direction.

Consistent paleocurrent measurements, between 210° and 250°, at two sites in the sand complex south of Big Lake may indicate that it was deposited by meltwater flow channeled southwestwards through the Upper Sturgeon Valley prior to the deposition and re-excavation of the valley infill discussed earlier. Formation of the anabranching channels and residuals over the eastern part of the sand complex clearly postdate deposition of the sand complex and may, therefore, have been temporally related to the flow that created the landforms found within the Sturgeon Valley.

The superposition of lacustrine rhythmites over the anabranching channel area east of the Big Lake Sand Complex suggest that formation of the channels preceded the onset of Glacial Lake Edmonton. Because each phase of this lake was said to be impounded against northeastward melting ice (St-Onge 1972), the formative flow event for the channels occurred subglacially, following deposition of the sand complex. A proglacial origin for the channels is ruled out because the meltwater would have had to flow uphill from the Sturgeon Valley onto the top of the existing sand complex, climbing nearly 20 m over as many kilometers. In other words, the minimum elevation of the channels over the Big Lake Sand Complex (690 m) far exceeds the maximum elevation of the Sturgeon Valley rim at its mouth (675 m). Given the paleocurrent directions, and the significant uphill trajectory of the anabranching channels, the water that deposited the complex must have flowed subglacially, under hydrostatic pressure.

The morphological and sedimentological characteristics of the sinuous ridge and channel that cross-cut the Big Lake Sand Complex west of Yekaw Lake reflects their subglacial fluvial genesis. The ridge is an esker within a tunnel channel formed during the last phase of meltwater evacuation through a larger subglacial meltwater conduit, perhaps from the the regional-scale cavity discussed by Holden and Shaw (in prep.). Let-down of an englacially or supraglacially formed esker into the channel requires coincidence and is not considered to be a reasonable hypothesis. Cross-cutting of the Big Lake Sand Complex by the channel unquestionably demonstrates that the complex was also deposited subglacially.

Zone 4. Oriented depressions

Landform morphology

To the east and southeast of the Upper Sturgeon Valley and south of the Lily Lake Valley, the arcuate ridges and troughs of Zone 1 are no longer present. In their place are elongate depressions oriented between 210 and 225° (H on Fig. 3.2). This 8 km swath of distinctly sculpted topography is located on, and south of, a topographic high composed mostly of glaciofluvial sand, silt and gravel. Analysis of DEMs of the area show that the high is continuous from the North Saskatchewan River, northward, to beyond the junction of the Redwater and Lily Lake Valleys. The processes forming the ridges northwest of the Sturgeon Valley were either not active in Zone 4, or the landforms produced by those processes were subsequently removed or reshaped.

This zone is subdivided by an esker (Shetsen 1990) oriented at 130° (I on Fig. 3.2). This relatively straight ridge can be traced below the northern edge of the topographic high from the Sturgeon Valley near New Lunnon to the North Saskatchewan River valley and does not appear to have marginal fans associated with it. It is consistently 125 m wide by 3 m high and follows the undulating topographic slope but does not cross-either of the valleys. The area to the north of the esker is largely covered by a local diamicton veneer and glaciofluvial sand and gravel, whereas the southern part is thinly mantled by fine silt and clay (Shetsen 1990).

Oriented depressions predominate to the south of the New Lunnon esker where lacustrine clays which only thinly mask the underlying topography. These depressions occupy an 8 by 8 km swath between the Sturgeon and North Saskatchewan River valleys. With few exceptions, the depressions are less than 500 m long and up to 7 m deep, and are not aligned along common troughs. Several of these are linked laterally and collectively form depressions transverse to the primary orientation (J on Fig. 3.2).

Discussion of Zone 4 landforms and sediments

The relationship of the esker to the oriented depressions and to the North Saskatchewan and Sturgeon River valleys is not fully understood. The existence of a hydraulic differential between larger sub- or englacial meltwater conduits could potentially drive flow between the two in a smaller conduit. The proximal location of the southeastern end of the Bon Accord Sand Complex relative to the New Lunnon esker may be indicative of such a process. As is often the case, the lack of exposures in the landform limits interpretation of its genesis.

Oriented depressions south of the esker are not inter-ridge lows as those seen in Zone 1 (Fig. 3.2), nor are they low areas between fluted ridges. The parallelism of individual depressions and their apparent lack of continuity does not favor genesis by lacustrine infilling of pre-existing channels or of fluted lows. The existence of several fairly well defined, non-parallel and unoriented channels within the isolated depression assemblage (Near J on Fig. 3.2) suggests that:

(i) meltwater forming the channels was not controlled by a strong potential gradient unlike the strongly controlled flow that created the depressions; and

(ii) lacustrine sedimentation was insufficient to cover these superimposed channels and, therefore, could not have preferentially covered other channels or fluted lows to produce isolated depressions.

Although erosion by a large proglacial flow of meltwater is possible, there must have been ice remaining after the formation of the oriented depressions to form the superimposed, unoriented meltwater channels. Furthermore, the esker would not have withstood erosion by such a proglacial meltwater flow and must have either been subsequently deposited from down-wasting ice, though it is morphologically pristine, or by late-stage subglacial meltwater. Thus the oriented depressions to the south are inferred to be subglacial erosional scours. It is possible that the scouring was by meltwater and was related to the same meltwater drainage events that overtopped the south wall of the Sturgeon Valley, producing flutings on the valley side and residuals within the valley.

Disturbance of the lower, oriented scour topography south of the Sturgeon Valley was probably inhibited by rapid melting, and possibly calving, of the ice front into a small proglacial lake, or deeper part of a larger lake as suggested by an isolated pocket of fine-grained lacustrine clay over the area.

Zone 5. The Lily Lake Valley

Description and discussion of landform morphology

The Lily Lake Valley is much larger than any other in the area and includes a large number of smaller channels on its southwestern slope, at progressively lower elevations. The highest of these is found at 693 m and is associated with sediments exposed at Site 1. Other channels on the same side of the valley are at roughly 685 m, 670 m, 660 m, and the lowest containing Lily Lake at 645 m. The highest and only major channel margin identifiable on the eastern side of the valley is just above 655 m.

Two sets of smaller, oriented channels to the southeast are between 655 m and 645 m, north of the esker in Zone 4, on the southwestern side of the valley (K on Fig. 3.2).

The first is composed of five short channels on a small diamicton high which trend toward the southwest. The other comprises a dozen or so southeasterly trending, long, narrow channels formed at lower elevations.

The meltwater history of the valley is unusual because of the occurrence of perched channels on only one of its two walls. Formation of the highest channels required ice marginal or subglacial conditions. The maximum continuous elevation of the northeastern valley wall is 660 m. North of the valley wall, the topography decreases in elevation to less than 640 m within 15 km. A non-glacial valley, therefore, could not possibly have confined water flowing more than 30 m higher than the local northern topography to form the highest channel on the southwestern wall.

The perched channels, therefore, were either formed in a subaerial environment at the ice margin, functioned as subglacial meltwater conduits, or the entire topography north of the Lily Lake southern valley wall was eroded after formation of the channels. The latter is immediately dismissed since the perched channels on the southern wall would have easily been eroded by such an extensive erosive flow. In the absence of landforms diagnostic of either subglacial or subaerial environments, an attempt to differentiate between these environments must focus on sediments associated with the Lily Lake Valley south wall channels.

Description and discussion of sedimentology

Site 1 Distributary mouth deposits

At the mouth of the highest and coincidentally the largest of the perched channels is a fairly wide, undulating plain, unlike the rest of the valley floor. A sand and gravel pit located at the outer edge of this plain provides several exposures in the underlying sediments. The pit is located 9 km north-northeast of Bon Accord on Range Road 234, just north of the Fifteenth Base Line. It extends about 300 m from north to south exposing numerous sections below a gently undulating surface. The western pit wall exposes 1.5 m of cross-bedded medium sand above 4 m of plane-bedded medium to coarse sand. Convex-upward cross-stratification above plane beds illustrate the succession from dominantly bedload transport to a significant increase in suspension deposition (A on Fig. 3.6). Although the Lily Lake Valley trends toward the southeast, sediments in this pit were deposited by an eastward flow, towards 75° to 95°, directed through one of two distributaries of the highest channel. The lack of southerly paleocurrents in the exposure suggests that the flow velocity within this part of the main valley was negligible compared to the velocity of the eastward flows which deposited the thick plane and cross-bedded units. As there was no significant change in the sedimentation regime indicative of deceleration on a fan or delta, the eastward-flowing meltwater did not empty into standing water.

The southeastern exposure of the eastern pit face offers interesting structural information (Fig. 3.6). Of the four 0.5 m thick cross-beds in this section, only the upper two are marked by fold and fault deformation. Folding occurs only in the upper cross-bed set and faulting within the underlying set. Displacement along the low-angle thrust faults rarely exceeds 20 cm and occasionally dies out laterally. Minor entrainment of sediment along the fault planes suggests that the unit was only partially liquified at the time of failure. The intra-formational and nearly horizontal nature of the faults also suggests a syn-depositional origin.

Allen and Banks (1972) describe recumbent folds in cross-bedded sands similar to those of the upper cross-bed. Though intra-formational faulting is also addressed by Allen and Banks (1972), they note its common restriction to wind-blown sands, citing Thompson (1969), Glennie (1970), and McKee et al. (1971), or suggest subaerial mass movement with high-angle inter-formational faulting.

Allen and Banks (1972) infer that the formation of soft-sediment recumbent folds results from a combination of fluidization and current drag. Leaving the intricacies of their

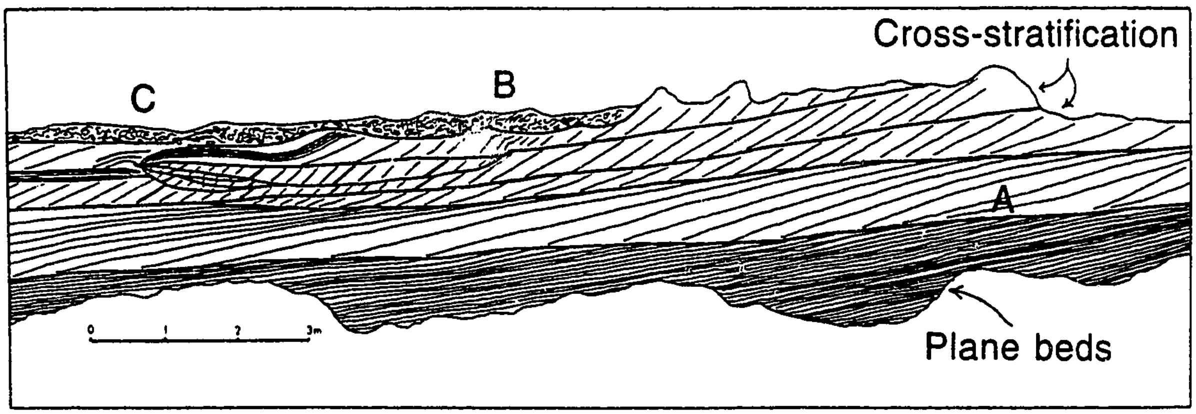

Fig. 3.6. Stratigraphic section through sediments exposed in a pit at the southern margin of a flat-surfaced landform that was formed at the down-flow end of a perched meltwater channel in Zone 5. The diagram shows a succession of 4 large-scale cross-beds in coarse sand over a thick plane-bedded unit. Convex-upward cross-stratification illustrates the succession from dominantly bedload transport to an increase in suspension deposition (A). A major horizontal fault in the third dune bed, Between (B) and C), is attributed to current drag and loading stresses imposed by the overlying dune unit. The primary faulting event caused secondary, near-horizontal faulting below the mobilized sediment and deformation of the upper unit (C). Note that this section is curved and that it faces northward at (A) and westward at (C).

"earthquake" fluidization process aside, the parallelism of the main fault to the depositional planes of this section supports deformation by tangential forces, that is, current drag.

The type of fold and fault deformation described by Allen and Banks (1972) is characterized by high-angle faults occurring with "much local blurring and destruction of the stratification". Although their model of liquifaction may be suited to the sediments they describe, the model might be modified to include low-angle faulting and intact cross-stratification in sand by examining shear stresses produced by current drag in special environments. They calculate that shear stress on a dune face in a river flowing at 1 m/s is quite small, between 10 & 20 dynes, and may not be substantial enough to deform (fold) a great thickness of sand unless it had lost much of its cohesion. Potential shear stresses imposed on sediments in special environments such as major ice-proximal meltwater channels, where sedimentation and flow velocities can be quite high, were not considered.

The upper surface of each cross-set in this section is neatly truncated except where, based on upward extrapolation of the fault plane, faults reach the surface of the unit along a bedding plane (B on Fig. 3.6). At these locations, the contact between the upper and lower cross-sets is usually displaced no more than 5 cm vertically.

Based on the horizontal and vertical displacement of the upper folded dune bed and the proximity of the upper unit folds to the lower unit fault planes, failure of a thin cross-stratified lens was caused by a strong current drag. Basal shearing caused by the mobilization of the sediment lens produced multiple secondary, near-horizontal, faults at the down-flow edge of the main lens (C on Fig. 3.6).

Though paleocurrents indicate deposition toward the east, the main flow was directed toward the southeast through a large channel or conduit and was redirected eastward by one of two smaller distributaries. These sediments, and structures therein, are consistent with deposition by fast-flowing water. In themselves, these sediments cannot be used to differentiate subaerial from subglacial deposits. Consequently, two depositional scenarios are presented.

If the channels were formed ice-marginally, a northeastward-melting ice front would permit the formation of successively lower channels down to an elevation less than 660 m where the northeastern Lily Lake Valley wall confined the meltwater. Alternatively, the formation of numerous subglacial meltwater channels at various elevations can be explained by avulsion of meltwater under hydrostatic pressure. The larger part of the valley would represent a single dominant channel, formed after the smaller perched channels, or, the entire valley may have operated as a subglacial conduit and the perched channels were formed at the base of a larger flow.

Site 2 Mega-dune

Glaciofluvial sediments 5 km north of Waugh, at the northern junction of the Redwater and Lily Lake valleys, illustrate high-flow-regime deposition near the base of the Redwater Valley (Fig. 3.7). This southeasterly-facing section is comprised of 10 m of plane beds in coarse sand with paleocurrents towards east-southeast. In the lower 4 m, the plane beds dip consistently at about 3°, whereas in the upper 6 m dips increase to more than 7°. The contact between the upper and lower plane-bed units is planar with only minor erosion.

The plane-bed sequence contains various sizes of armored till balls. Although not one is larger in intermediate diameter than 25 cm, some are cylindrical with long axes up to 60 cm. Clasts are concentrated in scours in front of and behind large till balls (Fig. 3.7). The sequence described occurs within a large landform 1.5 km long, 750 m wide, and 6 m high above the basal plane beds.

The consistency of grading and sorting of the coarse-sand plane beds suggests that they were deposited under a sustained high flow regime and that suspension deposition was negligible. Separation of the sediments into two distinct units, a lower gently-dipping plane bed unit of appreciable thickness, superimposed by an upper steeply-dipping unit, illustrates the down-flow migration and steepening of a large, prograding bedform. The

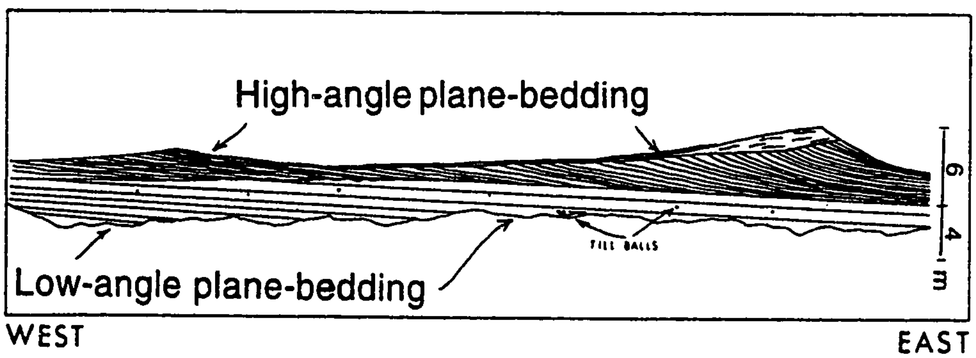

Fig. 3.7. A sequence of medium sand plane-beds greater than 4 m in thickness underlies steeply dipping coarse sand cross-strata reaching a maximum height of 10 m. The architecture of the section is representative of prograding bedforms in a sustained high-flow-regime. Till balls in the lower sediments illustrate the competance of the flow to transport large clasts.

bedform is 6 m high and has a minimum wavelength of 100 m. This is based on the distance between the foresets of two bedforms. Unfortunately, the area is well mined for its sand and gravel and the crest of the second bedform studied has since been removed.

The bedform is similar in morphology and scale to subglacially-formed gravel dunes described by Shaw and Gorrell (1991), but differs substantially in composition and structure. Dunes described by Shaw and Gorrell (1991) are up to 10 m in height with wavelengths up to 200 m, but are mostly composed of matrix- and clast- supported coarse gravels and cobbles. Numerous unconformities exist between successive gravel units, and secondary bedforms in sand were found on the lee face of a particular landform (Shaw and Gorrell 1991, Fig. 3.2). Such characteristics illustrate extremely variable subglacial meltwater flow conditions that were evidently not present during the formation of the mega-dune in the Redwater Valley.

The presence of large till balls in the lower plane beds indicates both erosion and transport of larger-sized clasts than coarse sand. Clasts up to 3 cm in diameter were found in a scour behind a large till ball within the lower unit. This indicates that coarser clasts were transported along the top of the plane beds and were only occasionally deposited in protected locations. This by-passing by both coarse material transported as bedload and the fine sediment transported in suspension resulted in the remarkable sorting of the plane beds. The flat plane-bedding would have facilitated the by-passing and coarse bedload would have been deposited to the lee of dunes just as it was to the lee of the till ball.

Site 3 Plane beds and diamicton remnants

A smaller pit less than 3 km to the south at an elevation of about 625 m exposes steeply-dipping plane beds deposited behind a diamicton 'high' by a southeasterly flow. The plane beds are in poorly sorted sediment: very coarse sand, pebbles and cobbles, with armored till balls up to 3 cm in diameter, likely derived from the local diamicton. The greenish-brown diamicton is fine grained but has numerous, dispersed clay ironstone, granite and quartzite pebbles within it. Deposition of the plane-bedded, coarse sediment and erosion of the diamicton were evidently synchronous. Thus, the diamicton 'high' is probably a residual.

Descriptions of the appearance and disappearance of "walls of stoney clay" by pit workers as the pit was excavated were supported by an exposure of undeformed, plane-bedded, coarse sand deposited against a vertical diamicton face. Deposition of the diamicton was evidently followed by its erosion and the formation of residuals. Subsequent deposition of coarse sand in plane beds buried the residuals, preserving the vertical walls of diamicton.

Site 4 Pendant bar

One of the last features deposited by meltwater flowing through the Lily Lake Valley was a pendant bar located 4 km north of Gibbons, at 631 m. The Mcleod pit, Site 4, is located on the bar to the lee of a bedrock residual near the floor of the valley. The lower 2 m of the down-flow facing section exposes regularly alternating high- and low- angle tabular cross-stratification (Fig. 3.8). Although no one type predominates, high-angle cross-stratification sets are consistently two to three times thicker such that individual couplets are easily differentiated. Low-angle cross-stratification is usually in coarser sand, and in some cases is found in troughs.

The section is situated through the central portion of the bar and exposes strata dipping away from the bar axis, creating convex-up architecture. Although stratification between the convex-up beds is in places similar to trough cross-stratification, formation of the troughs was solely dependent on the growth and forward migration of subaqueous linguoid dunes. Each node of convex-up strata (A on Fig. 3.8), bounded on either side by outward-migrating cross-beds, was formed by advance of the tip of a single linguoid bedform. Trough cross-stratification was in turn formed where depressions between such bedforms were filled by lateral and forward accretion.

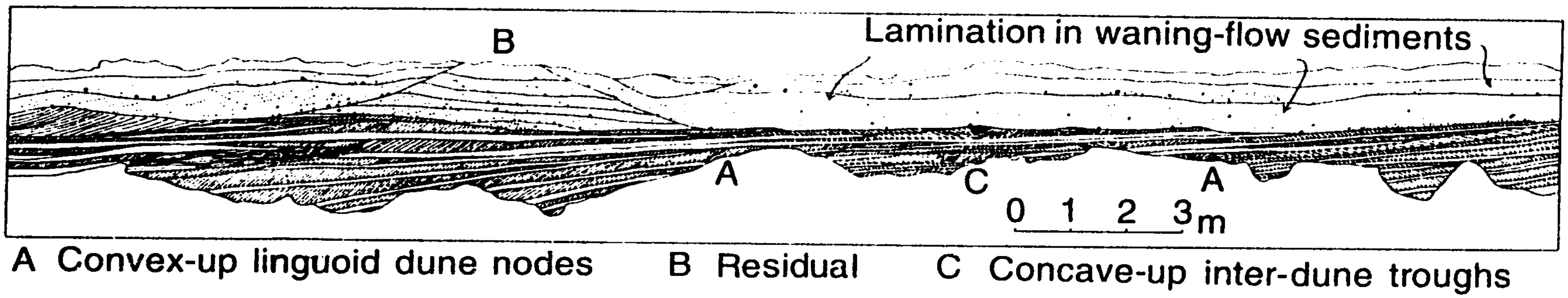

Fig. 3.8. Alternating successions of cross-stratified medium sand and coarse sand plane-beds (left white) capped by laminated waning-flow sediments in a pendant bar near the base of the Lily Lake Valley. Note the positions of flow centers within the cross-stratified units (convex-up darkened hemispheres) and the central positions of inter-dune trough stratification (concave-up hemispheres). The section faces southeastwards.

The upper meter of the section is composed of current-laminated, waning-flow silt, sand and pebbles with only few cross-beds present. A small, central portion of this sequence remains undisturbed between large scours at the base of the laminated sediments (B on Fig. 3.8). The scours suggest a period of reactivated discharge within the valley following abandonment of the bar. Eastward migration of bedforms is indicated by cross-lamination and cross-beds in the northeastern, fine-grained sediments, and westward migration indicated by the southwestern fine-grained sediments indicate deposition by a diverging flow towards either side of the pendant bar. Though the pendant bar occurs at 631 m, it was deposited subaqueously during the last stage of meltwater flow from the northwest. Capping waning-flow sediments illustrate the final decrease in flow from the north in the southern part of the Lily Lake Valley.

The Lily Lake Valley contains a variety of meltwater channels and sedimentary bedforms documenting a southeastward meltwater flow. Although several other channels and landforms in the immediate area are confidently interpreted to have formed subglacially, such genetic differentiation cannot be made based on the available information. Both ice-marginal and subglacial environments are, therefore, possible for the formation of the Lily Lake Valley and landforms therein.

A future study of the glaciofluvial complex that extends southeastwards from the mouth of the Tawatinaw Valley through the Redwater Valley, may provide the information necessary for a firm genetic interpretation of landforms and sediments in the Lily Lake Valley. The Tawatinaw complex was deposited at an elevation higher than that of Glacial Lake Jarvie which was located to the immediate west (St-Onge 1972; Rains 1993, pers. comm.). An examination of possible topographic constraints influencing sedimentation, and the establishment of a relative chronology for the lake phase and deposition of the glaciofluvial complex, will be necessary to determine whether the complex is of a sub or pro-glacial origin.

Regional Reconstruction

Deposition of the youngest deposits of the preglacial Empress Formation sands and gravels is C^{14} dated at between 40 and 22 ka B.P (Young et al. *submitted*). Incision of the ancestral Sturgeon Valley below the upper surface of these sediments, therefore, followed 22 ka B.P. The valley was subsequently partly infilled in the vicinity of Bon Accord with more than 30 m of sand and silt, with paleocurrents diverging around Bon Accord to the southeast and southwest, by subglacial meltwater sedimentation. Superimposition of a large esker over these sediments establishes the relative timing of the sedimentation event and esker formation. Moreover, the descent of the esker into the valley to the south dictates that the valley must have existed prior to deposition of the esker. The esker may, in fact, represent the final drainage of the cavity within which the Bon Accord Sand Complex was deposited. The fluted ridges traced from the base of the Sturgeon Valley to the adjacent upland were probably formed before deposition of the esker and most likely predated infilling of the valley to the north.

Late-stage drainage of the subglacial cavity south and west of Big Lake formed a 4 km wide channel, trending southeastwards from Spruce Grove, that cross-cut and eroded the western edge of the Big Lake Sand Complex. Final drainage through this channel was limited to a narrow conduit within which the Yekaw Lake esker was formed.

The oriented depressions between the Upper Sturgeon and North Saskatchewan valleys were scoured by southwesterly flowing turbulent, unchannelized subglacial meltwater before deposition of the Bon Accord Sand Complex and its associated esker, or afterwards, provided the latter were protected by grounded ice. This flow may have also occupied the Sturgeon Valley and produced the large flutes crossing the south valley wall. During the waning stages of this event, or in a subsequent phase of minor meltwater drainage, the New Lunnon esker was formed in a small conduit trending towards the southeast from the Sturgeon Valley toward the North Saskatchewan Valley. This conduit may have also been related to final drainage of the Bon Accord cavity.

At this late stage, the mouth of the Upper Sturgeon Valley may have been free of ice and, provided it was also free of impounded meltwater, may have allowed a burst of fast-flowing meltwater from the Sturgeon Valley to flow over the eastern margin of the Big Lake Sand Complex and erode several major channels. These large channels formed the trunk of a southward-expanding, anabranching, fan-like system of smaller channels and residuals which decrease in size southwards.

If St-Onge (1972) is correct, however, and the phases of Glacial Lake Edmonton were northeastward migrating impounded lakes, then a proglacial meltwater flow into deeper water could not have eroded this anabranching channel system. Alternatively, because the channels are draped by lacustrine sediments (Shetsen 1990), and are formed above the Big Lake Sand Complex, which is cross-cut by a channel and esker, the anabranching channels may have been formed by fast-flowing subglacial meltwater. Their formation may, therefore, have been contemporaneous with, or may predate, formation of the channel and esker.

A regressive series of parallel arcuate ridges and troughs were then formed over glaciofluvial sand and silt, stratigraphically, more than 15 m above the nearest diamicton. The ridges were formed in a northward retreating, ice-contact lacustrine environment where oscillating lake levels, in conjunction with northward melt of clean ice, caused deformation at the grounding line of proximal sediments into ice-marginal ridges.

Subglacial deformation into the interior perimeter of a large, expanding, subglacial cavity is presented as an alternative for the formation of these ridges. Landforms produced in a smaller confined environment such as the Yekaw Lake, Mink Lake (Holden and Shaw in prep.) and New Lunnon eskers, and the large anabranching channels south of Stony Plane (Holden and Shaw in prep.) suggest that ice was in contact with the substrate during the final, southward drainage of the proposed regional-scale cavity. This alternative is, therefore, dismissed.

Impounded water of Glacial Lake Edmonton followed the ice margin northward with a water level at approximately 695 m until an outlet through the southeastern portion of the Lily Lake Valley became free of ice. The sediment-constricted portion of the Upper Sturgeon Valley above which the esker formed, and the lower portion of the valley below the twin esker ridges, was deglacially re-excavated by water flowing toward the northeast through Sturgeon River. Continued drainage of the lake allowed the incision of a northwest to southeast-trending meltwater channel against the ice margin at about 693 m. Meltwater flow through the southeastern end of this channel split into two distributaries, the first of which allowed deposition of a fan at its mouth. A succession of impounded drainage channels were then formed down to the highest elevation of the northeastern Lily Lake Valley wall, at about 660 m, which was then capable of confining the flow. Progressively lowering water levels in the Lily Lake Valley by diversion of meltwater into the broader Redwater Valley caused the abandonment of a pendant bar at 631 m, immediately preceding abandonment of the valley. A large volume of fast-moving meltwater continued to flow through the Redwater Valley producing 6 m high prograding bedforms as large sand bars within the valley north of Waugh.

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CHAPTER 4

GENERAL CONCLUSIONS

This study of the late-Wisconsinan geology and geomorphology of the Edmonton region makes use of several technological advances that were not available during the mid 1960's and early 1970's when the local surficial geology was first mapped in detail. As a result, several new landforms were discovered that necessitate re-interpretation of older findings. These interpretations are presented in two articles, entitled "Subglacial Lacustrine and Fluvial Environments: The Stony Plain Area, Alberta" and "A Transgressive Model of Subglacial and Ice-Marginal Environments: Landforms North of Edmonton, Alberta".

The findings suggest that landforms west and north of Edmonton were largely formed by subglacial and ice-marginal processes. The progression between these two environments was traced by examining the complex relationships between specific landform types and sediments throughout the region. Because certain landforms are known to form under specific environmental conditions, firm genetic interpretations of these landforms makes possible a detailed glacial history of the region. For other landforms, exposures of internal structure were absent, and only landform associations and morphology could be used to decipher formative processes.

Geomorphic development of the Edmonton - Stony Plain area was dominated by subglacial fluvial processes, postdating the Livingstone Lake event proposed by Shaw (1983). The first stage in this development was the formation of a subglacial cavity, either by disrupting the flow of a Röthlisberger meltwater conduit (Fig 4.1 Stage 1), or by trapping subglacial flood-water in an existing basin. Progressive cavity development by ice closure, and partial reactivation of meltwater flow through the Sturgeon Valley conduit, is the preferred mechanism. Reactivation of meltwater drainage through a conduit, from the north, promoted increasing hydrostatic pressure at the point of closure, and progressively separated and raised the ice bed from its substrate (Fig. 4.1 Stage 2a). The resultant water-filled cavity may have been as large as is depicted in Stage 2b (Fig. 4.1), or may have been composed of smaller cavities which operated at different times, and fed by different sources. The Duffield Sand Complex was deposited in the largest of these cavities by meltwater flowing from the north-northeast (Fig. 4.1 Stage 3a), and permanently blocked drainage through the Sturgeon meltwater conduit, north of Stony Plain. Sedimentation within the cavity during its deepest phase created extensive and thick, cross-stratified and cross-laminated beds. Progressive filling of the cavity and, perhaps, an increased rate of drainage reduced the cross-sectional area between the ice and the sediment bed, causing increased flow velocities. The resulting mechanical and thermal erosion of the ice bed produced erosional scallops on the ice bed. Waning flow in the Duffield cavity deposited high flow-regime sediment in the small-scale cavities or scallops. With decreasing flow the glacier came to rest on a sediment bed that conformed to the its hummocky, sculpted base.The remaining meltwater was confined to conduits within which eskers were formed. The Mink Lake esker and its southern counterpart, the Jackfish Lake esker, record this final episode of subglacial sedimentation in the Duffield Sand Complex.

Deposition of the Big Lake Sand Complex (Fig. 4.2 Stage 3b), and formation of its hummocky topography, probably had the same origin as the Duffield Sand Complex, but may have formed in a smaller cavity closer to the subglacial Sturgeon Valley meltwater conduit. A large, partly channelized, southwestwardly meltwater flow event through the Sturgeon Valley eroded approximately 50 m of sediment from the southern valley wall (Fig 4.2 Stage 4). It also eroded numerous, arcuate, remnants in the eastern part of the Duffield

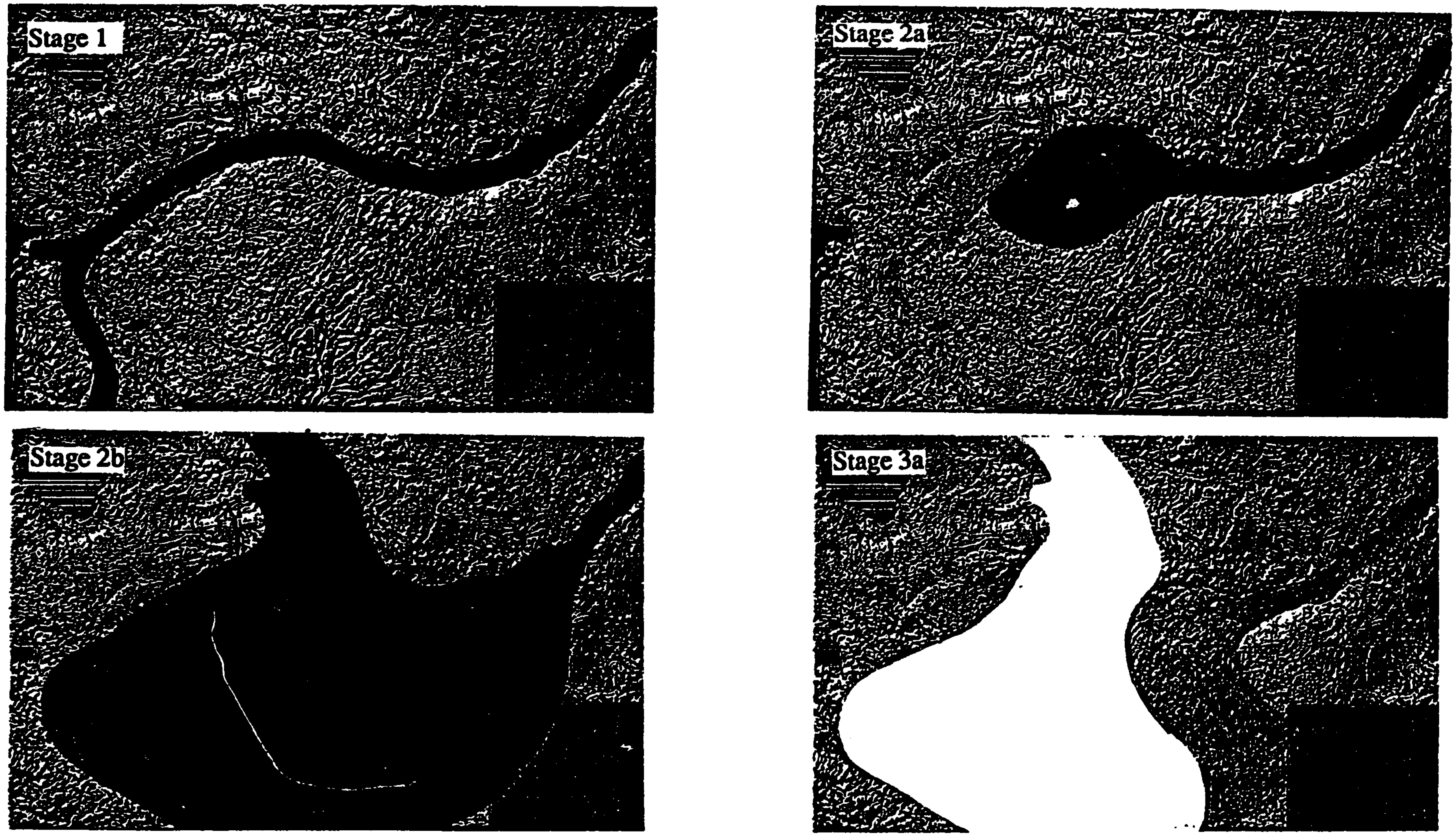

Fig. 4.1. Regional sequence of events. Stage 1: Blocking of subglacial meltwater conduit. Stage 2 and 2b: expansion of subglacial cavity. Stage 3a: deposition of the Duffield Sand Complex and formation of the hummocky topography.

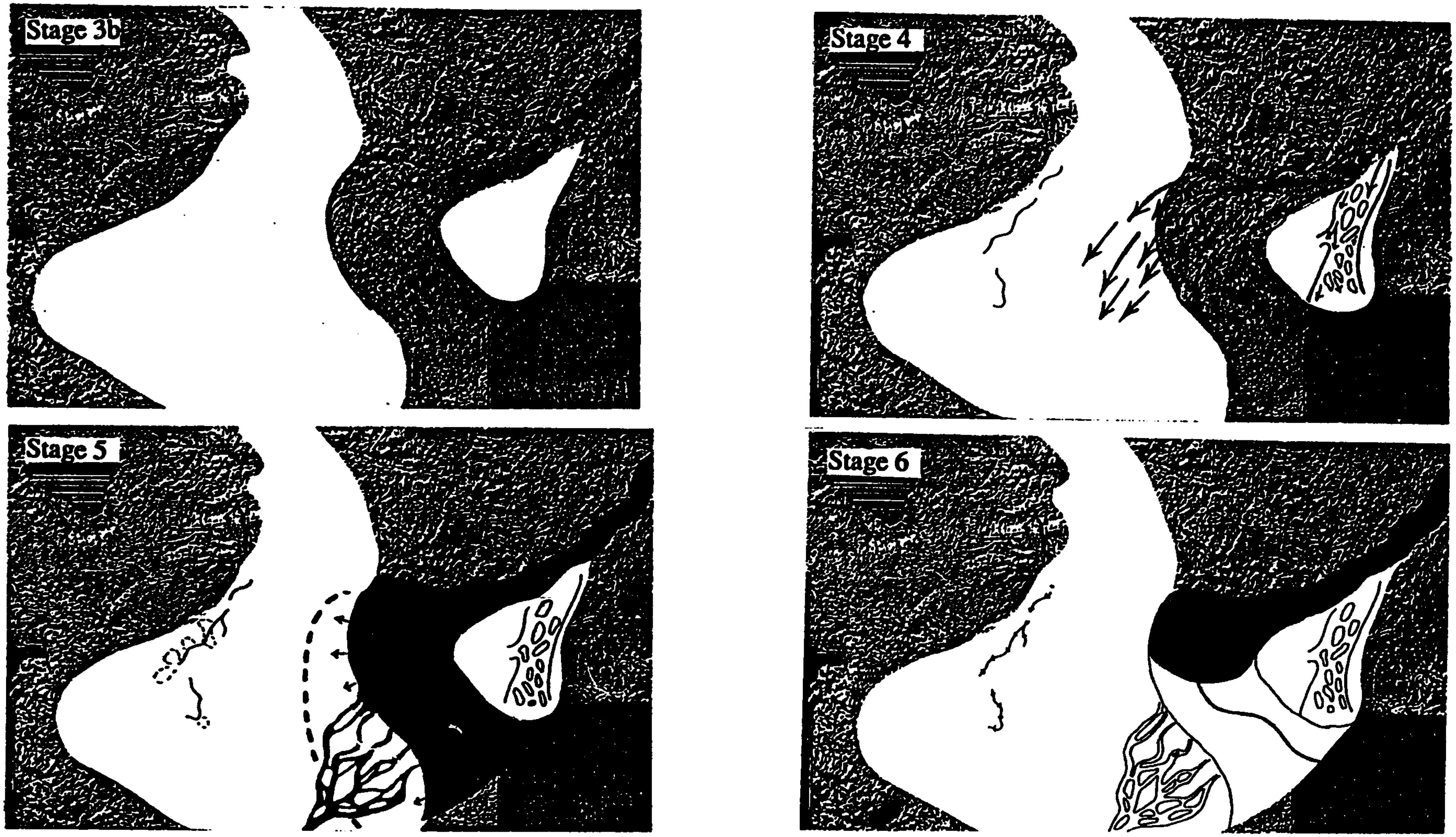

Fig. 4.2. Regional sequence of events (cont'd). Stage 3b: deposition of the Big Lake Sand Complex. Stage 4: a large, channelized flow through the Sturgeon Valley erodes anabranching channels and flutes east of the Big Lake Sand Complex and on the eastern side of the Duffield Sand Complex. Stage 5: the anabranching channels south of Stony Plain are formed by subglacial water under hydrostatic pressure. Stage 6: late-stage drainage of remaining meltwater in the cavity and formation of the Yekaw Lake esker.

Sand Complex and across the entire southern valley wall. These remnants extend more than 15 km southward. Erosion of the anabranching channels on the eastern side of the Big Lake Sand Complex was probably related to the same meltwater event.

Continued drainage of the subglacial basin between the Duffield and Big Lake sand complexes was southwestwardly, through a system of anabranching channels (Fig 4.2 Stage 5). These channels, which cut across the southeastern, subdued margin of the Duffield Sand Complex, record up-hill meltwater flow with reverse gradients approaching 1 m per km. Later, this drainage was southeastwardly, through a wide conduit which cross-cuts the Big Lake Sand Complex. Late-stage, southeastward drainage through a narrow conduit, produced the Yekaw Lake esker (Fig. 4.2 Stage 6).

Sediments and landforms in the northern part of the study area indicate flow events which were complementary to the subglacial meltwater events discussed above. An esker superimposed on the Bon Accord Sand Complex, above the northeastern wall of the Sturgeon Valley, indicates a subglacial genesis for the sand complex. The interpretation that the esker is superimposed on the sand complex is made without extensive sediment exposures in the landform. It rests on the reasonable assumption that the 29 m of the Bon Accord Sand Complex sediment would have buried the esker had the sand complex been deposited after esker formation. Furthermore, the esker continues from the surface of these deposits to the bottom of the Sturgeon Valley, which indicates that the valley pre-existed both deposition of the Bon Accord Sand Complex and the formation of the esker. The esker may, therefore, have been a product of the final drainage of the cavity within which the Bon Accord Sand Complex was formed. The New Lunnon esker is also interpreted to have formed during this final drainage episode.

Drainage through the Sturgeon Valley prior to deposition of the Bon Accord Sand Complex took the form of conduit flow which supplied meltwater to the cavity in the areas of the Duffield and Big Lake sand complexes. A major subglacial meltwater event, at this stage, flowed within and east of the valley, creating a scour field extending more than 50

km^2. The event also eroded numerous flutes along the southern rim of the Sturgeon Valley; some beginning near the valley floor may be traced for kilometers onto the adjacent upper plain. This event may have been contemporaneous with Stage 4 (Fig 4.2) in the southwestern study area.

Retreat and stagnation of ice over the region, perhaps at about 10 700 B.P (St-Onge 1972), produced landforms that cannot be explained by subglacial processes. Regressive, arcuate ridges over the Bon Accord Sand Complex are found in several other areas of glaciofluvial and glaciolacustrine sediment throughout the region. These ridges are interpreted to have formed by the raising and lowering of buoyant ice in an ice-marginal lake, with squeezing of sediment from beneath the ice at the grounding line. Progressive retreat, by calving and melting, produced the regular 300 to 400 m spacing between ridges.

Several perched meltwater channels on the southwestern wall of the Lily Lake Valley must have formed in proximity to ice. Meltwater channels of this type may have formed in either a subglacial or an ice-marginal environment. Subglacial meltwater flowing under hydrostatic pressure could have eroded channels on the valley wall, and could also have synchronously eroded all the perched channels within the Lily Lake Valley. Analyses of the channels show normal basal gradients, and there are no landforms indicative of a subglacial environment within the valley.

The relative timing of the events outlined in this thesis is primarily based on superimposition and cross-cutting relationships, and is better understood than their absolute chronology. Radio-carbon dating control, provided mostly by St-Onge (1972), helped establish the timing of events that occurred at deglaciation. The sediments and landforms interpreted to have been deposited or formed subglacially, however, are more difficult to date. Nevertheless, they may be placed in context with the position of the ice front at the time of their formation. Deposition and erosion by subglacial meltwater activity may well have occurred throughout deglaciation, extending to the ice margin and beyond. It is not possible to judge when formation of the landforms discussed here was first initiated, but this must have occurred after the Livingstone Lake event (Shaw 1983; Shaw et al. 1989) which is estimated to have been between 18 and 15 ka B.P. (Rains et al. 1993). Because subglacial fluvial alteration of the Edmonton region landforms by small episodic meltwater drainage events may have continued up to deglaciation, their formation may have only been complete when the ice margin was immediately southwest of Edmonton, circa 13 000 B.P.

It must be stressed that the hypotheses presented here as alternatives to earlier interpretations of the glacial environments and late-Wisconsinan history of the Edmonton region are by no means complete. The interpretations given appear to best explain the genesis of various landforms and the relationships between landforms. The Duffield Sand Complex, for instance, was previously described as a pitted delta formed in an ice-marginal lake dammed by the Laurentide ice sheet. New evidence, such as the discovery of eskers on the complex and the conclusion that a hummock within the complex was probably formed by subglacial meltwater, calls for reinterpretation. Evidence now exists for a much more complex suite of sediments and landforms in the Edmonton region. Continued use of new data sources, technology, and specialized software to analyze and interpret the geomorphology of the Edmonton region will provide new insight to old problems.