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Deglacial meltwater drainage and glaciodynamics: inferences from Laurentide eskers, Canada

Dr. Tracy A. Brennand *

Department of Geography, Simon Fraser University, 8888 University Drive, Burnaby, BC, CanadaV5A 1S6

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Abstract

This paper evaluates current knowledge of Laurentide eskers in Canada in the light of developments in glacier hydrology and glacial sedimentology. Questions regarding the morpho-sedimentary relations of eskers, the synchroneity and operation of R-channel systems, the role of supraglacial meltwater input and proglacial water bodies, the controls on esker pattern, and the glaciodynamic condition of the ice sheet at the time of esker formation are discussed. A morphologic classification of eskers is proposed. Five types of eskers are identified and investigated. Type I eskers likely formed in extensive, synchronous, dendritic R-channel networks under regionally stagnant ice that terminated in standing water. Type II eskers likely formed in short, subaqueously terminating R-channels or reentrants close to an ice front or grounding line that may have actively retreated during esker sedimentation. Type III eskers plausibly formed in short R-channels that drained either to interior lakes in, or tunnel channels under, regionally stagnant ice. Type IV eskers may have formed as time-transgressive segments in short, subaerially terminating R-channels (or reentrants) that developed close to the ice margin as the ice front underwent stagnation-zone retreat or downwasted and backwasted regionally (stagnant ice); however, formation in synchronous R-channels cannot be discounted on the basis of reported observations. Type V eskers may have formed in H-channels that terminated subaerially. The spatial distribution of these esker types is discussed. The factors that determined Laurentide R-channel pattern and operation were likely a complex combination of (i) supraglacial meltwater discharge, (ii) the number and location of sink holes, (iii) the ice surface slope, thickness and velocity, and (iv) the permeability, topography and rigidity of the bed. These factors cause and respond to changes in ice dynamics and thermal regime over the glacial cycle.

Keywords: glacier hydrology; subglacial drainage systems; ice dynamics; eskers; Laurentide Ice Sheet

1. Introduction

The Laurentide Ice Sheet (Fig. 1a) covered most of Canada and the northern United States in the Late Wisconsinian (23–10 ka BP; Fulton, 1989). The melting of this ice sheet produced large volumes of meltwater which likely had important implications for (i) ice motion by basal sliding (regelation and slippage; Weertman, 1957; Iken, 1981) and bed deformation (e.g., Boulton and Jones, 1979), (ii) ice sheet profile (Arnold and Sharp, 1992; Shoemaker, 1992a,b), and (iii) glacial landform genesis (Boulton and Hindmarsh, 1987; Shaw, 1996).

* Corresponding author. Tel.: +1-604-291-3617; Fax: +1-604-291-5841.

E-mail address: tabrenna@sfu.ca (T.A. Brennand).

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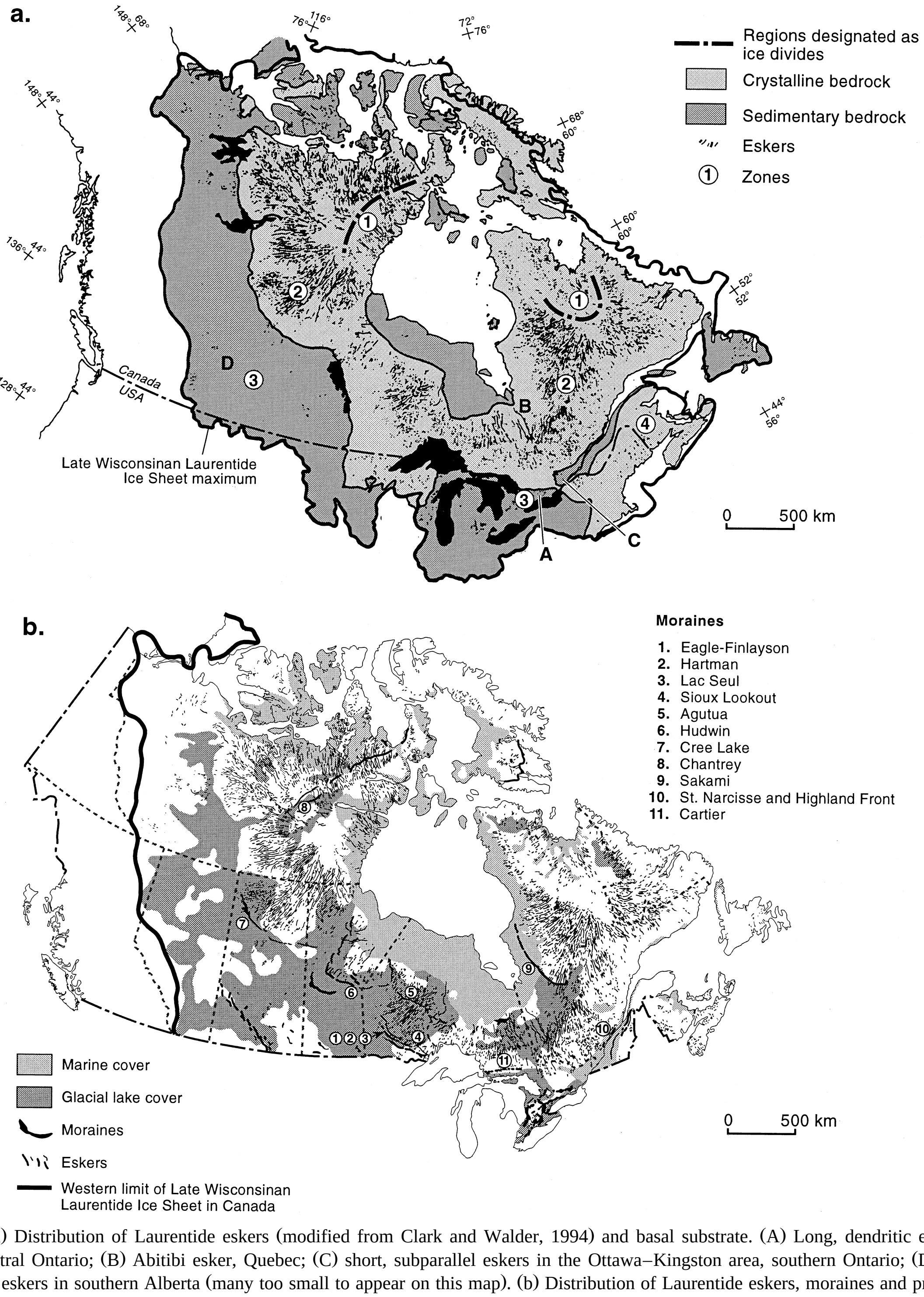

Fig. 1. (a) Distribution of Laurentide eskers (modified from Clark and Walder, 1994) and basal substrate. (A) Long, dendritic eskers in south-central Ontario; (B) Abitibi esker, Quebec; (C) short, subparallel eskers in the Ottawa–Kingston area, southern Ontario; (D) short, deranged eskers in southern Alberta (many too small to appear on this map). (b) Distribution of Laurentide eskers, moraines and proglacial water bodies (Prest et al., 1968).

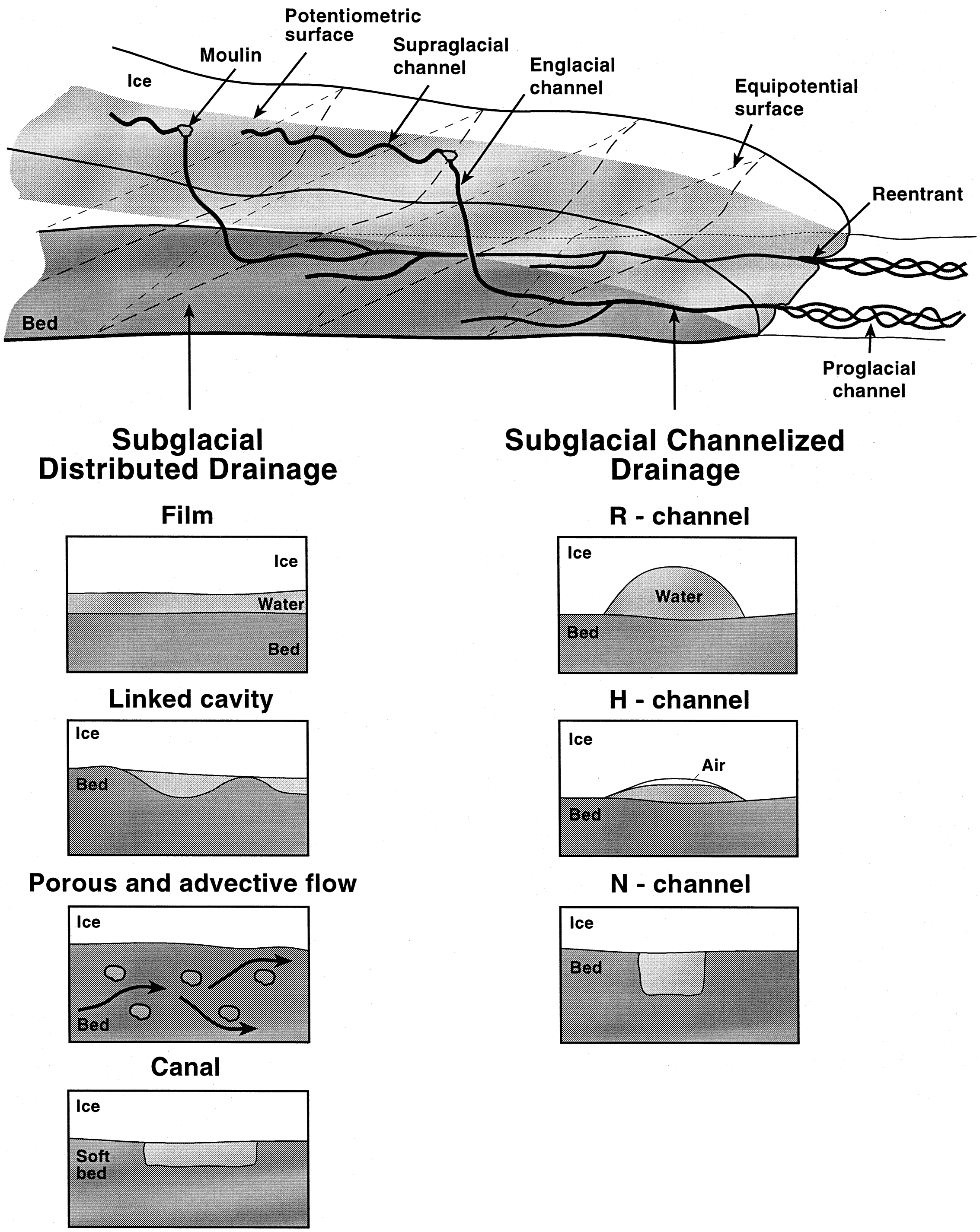

Fig. 2. Karstic meltwater drainage in temperate ice as inferred from field-based glacier hydrology and glaciological theory. See text for explanation.

How can an understanding of Laurentide meltwater drainage be best developed? The way forward lies in a dialog between three bodies of research. (1) Field studies of the hydrology of contemporary glaciers provide insight into the spatial and temporal variability of glacier plumbing (e.g., Hubbard and Nienow, 1997). (2) Theoretical glaciology describes and investigates physically possible drainage scenarios (e.g., Röthlisberger and Lang, 1987). (3) Glacial geomorphology seeks to identify, describe and explain the genesis of glaciofluvial landforms such as eskers (e.g., Banerjee and McDonald, 1975).

Eskers are stratified ridges of sand and gravel that are inferred to be the casts of subglacial channels (R-channels, Röthlisberger, 1972) or reentrants into the ice front (Banerjee and McDonald, 1975; Fig. 2). It has been 45 years since Banerjee and McDonald's (1975) seminal paper on esker genesis. Since that time there have been significant developments in glacier hydrology (below) and in glacial sedimentology (e.g., Rust and Romanelli, 1975). What can eskers usefully tell us about the R-channel drainage of the Laurentide Ice Sheet? Do eskers deposited in R-channels that terminated subaerially exhibit different morpho-sedimentary relations from those deposited in R-channels that terminated in standing water? Did such R-channels function differently? Do eskers record synchronous R-channel systems or are they composed of segments deposited time-transgressively in R-channels that developed close to the ice front as the ice front retreated? What were the hydraulic processes responsible for esker formation? Does esker formation require supraglacial meltwater input? What determined Laurentide esker pattern and distribution? Was the ice sheet regionally active or stagnant at the time of esker formation? This contribution evaluates current knowledge of Laurentide eskers in Canada (Fig. 1) in the light of developments in glacier hydrology and glacial sedimentology. Developments in glacier hydrology are discussed first.

2. Spatial and temporal variability in glacier plumbing

A productive dialog between field hydrologists and glacial theorists has produced a picture of spatial (Fig. 2) and temporal variability in glacier plumbing (e.g., Sharp and Richards, 1996; Hubbard and Nienow, 1997). Field research has included bulk meltwater discharge and chemistry measurements, borehole measurements, tracer studies and investigation of proglacial bedrock exposures (Hubbard and Nienow, 1997). It has been mostly conducted on active, temperate alpine glaciers with steep ice surface slopes that flow over bedrock or bedrock with a thin (<1 m) or patchy sediment cover (e.g., Brand et al., 1987; Hubbard et al., 1995) and that terminate subaerially. Comparable research on non-temperate glaciers is currently lacking (Hodgkins, 1997), and soft-bed drainage scenarios are mainly theoretical as yet (Alley, 1989, 1992; Walder and Fowler, 1994). Field research on the subglacial hydrology of modern ice sheets is limited due to inaccessibility.

Theory and observation suggest that temperate, alpine glacier plumbing is often karstic in form (Shreve, 1972), and functions to route supraglacially, englacially and subglacially generated meltwater to the ice front (Fig. 2). Supraglacial lakes may store water where surface ice is cold (Hodgkins, 1997). Non-temperate ice is mainly drained by supraglacial streams, but subglacial delivery of water via moulins (Iken, 1972) and the passage of channels laterally through non-temperate ice (Pfirman and Solheim, 1989) have also been inferred. The persistence of channels through cold ice may suggest either (i) that such pathways were inherited, perhaps from earlier warm-ice conditions (Hodgkins, 1997) or (ii) that supraglacial meltwater input exceeds the threshold required to keep such channels open. The discussion below focuses on the more extensively researched drainage of alpine, temperate ice (Fig. 2) as a context for understanding Laurentide glaciofluvial landforms. However, development of such landforms under non-temperate ice cannot be discounted and alpine glacier hydrology need not be directly analogous to Laurentide Ice Sheet hydrology.

Water delivered by rainstorms or generated by surface melting of temperate alpine glaciers may percolate through snow and firn or flow as a sheet, but has an incipient tendency to channelize, forming meandering supraglacial streams (Parker, 1975; Fig. 2). Surface water enters the glacier either through a three-dimensional network of veins (Nye and Frank, 1973) and tubes connecting to conduits (Raymond and Harrison, 1975), or through crevasses (Robin, 1974; Röthlisberger and Lang, 1987) or moulins (Holmlund, 1988; Holmlund and Hooke, 1983; Fig. 2). When such sink holes become water-filled the density difference between ice and water (Robin, 1974; Weertman, 1974), and the heat released by the water (Hooke, 1989) can result in the propagation of sink holes down through the glacier to the glacier bed. Englacial channels are thus formed (Fig. 2). Water in such channels may initially drain to a distributed drainage system, but once a threshold discharge is exceeded it will drain to subglacial channels when the glacier bed is rigid (Willis et al., 1990; Fig. 2).

Shreve (1972) has argued (i) that englacial channels form an upward-branching, arborescent system normal to upglacier-dipping equipotential surfaces and (ii) that subglacial channels follow paths normal to equipotential contours (Fig. 2). He assumed that water pressure in such channels is similar to the pressure in confining ice and that the bed slope is less than ~ 11 times the ice surface slope. Equipotential surfaces are defined as surfaces of equal water-pressure potential and equipotential contours are defined as the intersection of the equipotential surfaces with the bed (Shreve, 1972; Fig. 2). As equipotential contours are mainly controlled by ice surface slope and secondarily by bed topography, subglacial channels may also follow topographic lows, be routed mid-way up valley sides (Hooke, 1998, fig. 8.23) and cross bumps (Shreve, 1972). However, during conditions of variable flow discharge, Shreve's (1972) assumption of balanced water and ice pressure is unlikely to be valid. Under such conditions, the slope of the potentiometric surface (as defined for each conduit; Fig. 2) or hydraulic head (i.e., the elevation difference between the zone of recharge at the surface and discharge at the portal) may be more important in controlling subglacial water flow. Rapidly increasing supraglacial meltwater discharge may cause the potentiometric surface to become steeper than the ice surface slope because small channels throttle flow (Röthlisberger, 1972, Fig. 5f; Menzies and Shilts, 1996). Consequently, water may flow to the ice margin even if the ice surface slope is flat.

Channels are only one of many subglacial drainage configurations. The subglacial drainage system develops spatially and temporally variable channelized and distributed systems (e.g., Hubbard and Nienow, 1997; Fig. 2) depending on such factors as water discharge, ice surface slope and thickness, ice velocity, and the permeability, topography and rigidity of the bed (e.g., Arnold and Sharp, 1992). Channels form efficient drainage pathways, stable at relatively low steady-state water pressures. In contrast, distributed systems form inefficient drainage pathways, stable at relatively high steady-state water pressures.

2.1. Subglacial drainage on rigid beds

The configuration of the subglacial drainage system of temperate alpine glaciers flowing over a rigid bed often takes the form of channels and films and/or linked cavities (e.g., Hubbard and Nienow, 1997). Channels may exhibit one of three forms (Fig. 2) and may only drain a fraction of the bed (Hubbard and Nienow, 1997). (i) R-channels (Röthlisberger, 1972) are semi-circular channels cut upward into the ice. Their size and water pressure reflects a balance between channel enlargement by viscous heating and closure by ice deformation when the channels are water-filled. Röthlisberger's theory predicts that steady-state (instantaneous heat transfer) water pressures should be lower when discharges are higher and consequently R-channels should form dendritic networks. Walder (1982) has argued that supraglacial meltwater input is required for R-channels to be persistent at relatively high discharges. (ii) H-channels (Hooke, 1984) are broad, flat channels cut upward into the ice that tend to follow local bed slope. Such channels form where water flows at atmospheric pressure beneath thin ice and on steep downglacier bedslopes. (iii) N-channels (Nye, 1973) are incised into bedrock, perhaps suggesting long-term channel stability under some alpine glaciers (Walder and Hallet, 1979).

The presence of linked cavities and films (Fig. 2) have been inferred from proglacial bedrock mapping (e.g., Walder and Hallet, 1979), slow dye returns (e.g., Willis et al., 1990) and high subglacial water pressures (e.g., Kamb, 1987). This distributed system is mainly responsible for transferring interchannel, subglacially generated meltwater (basal melt from geothermal and viscous heating $\sim 10^{-2}$ m a^{-1}; Paterson, 1994), or temporarily trapped meltwater from other sources, to channels (e.g., Walder and Hallet, 1979; Willis et al., 1990). Water films have been attributed to glacier sliding by regelation (Weertman, 1957, 1964). Water-filled cavities linked by small channels form in the lee of bed bumps when glacier sliding velocity exceeds a critical value (Lliboutry, 1968).

2.2. Subglacial drainage on ''soft'' beds

Where glaciers flow over permeable substrates, either distributed drainage (Walder and Fowler, 1994) or coexisting distributed and channelized drainage (Alley, 1992) has been proposed. Water may drain through the voids between substrate clasts (Darcian flow; e.g., Boulton and Jones, 1979) or through macropores (Clarke et al., 1984; Murray and Clarke, 1995). Such porous flow (Fig. 2) has been inferred from borehole measurements of water pressure and turbidity across subglacial channels (e.g., Hubbard et al., 1995; Gordon et al., 1998). It is rarely sufficient to evacuate all water flowing at the bed (Alley, 1989). Where subglacial pore water pressure exceeds a critical value, dependent on substrate rheology and water flux, the bed is considered to be ''soft'' and drainage may occur by advection in mobilized sediment (Alley et al., 1986), a thin film (Alley, 1989) or canals (broad, shallow, ''braided'' drainage pathways; Walder and Fowler, 1994) (Fig. 2). Canals differ from N-channels in that they are proposed to be stable at high water pressures, and form by a combination of bed deformation and fluvial erosion; they have yet to be observed (Walder and Fowler, 1994). In contrast, coexisting low pressure channels and deforming beds have been reported (e.g., Engelhardt et al., 1978) and theoretically explained on the grounds that substrate creep toward such channels is limited by substrate rheology (Alley, 1992).

2.3. Temporal variability and evolution of glacier drainage

The availability of supraglacially and englacially generated meltwater varies according to diurnal, seasonal and annual melt cycles; the supply is unsteady (e.g., Østrem, 1975). Subglacially generated meltwater is available throughout the year, but may increase in summer as ice motion increases (Willis et al., 1990). Jökulhlaups (e.g., Nye, 1976) and cavity drainage events (e.g., Kamb et al., 1985) can also affect episodic variability in meltwater discharge.

Where temperate, alpine glaciers terminate subaerially, drainage evolves annually as the melt zone expands and contracts (e.g., Willis et al., 1990; Richards et al., 1996). This evolution involves competition and replacement of drainage modes. In the winter, distributed systems (Fig. 2) dominate subglacial drainage of water supplied by basal melt or storage release (Willis et al., 1993). During the melt season, channels dominate drainage as supraglacial meltwater input causes discharge to increase beyond a critical value (Willis et al., 1990). Channels develop by coalescence of linked cavities (Kamb, 1987) or by downglacier extension (melt) of individual channel segments (Gordon et al., 1998; Nienow et al., 1998). The path of such water-filled channels is mainly determined by ice surface slope (Shreve, 1972). This path may persist despite flow at atmospheric pressure (in H-channels) later in the melt season because channel migration to bed lows may be limited by local back-pressure effects, the position of the supraglacial sink point and the time available (Sharp et al., 1993). The upglacier expansion of the melt zone during the melt season gives the impression that channel systems grow headward (Gordon et al., 1998; Nienow et al., 1998). Diurnal pressure fluctuations may drive porous flow into and out of channels (e.g., Hubbard et al., 1995). In the fall, channels are replaced by distributed systems; reduced meltwater discharge allows ice deformation to constrict channels, eventually reversing the subglacial water pressure gradient (e.g., Hooke, 1989). Channels may completely collapse in the winter unless they remain water-filled (Haefeli, 1970). Such collapse is most likely to start where ice is thickest, channel size smallest and meltwater discharge least (Hubbard and Nienow, 1997). Channels may remain water-filled in winter if (i) channel collapse (Hubbard and Nienow, 1997) or sediment plugging (Boulton and Hindmarsh, 1987) is initiated in a downflow location, (ii) a threshold discharge persists through winter, perhaps supplied from stored sources (Willis et al., 1993), or (iii) channels end in a standing water body, in which case they are hydraulically dammed and will remain water-filled to the level of the winter potentiometric surface (e.g., Powell, 1990).

3. Laurentide meltwater drainage systems

It seems reasonable to expect that subglacial drainage configurations of the Laurentide Ice Sheet included both distributed and channelized systems (e.g., Alley, 1996; Fig. 2), and that such systems were likely spatially and temporally variable (e.g., Arnold and Sharp, 1992). Glacial geomorphology provides one approach to elucidating this variability, because morpho-sedimentary observations must be explained by models of ice sheet hydrology. However, this approach is not without limitations; for example (i) observations are mostly biased toward subglacial and proglacial drainage systems and deglacial snap-shots of Laurentide Ice Sheet hydrology, (ii) absolute chronological control is generally lacking for subglacial landforms and sediments, and (iii) subglacial sedimentary evidence is often biased toward high magnitude flow events, but sediments in proglacial water bodies contain a more complete record of ice sheet drainage.

To date, most geologic and geomorphic corroboration of subglacial distributed drainage systems has been viewed as equivocal (e.g., Benn and Evans, 1998). Roches moutonnées, with their abraded stoss sides and plucked lee slopes, suggest drainage by linked cavities (e.g., Lliboutry, 1968; Iken and Bindschadler, 1986). Morpho-sedimentary observations from drumlins have been interpreted as evidence of both catastrophic sheet flow (e.g., Shaw, 1996) and advective flow (bed deformation; e.g., Boulton and Hindmarsh, 1987). Linear mineral dispersal trains may suggest advective flow (e.g., Dyke and Morris, 1988). Tunnel channels with upslope thalwegs have been attributed to canal drainage (Walder and Fowler, 1994) and subglacial deformation (Boulton and Hindmarsh, 1987), but catastrophic channel drainage (e.g., Wright, 1973) and more ‘‘normal’’ N-channel drainage (Mooers, 1989) have also been proposed. Geomorphic corroboration of subglacial channelized drainage systems has been less controversial (with the exception of tunnel channels; O'Cofaigh, 1996). Eskers may form in either R- or H-channels, but such a distinction cannot be easily made from morpho-sedimentary observations.

4. Classification of Laurentide eskers

Eskers have been mapped throughout the area covered by the Laurentide Ice Sheet (Prest et al., 1968; Fig. 1). The longest Laurentide eskers are over 800 km long (Craig, 1964) including gaps, and may be nearly unbroken for up to 300 km (Shilts et al., 1987). They attain heights of a few meters to over 50 m at the land surface, but deposits in excess of 200 m may occur below the surface (Munro esker, Ontario; Banerjee and McDonald, 1975). Eskers are generally less than 150 m in width, but may exceed several kilometers (e.g., Banerjee and McDonald, 1975). In planform, eskers can be single, sinuous, undulatory ridges with broad or sharp crests (e.g., Shreve, 1985). Alternatively, they may exhibit periodic enlargements or ‘‘beads’’ (e.g., Banerjee and McDonald, 1975; Brennand, 1994), or subdivide and recombine downflow (anabranched planform, e.g., Shaw et al., 1989). Some eskers are relatively isolated, others form integrated dendritic networks with up to fourth-order tributaries (e.g., Shilts et al., 1987). Eskers may occur in association with both subaerial and subaqueous ice-marginal landform assemblages such as pitted subaerial outwash, subaerial or subaqueous fans (although not all subaqueous fans need be ice marginal, Gorrell and Shaw, 1991), deltas and end moraines, and glaciolacustrine and glaciomarine sediments (e.g., Shilts et al., 1987; Fig. 1b). They often occur in association with subglacial landforms such as drumlins, rogen moraine, hummocky terrain and tunnel channels (e.g., Shilts et al., 1987; Shaw et al., 1989; Munro and Shaw, 1997).

Banerjee and McDonald (1975) chose a genetic classification of eskers which identified esker formation in tunnels (R-channels) and open channels (H-channels or reentrants into the ice front). They emphasized that these channels could terminate subaqueously or subaerially. Shilts et al. (1987) describe zonal assemblages of sediments and landforms, including eskers, on the Canadian Shield. Menzies and Shilts (1996) describe esker patterns associated with North American ice sheets. The zones described below and shown on Fig. 1a are based on these zonal descriptions.

In zone 1 eskers are rare; these are regions designated as ice divides (Fig. 1; Shilts et al., 1987). In zone 2 (Fig. 1a) and generally in areas underlain by crystalline bedrock, esker density is high (esker spacing $\sim$ 10 km) except in drift-poor areas. Most eskers form long, partially discontinuous, dendritic networks (Fig. 3a) which either (i) radiate away from regions designated as ice divides (Fig. 1a), or (ii) terminate at major arcuate moraines (Fig. 1b). Gaps along eskers may be occupied by channels eroded into the basal substrate (Klassen, 1986; Thorleifson and Krystjanssen, 1993). Short eskers occur in some places between major dendritic esker systems. In zone 3, esker density is low downflow from major arcuate moraines and in areas generally underlain by sedimentary bedrock (Fig. 1). Most eskers are short, forming subparallel or deranged patterns (Fig. 3b,c), or are isolated. Several longer, dendritic esker systems are located in tunnel channels (A in Fig. 1a; Brennand, 1994). Zone 4 (Fig. 1a) encompasses regions of high relief, such as the Appalachians. Here, eskers are generally confined to major valleys but may cross drainage divides (Rampton et al., 1984).

Some authors suggest that this esker pattern was substrate-controlled; eskers are present over rigid beds (zone 2) and rare over soft beds (zone 3) (Clark and Walder, 1994; Walder and Fowler, 1994; Fig. 1a). Other authors have noted that mapping of the complete esker pattern is likely confounded by later burial or reworking in glaciolacustrine or glaciomarine environments (e.g., Allard, 1974; Shilts et al., 1987; Fig. 1b). Shilts et al., (1987) and Menzies and Shilts (1996) have argued that sediment availability and changes in ice dynamics during the deglaciation cycle are causal factors; zone 3 eskers formed at an actively retreating ice margin and zone 2 eskers formed under regionally stagnant ice (Fig. 1a).

This paper introduces a morphologic classification of Laurentide eskers in an attempt to reconcile the importance of both individual esker genesis and regional esker pattern to an understanding of Laurentide R-channel drainage. Three esker morphologies are identified: (i) long, dendritic eskers, (ii) short, subparallel eskers, and (iii) short, deranged eskers (Fig. 3). Each of these esker morphologies may have formed in channels which terminated subaqueously or subaerially (Banerjee and McDonald, 1975). Five combinations have been reported in the literature: (i) long, dendritic eskers which formed in R-channels that terminated in standing water (type I); (ii) short, subparallel eskers which formed in R-channels or reentrants that terminated in standing water (type II);

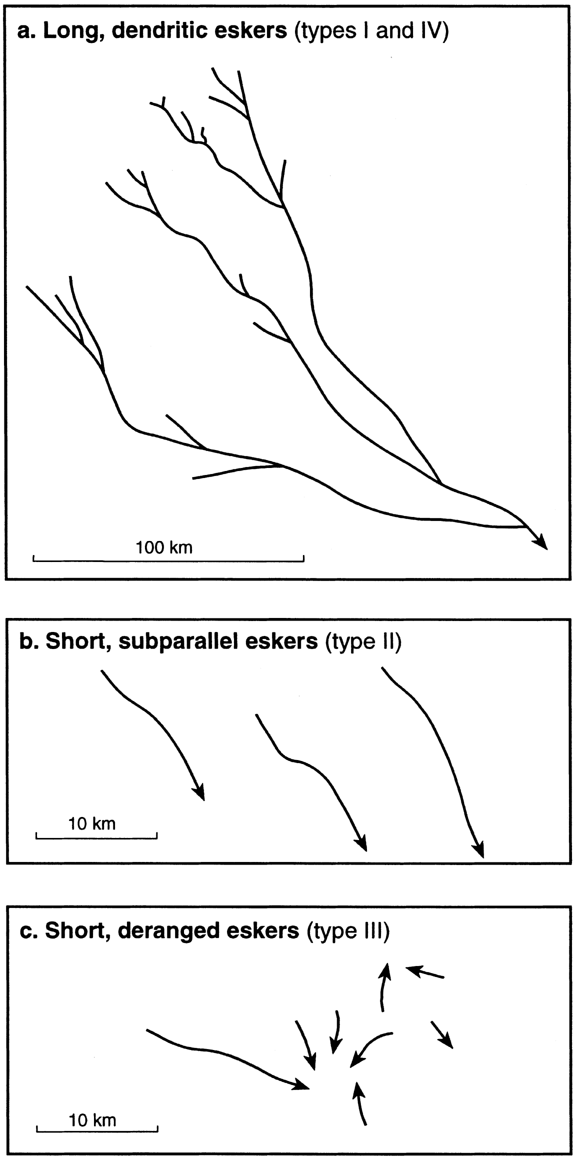

Fig. 3. A morphological esker classification. Each of these esker morphologies may have formed in R-channels terminating in standing water (types I, II and III). Long, dendritic eskers may also have formed in R-channels terminating subaerially (type IV). Deranged eskers lack regional alignment.

(iii) short, deranged eskers which formed in R-channels that terminated in standing water (type III); (iv) long, dendritic eskers which formed in R-channels that terminated subaerially (type IV); and (v) short eskers which formed in R-channels or reentrants that terminated subaerially. Different morpho-sedimentary relations (below) for types I, II and III suggest that distinct drainage systems and glaciodynamic conditions are associated with each of these esker types. Lack of observations and contradicting and equivocal interpretations for types IV and V make the discussion of subaerially terminating R-channels inconclusive.

5. Eskers deposited in R-channels that terminated in standing water

Eskers that terminate in subaqueous fans (identified on morphologic and sedimentologic grounds) or that exhibit subaqueous fans along their length most likely formed in R-channels which terminated in standing water. Often such landforms are transitional downflow into glaciolacustrine or glaciomarine sediments. Most of the research on Laurentide eskers in Canada has been concerned with such conditions.

5.1. Long, dendritic eskers formed in R-channels that terminated in standing water (type I)

5.1.1. General depositional environment

Type I esker systems are tens to hundreds of kilometers long (Fig. 3a). Were such long eskers deposited in extensive, synchronous R-channel systems or were they deposited as time-transgressive segments in less extensive R-channels or reentrants that developed close to the ice front as the ice front retreated? This question is difficult to answer with certainty as determination of geomorphic and sedimentary continuity are generally confounded by gaps along esker ridges, esker burial or removal (for aggregates), limited sedimentary exposures and the non-uniform nature of fluvial sedimentation. Where research has been designed to answer this question, esker sedimentation in extensive, synchronous R-channels has been inferred (Brennand, 1994; Brennand and Shaw, 1996). This conclusion is contrary to the widespread assertion that long eskers were formed in successive segments time-transgressively (e.g., Banerjee and McDonald, 1975; Dyke and Dredge, 1989). Morpho-sedimentary diagnostic criteria for eskers formed (i) in extensive (long), synchronous R-channels and (ii) as time-transgressive segments in (short) R-channels or reentrants close to the ice front are presented in Fig. 4.

The basis for inferring extensive, synchronous R-channels is elaborated below, based on observations from four eskers (40–70 km long) in south-central Ontario (Brennand, 1994; A in Fig. 1a), and one esker (at least 300 km long) in Quebec (the Abitibi esker, or Harricana Moraine, Veillette, 1986; Brennand and Shaw, 1996; B in Fig. 1a). (1) Esker ridges are relatively continuous. On surficial geology maps, long ridges are separated by short gaps, but these ridges are not punctuated by subaqueous fans or deltas. Subaqueous fans do occur at the downflow ends of these eskers; some are superimposed over esker ridges near their downflow ends. (2) Paleoflow variability measured from both sand structures and gravel fabrics at points along tens to hundreds of kilometers of esker ridge is invariably low; higher paleoflow variability is observed in subaqueous fans at the downflow ends of eskers. This implies deposition within a constrained environment, most likely an extensive, synchronous R-channel, terminating in standing water. For low paleoflow variability to be recorded in a segmental esker, either the reentrant must have remained invariably narrow, or subaqueous fan formation at the ice front must have been inhibited; both conditions seem unlikely. (3) Regional trends in clast characteristics are observed. In south-central Ontario, the increase in clast mean roundness downflow along a 70-km long esker suggests relatively long transport distances, vigorous flows and low sedimentation rates. In Quebec, the upflow portion of the Abitibi esker is relatively narrow; there is poor preservation of metabasalt clasts; most clasts are well rounded and paleoflow variability is low. High flow power and vigorous transport in a narrow R-channel are consistent with these observations. Downflow the esker is wider; metabasalt clasts are better preserved; clasts are mainly subrounded and paleoflow variability is higher. Assuming constant sediment supply along the R-channel, this implies lower flow power, less vigorous transport and higher sedimentation rates downflow. Together, these regional trends, over

SEDIMENTATION IN EXTENSIVE, SYNCHRONOUS R-CHANNELS	SEGMENTAL SEDIMENTATION IN R-CHANNELS OR REENTRANTS CLOSE TO THE ICE FRONT AS THE ICE FRONT RETREATED
Esker Morphology: • relatively continuous ridge with gaps • fans in a terminal position or superimposed **Esker Sedimentology:** • coarse ridge with fines mostly in terminal or superimposed fans • low paleoflow variability except in terminal superimposed fans • regional trend in clast characteristics (e.g. lithology and roundness)	**Esker Morphology:** • ridge segments punctuated by fans **Esker Sedimentology:** • segmental trends in sediment texture and structure consistent with flow deceleration on entry into a standing water body • higher overall paleoflow variability due to presence of fans • segmental trends in clast characteristics (e.g. roundness)

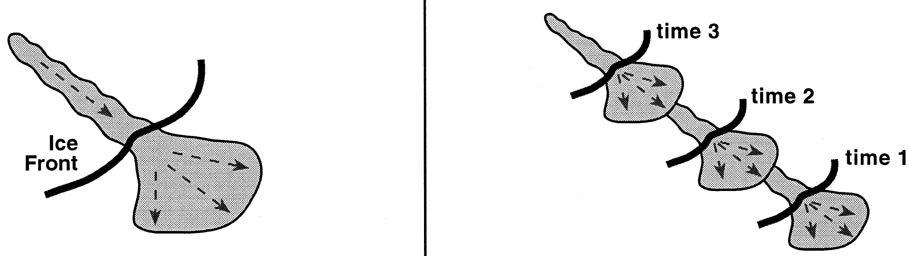

Fig. 4. Morpho-sedimentary diagnostic criteria for determining the genesis of long, dendritic eskers which formed in R-channels or reentrants that terminated in standing water (derived from Banerjee and McDonald, 1975; Shreve, 1985; Brennand, 1994; Brennand and Shaw, 1996).

~ 250 km, are consistent with a synchronous R-channel that increases in size downflow as the ice sheet thins toward its margin. The presence of deep, proglacial lake Barlow-Ojibway likely also favoured higher sedimentation rates and reduced R-channel closure rates downflow in the Abitibi esker.

Extensive, synchronous R-channels terminating in standing water must have been water-filled at least to the level of the potentiometric surface defined by the water body. This surface must have been steep in summer to provide sufficient head to drive water to the ice front, often against topographic gradient (Brennand and Shaw, 1996). Gaps along long eskers formed under these conditions may be attributed to contemporaneous erosion or non-deposition in non-uniform conduits (Shreve, 1972; Banerjee and McDonald, 1975; Brennand, 1994), as well as post-depositional erosion and burial (Allard, 1974).

5.1.2. Hydraulic processes

The building blocks of type I eskers are gravel facies (e.g., Brennand, 1994; Fig. 5; Table 1). Sand facies (massive, plane-bedded, cross-bedded, cross-laminated and in-phase wave structures) alternate with these, but generally form thin, discontinuous beds (Brennand, 1994; Fig. 5c). Limited experimentation has been reported on the sedimentary characteristics expected for coarse sediment in pipe flow (e.g., Acaroglu and Graf, 1968; McDonald and Vincent, 1972). On the basis of sedimentary descriptions, grain-size and imbrication measurements, several gravel facies have been identified and interpreted (Fig. 5; Table 1; the basis for these interpretations is presented in Brennand, 1994). They record deposition from powerful fluid flows and hyperconcentrated dispersions. Gravel clasts range in size from granules to boulders, and boulder facies are not

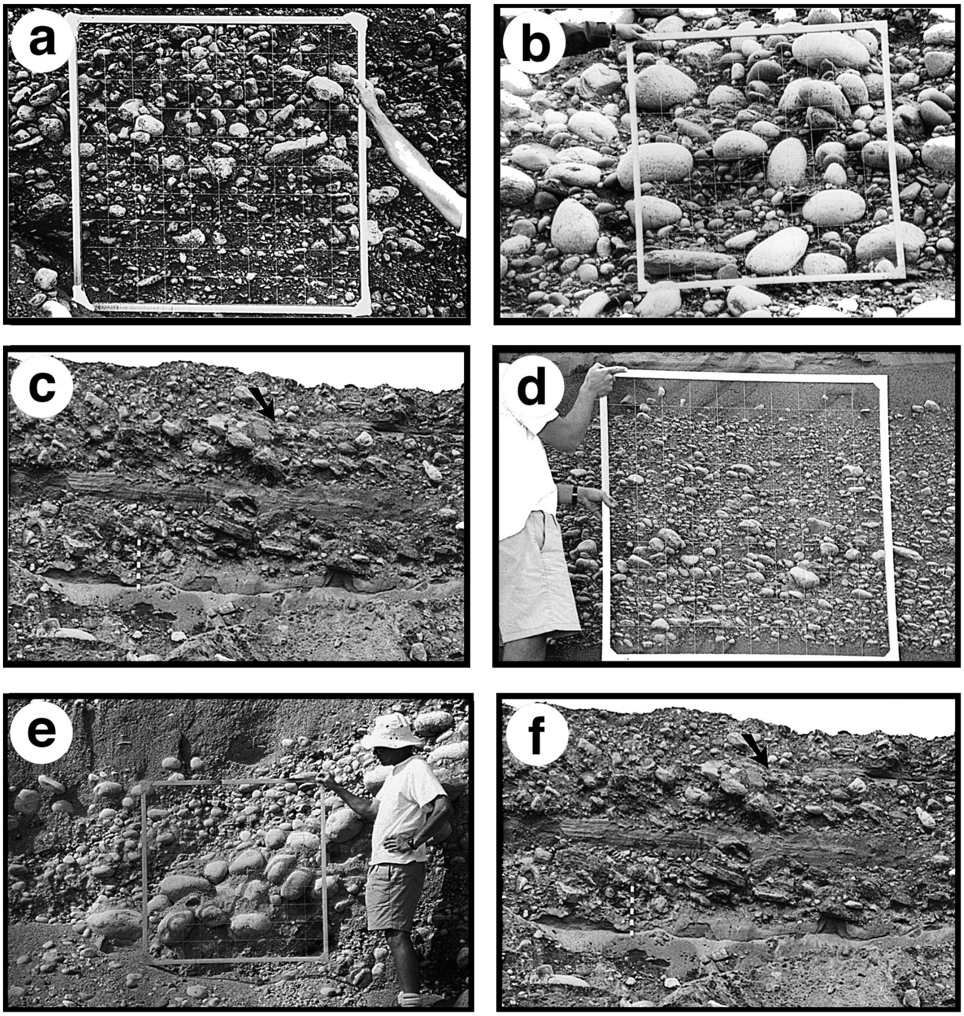

Fig. 5. Gravel facies observed in type I eskers. (a) Heterogenous, unstratified gravel with poorly delineated lenses of bimodal, clast-supported, polymodal and openwork gravel, Campbellford esker (Brennand, 1994). Grid is 1 m^2. (b) Heterogenous, unstratified gravel with imbricate clast clusters, Abitibi esker (Brennand and Shaw, 1996). (c) Massive, imbricate, clast-supported gravel alternating with sand (truncated at arrow), Tweed esker (Brennand, 1994). Meter rod for scale. (d) Imbricate, polymodal, matrix-rich gravel, Norwood esker (Brennand, 1994). (e) Grading in cross-bedded gravel, Abitibi esker (Brennand and Shaw, 1996). (f) Trough cross-bedded gravel, Tweed esker (Brennand, 1994). See Table 1 for description and interpretation. Reprinted with permission from Elsevier Science.

uncommon. Mean flow velocities required to move boulders up to 1 m in diameter may exceed 5 m s^{-1} (after Costa, 1983).

Gravel facies are arranged into ridge-scale bedforms or macroforms: composite, oblique-accretion avalanche bed, and pseudoanticlinal (Fig. 6). Macroforms are sedimentation zones that record powerful flows down geometrically non-uniform R-channels, the macroform style likely being controlled by R-channel geometry (Brennand, 1994; Brennand and Shaw, 1996). Non-uniform R-channel geometry is implied by the undulatory crest-line and variable width of eskers. Some esker enlargements (wider and/or higher reaches), interpreted as deltas in earlier literature (e.g., Allard, 1974; Ringrose and Large, 1977; Rondot, 1982; Lindström, 1993), may be reclassified as macroforms.

Composite macroforms are architecturally complex, generally being composed of stacked gravel and sand facies (Fig. 6a; Table 1). They may be

Table 1
Gravel facies observed in type I eskers[a]

Facies	Structural characteristics[b]	Main sediment support mechanisms	Origin of facies structure	Interpretation
Heterogeneous, unstratified gravel	vaguely lenticular textural organization in heterogeneous gravel; ungraded; framework supported; cluster imbrication with a(t) b(i) and a(p) a(i) and high dips; tabular or pseudoanticlinal bed geometry	fluid turbulence; bed	longitudinal sediment sorting during tractional transport; sorting associated with the development of cluster bedforms	facies within a composite or pseudoanticlinal macroform; deposition during the waning stages of floods
Massive, imbricate, clast-supported gravel	massive; relatively bimodal and clast-supported; generally ungraded; pervasive imbrication with a(t) b(i) and a(p) a(i) and high dips; tabular or pseudoanticlinal bed geometry	fluid turbulence; bed	deposition primarily from traction transport with minor suspension and saltation transport	facies within a composite or pseudoanticlinal macroform
Plane-bedded gravel	plane bedded, becoming more massive in coarser units; polymodal, graded or longitudinally sorted beds; a(t) b(i) dominant; tabular bed geometry	bed; fluid turbulence	deposition primarily from traction transport	facies within a composite macroform
Imbricate, polymodal gravel	texturally polymodal although some may be more matrix-rich; un-graded or weak normal or inverse-to-normal grading; a(p) a(i) and a(t) b(i); tabular or lenticular bed geometry	fluid turbulence; dispersive pressure; buoyancy; hindered settling	deposition from hyper-concentrated dispersion (homogenous or hetero-genous)	may form in-phase wave structures associated with the establishment of a density interface at points of flow expansion
Cross-bedded gravel	graded foreset beds: (A) base — bimodal, clast-supported gravel occasionally with convoluted laminae in sand; (B) middle — openwork gravel; (C) upper — openwork gravel (smaller grain size)	fluid turbulence; bed	bedform migration or macroform progradation; longitudinal sediment sorting during transport of hetero-geneous gravel, and lee-side deposition of suspended load in return flow beneath a separation eddy	large gravel dunes; facies within composite macroform, associated with bedform migration or macroform progradation

[a] Modified from Brennand and Shaw (1996).
[b] a(t) b(i): clast *a*-axis transverse to flow direction, *b*-axis imbricate; a(p) a(i): clast *a*-axis parallel to flow direction and imbricate.

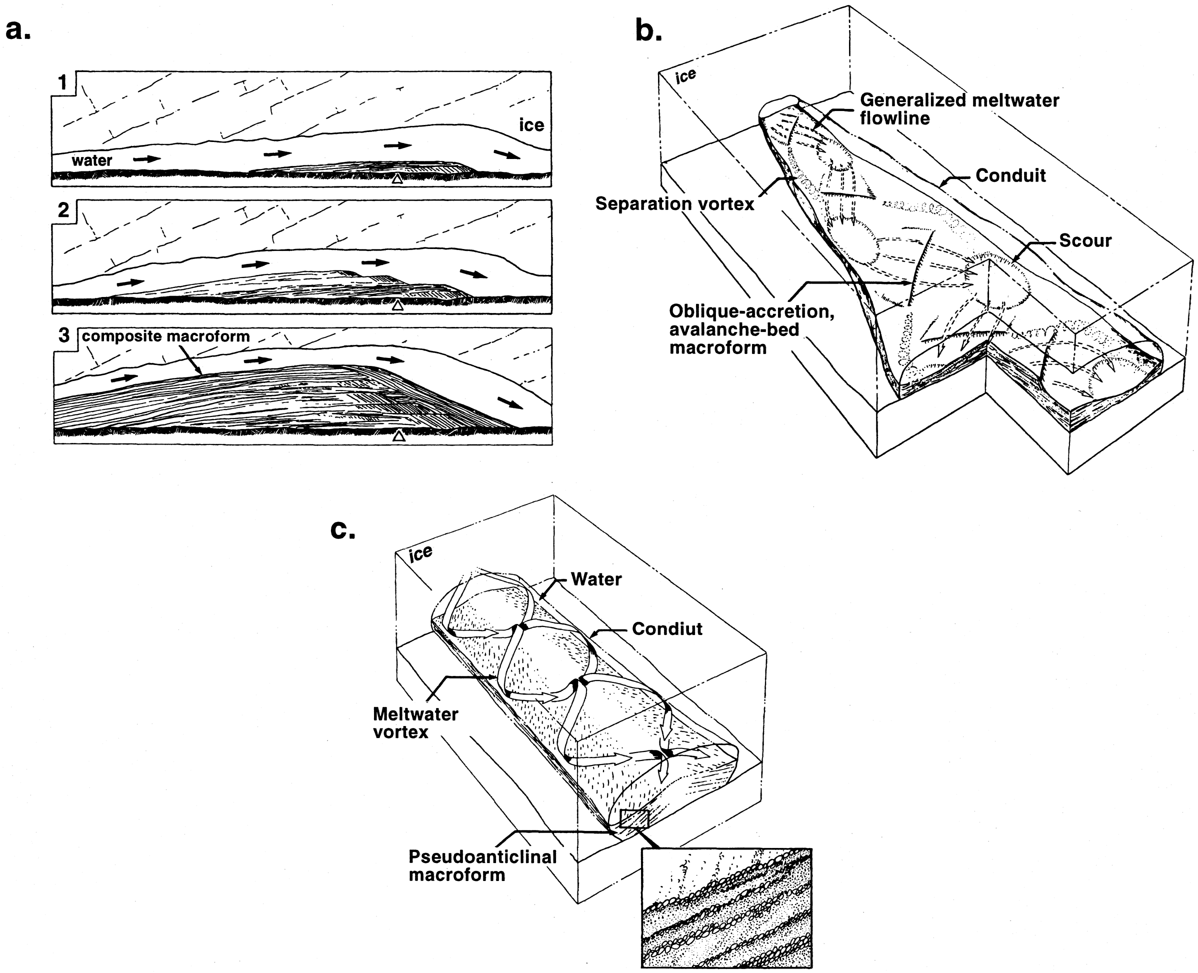

Fig. 6. Models of macroform development in type I eskers (after Brennand and Shaw, 1996). (a) Incremental formation (stages 1–3) of a composite macroform in an R-channel enlargement. Over time and with sedimentation the ice roof melts up and back. (b) Formation of oblique macroforms resulting in oblique-accretion, avalanche beds. These forms are related to the combined effects of separation vortices generated to the lee of the bedform and convergence scours. Alternating, oblique-accretion avalanche beds are expected where oblique macroforms migrate into R-channel enlargements. (c) Formation of pseudoanticlinal macroforms related to paired vortices of similar power in R-channel reaches of uniform geometry. Reprinted with permission from Elsevier Science.

more than 10 m in height and over 1 km long. They have gently sloping stoss-sides (5°–10°), and may have higher-angled lee slopes. Some composite macroforms exhibit downflow and lateral fining. Morphologically, such macroforms are associated with esker enlargements. These characteristics have been used to infer that composite macroforms were formed at positions of R-channel enlargement and concomitant flow expansion within a synchronous, non-uniform R-channel (Brennand, 1994; Brennand and Shaw, 1996). The location and sedimentary details of composite macroforms were most likely initially controlled by local R-channel geometry and sediment supply, and later controlled by the interaction between macroform and R-channel geometries (Fig. 6a). Oblique-accretion avalanche bed (OAAB) macroforms exhibit large-scale cross-beds dipping obliquely to the general flow direction (Fig. 6b) indicated by the esker axis and other bedforms. They occur at esker enlargements. These macroforms record alternating bars that likely developed downflow from flow convergence scours under conditions of high flow power (Brennand and Shaw, 1996; Fig. 6b). Pseudoanticlinal macroforms (q.v. Stone, 1899) exhibit broad, low-angled, arched or anticlinal beds with crest-convergent, downflow gravel fabrics (Fig. 6c). These macroforms occur along geometrically uniform, narrow reaches of esker ridges. These observations are most simply explained by the operation of secondary currents or vortices of similar power within narrow reaches of a synchronous R-channel (Fig. 6c; Shreve, 1972).

Minor, subaqueous fan sediments may occur alongside bends in type I eskers (Brennand, 1994). Anabranched reaches and extended, hummocky zones are associated with some eskers of this type (Shaw et al., 1989; Brennand, 1994). These features contain intensely folded and faulted sediment. They are best explained as depositional products of episodic flood events (Shaw et al., 1989; Brennand, 1994). Such flood events were most likely the result of meltwater storage release, perhaps from supraglacial lakes or subglacial or englacial water-bodies (Shaw, 1994), but may have been triggered by seasonal melting and rainstorms. During such high discharge events, water pressure within the R-channel was sufficient to breach the high pressure seal at the R-channel margin or cause it to migrate outwards (Gordon et al., 1998). Three effects are proposed: (i) R-channel — cavity connection for subaqueous fan sediments; (ii) the establishment of a broad zone of minor R-channels for anabranched reaches; or (iii) localized hydraulic lifting over a broad zone on either side of the R-channel for extended, hummocky zones (Brennand, 1994). Decreased flow competence due to flow expansion resulted in rapid deposition. Sedimentation ceased when discharge declined and the high pressure seal to the R-channel margin was reestablished. Sediments were deformed during R-channel closure and ice-bed recoupling. Alternatively, anabranched reaches may be the product of sediment plugging (Shilts et al., 1987) or hydraulic damming (M. Sharp, 1998, personal communication) of R-channels and their subsequent relocation.

5.1.3. Hydrologic regime

Vertically stacked, gravel–sand couplets are observed in many type I esker exposures (e.g., Brennand, 1994; Brennand and Shaw, 1996). Whereas some may be attributed to spatial changes in transport and deposition along a non-uniform R-channel, where laterally extensive they are more likely to be a product of abrupt temporal changes in flow competence (Brennand, 1994; Brennand and Shaw, 1996). Such unsteady flow may have occurred on daily, weather-related, seasonal, annual or episodic (e.g., jökulhlaups or other forms of storage release; Nye, 1976; Willis et al., 1993) time scales. As varves are often transitional to subaqueous fans at the downflow ends of type I eskers, it is most likely that gravel–sand couplets record supraglacial input to subglacial R-channels (q.v. Banerjee and McDonald, 1975). Supraglacial meltwater input is likely also necessary to maintain powerful flows down discrete R-channels (Walder, 1982); intrachannel meltwater, generated by geothermal and frictional heating or including supraglacial water that has been stored, likely passes though an inefficient distributed system before entering R-channels (Hubbard and Nienow, 1997).

The implications of supraglacial meltwater input are important. The downflow transition of multiple gravel–sand couplets in eskers to varves raises the probability that esker sedimentation occurred over many years. Type I esker sediments are remarkably undeformed, suggesting that R-channels did not close during esker deposition, despite the normal winter tendency (q.v. Hooke, 1989). R-channels may remain water-filled in winter if (i) channel collapse (Hubbard and Nienow, 1997) or sediment plugging (Boulton and Hindmarsh, 1987) was initiated in a downflow location, (ii) a threshold discharge persisted through the winter (Willis et al., 1993), or (iii) the channel terminated in a standing water body (Powell, 1990; Fig. 7a).

5.1.4. Deglacial drainage and glaciodynamics associated with type I esker systems

Although the above discussion focuses on particular eskers within extensive dendritic systems, it seems reasonable to propose that many type I eskers likely formed in synchronous, non-uniform R-channel networks. Such R-channels were likely maintained for many years by powerful, unsteady flows delivered mainly from supraglacial sources during the melt season, and by the hydraulic damming effects of

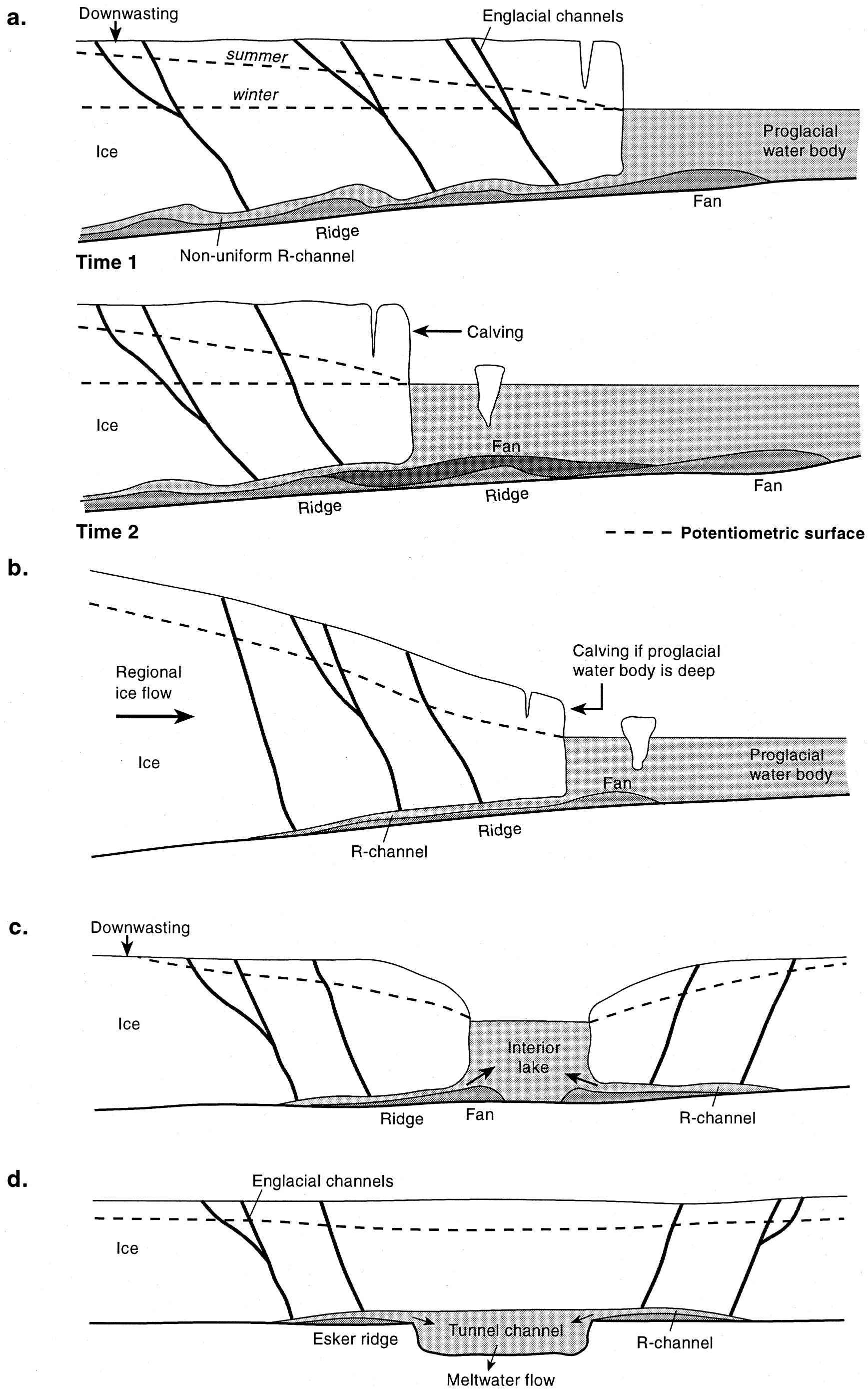

Fig. 7. Plausible conditions for the formation of eskers in R-channels that terminated in standing water. (a) Type I eskers (dendritic) formed in extensive, synchronous R-channels draining regionally stagnant ice, time 1 and 2. (b) Type II eskers (subparallel) formed in short R-channels draining actively retreating ice. (c) Type III eskers with fans formed in short R-channels draining to interior lakes in regionally stagnant ice. (d) Type III eskers without fans formed in short R-channels draining to tunnel channels in regionally stagnant ice. Note, (a) is illustrated at a much smaller scale than (b), (c) and (d). See text for further explanation.

large proglacial water bodies in the winter (Figs. 1b and 7a).

Was the Laurentide Ice Sheet regionally active or regionally stagnant during the operation of such extensive, synchronous R-channel systems? By ''regionally active'' I mean that ice flow was driven by a regional ice surface slope. By ''regionally stagnant'' I mean that the ice sheet had a flat ice surface and flow occurred only locally toward channels or was induced at the ice front by calving. Shreve (1972, 1985) argues for the synchronous operation of extensive, dendritic R-channel systems under sluggish ice. He defines ''sluggish'' as a condition where ice flowed forward everywhere driven by a gentle ice surface slope, yet flowed more slowly than present-day ice sheets (Shreve, 1985, p. 644). He suggests that R-channel path was primarily determined by ice surface slope and secondarily by bed topography. However, the underlying assumption of his analysis, that water pressure approximates ice confining pressure, is unlikely to be valid for conditions of unsteady discharge as inferred from type I eskers. In contrast, Röthlisberger (1972, p. 200) states that R-channels are likely to continuously change their position under active ice, a condition not conducive to preserving intricate dendritic esker systems that required R-channel stability for many years. Such preservation seems more likely if the R-channel systems developed under regionally stagnant ice (q.v. Mannerfelt, 1981). Under these conditions the dendritic pattern may have resulted from some combination of the pattern of sink points, water pressure gradients between large and small R-channels (Röthlisberger, 1972), and bed topography (Brennand, 1994).

The confining pressure of the ice and the input of water from supraglacial sources together determine the hydraulic gradient and the potentiometric surface within an ice sheet (Röthlisberger, 1972). Where supraglacial meltwater input is high and increasing, the potentiometric surface may become steeper than the ice surface slope, as the channel system lacks the capacity to efficiently drain surface water (Fig. 7a). Consequently, water could be routed through the R-channel system to the ice front even though the ice surface was flat (Röthlisberger, 1972, Fig. 5f). The tendency for wall melt to be balanced by creep closure in R-channels (Röthlisberger, 1972) causes local ice flow toward R-channels and explains both esker-convergent striae (Veillette, 1986) and observations of clasts derived from areas lateral to type I eskers (Brennand and Shaw, 1996); these observations do not necessitate regionally active ice. Lateral transport of sediment into an R-channel is likely to occur so long as ice exceeds a thickness of ~ 50 m (q.v. Nienow et al., 1998).

Is there independent evidence of regionally stagnant ice? Many type I eskers are located upflow from major arcuate moraines and converge on so-called interlobate moraines (Fig. 1b). Ice-frontal landforms such as end moraines are generally lacking upflow from arcuate moraines (Prest et al., 1968). This observation is consistent with regional ice stagnation (Shaw, 1996), but has also been attributed to rapid, active ice retreat (Veillette, 1986). Active ice retreat is difficult to rationalize with the preservation of extensive and complex esker systems (Shilts et al., 1987).

How could regionally stagnant ice have developed? What is the significance of major arcuate moraines? Why do eskers appear to radiate out toward them and converge on ''interlobate'' positions? Arcuate moraines have been inferred to record the terminal positions of ice streams or surging lobes (e.g., Dyke et al., 1989). This conclusion is mainly based on the radiation of streamlined forms and eskers out toward arcuate moraines and their convergence on ''interlobate'' positions. Some arcuate and interlobate moraines also enclose thick sequences of exotic till (e.g., Shilts et al., 1987). An assumption is that streamlined forms were formed during surging by a subglacial deformation process (Boulton and Hindmarsh, 1987). However, (1) the alignment of streamlined forms is often consistent upflow and downflow of arcuate moraines (Prest et al., 1968). (2) Streamlined forms converge on but do not cross ''interlobate'' positions (Brennand et al., 1996). (3) Some interlobate moraines are eskers (e.g., Ringrose and Large, 1977; Kaszycki and Dilabio, 1986; Brennand and Shaw, 1996). (4) Eskers are mainly a passive record of ice dynamics and morphology. It seems more likely that esker convergence on ''interlobate'' positions and arcuate moraine geometry are a consequence of events that formed the regional streamlined fields, rather than coeval with them (Brennand and Shaw, 1996). Synchronous formation of streamlined forms by catastrophic meltwater floods is consistent with current observations for these areas (Shaw, 1996). Vigorous basal melt and ice sheet extension during these ice-bed decoupling events would likely have resulted in an intensely crevassed, thin, flat ice sheet (Shoemaker, 1992a,b) with local saddles along axes of flood path convergence (Brennand and Shaw, 1996). Esker convergence may record such saddles. Thinner ice at these saddles and the thermomechanical erosion of later meltwater funneling below them may have resulted in reentrants in the ice front and an overall lobate appearance to it (Shaw, 1996). Large eskers may have formed beneath these saddles (Brennand and Shaw, 1996); such eskers have been interpreted as interlobate moraine (e.g., Shilts et al., 1987).

Laurentide arcuate moraines are mainly composed of subaqueous outwash sediments, occasionally with proximal till deposits (Dubois and Dionne, 1985); they mainly record grounding-line positions (Hillaire-Marcel et al., 1981; Sharpe and Cowan, 1990). These grounding-line deposits may record a re-equilibration of the potentiometric surface within the ice sheet, precipitated by a drop in proglacial water level (Hillaire-Marcel et al., 1981; Sharpe and Cowan, 1990). However, the similarity of the configuration and occurrence of Laurentide arcuate moraines with those of the Eurasian ice sheet (Punkari, 1997) is striking. The Salpausselkä moraines mainly record grounding lines (Fyfe, 1990) of Younger Dryas age (Rainio et al., 1995). Younger Dryas cooling was likely a global event (Peteet, 1995). The Hartman Moraine is of late Younger Dryas age (~10,400 B.P.; Sharpe et al., 1992; Fig. 1b). Should many of the Laurentide arcuate moraines (excluding the Sakami moraine; Hillaire-Marcel et al., 1981) be of similar age, they may record Younger Dryas grounding-line re-equilibration positions. Following the Younger Dryas, continued climatic warming coupled with thin, flat ice (Shaw, 1996) may have effected the regional ice stagnation which seems most likely for type 1 esker systems (zone 2 in Fig. 1a).

Plausible conditions for type 1 esker formation are presented in Fig. 7a. A climatically forced regional melt zone caused ice downwasting and produced high volumes of supraglacial meltwater. On flat, stagnant ice this melt zone may have extended well toward the regions designated as ice divides (zone 1, Fig. 1a) and resulted in the formation of numerous supraglacial lakes so long as near-surface ice remained cold (Hodgkins, 1997). Heat from supraglacial meltwater and the density differences between water and ice caused moulins and crevasses to propagate downward, forming englacial channels and finally connecting those channels to the bed (Weertman, 1973, 1974). Extensive crevassing may have been inherited from earlier meltwater flood events (Shoemaker, 1992a,b). At the bed, meltwater may have connected to a distributed system early in the melt season. As discharge increased over the melt season, an integrated R-channel system likely developed by cavity coalescence (Kamb, 1987) and downglacier channel extension (Gordon et al., 1998; Nienow et al., 1998). These R-channels likely grew headward as the melt zone expanded up-ice over the melt season (Gordon et al., 1998). Summer flow in this integrated drainage system was driven by a steep hydraulic gradient (potentiometric surface, Fig. 7a). Episodic drainage of supraglacial lakes through these systems may have been responsible for some gravel–sand couplets in type I eskers (Shaw, 1994) and for anabranched reaches, subaqueous fans at bends and hummocky, extended zones (Brennand, 1994). The proglacial water body kept R-channels open in the winter at least to the level of the potentiometric surface defined by the water body (Fig. 7a), and permitted multi-year operation of the R-channels. If the proglacial water body was deep, back-pressure effects (Shoemaker, 1992b) on the ice mass may have caused ice to float and a grounding line to develop (Brennand and Shaw, 1996). Such conditions favour both the formation of subaqueous fans superimposed on esker ridges as the grounding line vacillated (Brennand and Shaw, 1996), and calving at the ice front (Fig. 1a; Veillette, 1986). Calving may have caused some reactivation of ice close to the ice margin.

5.2. Short eskers formed in R-channels or reentrants that terminated in standing water (types II and III)

5.2.1. General depositional environment

These eskers are tens of meters to tens of kilometers long, and exhibit subparallel (type II eskers) and deranged (type III eskers) regional patterns (Fig. 3b,c). They are composed of (i) a short ridge (southern Alberta, J. Shaw, 1998, personal communication), (ii) a short ridge with a subaqueous fan or a fan complex extending from, or superimposed over, the ridge (southern Ontario and Quebec; Banerjee and McDonald, 1975; Sharpe, 1988; Gorrell and Shaw, 1991; Figs. 8 and 9), or (iii) alternating ridges and fans (Banerjee and McDonald, 1975). The discussion below draws mostly on research from subparallel or

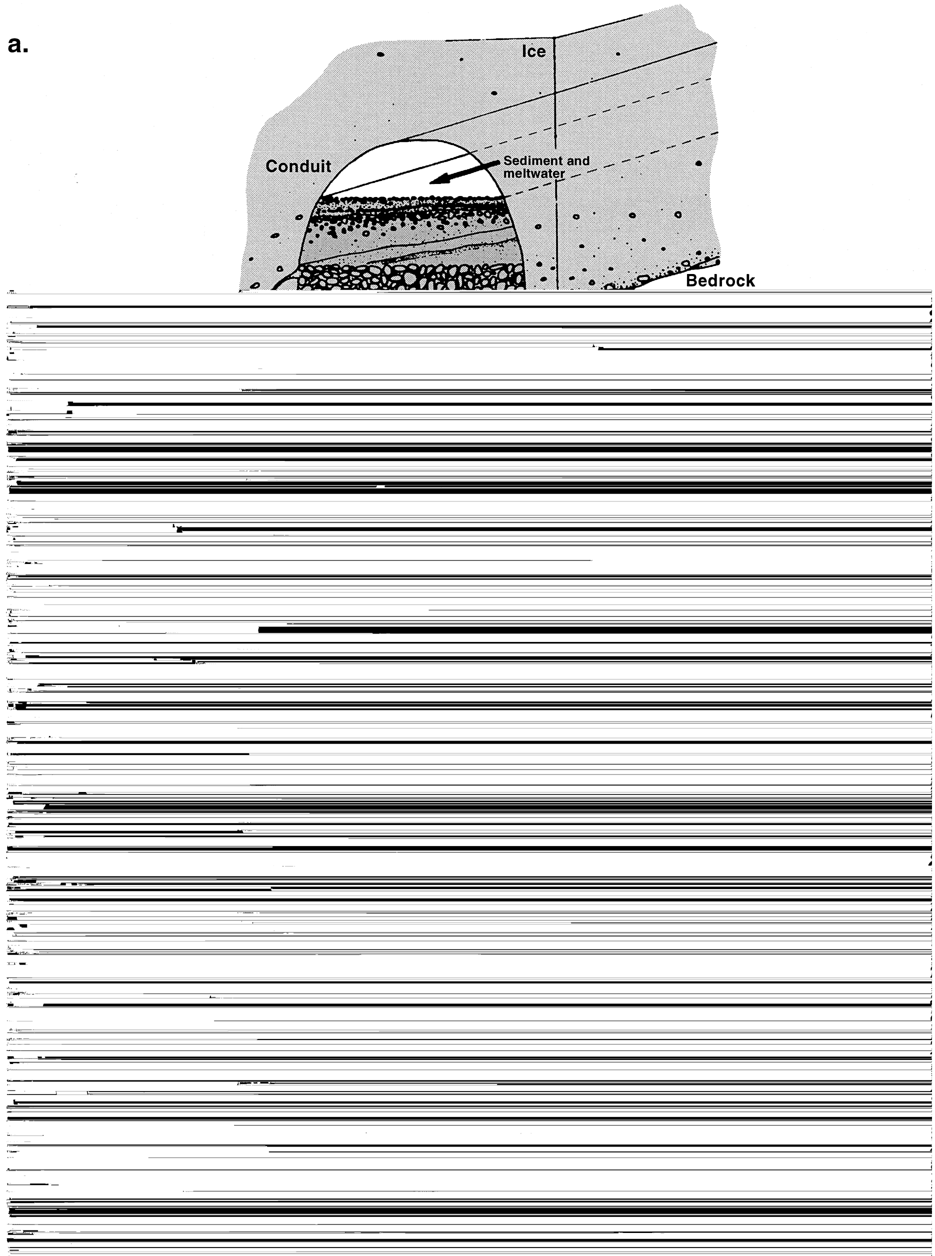

Fig. 8. Formation of a type II esker-fan sequence. (a) Formation of an esker ridge in an R-channel. (b) Ice retreat, ridge collapse and superimposition of subaqueous fan sediments. (c) Glaciomarine sedimentation and littoral reworking. Diagrams modified from Sharpe (1988; original drawings by R. Gilbert) for the Twin Elm esker, deposited in an R-channel that terminated in the Champlain Sea.

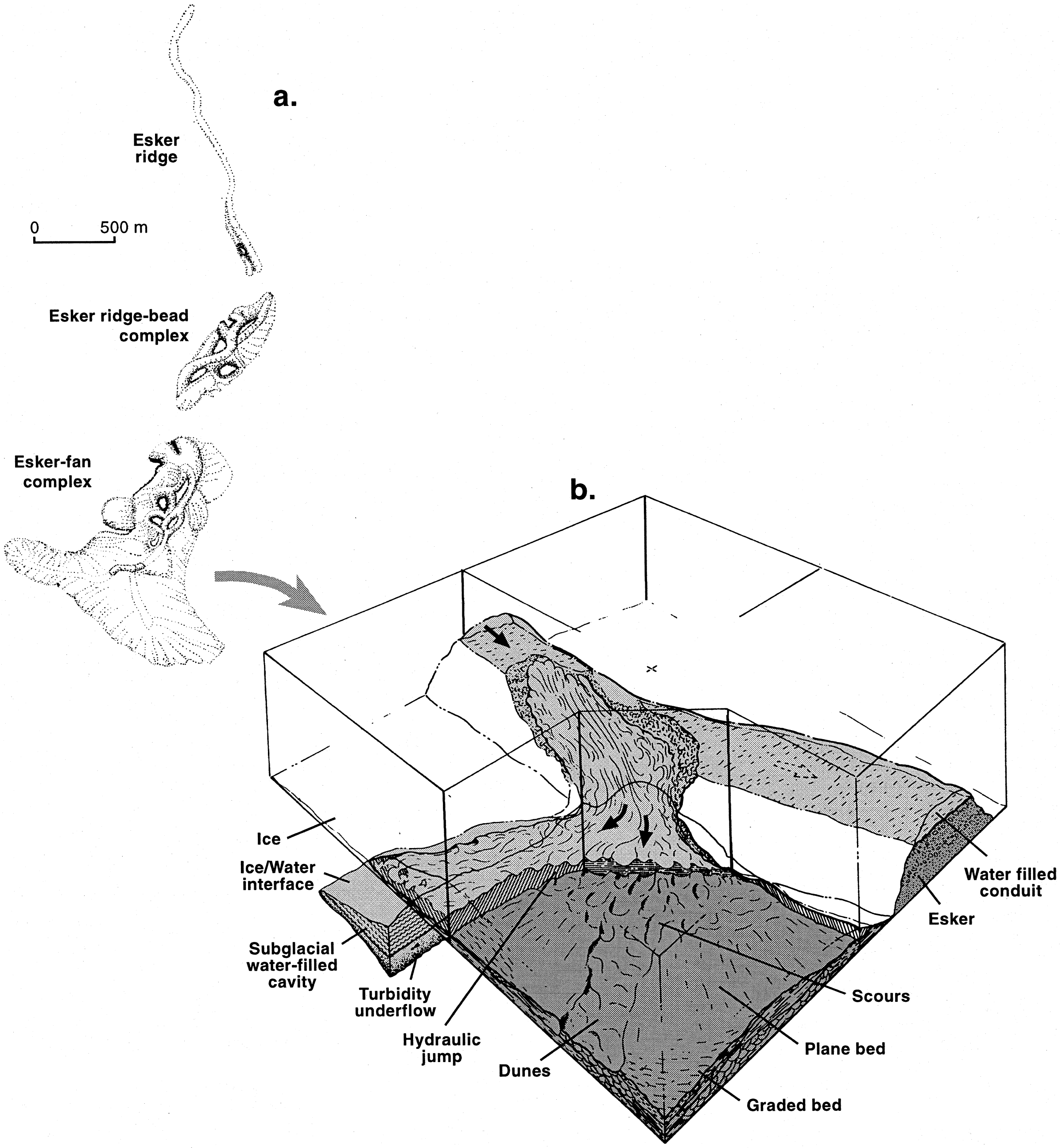

Fig. 9. (a) Type II esker–bead–fan complex, Lanark, Ontario. (b) At a grounding line close to flotation, small increases in subglacial water pressure result in hydraulic lifting and the formation of lateral subglacial fans (modified from Gorrell and Shaw, 1991). Reprinted with permission from Elsevier Science.

isolated eskers in southern Ontario and Quebec (Banerjee and McDonald, 1975; Saunderson, 1977; Cheel, 1982; Cheel and Rust, 1982, 1986; Henderson, 1988; Rust, 1988; Rust and Romanelli, 1975; Sharpe, 1988; Gorrell and Shaw, 1991; Spooner and Dalrymple, 1993). Deranged eskers are most clearly observed in southern Alberta (Shetsen, 1987), but detailed morpho-sedimentary observations are lacking.

Short esker ridges differ in sedimentary detail and thus depositional environment. Some are dominated by gravel facies, exhibit low paleoflow variability, lateral faulting, proximal folding, and minor draped or intercalated diamicton beds (Fig. 8; McDonald and Shilts, 1975; Henderson, 1988; Sharpe, 1988; Gorrell and Shaw, 1991). Diapirs and upslope paleoflows in eskers draped over hummocky terrain in the Prairies suggest that they were subglacial forms (Stalker, 1959, 1960; Munro and Shaw, 1997). Short esker ridges that terminate in subaqueous fans where deposited in water-filled R-channels (Banerjee and McDonald, 1975; Henderson, 1988; Gorrell and Shaw, 1991). It has been suggested that water-filled R-channels may also be inferred from the presence of a sliding bed gravel facies (Saunderson, 1977; Ringrose, 1982), whereas open channel flow (flow at atmospheric pressure) was assured by the presence of antidunes (e.g., Banerjee and McDonald, 1975). The logic linking sedimentary descriptions of the sliding-bed facies to its hydraulic interpretation is incomplete (q.v. Saunderson, 1977; Brennand, 1994). Furthermore, antidunes may form at a range of density interfaces (e.g., Hand, 1974; Table 1). Consequently, sedimentary facies are not good diagnostic indicators of depositional environment, yet they must be consistent with it. Some short esker ridges are composed of faulted, sandy subaqueous fan sediments or offlapping subaqueous fans (Rust and Romanelli, 1975; Cheel, 1982). These eskers were deposited in laterally constrained environments, most likely reentrants into the ice front (Saunderson, 1975; Cheel, 1982; Cheel and Rust, 1982).

Subaqueous fans record ice-frontal or grounding-line sedimentation (e.g., Gorrell and Shaw, 1991; Fig. 9). Flow decelerated as channelized meltwater exited the portal and entered a standing water body. This is confirmed by downflow fining, increased paleoflow variability, and changes in sedimentary structures generally consistent with weakening flow. The upward-fining of subaqueous fan sediments indicates decreased flow intensity due to either local ice retreat (e.g., Sharpe, 1988; Sharpe et al., 1992) or a change in portal location. Prolonged ice front stability allowed some fans to build to the water surface and form deltas (Martini, 1990). The channels supplying meltwater and sediment to the delta thus became subaerial and may have taken the form of H-channels (Hooke, 1984) close to the margin. Some esker sediments were subsequently modified by wave action as the level of the proglacial water body fell (Fig. 8; Matile, 1984; Veillette, 1986; Sharpe, 1988).

Occasionally, the ridge-subaqueous fan sequence appears to repeat upflow (Banerjee and McDonald, 1975; Ringrose, 1982; Henderson, 1988; Spooner and Dalrymple, 1993). In this case, either the esker formed (i) in time-transgressive segments in R-channels close to the ice front as the ice front retreated (Banerjee and McDonald, 1975; Fig. 4), or (ii) unsystematically in R-channels at a fluctuating grounding line where the ice front was close to flotation (Gorrell and Shaw, 1991; Fig. 9).

5.2.2. Hydraulic processes

Esker ridges (cores) are mainly composed of massive to cross-bedded gravel (pebble–boulder) with low paleoflow variability (Saunderson, 1977; Henderson, 1988). Clasts may exhibit *b*- or *a*-axis imbrication, suggestive of powerful traction or transport as a hyperconcentrated dispersion, respectively (Rust, 1988; Rust and Romanelli, 1975; Gorrell and Shaw, 1991). For the Guelph esker, southern Ontario (Saunderson, 1977) and the Windsor esker, Quebec (Banerjee and McDonald, 1975) powerful flow is further suggested by the likely presence of macroforms (q.v. Brennand, 1994). Lower meltwater discharge in winter may be indicated by the presence of diamicton beds and lenses (Gorrell and Shaw, 1991).

Most of the sedimentary facies and facies associations described from short eskers formed in R-channels that terminated in standing water actually record grounding-line or subaqueous fan sedimentation (Rust and Romanelli, 1975; Powell, 1990). This may be superimposed over, or extend from, the gravelly esker core (Sharpe, 1988; Gorrell and Shaw, 1991). In general, the sediments in subaqueous fans fine, and exhibit lower-flow-regime structures and higher paleoflow variability, radially outward from the fan apex (Cheel, 1982; Gorrell and Shaw, 1991; Spooner and Dalrymple, 1993). These characteristics are due to flow expansion and deceleration away from the R-channel portal; traction transport is replaced by deposition from suspension downflow (Brennand and Sharpe, 1993), and ultimately glaciolacustrine or glaciomarine rhythmites replace and onlap subaqueous fan deposits (Banerjee and McDonald, 1975; Henderson, 1988). High rates of deposition from suspension are inferred from thick sand beds and dewatering structures (Cheel and Rust, 1982, 1986; Brennand and Sharpe, 1993). Despite these generalities, the sedimentary character of any one fan may be determined by: (i) the relative density of the effluent jet (sediment load, temperature) and ambient water body (sediment load, temperature, salinity), and (ii) unsteady meltwater discharge and sediment transport (discussed below) (Powell, 1990). Consequently, subaqueous fan sediments generally show no strong vertical textural or structural arrangement (Gorrell and Shaw, 1991).

In glaciolacustrine or brackish environments the high sediment load carried by the effluent jet renders it relatively more dense than the ambient water and consequently underflows are common (Figs. 8 and 9; Sharpe, 1988; Gorrell and Shaw, 1991; Spooner and Dalrymple, 1993). Proximal imbricate gravel is transitional downflow to trough cross-bedded gravel (pebbles) then massive to cross-laminated sand often with dropstones (interchannel sands of Rust, 1988). Bouma sequences record turbidity flows (Brennand and Sharpe, 1993). Mid-fan sands may exhibit trough-shaped scours filled with diffusely graded sand (channel sands of Rust, 1988). These structures may record channels filled with sediment flows generated by slump failures (Rust, 1977, 1988) or local scour and deposition beneath supercritical submerged jets at hydraulic jumps (Fig. 9; Saunderson, 1977; Gorrell and Shaw, 1991). Fan facies may pass laterally into glaciolacustrine rhythmites (Henderson, 1988). Many authors provide extensive descriptions of the subaqueous fan facies and facies associations of short eskers, for example Cheel (1982), Cheel and Rust (1982, 1986), Henderson (1988) and Gorrell and Shaw (1991).

In modern glaciomarine environments the effluent jet may be less dense than the ambient sea water, and buoyant forces create turbid plumes (Powell, 1990). Gravel is generally deposited very close to the portal, may build to form a barchanoid dune and may be subject to mass movement due to oversteepening (Powell, 1990). Sand is carried by rising turbid plumes to the sea surface, thence laterally as overflow away from the glacier front. It settles back through the water column, generally forming massive beds when plume velocities fall below that critical to counteract grain weight. Suspension settling of silt and clay (generally flocculated) is controlled by the dynamic interaction between the oceanward plume movement and migrating tidal wedge over the tidal cycle; rhythmically laminated cyclosams and cyclopels are deposited (Powell, 1990). In Laurentide glaciomarine environments around R-channel portals, the effluent jet was likely more dense than the ambient seawater (sediment concentration greater than 35 g l^{-1}) and turbid underflows likely carried coarser sediment further from the portal than in modern glaciomarine environments (Gilbert, 1983). R-channels terminating in saline water have been proposed from the presence of *Hiatella arctica* (Sharpe, 1988), although this was likely a later colonizer (R. Gilbert, 1998, personal communication). The flocculating effect of saline water may have been responsible for thick clay beds in fine-grained esker ridges and fans on Victoria Island (Brennand and Sharpe, 1993).

5.2.3. Hydrologic regime

Vertical variations in sedimentary structures and textures in short esker ridges and subaqueous fans imply fluctuations in meltwater discharge and sediment load over time. Short esker ridges may contain several gravel–sand couplets (Banerjee and McDonald, 1975; Gorrell and Shaw, 1991). Fans are mainly composed of stacked packages of fining-upward sand–silt–clay (Henderson, 1988; Gorrell and Shaw, 1991; Brennand and Sharpe, 1993). Seasonal fluctuations in meltwater discharge and sediment load are inferred from clay caps and gravel–sand couplets, and diurnal and weather-related events are invoked for rhythmic textural and structural variations between clay caps in fans (Henderson, 1988; Gorrell and Shaw, 1991; Brennand and Sharpe, 1993; Spooner and Dalrymple, 1993). Such unsteadiness in meltwater discharge and thus sediment transport and deposition requires a supraglacial meltwater input to the subglacial drainage system. This input is confirmed where esker gravels are traceable downflow into rhythmites (Fig. 8; Banerjee and McDonald, 1975; Sharpe, 1988).

Unsteadiness in meltwater discharge is coupled to variability in subglacial water pressure and results in sedimentary complexity at grounding lines where the ice is close to flotation. This situation is well documented by Gorrell and Shaw (1991) for an esker–bead–fan complex in southern Ontario (Fig. 9). High discharge and water pressure events locally breached the high pressure seals at the margin of an R-channel; this occurred preferentially at bends where water pressure was highest (Streeter and Wylie, 1979). Lateral cavities, perhaps part of a distributed drainage system, became temporarily connected to the R-channel by small distributary and contributory channels (Fig. 9). Flow escaping from the R-channel expanded in the cavities, resulting in subglacial fan sediments (beads): pronounced downflow fining and structural change are consistent with rapid flow velocity reduction and loss of competence (Fig. 9). Ice-bed recoupling when discharge and water pressure declined resulted in faulting and overfolding in bead sediments, confirming a subglacial depositional environment. Distally, a series of overlapping fans emanate from anastomosing minor ridges lateral to the main esker ridge (Fig. 9). The fans display complex sediment sequences suggestive of highly variable deposition in time and space at a grounding line close to flotation. The fans are considered to be subglacial as they show limited development of collapse structures, occupy a position lateral to the main ridge, and exhibit thrust faulting in their upper beds. A similar hydraulic-lift mechanism has been invoked to explain extended deposits and fans lateral to a longer esker ridge in the low Arctic (Brennand and Sharpe, 1993). This dynamic valve mechanism at a grounding line may also be applicable to other beaded eskers thus far explained by the segmental, time-transgressive model (e.g., Banerjee and McDonald, 1975; Henderson, 1988; Fig. 4); they exhibit similar sedimentary characteristics such as pronounced downflow fining, folds, faults and high paleoflow variability. Where R-channels terminated in a marine environment, tidal forcing may have had similar effects.

5.2.4. Deglacial drainage and glaciodynamics associated with type II and III eskers

Type II and III eskers formed in subaqueously terminating R-channels or reentrants close to the ice front or a grounding line. Sedimentation was by unsteady flows delivered mainly from supraglacial sources. Why are these eskers short? Short eskers may form when (i) there is an insufficient supply of supraglacial meltwater to maintain long R-channels (Walder, 1982), (ii) the supraglacial catchment area of sink holes is constrained by the ice surface topography, (iii) sink holes (crevasses and moulins), which allow supraglacial water access to the glacier bed, are limited to positions close to the ice front (q.v., Nienow et al., 1998), (iv) cold, non-crevassed ice behind a narrow ice marginal zone prevents supraglacial meltwater from reaching the bed, (v) the ice is active and tends to cause hydrologic switching (Röthlisberger, 1972), or (vi) there is limited sediment supply for esker building. Different combinations of these factors may be responsible for the development of short, subparallel (type II) and short, deranged (type II) esker patterns.

Type II eskers are often few in number but regionally aligned (Fig. 3b; Rust, 1988). Such eskers generally follow level or downslope paths and valleys (Gorrell and Shaw, 1991; Spooner and Dalrymple, 1993). They are common in the Ottawa–Kingston area of southern Ontario (C in Fig. 1a; Rust and Romanelli, 1975; Cheel, 1982; Cheel and Rust, 1982; Henderson, 1988; Rust, 1988; Sharpe, 1988; Gorrell and Shaw, 1991; Spooner and Dalrymple, 1993).

The regional alignment of type II eskers and their preferential location in valleys suggests that the R-channel path was controlled by a regional ice surface slope (Shreve, 1972) or an upglacier-rising potentiometric surface (Röthlisberger, 1972) and bed topography (Shreve, 1972). Alternating esker ridge–subaqueous fan sequences suggest systematic, active ice retreat (Banerjee and McDonald, 1975), driven by a regional ice surface slope and perhaps facilitated by calving at the ice front. The sparsity of end moraines associated with type II eskers may suggest that retreat was very fast (q.v. Veillette, 1986). Active ice retreat necessitates a negative mass balance, with the dominance of melting over accumulation. Such conditions favour a regional melt zone, yet eskers are short. Under such conditions esker length and number may have been inhibited by (i) the location of sink holes only close to the ice front, a condition perhaps resulting from extensional flow behind a calving margin, (ii) the tendency for R-channels to close and relocate under active ice (Röthlisberger, 1972; Shilts et al., 1987), (iii) the presence of cold ice upflow from a narrow marginal zone, or (iv) sediment supply. On balance, type II eskers were most likely formed in short R-channels (or reentrants) under active ice exhibiting a regional ice surface slope. Some were likely formed in time-transgressive segments as the ice front or grounding line retreated during deglaciation. Plausible conditions for the formation of type II eskers are presented in Fig. 7b.

Type III eskers lack regional alignment (Fig. 3c), may be few in number and may follow paths irrespective of bed topography (Stalker, 1959, 1960; Munro and Shaw, 1997). They are common in southern Alberta (Shetsen, 1987; D in Fig. 1a). Here subaqueous fans may or may not be present (Stalker, 1960; J. Shaw, 1998, personal communication).

The lack of a regional alignment in type III eskers argues against R-channel path being determined by a regional ice surface slope on active ice (Shreve, 1985). It seems more likely that R-channel path was determined by local potentiometric surfaces or hydraulic head in stagnant ice (Fig. 7c,d). Stagnant ice is also consistent with a lack of associated end moraines on the Prairies (Shetsen, 1987, 1990), and the meltwater model of sub-ice landscape evolution proposed for that region (Rains et al., 1993). Two plausible conditions for the formation of type III eskers are presented in Fig. 7; Fig. 7c shows conditions that may have formed short eskers with fans, and Fig. 7d shows conditions which may have formed short eskers without fans. Interior lakes may have developed within regionally stagnant ice in an extensive melt zone during deglaciation (Fig. 7c). The catchment area for these lakes would have been determined by ice surface topography. The paths of R-channels draining to such lakes would have been determined by local ice surface slopes and potentiometric surfaces. Such R-channels are likely to form a deranged pattern when viewed at a regional scale. Subaqueous fans formed downflow of R-channel portals in interior lakes. Esker length was most likely limited by the supraglacial catchment area for interior lakes and sediment supply. Debris-poor ice on the Prairies is consistent with the pristine preservation of subglacial landforms there (Rains et al., 1993). Local ice activity may have occurred in the vicinity of interior lakes. In contrast, type III eskers lacking subaqueous fans may have formed in R-channels connected to broader tunnel channels in regionally stagnant ice (Fig. 7d). Higher subglacial water pressures in R-channels than in tunnel channels would have cause R-channels to become connected to tunnel channels. Eskers formed in such R-channels may lack subaqueous fans as sediment debouched from R-channels may have been transported away by meltwater flowing down tunnel channels (J. Shaw, 1998, personal communication).

6. Eskers deposited in R-channels that terminated subaerially

Eskers that end in subaerial outwash or deltas (Shilts, 1984; Aylsworth and Shilts, 1989; Thorleifson and Krystjanssen, 1993) or those that are located in regions which exhibit little or no independent evidence of the presence of standing water (e.g., no rhythmic glacial lake sediments or raised beaches) were most likely deposited in R-channels that terminated subaerially. Associations between eskers and subaerial outwash have been reported for modern eskers (Gustavson and Boothroyd, 1987; Syverson et al., 1994). Subaerially terminating R-channels have been inferred for (i) some long, dendritic esker systems in the Mackenzie–Keewatin area (Shilts, 1984; Shilts et al., 1987; St-Onge, 1984) and the Beardmore–Geraldton area (Thorleifson and Krystjanssen, 1993) (zone 2 in Fig. 1a), and (ii) several short, isolated eskers in southern Ontario (Banerjee and McDonald, 1975; Saunderson, 1975; zone 3 in Fig. 1a). In general though, subaerially terminating R-channel systems have rarely been inferred from Laurentide eskers. This situation may reflect (i) that such systems rarely existed, (ii) that eskers were rarely produced in such systems, (iii) that eskers produced in such systems are difficult to identify (Banerjee and McDonald, 1975), or (iv) that these systems existed in regions where research is generally lacking.

6.1. Long, dendritic eskers formed in R-channels that terminated subaerially (type IV)

6.1.1. General depositional environment

Type IV esker systems are tens to hundreds of kilometers long and may be unbroken for up to 75 km (Shilts et al., 1987) (Fig. 3a). Were such long eskers deposited in extensive, synchronous R-channel systems or were they deposited as time-transgressive segments in less extensive R-channels or reentrants that developed close to the ice margin as the ice front retreated? Some type IV eskers are flanked by a series of small, sequential end moraines (Gadd, 1973); others exhibit repeated downflow transitions to subaerial outwash (Thorleifson and Krystjanssen, 1993). Such transitions have been defined on morphologic grounds (aerial photographs) but have yet to be confirmed by sedimentologic research. However, these observations make it likely that some type IV eskers were deposited as time-transgressive segments in short R-channels or reentrants that developed close to the ice margin as the ice front retreated or downwasted and backwasted. Such R-channels may have been as short as 750 m based on esker-subaerial outwash spacing (Thorleifson and Krystjanssen, 1993).

Other observations and arguments have also been presented in support of this depositional environment, yet are perhaps more equivocal. (1) Esker ridges alternate with channels eroded into the substrate and anastomosing reaches inferred to record ice-frontal positions (Shilts, 1984). However, channels may be expected to alternate with eskers that were deposited in long R-channels (Shreve, 1985) and anabranched reaches need not record ice-frontal positions (Brennand, 1994). (2) Eskers exhibit downflow repetition of boulders and coarse sand along their crests (St-Onge, 1984). However, crest-line textures may not reflect grain sizes throughout an esker ridge and, even if they did, textural patterns in a continuous ridge may record flow dynamics along synchronous, non-uniform R-channels (Brennand, 1994). (3) Provenance studies suggest relatively short transport distances for esker clasts (Bolduc et al., 1987). Although such observations may favour short R-channels, they may also result from sedimentation zones along long R-channels (q.v. Brennand, 1994). (4) Eskers are flanked by time-transgressive terraces formed in supraglacial, subaerial outwash that was initially superimposed on eskers (Shilts, 1984; Fig. 10). The fact that subaerial outwash was superimposed on eskers suggests that it was deposited and terraced after esker formation as the ice front retreated or backwasted (Fig. 10); it does not inform as to the conditions of esker formation. However, Gustavson and Boothroyd (1987, Fig. 3) describe a situation at the margin of Malaspina Glacier where subaerial outwash deposits are superimposed on an esker, yet were supplied by sediment from the same R-channel that formed the esker. (5) On the grounds that thick ice near ice divides would tend to close R-channels and thin ice would have resulted in greater topographic control on esker path (Shreve, 1985), Shilts (1984) argues against formation in extensive, synchronous R-channel systems for type IV eskers in the Mackenzie–Keewatin area. However, R-channels are only likely to have closed completely if they drained of water, and the relation between eskers and bed topography has yet to be fully documented in this area. These uncertainties leave open the possibility that some eskers in the Mackenzie–Keewatin area may have developed in extensive, synchronous R-channel systems prior to ice retreat or backwasting. However, if such R-channels were to remain water-filled for many years, then they must (i) terminate in standing water (Powell, 1990), (ii) become plugged with sediment or collapse downflow (Boulton and Hindmarsh, 1987; Hubbard and Nienow, 1997) thus trapping water over the winter, or (iii) experience a discharge sufficient to maintain them over the winter (Willis et al., 1993).

6.1.2. Hydraulic processes

The presence of rounded boulders on the surface of some eskers indicates powerful flows (St-Onge, 1984). Subaerial outwash exhibits the traces of braided channels at its surface (Shilts et al., 1987). As additional sedimentological observations are lacking, investigation of the hydraulic processes associated with type IV eskers awaits further research.

6.1.3. Hydrologic regime

Supraglacial meltwater input to R-channels has been proposed to account for the dendritic pattern of type IV eskers in the Mackenzie–Keewatin area (Shilts, 1984). This pattern is attributed to the imprint of a supraglacial drainage system; the headward growth of short R-channels followed the traces of a connected dendritic supraglacial channel system (Shilts, 1984). The development of this supraglacial system required a regional melt zone, cold near-surface ice (q.v. Gustavson and Boothroyd, 1987)

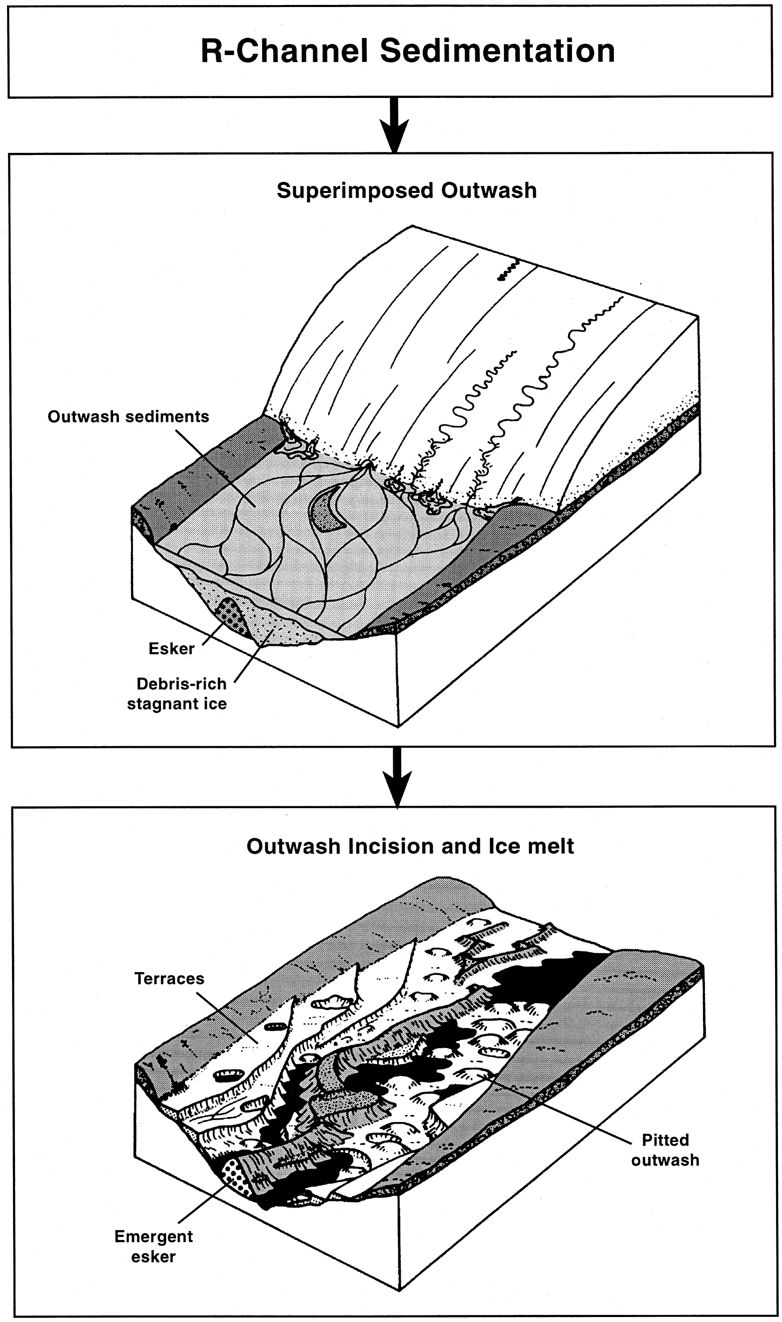

Fig. 10. Formation of type IV eskers as proposed by Shilts et al. (1987) (adapted from Menzies and Shilts, 1996; original drawings by J.A. Aylsworth). The nature of R-channel sedimentation is unknown. Outwash sediments were laid down over stagnant ice containing the esker as the ice backwasted and downwasted. Meltout of buried ice and outwash incision formed pitted and terraced outwash; a partially eroded esker emerged. Reprinted with permission from Butterworth Heinemann Publishers, a division of Reed Educational and Professional Publishing Ltd.

and sink points preferentially located at the ice front. Invoking the Coriolis effect, Aylsworth and Shilts (1989) have argued that the fact that many esker tributaries join trunk eskers from the left, confirms the operation of an extensive, dendritic supraglacial channel system on nearly flat ice. St-Onge (1984) proposes supraglacial meltwater input on the basis of sediment texture along an esker crest. He suggests that surface boulders record points of supraglacial to subglacial connection, sediments fining toward the proposed ice margin; the origin of the boulders is not addressed. Thorleifson and Krystjanssen (1993) describe long glaciofluvial corridors containing discontinuous eskers in the Beardmore–Geraldton area, Ontario. They infer both subaerially and subaqueously terminating R-channels controlled by bed topography. As esker ridges are transitional to subaqueous fans and varved lacustrine sediment (Thorleifson and Krystjanssen, 1993), supraglacial meltwater input was likely for eskers throughout the region.

6.1.4. Deglacial drainage and glaciodynamics of type IV eskers

Although the above discussion highlights many uncertainties, some type IV eskers likely formed as segments, time-transgressively in R-channels close to the ice margin as the ice front retreated or downwasted and backwasted. Such R-channels were likely supplied by water from supraglacial sources. Was the ice sheet regionally active or stagnant during the operation of these R-channels? St-Onge (1984) states that eskers show no sign of overriding. Shilts et al., (1987) highlight the excellent preservation of extensive, complex esker systems. Both argue that these observations suggest that the ice was largely stagnant. The segmental nature of the eskers was a product of downwasting and backwasting of this ice mass. Thorleifson and Krystjanssen (1993) argue, based on the spacing of esker ridges and outwash deposits and on the presence of kettle holes adjacent to eskers, that the ice underwent stagnation-zone retreat (actively retreating ice with a stagnant fringe within which eskers developed, e.g., Hebrand and Åmark, 1989). A model of plausible conditions for the formation of type IV eskers is not presented due to lack of consensus and evidence on which to ground it.

6.2. Short eskers formed in R-channels or reentrants that terminated subaerially (type V)

Type V eskers are tens of meters to tens of kilometers long. The termination of short eskers in deltas (Banerjee and McDonald, 1975; Saunderson, 1975) indicates that at least the latter stage of channel flow was at atmospheric pressure and may suggest that some esker sedimentation occurred in H-channels (Hooke, 1984). The facies associations and arguments presented by Banerjee and McDonald (1975) and Saunderson (1975) in support of deltaic sedimentation at the end of a subaerially terminating H-channel include (i) downflow textural and structural trends consistent with weakening flow, (ii) thick, gravel cross-beds interpreted as topset braid bars (Saunderson, 1975), or as delta or bar front deposits (Banerjee and McDonald, 1975), (iii) lateral faults consistent with removal of lateral ice support, and (iv) backset beds or antidunes interpreted as evidence of subaerial sedimentation (Banerjee and McDonald, 1975). This evidence is equivocal as similar observations are made in subaqueous fans deposited in R-channels or reentrants close to the ice front (Rust and Romanelli, 1975; Gorrell and Shaw, 1991). Yet short eskers terminating in deltas are observed (Martini, 1990). It is also plausible that short eskers terminating in subaerial outwash may be expected from subaerially terminating R-channels, but such eskers have not been reported, perhaps because they are difficult to recognize (Banerjee and McDonald, 1975). Due to lack of evidence on which to ground it, a model of possible conditions for the formation of type V eskers is not presented.

7. Controls on the spatial distribution of Laurentide esker types

Five types of eskers have been identified and discussed in terms of their likely implications for deglacial drainage and glaciodynamics. How are these esker types spatially distributed? The answer to this question is by no means certain. Referring to the zones in Fig. 1a: zone 2 is likely dominated by esker types I and IV, and zone 3 mostly includes esker types II and III, although esker type I occurs in tunnel channels. The distribution of esker type V is unknown, and few observations are reported for zone 4 in Canada (Rampton et al., 1984).

What determined this spatial distribution? Clark and Walder (1994) have argued that esker distribution is substrate-controlled: eskers being present on rigid beds (zone 2 in Fig. 1a) and largely absent on soft beds (zone 3 in Fig. 1a). This argument is based on Walder and Fowler's (1994) theory which predicts that R-channels are replaced by canals on pervasively deforming beds. While elegant, this theory may be mute for explaining Laurentide esker distribution. Eskers are present in regions designated as soft bed and assumed to be pervasively deforming during ice retreat by Clark and Walder (1994). This contradiction to theoretical prediction may suggest that (i) R-channels were present over deforming beds (q.v. Alley, 1992), or (ii) pervasive bed deformation never occurred, or had ceased, prior to esker formation. In south-central Ontario, the presence of drumlins on a regional till sheet has been interpreted as evidence of pervasive bed deformation (Boyce and Eyles, 1991). Eskers were deposited in tunnel channels incised into this landscape. However, the regional till sheet (Newmarket Till, Sharpe et al., 1996) contains horizontal sorted beds which argue against pervasive bed deformation, and the eskers where formed after till deposition and drumlin formation. Consequently, these eskers likely formed in R-channels on a deformable bed that was not pervasively deforming during esker formation, if indeed it ever had. Consequently, although R-channels may only be stable on hard beds, deformable beds need not deform. Perhaps large volumes of supraglacial meltwater delivered at point sources into the subglacial environment forced the development of an efficient R-channel system on the deformable bed; the effects of point-source water delivery to the bed is not considered by Walder and Fowler (1994). Basal substrate certainly has a role in esker distribution, at least as a sediment source (Shilts et al., 1987), but it is unlikely to be the sole determinant.

Shilts et al. (1987) argue that sediment availability and stagnant ice upflow from major arcuate moraine in zone 2 (Fig. 1a) was responsible for extensive, dendritic esker systems, whereas active ice retreat was largely responsible for the sparsity of eskers in zone 3 (Fig. 1a). The discussion presented here tends to confirm ice stagnation in zone 2 on the basis of type I eskers, but is equivocal on the basis of type IV eskers. It seems probable that the extensive esker systems in zone 2 were a response to the late deglacial coupling of thin, mostly stagnant ice with rapid climate warming, possibly following the Younger Dryas. Thin ice was likely inherited from earlier catastrophic events such as surges and subglacial sheet floods (Shoemaker, 1992a; Shaw, 1994). Zone 3 (Fig. 1a) mainly contains short eskers which were likely formed under actively retreating ice in parts of southern Ontario (C in Fig. 1a), but under stagnant ice in the Prairies (D in Fig. 1a). In other parts of southern Ontario (A in Fig. 1a) type I eskers may suggest stagnation-zone retreat (q.v. Shilts et al., 1987) in parts of zone 3.

This investigation of Laurentide eskers suggests that the factors that determined Laurentide R-channel pattern and operation were likely a complex combination of (i) supraglacial meltwater discharge, (ii) the number and location of sink holes, (iii) the ice surface slope, thickness and velocity, and (iv) the permeability, topography and rigidity of the bed. These factors cause and respond to changes in ice dynamics and thermal regime over the glacial cycle. Clearly, there is much still to be learned about eskers and their implications for R-channel drainage of the Laurentide Ice Sheet.

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