Macroforms, large bedforms and rhythmic sedimentary sequences in subglacial eskers, south-central Ontario: implications for esker genesis and meltwater regime

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Abstract

Eskers of south-central Ontario were deposited in closed, subglacial conduits which were continuous (main conduits). This interpretation is supported by: the intimate association of eskers with an anastomosing network of tunnel channels; relatively continuous esker ridges; minimal post-formational disturbance of esker sediments; intercalation of till and stratified sand and gravel; diapirc folding at an esker core; low variability in palaeocurrent direction; and upslope flow paths. Down-esker trends in clast lithology, roundness and sphericity indicate continuous conduits with sediment supply by subglacial deformation of adjacent material into the conduits, melt-out of sediment from the conduit walls, and fluvial resedimentation.

Esker ridge morphology is attributed to synchronous erosion, transportation and deposition along the main conduits, such that ridge discontinuities are primarily explained as zones of non-deposition during esker formation. Composite, pseudoanticlinal, and oblique accretion avalanche bed macroforms are identified. Gravel facies within these macroforms were deposited from fluidal flows or hyperconcentrated dispersions. Progradation of macroforms or migration of constituent large bedforms through the main conduits may have temporarily blocked constricted portions of those conduits and acted as possible internal (autogenic) controls on sediment availability and flow resistance. The location of macroforms within conduits was primarily controlled by conduit geometry and sediment availability, and later by feedback between macroform and conduit geometries.

Sand and gravel units alternate rhythmically in vertical section. Rhythmcity is interpreted as a response to episodic flood flows, controlled by seasonal changes in water supply, but not necessarily representing annual events. Fans, beads, anabranching reaches and extended, hummocky deposits are intimately associated with the main esker ridges. They are interpreted in terms of subglacial cavities and localized flotation zones, connected to the main conduits during flood events. In-phase wave structures are the products of hydraulic jumps in hyperconcentrated flood flows at flow expansions into swells within the main conduits, into cavities connected to the main conduits, or at a grounding line. Esker ridges record only the most powerful flood events, whereas fans and beads record flow events in finer detail, being primarily depositional reservoirs. Laterally fining deposits record the waning stages of conduit operation.

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1. Introduction

Previous studies on eskers have been conducted at two levels: (i) genetic interpretation based on external morphology (e.g. Shilts, 1984; St-Onge, 1984); and (ii) detailed site observation (e.g. Allen, 1971; Denny, 1972; Banerjee and McDonald, 1975; Saunderson, 1975; Diemer, 1988). A number of models of esker genesis have been put forward (cf. Banerjee and McDonald, 1975). Banerjee and McDonald's (1975) seminal paper related esker morphology and sedimentology to the site of deposition and the nature of the conduit (open or closed). However, their interpretation of fans and deltas as ice-marginal indicators only, led them to conclude that all eskers were time-transgressive. This notion was echoed by Hebrand and Åmark (1989) for some Swedish eskers.

The literature to date describes esker formation in terms of glacial hydrologic theories derived from observations on contemporary glaciers and small eskers, often in the process of formation (cf. Lewis, 1949; Stokes, 1958). However, esker systems produced by Pleistocene ice sheets tend to be much larger in dimensions than those forming today (cf. Banerjee and McDonald, 1975). It is debatable, therefore, whether a modern ana-

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**LEGEND:**
- Precambrian Rocks
- Paleozoic Limestone
- Granitoid (Felsic)
- Eskers

**Fig. 1.** Bedrock geology of the study area (modified from Ontario Geological Survey, 1991), and esker distribution (modified from Barnett et al., 1991). \(A\) = Actinolite.
logue for the formation of Pleistocene eskers exists and "to what degree modern subaerially exposed eskers should constrain the interpretation of sedimentary sequences in large 'fossil eskers'" (Banerjee and McDonald, 1975, p. 133).

The present discussion focuses on the landform associations, morphology and sedimentology, including clast lithology, sphericity and roundness, of late Wisconsinan eskers in south-central Ontario, Canada. Four main eskers passing through Tweed, Marlbank, Campbellford and Norwood are examined in detail (Figs. 1, 2). Large bedforms, macroforms and sedimentary sequences are identified, described and interpreted with respect to esker morphology and the meltwater regime responsible for their formation. An attempt is made to relate esker sedimentology systematically to external morphology by invoking formative processes in a uniformitarian manner (cf. Baker, 1988a, b); processes observed in modern environments are preferred as explanation, but these processes may be invoked in different combinations or situations and at very different magnitudes from those observed today. A model of synchronous subglacial sedimentation for large Pleistocene eskers, where ice-marginal sedimentation was, at most, a minor (later) component of esker sedimentation, is presented. Sediment sup-

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**Fig. 2.** Morphologic elements of main south-central Ontario eskers. Tributary ridges are not shown. The only obvious bead is mapped at pit N35.
ply and the dynamics of the glacial hydrologic system necessary to maintain a continuous conduit are investigated.

2. Theories of esker genesis

There appears to be general agreement that large Pleistocene eskers were formed subglacially or ice-marginally in deltaic, fan or re-entrant environments (Table 1). However, many Pleistocene eskers may have undergone complex genesis (cf. Lundqvist, 1979). The eskers of south-central Ontario exhibit little post-depositional sedimentary disturbance. Consequently, supraglacial and englacial depositional environments are not discussed here. Evidence used to determine subglacial and ice-marginal sites of esker deposition in the literature is presented (Table 2). There are, however, disagreements as to whether subglacial conduits were flowing full (closed) or partly full (open, at atmospheric or triple-point pressure) at the time of esker sedimentation (cf. Shreve, 1972 versus Hooke, 1984), and as to whether eskers were deposited in time-transgressive segments (Table 1) or synchronously (cf. Shreve, 1985; Garbutt, 1990). In addition, although conduit drainage may have been continuous along its length, deposition may have occurred in time-transgressive segments (Ashley et al., 1991).

To add to this confusion, the evidence listed for ice-marginal esker formation (Table 2) has also been documented for associated esker beads and fans inferred to have been deposited in a subglacial, grounding line position (Gorrell and Shaw, 1991). Conversely, the path of an esker up an adverse slope has been explained not only by a

Table 1
Literature on site of esker deposition

<table>
<thead>
<tr>
<th>Subglacial</th>
</tr>
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<tbody>
<tr>
<td>Pleistocene eskers:</td>
</tr>
<tr>
<td>Woodworth (1894); Stone (1899); Deane (1950) a; Gravenor (1957) a; Mirynech (1962) a; Lobanov (1967); Frakes et al. (1968); Denny (1972); McDonald and Vincent (1972); Banerjee and McDonald (1975) b; Shilts and McDonald (1975); Sauderson (1977, 1982) a; Ringrose (1982); Shilts (1984) b; St-Onge (1984) a; Lindström (1985); Shilts (1985); Shreve (1985); Térwindt and Augustin (1985); Shilts and Aylsworth (1987) b; Diemer (1988) b; Visser et al. (1987); Jensen (1988); Aylsworth and Shilts (1989a, b); Garbutt (1990); Ashley et al. (1991) b; Gorrell and Shaw (1991); Brennand and Sharpe (1994).</td>
</tr>
<tr>
<td>Contemporary eskers:</td>
</tr>
<tr>
<td>Stokes (1958); Jawtuchowicz (1965); Szuprynzyński (1965); Price (1966, 1969); Gustavson and Boothroyd (1982, 1987).</td>
</tr>
</tbody>
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<tr>
<th>Englacial</th>
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<tbody>
<tr>
<td>Pleistocene eskers:</td>
</tr>
<tr>
<td>Alden (1918); Tanner (1932); Shulmeister (1989).</td>
</tr>
<tr>
<td>Contemporary eskers:</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Supraglacial</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pleistocene eskers:</td>
</tr>
<tr>
<td>Crosby (1902); Tanner (1932).</td>
</tr>
<tr>
<td>Contemporary eskers:</td>
</tr>
<tr>
<td>Lewis (1949); Szuprynzyński (1965); Petrie and Price (1966); Price (1966, 1969); Fitzsimons (1991).</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Ice-marginal delta, fan or re-entrant environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pleistocene eskers:</td>
</tr>
<tr>
<td>De Geer (1897); Banerjee (1969); Sauderson and Jopling (1970); Aario (1971a, b); Denny (1972); Shaw (1972); Rust and Romanelli (1975); Banerjee and McDonald (1975); Sauderson (1975); Sauderson and Jopling (1980); Cheel (1982); Thomas (1984); Cheel and Rust (1986); Burbidge and Rust (1988); Christian (1988); Diemer (1988); Henderson (1988); Sharpe (1988); Hebrard and Amark (1989).</td>
</tr>
<tr>
<td>Contemporary eskers:</td>
</tr>
<tr>
<td>Szuprynzyński (1965); Howarth (1971).</td>
</tr>
<tr>
<td>Experimental:</td>
</tr>
<tr>
<td>Hanson (1943).</td>
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</tbody>
</table>

<table>
<thead>
<tr>
<th>Crack- or crevasse-fill in stagnant ice</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pleistocene eskers:</td>
</tr>
<tr>
<td>Upman (1910); Flint (1928, 1930); King and Buckley (1969).</td>
</tr>
</tbody>
</table>

a Authors specifically infer time-transgressive sedimentation and a subglacial site of deposition.
hydraulic head in a subglacial environment, but also, theoretically, by time-transgressive sedimentation at the ice-margin as the ice retreated (e.g. Banerjee and McDonald, 1975; Shilts, 1984). The upslope path, in the latter case, was inferred to be a result of segmental sedimentation over low gradients rather than an active upslope water flow path.

3. Geologic setting

The study area is divided into two major geologic regions: the Precambrian Shield in the north, and the Palaeozoic limestones in the south (Fig. 1). The north-facing Black River–Trenton escarpment marks the Shield margin. Precambrian rocks (Grenville Province) are felsic to ultramafic plutonic, and metasedimentary, and include: granites, granodiorites, migmatites, gabbros, marbles, conglomerates and breccias (Ontario Geological Survey, 1991). Palaeozoic outliers (Fig. 1) indicate that the Shield margin was north of its present location prior to the last glaciation (Mirynech, 1962). Some eskers lie on the Shield and extend onto the Palaeozoic carbonate (Marlbank and Tweed eskers; Fig. 1), while others appear to start at or near the Shield margin (Norwood and Campbellford eskers; Fig. 1). Consequently, the Shield margin provides a means to test the provenance of esker sediments.

All of the main eskers are located in channels cut into glaciogenic sediment or bedrock, while some minor eskers (< 3 km long) occur on interfluves (Brennand and Shaw, 1994). These channels are difficult to follow north of the Shield margin. However, Wilson (1904) suggested that some do extend headward onto the Shield. The channels appear to have been occupied and enhanced by subglacial meltwater and follow a path which is against the regional gradient (cf. Gilbert, 1990; Brennand and Shaw, 1994). Consequently, they are inferred to be tunnel channels formed by catastrophic subglacial meltwater flows (cf. Gilbert, 1990; Shaw and Gilbert, 1990; Shaw and Gorrell, 1991; Gilbert and Shaw, 1992; Brennand and Shaw, 1994). This does not rule out the possibility that some of the channels may have pre-dated the last glaciation (Wilson, 1904), but it does imply that they were enhanced by subglacial meltwater during the late Wisconsinan (cf. Gilbert, 1990; Brennand and Shaw, 1994).

4. Gross morphology

The four eskers studied in detail (Fig. 2) exhibit four main morphologic elements: (1) main ridge, (2) fans, (3) beads with minor ridges, and (4) extended, hummocky deposits.
4.1. Main esker ridges

In general, these are single sinuous ridges, but bifurcation is not uncommon (Fig. 2). Obvious anabranching reaches along the Marlbank and Campbellford eskers are 2.0 km and 0.7 km in length, respectively (Fig. 2). The width of the main ridges increases from east to west across the study area. The Tweed and Marlbank esker ridges are generally 0.03–0.10 km wide, while the Campbellford and Norwood esker ridges are up to 0.15 km and 0.40 km wide, respectively. Major and minor (< 1 km long) tributary ridges join the main esker ridges from upflow (Fig. 1).

Eskers in this region are relatively sharp-crested. The Norwood ridge broadens in places, but is not flat-topped. The term “broad-crested” is not applied here, as this term has dynamic implications which may not be directly equivalent to those proposed by Shreve (1985) for the Katahdin esker system. In plan view (Fig. 2), the ridges do not exhibit a consistent width along their length. In some places (pit M6*, Fig. 2), the main ridge is composed of a number of wider bulbous areas (O'Donnell, 1966) joined by narrower sharp-crested ridges.

Crest long-profiles of the main ridges are irregularly undulatory (Fig. 3). These profiles have been rotated to bring the high-stand shoreline of Glacial Lake Iroquois (750’ or 228.6 m) back to a horizontal plane, using isobase data (Mirynech, 1962). This adjustment (raising the southwest end) was necessary to compensate for greater rebound in the northeast following deglaciation. While it cannot compensate for all isostatic adjustments, that is those prior to the Iroquois high stand, it gives the best approximation at present. The long-profiles (Fig. 3) show that crestslines trend upslope. The elevational range (climb) of the eskers is between 92 m (Tweed esker) and 53 m (Norwood esker). Since the thickness of aggregate deposits in the Tweed esker is 3–25 m (Ministry of Natural Resources, 1987), and this is considerably less than the crestline elevational change, the upslope paths of the main eskers in this study are considered to be real; upslope paths are not a result of changes in the thickness of aggregate deposits downstream.

Ridge continuity is also demonstrated by crest long-profiles (Fig. 3). Most of the discontinuities in the ridges are occupied by underfit creeks, Trent River or West Ouse River (Fig. 3). The Campbellford esker is more continuous northeast of Campbellford to the Shield margin, and relatively discontinuous southwest of Campbellford (Fig. 3b). Mirynech (1962) suggested that the southern portion of the Campbellford esker is wave modified. The Marlbank esker is the most discontinuous of the four eskers studied (consequently, a long-profile was not plotted; Fig. 2); it is quite discontinuous on the Shield and for 32 km above its junction with the Tweed esker (Fig. 1). Some of its inferred path is now occupied by the Moira River.

4.2. Fans

Both major and minor fans extend from the main esker ridges in a downflow direction. Major fans are en echelon, or overlapping, towards the southern ends of the main eskers (Tweed, Campbellford and Norwood; Fig. 2). They reach a maximum of 2.5 km long and 1 km wide. Mirynech (1962) reported a kame–esker complex, interpreted here as a fan complex, at the southern end of the Tweed esker. Minor fans are connected laterally to the main esker ridges. They are mapped on both sides of each ridge and preferentially at bends. They also occur along straighter portions of the ridges. Minor fans are generally less than 1 km long and 0.2–0.3 km wide. The Tweed esker exhibits at least 18 minor fans (Fig. 2).

4.3. Beads with minor ridges

A single bead is located ~1 km southwest of Norwood (pit N35; Fig. 2) on Norwood esker. The bead is ~0.25 km wide and ~0.3 km long. This bead is joined to the main ridge by narrow minor ridges. Although it was the only bead mapped in this study, some of the minor fans which run alongside the main ridges may represent a morphological (and functional?) continuum with beads.
Fig. 3. Crest long-profiles rotated to bring the high-stand shoreline (750 ft or 228.6 m) of Glacial Lake Iroquois back to a horizontal plane. (a) Tweed esker. (b) Campbellford esker. (c) Norwood esker.
4.4. Extended, hummocky deposits

Long (up to 8.5 km), wide (up to 0.5 km) bands of hummocky deposits, previously mapped as proximal sand, gravelly sand and gravel (Leyland and Mihychuk, 1983, 1984a, b; Leyland and Russell, 1984), are observed on one or both sides of the main ridges towards the headward (northern) ends of the Tweed, Campbellford and Norwood eskers (Fig. 2). These deposits are at lower elevations than the main ridges. This morphologic element is not to be confused with the extended deposits of Hebrand and Åmark (1989). They considered these deposits to include all non-linear morphologic components: hummocks, plateaus, and terraces. Plateaus and terraces are not recognized in this study. In contrast to the influence of Hebrand and Åmark (1989) that extended deposits occur as downflow extensions of the main ridges, the hummocky, extended deposits in this study are lateral to, and run parallel to, the main ridges.

5. Down-esker trends

5.1. Clast lithology, sphericity and roundness

Clast lithology, sphericity and roundness were recorded along the length of the esker ridges. Distance from a datum was determined planimetrically for each ridge. Data collection from the Tweed esker was the most systematic, and is the focus of attention here. Samples were collected from two clast populations: (i) unit samples, or in situ clasts, from vertical sections; and (ii) oversize samples gathered from the base of slopes under pit faces and from piles of oversize material within the pit. Unit samples include pebbles and cobbles (from 4.5φ to 7.5φ diameter). Oversize samples include cobbles and boulders (from 5.5φ to 11.0φ diameter). The separate treatment of the two samples is a crude attempt to control grain size (cf. Sneed and Folk, 1958). Sample size at each location varied from 60 to 120 clasts (Table 3).

Clast roundness was determined visually in the field (cf. Powers, 1953). For statistical manipulation, the geometric mean of the visual roundness class (Powers, 1953) exhibited by each clast was assigned as its roundness value. Maximum projection sphericity (Ψp) was calculated from clast axial lengths (cf. Sneed and Folk, 1958). The results of the analyses are presented in total (Table 3) and graphed selectively (Figs. 4, 5, 6).

With distance from the Shield margin, there is an increase the proportions of limestone clasts and a concomitant decrease in the proportions of Shield clasts for the unit samples taken from Tweed esker (Fig. 4a; Table 3). Limestone clasts dominate unit sample clast counts at each location. Trends in the oversize samples are almost diametrically opposite to those exhibited by the unit samples (Fig. 4b; Table 3). Although Shield lithologies dominate oversize clast counts, downflow (south) from the Shield margin the percentage of Shield clasts increases and that of limestone clasts decreases. The percentage of gabbro clasts in both unit and oversize samples, increases away from the Shield margin.

Mean Ψp (Sneed and Folk, 1958) of unit sample clasts decreases with distance down-esker for most lithologies (Fig. 5a; Table 3). Similar trends are observed in the oversize samples (Fig. 5b; Table 3). Excluding lithologies only represented by one clast, grand mean Ψp ranges (Figs. 5c, 5d; Table 3) are similar for both unit and oversize samples. Limestone has one of the lowest grand mean Ψp in both oversize and unit samples (Table 3).

Grand mean roundness classes for most clast lithologies, in both unit and oversize samples, are rounded to subrounded (Figs. 6c, 6d; Table 3; cf. Powers, 1953). Unit samples exhibit a general increase in the mean roundness of Shield clasts with distance downflow, while the mean roundness of limestone clasts remains relatively constant (Fig. 6a; Table 3). Oversize samples show no consistent downflow trends (Fig. 6b; Table 3) and large confidence intervals on closely grouped means (Table 3) restrict further generalization.

The divergence of clast lithology trends down-esker for unit and oversize samples suggests that these groups may be recording two different processes. Unit samples exhibit a decrease in Shield clasts with distance from the Shield margin (Figs. 1, 4a). This is expected fluvial transport be-
Table 3
Lithologic frequency, mean roundness and mean sphericity, by lithology for clast samples from the Tweed esker

<table>
<thead>
<tr>
<th>Pit</th>
<th>Distance (km)</th>
<th>N (unit)</th>
<th>Limestone (L)</th>
<th>Mudstone (M)</th>
<th>Granite (GT)</th>
<th>Granodiorite (GD)</th>
<th>Gabbro (GB)</th>
<th>Metasedimentary a</th>
<th>Migmatite (MG)</th>
<th>Basalt (B)</th>
<th>Granitoid b (GRAN)</th>
<th>Shield c (SHEL)</th>
<th>Mean sphericity d:</th>
<th>Mean roundness e:</th>
</tr>
</thead>
<tbody>
<tr>
<td>T14</td>
<td>21.56</td>
<td>60</td>
<td>50.0 ± 12.7</td>
<td>5.0 ± 5.5</td>
<td>16.7 ± 9.4</td>
<td>8.3 ± 7.0</td>
<td>1.7 ± 3.2</td>
<td>18.3 ± 9.8</td>
<td>0.0</td>
<td>0.0</td>
<td>25.0 ± 11.0</td>
<td>45.0 ± 12.6</td>
<td>0.655 ± 0.019</td>
<td>0.496 ± 0.065</td>
</tr>
<tr>
<td>T23</td>
<td>58.95</td>
<td>60</td>
<td>58.3 ± 12.5</td>
<td>0.0</td>
<td>10.0 ± 7.6</td>
<td>11.7 ± 8.1</td>
<td>18.3 ± 9.8</td>
<td>1.7 ± 3.2</td>
<td>0.0</td>
<td>0.0</td>
<td>21.7 ± 10.4</td>
<td>41.7 ± 12.5</td>
<td>0.742 ± 0.025</td>
<td>0.320 ± 0.056</td>
</tr>
<tr>
<td>T24</td>
<td>63.57</td>
<td>60</td>
<td>73.3 ± 11.2</td>
<td>0.0</td>
<td>1.7 ± 3.2</td>
<td>3.3 ± 4.5</td>
<td>21.7 ± 10.4</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>5.0 ± 5.5</td>
<td>26.7 ± 11.2</td>
<td>0.892 ± 0.020</td>
<td>0.479 ± 0.048</td>
</tr>
<tr>
<td>T26</td>
<td>70.13</td>
<td>60</td>
<td>58.3 ± 12.5</td>
<td>0.0</td>
<td>5.0 ± 5.5</td>
<td>6.7 ± 6.3</td>
<td>20.0 ± 10.1</td>
<td>1.7 ± 3.2</td>
<td>1.7 ± 3.2</td>
<td>0.0</td>
<td>11.7 ± 8.1</td>
<td>41.7 ± 12.5</td>
<td>0.713 ± 0.029</td>
<td>0.479 ± 0.048</td>
</tr>
</tbody>
</table>

Mean sphericity d: unit samples
- T14: 0.665 ± 0.019, 0.585 ± 0.108, 0.733 ± 0.046, 0.742 ± 0.036
- T23: 0.742 ± 0.025, 0.777 ± 0.025, 0.707 ± 0.062
- T24: 0.835 ± 0.061, 0.707 ± 0.062
- T26: 0.835 ± 0.061, 0.707 ± 0.062

Mean roundness e: unit samples
- T14: 0.320 ± 0.056, 0.532 ± 0.166, 0.532 ± 0.166
- T23: 0.532 ± 0.166, 0.532 ± 0.166
- T24: 0.532 ± 0.166, 0.532 ± 0.166
- T26: 0.532 ± 0.166, 0.532 ± 0.166

Mean sphericity d: oversize samples
- T14: 0.662 ± 0.026, 0.731 ± 0.054, 0.737 ± 0.031
- T23: 0.636 ± 0.026, 0.731 ± 0.054, 0.737 ± 0.031
- T24: 0.636 ± 0.026, 0.731 ± 0.054, 0.737 ± 0.031
- T26: 0.636 ± 0.026, 0.731 ± 0.054, 0.737 ± 0.031

Mean roundness e: oversize samples
- T14: 0.329 ± 0.050, 0.432 ± 0.071, 0.432 ± 0.071
- T23: 0.329 ± 0.050, 0.432 ± 0.071, 0.432 ± 0.071
- T24: 0.329 ± 0.050, 0.432 ± 0.071, 0.432 ± 0.071
- T26: 0.329 ± 0.050, 0.432 ± 0.071, 0.432 ± 0.071

a Metasedimentary abbreviated to MS in Figs. 5 and 6.

b Granitoid class includes granites and granodiorites.

c Shield class includes granites, granodiorites, gabbros, migmatites, basalts and metasedimentary rocks.

d Lithologic frequency data presentation: % frequency ± sampling error at 95% confidence interval. Calculation modified from Dryden (1931) using 95% confidence interval.

e Mean roundness and sphericity data presentation: mean ± 95% confidence interval for the sample mean.

NS: insufficient data for calculation of confidence limits.
haviour (cf. Sneed and Folk, 1958). However, the increased percentage of Shield clasts in oversize samples and of gabbro in both samples, away from their source (Figs. 1, 4), cannot be explained fully by fluvial processes. Possible sources of sediment deposited in an esker include allochthonous sediment, sediment melted out from debris-rich ice, bedrock eroded by flowing water, adjacent deformable sediment squeezed into the conduit (cf. Shoemaker, 1986; Shoemaker and Leung, 1987; Alley, 1991), sediment flushed from cavities into the conduit, and/or fluviially reworked or episodically transported sediment from within the conduit system. The most likely cause of an increase in exotic lithologies with distance from source in the oversize samples, is the squeezing of deformable substrate (containing both exotic and local lithologies) into the conduit or melt-out of sediment from debris-rich ice. Moreover, gabbro is potentially the most far-travelled lithology (Fig. 1), and towards the southern end of Tweed esker the percentage of gabbro clasts increases while that of granitoid clasts declines (Fig. 4). This observation is consistent with a spatially controlled, incremental freeze-on and melt-out of debris in the subglacial environment. Such material may be consequently delivered to the conduit through subglacial deformation or ice transport and melt-out processes. The appearance of limestone clasts north (upflow) of the Shield margin in an oversize sample (Fig. 4b) is explained by the presence of a limestone outlier (Fig. 1). The lower percentage of large limestone clasts within these oversize samples (Fig. 4b) may be explained by the poor preservation of large, relatively soft limestone clasts.

Past research has suggested that clast sphericity is primarily controlled by lithology and grain size (cf. Sneed and Folk, 1958). An overall decrease in sphericity with distance for the Tweed unit samples (Fig. 5a) may be explained by the mode of fluvial transport. Clast fabric data (discussed later) from gravels within the esker ridge records imbricate clasts with a-axes dominantly transverse to flow direction. It is inferred from this that clasts were primarily transported by tractional rolling as bedload (cf. Johansson, 1963, 1965, 1976; Rust, 1972). Consequently, a decrease in sphericity downdownflow is interpreted as the effect of clast abrasion by rolling along the bed, thus producing prolate clast forms. In addition, limestone clasts increase in frequency down-esker in the unit samples (Fig. 4a) and exhibit the most pronounced increase in sphericity with distance along the ridge (Fig. 5a). Limestone tends to be removed from outcrop in tabular fragments, preferentially breaks parallel to bedding planes (cf. Sneed and Folk, 1958), and forms discs or rods during the abrasion process. Similar trends are also evident for the oversize samples and deserve comment. Trends in oversize clast lithology suggest that these clasts were transported primarily
by ice, and only secondarily by water. If the continuous decrease in sphericity with distance (Fig. 5b) is real, it must be explained by coherent processes rather than stochastic inference. While it is possible that a decrease in sphericity may be explained by fluvial transportational vigour over short distances, this does not explain the apparent consistent down-esker trend. It is possible that grain size may have been a controlling factor. The data set was inadequate to investigate this hypothesis as confidence limits are too large (Table 3). Mean sphericity trends in the Tweed oversize sample remain enigmatic.

The dominance of rounded to subrounded clasts, in both the unit and oversize samples (Figs. 6c, 6d), is consistent with fluvial transport processes (cf. Sneed and Folk, 1958). Clast roundness in a fluvial system is primarily controlled by transportational distance or vigour (cf. Sneed and Folk, 1958). An increase in the mean roundness of far-travelled clasts (Shield clasts) with distance from source for unit samples (Fig. 6a) suggests that this is primarily a function of transport distance in a continuous meltwater conduit. A constant mean roundness for limestone clasts downflow (~ 0.48) in the same data set (Fig. 6a) may indicate the effects of preferential cleavage parallel to bedding planes, thus cancelling out the expected effects of attrition (cf. Sneed and Folk, 1958). Alternatively, it could indicate a continual addition of limestone clasts to the system by melt-out or squeezing of deformable substrate into the conduit. The grand mean roundness of limestone clasts in the unit samples (0.478 ± 0.026, Table 3) is significantly less than the asymptotes reported elsewhere (e.g. 0.63, Sneed and Folk,

---

Fig. 5. Mean and grand mean maximum projection sphericity ($\Psi_m$, Sneed and Folk, 1958) for the Tweed unit (a, c), and oversize (b, d) samples. The vertical line is the location of the Shield margin. For lithologic abbreviations see Table 3. Actinolite is located on Fig. 1.
lithology, sphericity and roundness suggest sediment supply to the Tweed esker was mainly from squeezing of adjacent sediment into the main conduit or tributary conduits, melt-out of sediment from conduit walls, and fluvial reworking of that sediment within the conduit system. Fluvial transport processes within the continuous, closed conduit are suggested to have been dominated by tractional rolling as bedload.

5.2. Palaeocurrents

Palaeocurrent measurements were taken from gravel fabrics, cross-beds in gravel and sand, and cross-laminations in sand at a number of localities along the length of the eskers (Fig. 7; Tables 4, 5). Bulk gravel fabrics (azimuth and plunge of...
Fig. 7. Palaeocurrent observations for the Tweed (T), Campbellford (C) and Norwood (N) eskers. TJS1, C43, etc. are pit locations. Gravel fabrics presented as lower-hemisphere projections (contour interval is 2 standard deviations), and cross-lamination and cross-bed measurements as rose diagrams. Rose diagrams are plots of actual number of observations. Longest segment at pit C31 represents 37 observations; the same scale is used for all rose diagrams except at pit N36 (displayed at 50%; longest segment represents 71 observations).
Table 4  
Paleocurrent statistics from gravel fabric measurements

<table>
<thead>
<tr>
<th>Pit number</th>
<th>Sedimentary facies/structures</th>
<th>Sample number</th>
<th>Flow azimuth (°)</th>
<th>Mean plunge (°)</th>
<th>Vector strength ($S_1$)</th>
<th>Significance level (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>TJS1</td>
<td>Horizontally bedded; pseudoanticlinal macroform a</td>
<td>180</td>
<td>198</td>
<td>52</td>
<td>0.5887</td>
<td>99.0</td>
</tr>
<tr>
<td>T14</td>
<td>Eastern side of pseudoanticlinal macroform only</td>
<td>60</td>
<td>277</td>
<td>55</td>
<td>0.6203</td>
<td>99.0</td>
</tr>
<tr>
<td>T24</td>
<td>Heterogeneous, unstratified gravel</td>
<td>60</td>
<td>221</td>
<td>49</td>
<td>0.5137</td>
<td>99.0</td>
</tr>
<tr>
<td>T26</td>
<td>Heterogeneous, unstratified gravel</td>
<td>110</td>
<td>178</td>
<td>36</td>
<td>0.5603</td>
<td>99.0</td>
</tr>
<tr>
<td>C43</td>
<td>Heterogeneous, unstratified gravel</td>
<td>60</td>
<td>260</td>
<td>57</td>
<td>0.6520</td>
<td>99.0</td>
</tr>
<tr>
<td>C30</td>
<td>Heterogeneous, unstratified gravel</td>
<td>133</td>
<td>246</td>
<td>27</td>
<td>0.5946</td>
<td>99.0</td>
</tr>
<tr>
<td>C32</td>
<td>Heterogeneous, unstratified gravel</td>
<td>25</td>
<td>206</td>
<td>28</td>
<td>0.8665</td>
<td>99.0</td>
</tr>
<tr>
<td>C31</td>
<td>Horizontally bedded; massive, imbricate, clast-supported gravel</td>
<td>120</td>
<td>189</td>
<td>49</td>
<td>0.6804</td>
<td>99.0</td>
</tr>
<tr>
<td>N54</td>
<td>Heterogeneous, unstratified gravel; massive, imbricate, clast-supported gravel</td>
<td>50</td>
<td>218</td>
<td>46</td>
<td>0.7895</td>
<td>99.0</td>
</tr>
<tr>
<td>N35</td>
<td>Heterogeneous, unstratified gravel</td>
<td>30</td>
<td>209</td>
<td>39</td>
<td>0.8004</td>
<td>99.0</td>
</tr>
<tr>
<td>N36</td>
<td>Heterogeneous, unstratified gravel; massive, imbricate, clast-supported gravel</td>
<td>100</td>
<td>109</td>
<td>33</td>
<td>0.5281</td>
<td>99.0</td>
</tr>
</tbody>
</table>

a Significance level of sample being non-random, according to test statistic ($S_1/S_2$) of Woodcock and Naylor (1983).

b Both sides of pseudoanticlinal macroform included.

clast ab-planes) were primarily recorded in heterogeneous, unstratified gravel and pseudoanticlinal macroforms in the esker ridges (Fig. 7; Table 4). The statistics for pit TJS1 (Fig. 7; Table 4) include one sample from each side of a pseudoanticlinal macroform which together produce an inferred palaeoflow down the esker ridge. Data for pit T14 (Fig. 7) is from only the eastern side of a pseudoanticlinal macroform. Because secondary vortices responsible for the formation of the pseudoanticlinal macroform produced imbrication downflow and towards the crest of the anticline (discussed later), the stereonet and statistics for pit T14 (Fig. 7; Table 4) are inter-

Table 5  
Paleocurrent statistics from cross-bed and cross-lamination measurements

<table>
<thead>
<tr>
<th>Pit number</th>
<th>Morphologic element</th>
<th>Sample number</th>
<th>Vector mean ($\bar{\theta}$ (°))</th>
<th>Mean resultant magnitude ($\overline{R}$)</th>
<th>Standard error ($S_2$ (°))</th>
<th>Significance level of $\overline{R}$ (%) a</th>
<th>Deviation of $\bar{\theta}$ from main ridge axis (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>TJS1</td>
<td>Laterally fining deposit</td>
<td>23</td>
<td>154</td>
<td>0.9187</td>
<td>4.8743</td>
<td>99.0</td>
<td>22</td>
</tr>
<tr>
<td>T16</td>
<td>Laterally fining deposit</td>
<td>35</td>
<td>210</td>
<td>0.7168</td>
<td>7.8130</td>
<td>99.0</td>
<td>20</td>
</tr>
<tr>
<td>T15</td>
<td>Minor fan</td>
<td>21</td>
<td>276</td>
<td>0.9199</td>
<td>5.0978</td>
<td>99.0</td>
<td>76</td>
</tr>
<tr>
<td>T12</td>
<td>Laterally fining deposit</td>
<td>43</td>
<td>229</td>
<td>0.9504</td>
<td>2.7965</td>
<td>99.0</td>
<td>1</td>
</tr>
<tr>
<td>T11</td>
<td>Minor fan</td>
<td>76</td>
<td>254</td>
<td>0.1661</td>
<td>27.4506</td>
<td>70.0</td>
<td>25</td>
</tr>
<tr>
<td>T19</td>
<td>Minor fan</td>
<td>14</td>
<td>256</td>
<td>0.9028</td>
<td>6.9972</td>
<td>99.0</td>
<td>26</td>
</tr>
<tr>
<td>C43</td>
<td>Laterally fining deposit</td>
<td>24</td>
<td>273</td>
<td>0.8426</td>
<td>6.8309</td>
<td>99.0</td>
<td>23</td>
</tr>
<tr>
<td>C30</td>
<td>Laterally fining deposit</td>
<td>6</td>
<td>171</td>
<td>0.7162</td>
<td>18.8783</td>
<td>95.0</td>
<td>59</td>
</tr>
<tr>
<td>C32</td>
<td>Laterally fining deposit</td>
<td>61</td>
<td>212</td>
<td>0.8166</td>
<td>4.5793</td>
<td>99.0</td>
<td>12</td>
</tr>
<tr>
<td>C31</td>
<td>Laterally fining deposit</td>
<td>62</td>
<td>182</td>
<td>0.9709</td>
<td>1.7950</td>
<td>99.0</td>
<td>18</td>
</tr>
<tr>
<td>N35</td>
<td>Lateral beak</td>
<td>31</td>
<td>221</td>
<td>0.9694</td>
<td>2.5405</td>
<td>99.0</td>
<td>5</td>
</tr>
<tr>
<td>N56</td>
<td>Laterally fining deposit</td>
<td>54</td>
<td>221</td>
<td>0.8581</td>
<td>4.2564</td>
<td>99.0</td>
<td>5</td>
</tr>
<tr>
<td>N36</td>
<td>Major fan or fan complex</td>
<td>352</td>
<td>201</td>
<td>0.6309</td>
<td>2.9777</td>
<td>99.0</td>
<td>0–60</td>
</tr>
</tbody>
</table>

a Significance level determined from critical values of $\overline{R}$ for Rayleigh's test for the presence of a preferred trend (cf. Curray, 1956; Davis, 1986).
preted to be the result of deposition within a constrained conduit. In most cases, flow azimuths deviate very little from the ridge axis and are significant at the 99% level (Table 4). The flow azimuth for pit N36 (Fig. 7; Table 4) appears to be at ~90° to the axis of the main ridge. This data is from a major fan or fan complex (Fig. 2) towards the distal end of the esker.

Palaeoflows inferred from palaeocurrent measurements on cross-beds and cross-laminations exhibit a much higher degree of deviation from the ridge axis (Fig. 7; Table 5). This is expected

Fig. 8. (a) Normal faults lateral to an undisturbed ridge core, Campbellford esker (pit C32; Fig. 2). Note alternating sand and gravel units. (b) Diapiric fold at the core of Tweed esker (pit T14; Fig. 2). Metre rods for scale.
given the smaller scale of the structures, the lower flow regime inferred for their formation, and the greater influence of local bed topography on their formative flow direction (Allen, 1966). Statistically, most are unidirectionally significant at the 95% level (Table 5). Data for pit T11 (Fig. 7) includes regressive, type B, ripple-drift cross-lamination (cf. Jopling and Walker, 1968) in a minor fan. Fans are characterised by inferred palaeoflows with over 25° deviation from the main ridge axis (Table 5). In general, palaeoflows inferred from laterally fining deposits within the ridge deviate by less than 25° from the ridge axis. A possible exception is pit C30 (Fig. 7), where the sample size is so small that its deviation of 59° (Table 5) cannot be accepted with confidence. Data for pit N36 (Fig. 7) may include measurements from more than one fan. However, up to 360° variation in palaeocurrent measurements could be consistent with deposition in a single subaqueous fan (cf. Cheel, 1982).

6. General depositional environment

The general environmental constraints on esker sedimentation must first be determined before sedimentological interpretation of esker deposits, in terms of detailed genesis and hydrodynamic controls on sedimentation, can be attempted. This first step is vital as many sedimentary structures within eskers are common to both open-channel and closed-conduit systems (cf. Saunderson, 1977, 1982; Ringrose, 1982). Landform associations, esker morphology, and downslope trends in clast characteristics and palaeoflow estimates, in combination, suggest a continuous, subglacial, closed-conduit depositional environment. Location of esker ridges within tunnel channels, upslope paths and low variability in palaeocurrent directions are characteristic of eskers deposits in subglacial conduits (Table 2). Ridge continuity, combined with upslope paths, imply that water must have been under considerable pressure to flow against the topographic gradient. This would have necessitated closed-conduit conditions, where water was driven to the ice margin by excess pressure over hydrostatic pressure (cf. Shreve, 1972). Discontinuities in esker ridges have been attributed to post-depositional erosion or time-transgressive sedimentation (e.g. De Geer, 1897; Henderson, 1988). For closed-conduit, steady-state conditions, Shreve (1985) stated that equipotential contours would be closest over bump crests. This would have resulted in high transportational capacities in the subglacial streams crossing them. Therefore, discontinuities may have been zones of non-deposition that were synchronous with esker sedimentation elsewhere in continuous conduits. Saunderson (1977) suggested that gaps were erosional zones that were contemporaneous with sliding bed deposition. The notion of non-depositional zones within continuous conduits will be discussed later.

7. Sedimentology of the main esker ridges

The sedimentologic characteristics of 37 pits were investigated (Fig. 2). Each observed unit was given a number in the form: pit number/face number–unit number (e.g. T24/1–1). Palaeoflow estimates from cross-bedded sand and gravel, cross-laminated sand, and imbricate clast ab-planes, were recorded together with clast a-axis orientations (Table 7). An architectural approach towards esker ridge sedimentology was taken (cf. Miall, 1985).

On the scale of the esker ridge, gravel architecture may be tabular or pseudoanticlinal. Normal faults are common towards the lateral flanks of the main ridges, but there is minimal post-formational disturbance at the ridge cores (Fig. 8a). This corroborates the inference of a subglacial environment of deposition with lateral ice support (cf. McDonald and Shilts, 1975) for the esker ridges. At one location along the Tweed esker, a diapiric fold is observed at the ridge core (Fig. 8b), and a seismic investigation (reflection method supplemented by refraction measurements) north of pit T24 (Fig. 2) revealed diamicton, possibly till, intercalated with glaciofluvial sand and gravel (G. Gorrell, personal communication, 1989). The formation of a diapiric fold would have required relatively low pressure conditions to have been
established within the conduit. Under similar meltwater discharge \(Q_w\) conditions, large conduits would have had relatively lower pressures than small conduits (cf. Röthlisberger, 1972). Since a pressure gradient must have existed between adjacent ice and the water within the conduit, till creep or injection into the conduit would have occurred (cf. Hoppe, 1952; Shoemaker, 1986; Shoemaker and Leung, 1987; Alley, 1991). Sediment may have been squeezed into conduits during low-pressure conditions in the winter (cf. Alley, 1991; Gorrell and Shaw, 1991) or during transient very low pressure conditions during a rise in \(Q_w\) (cf. Röthlisberger, 1972). In the latter case, invoking Bernoulli's Principle, basal sediment and ice would have been sucked into the conduit, the sediment forming the diapiric fold and the ice being removed by melting.

No diamicton drape was observed over the esker ridges, but this is not problematic to a subglacial interpretation as clast lithology, sphericity and roundness trends, and diamicton, possibly till, intercalated with ridge sand and gravel, suggest that the primary source of esker sediment was from adjacent basal sediment. This implies that the ice was relatively clean at the time of esker formation. These sediment supply considerations may explain the relatively narrow ridge and discontinuous nature of the Marlbank esker (Fig. 2), which is located in a bedrock channel.

Within the main esker ridges, three types of macroforms, or ridge-scale sedimentary structures, are identified: composite, pseudoanticlinal, and oblique accretion avalanche bed. The sedimentology of these macroforms are discussed next.

### 7.1. Composite macroforms

Composite macroforms (cf. Hoey, 1992; e.g. expansion bars, Baker, 1978) on the scale of the ridge exhibit a number of different sand and gravel facies (described and interpreted below), including climbing dunes in gravel (Fig. 9). These macroforms are over 10 m in height. They are interpreted to be products of deposition in zones of flow expansion where conduits widen. This mechanism is discussed later.

#### 7.1.1. Heterogeneous, unstratified gravel

Heterogeneous, unstratified gravel (Fig. 10) is the dominant sedimentary facies in the esker ridges and is the primary gravel facies at their cores. Facies architecture can be tabular or pseudoanticlinal. Lower contacts are erosional. Units are 0.5–5.0 m thick, and may exhibit a maximum grain size of boulders (generally \(< -10\phi\) diameter), cobbles or pebbles. This facies is polymodal (Figs. 11a, 11c), ungraded, and framework-supported by clasts over \(-1\phi\) diameter (Fig. 10; cf. Rust and Koster, 1984). Material finer than \(4\phi\) constitutes less than 5% by weight, of each unit (Fig. 11).

Standard deviations calculated from grain size data on some of the finer (pebble-dominated) heterogeneous, unstratified gravels (Table 6), suggest that, *en masse*, this facies is very poorly sorted (standard deviation \(> 2\phi\); Table 6) (cf. Folk and Ward, 1957). This is misleading as gravel structure includes vaguely delineated, lenticular packages of bimodal clast-supported, bimodal matrix-supported, polymodal, and unimodal/openwork gravel (Fig. 10). The arrangement of these packages vertically or laterally follows no consistent pattern. This structure is problematic when collecting representative samples for grain size analysis. For example, samples T23/1–1a and T23/1–1b (Table 6) were taken from the same unit in close proximity. Their grain size distributions are markedly different (Fig. 11b). Sample T23/1–1a* (Figs. 11a, 11b; Table 6) is an amalgamation of samples T23/1–1a and T23/1–1b.

Larger clasts (boulders, cobbles and/or pebbles) often occur in imbricate clusters, within gravel with an otherwise visually random fabric (Fig. 10). Although the dominant orientation of imbricate clasts is with their \(a\)-axis transverse to flow direction \((a(r))\), a high proportion of clasts are oriented with their \(a\)-axis parallel to flow direction \((a(p))\) (Table 7). In most cases, the dominant orientation of both cobbles and pebbles is \(a(t)\) (Table 7). The mean plunge of the imbr-
Fig. 9. Part of a composite macroform, Norwood esker (pit N36; Fig. 2). Note climbing gravel dunes (right). Flow to the left. Metre rods for scale.

Fig. 15. Pseudoanticlinal macroform composed of heterogeneous, unstratified gravel with a crest-convergent fabric, Tweed esker (pit TJS1; Fig. 2). Fabric statistics in Table 7. Metre rods for scale; Primary flow was into the photograph (southward).
cate plane of clasts ranges from 14° to 57°, with
most over 30°. The unit (unit N36/1–1; Table 7)
exhibiting a mean plunge of 14° has a higher
proportion of fines than the other heterogeneous,
unstratified gravels. Vector strength (eigenvalue
Sv) is variable (Table 7).
Deposition from fluidal flows is inferred from
the presence of imbricate clast clusters with a-axis
orientations dominantly transverse to flow direc-
tion. Large clasts were primarily transported by
tractional rolling as bedload (cf. Johansson, 1963,
1965, 1976; Rust, 1972). The a-axis parallel clasts
suggest that transport by suspension and saltation
also occurred prior to deposition (cf. Johansson,
1963). High plunge angles are attributed to the
high frequency of clast-to-clast contacts during
deposition of this framework-supported facies (cf.
Fig. 10. Heterogeneous, unstratified gravel facies. (a) Poorly
delineated lenses of bimodal clast-supported, polymodal and
openwork gravel. Visually chaotic fabric, Campbellford esker
(pit C30; Fig. 2). (b) Vaguely delineated lenticular structure
and clast imbrication of larger clasts, Norwood esker (pit
N36; Fig. 2). Metre grid for scale.

Fig. 11. Heterogeneous, unstratified gravel facies. (a) Grain
size distributions. (b) Partitioning of facies. Grain size data for
sample T23/1–1° is an amalgamation of samples T23/1–1a
and T23/1–1b. (c) Grain size histogram demonstrating poly-
modal texture (unit C43/1–1; Tables 6, 7).
Rust, 1972). This is consistent with lower plunge angles recorded for more matrix-rich members of this facies, where clast-to-clast contacts would have been less frequent (unit N36/1–1, Table 7).

Sediment support mechanisms in a fluid flow with traction, saltation and suspension transport would have included support from the bed and from fluid turbulence. The absence of grading and the dominance of clast-to-clast contacts indicate that dispersive pressure within a highly concentrated sediment dispersion (cf. Smith, 1986; Costa, 1988) or a sliding bed (Saunderson, 1977, 1982) was not important.

Deficiency of very coarse sand to granule grain sizes (from 0φ to 2φ diameter) in open-channel fluvial sediments, imparts a natural bimodality to gravel deposits (cf. Shaw and Kellerhals, 1982). This deficiency may be the result of preferential transport of this size range (cf. Russell, 1968) or, more likely, of selective crushing and abrasion of the smallest sizes carried as bedload (cf. Shaw and Kellerhals, 1982). Obvious deficiency in this size range is not exhibited in the frequency histograms for the heterogeneous, unstratified gravel within the eskers (Fig. 11). The preservation of this grain size range in esker and subaqueous fan sediments may have been the result of relatively rapid rates of sedimentation in these systems.

In flume experiments, longitudinal sediment sorting was observed in poorly sorted gravel (cf. McEvoy et al., 1975; Iseya and Ikeda, 1987; Whiting et al., 1988). Here, gravel travelled as sheets with distinctive longitudinal sorting related to differential transport velocities. Large clasts had higher velocities than smaller ones, where shear velocities were well above critical values for all grain sizes present (cf. Meland and Normman, 1969; Shaw, 1969). Gravel sorting with a congested, openwork zone, a smooth, matrix-rich zone and a transitional, half matrix-filled zone resulted (Iseya and Ikeda, 1987). This sequence was produced in flume experiments under steady flow discharge but involving non-uniform and unsteady bedload transport. If the rate of sediment deposition was relatively rapid, such longitudinal sorting may have been preserved in the sedimentary record. In vertical section, such a mechanism may have produced a vaguely lenticular organisaton in the resulting gravel facies. In addition, the structure may have been complicated by pulses in flow velocity during a depositional event. This is perhaps corroborated by the transport data for unit C30/7–1 (Table 7), where cobbles have dominant a(p) orientations and pebbles have dominant a(t) orientations. This suggests that cobbles were primarily transported in suspension, while pebbles were transported as traction bedload. Clearly, if flow discharge was steady, this would not be sensible. The apparent anomaly may be accounted for by unsteady flow or, alternatively, by reorientation of clasts about obstacles as they came to rest (cf. Johansson, 1963, 1976).

An alternative to the relatively regular sorting produced by the above mechanism is reported in the literature on cluster bedforms (cf. Brayshaw et al., 1983; Brayshaw, 1984, 1985). Here, obstacle clasts act as the focus for entrapment of clasts which become tightly packed and imbricate on the stoss side of obstacles. Finer material "hides" in the wake of obstacle clasts (downflow separation bubble on their lee side; Brayshaw et al., 1983). Such cluster bedforms have been reported to vary in length from 0.1–1.2 m, and be twice as wide as they are long (Brayshaw, 1984). Clast clusters are generally offset from one another downflow, less frequent towards channel walls, more frequent on bars, and are reported to vary in frequency downflow from zero to over 4 clusters per m² (Hassan and Reid, 1990). Such clusters tend towards an equilibrium spacing for the prevailing flow conditions and are perhaps a gravel-bed analog for the ripple and dune bedforms of sand-bed streams (Hassan and Reid,

<table>
<thead>
<tr>
<th>Table 6</th>
<th>Grain size statistics from heterogeneous, unstratified gravel</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sample number</td>
<td>Graphic mean (φ)</td>
</tr>
<tr>
<td>T19/6–7</td>
<td>-3.90</td>
</tr>
<tr>
<td>T19/9–1</td>
<td>-3.53</td>
</tr>
<tr>
<td>T23/1–1a</td>
<td>-4.72</td>
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<tr>
<td>T23/1–1b</td>
<td>-0.07</td>
</tr>
<tr>
<td>T23/1–1c</td>
<td>-2.28</td>
</tr>
<tr>
<td>C43/1–1</td>
<td>-2.12</td>
</tr>
</tbody>
</table>

Statistics after Folk and Ward (1957).
<table>
<thead>
<tr>
<th>Unit number</th>
<th>Sedimentary facies/structure</th>
<th>Sample number</th>
<th>Flow azimuth (°)</th>
<th>Mean plunge (°)</th>
<th>Vector strength (S1)</th>
<th>Significance level (%)</th>
<th>Transport orientation data</th>
<th>pebbles</th>
</tr>
</thead>
<tbody>
<tr>
<td>T24/1-1</td>
<td>Heterogeneous, unstratified gravel</td>
<td>60 221 49</td>
<td>0.5137</td>
<td>99.0</td>
<td>48.3 43.3 a(t) a(p)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>C43/1-1</td>
<td>Heterogeneous, unstratified gravel</td>
<td>60 260 57</td>
<td>0.6520</td>
<td>99.0</td>
<td>53.3 41.7 a(t) a(t)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>C30/7-1</td>
<td>Heterogeneous, unstratified gravel</td>
<td>59 230 37</td>
<td>0.4708</td>
<td>97.5</td>
<td>47.5 50.9 a(t) a(t)</td>
<td></td>
<td></td>
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<tr>
<td>C32/3-5</td>
<td>Heterogeneous, unstratified gravel</td>
<td>25 206 28</td>
<td>0.8665</td>
<td>99.0</td>
<td>N/A N/A N/A N/A N/A</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>N54/1-7</td>
<td>Heterogeneous, unstratified gravel</td>
<td>25 202 42</td>
<td>0.8112</td>
<td>99.0</td>
<td>N/A N/A N/A N/A N/A</td>
<td></td>
<td></td>
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</tr>
<tr>
<td>N35/2-2</td>
<td>Heterogeneous, unstratified gravel</td>
<td>30 209 39</td>
<td>0.8004</td>
<td>99.0</td>
<td>N/A N/A N/A N/A N/A</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>N36/1-1 (1990)</td>
<td>Heterogeneous, unstratified gravel (with higher% fines)</td>
<td>60 302 14</td>
<td>0.4233</td>
<td>90.0</td>
<td>53.3 43.3 a(t) a(t)</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>C31/6-2</td>
<td>Massive, imbricate, clast-supported gravel</td>
<td>60 188 50</td>
<td>0.8312</td>
<td>99.0</td>
<td>56.7 40.0 a(t) a(t)</td>
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<td></td>
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<tr>
<td>N54/1-5</td>
<td>Massive, imbricate, clast-supported gravel</td>
<td>25 215 51</td>
<td>0.7777</td>
<td>99.0</td>
<td>N/A N/A N/A N/A N/A</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>N36/18-1</td>
<td>Massive, imbricate, clast-supported gravel</td>
<td>40 115 42</td>
<td>0.8163</td>
<td>99.0</td>
<td>55.0 45.0 N/A N/A N/A</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>TJS1/1-3</td>
<td>Horizontally bedded gravel</td>
<td>60 194 35</td>
<td>0.6306</td>
<td>99.0</td>
<td>63.3 25.0 N/A N/A N/A</td>
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<td></td>
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<tr>
<td>C31/10-1</td>
<td>Horizontally bedded gravel</td>
<td>60 191 47</td>
<td>0.5303</td>
<td>99.0</td>
<td>41.7 55.0 N/A N/A N/A</td>
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<tr>
<td>TJS1/2-3</td>
<td>Pseudooticinal macroform (west)</td>
<td>60 155 50</td>
<td>0.7955</td>
<td>99.0</td>
<td>73.3 21.7 N/A N/A N/A</td>
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<td></td>
<td></td>
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<tr>
<td>TJS1/2-1</td>
<td>Pseudooticinal macroform (east)</td>
<td>60 255 45</td>
<td>0.7258</td>
<td>99.0</td>
<td>61.7 31.7 N/A N/A N/A</td>
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<tr>
<td>T14/1-3</td>
<td>Pseudoocticinal macroform (east)</td>
<td>60 277 55</td>
<td>0.6208</td>
<td>99.0</td>
<td>55.0 35.0 a(t) a(t)</td>
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<tr>
<td>N36/22-1</td>
<td>Pseudoocticinal macroform (east)</td>
<td>30 297 37</td>
<td>0.5247</td>
<td>90.0</td>
<td>40.0 40.0 a(t) a(p)</td>
<td></td>
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<td></td>
</tr>
<tr>
<td>N36/22-5</td>
<td>Pseudoocticinal macroform (west)</td>
<td>100 120 39</td>
<td>0.5760</td>
<td>99.0</td>
<td>52.0 44.0 none a(t)</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>M6* /1-centre</td>
<td>Oblique accretion avalanche bed macroform</td>
<td>60 270 13</td>
<td>0.4078</td>
<td>90.0</td>
<td>38.3 55.0 N/A N/A N/A</td>
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<td></td>
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<tr>
<td>M6* /1-LHS</td>
<td>Oblique accretion avalanche bed macroform</td>
<td>59 236 44</td>
<td>0.4565</td>
<td>95.0</td>
<td>55.0 41.7 N/A N/A N/A</td>
<td></td>
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<td></td>
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<tr>
<td>N36/7-6 (1989)</td>
<td>In-phase wave structure, gravel</td>
<td>50 88 25</td>
<td>0.7529</td>
<td>99.0</td>
<td>N/A N/A N/A N/A N/A</td>
<td></td>
<td></td>
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<tr>
<td>N36/7b-1</td>
<td>In-phase wave structure, gravel</td>
<td>60 107 25</td>
<td>0.4366</td>
<td>99.0</td>
<td>50.0 40.0 N/A N/A N/A</td>
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<tr>
<td>N36/7b-9</td>
<td>In-phase wave structure, sand</td>
<td>30 53 27</td>
<td>0.5737</td>
<td>99.0</td>
<td>39.1 52.2 a(t) a(p)</td>
<td></td>
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<tr>
<td>N36/17-8</td>
<td>In-phase wave structure, gravel</td>
<td>100 122 11</td>
<td>0.4707</td>
<td>99.0</td>
<td>49.0 38.0 a(p) a(t)</td>
<td></td>
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<td></td>
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<tr>
<td>N36/19-5</td>
<td>In-phase wave structure, gravel</td>
<td>60 116 29</td>
<td>0.5508</td>
<td>99.0</td>
<td>62.7 23.7 equal a(t)</td>
<td></td>
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</tr>
<tr>
<td>N36/17-10</td>
<td>Water escape structure in a gravel in-phase wave structure</td>
<td>30 15 29</td>
<td>0.6191</td>
<td>99.0</td>
<td>30.0 70.0 N/A N/A N/A</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

N/A: no data available.

* Significance level of sample being non-random, according to test statistic ($S_1 / S_2$) of Woodcock and Naylor (1983).

b Dominant a-axis orientation for cobble and pebble grain sizes.
Clusters are believed to be deposited during the waning stages of floods (Brayshaw, 1984). Their break-up can produce short-term pulses in bedload transport (Brayshaw, 1985), as they tend to delay incipient motion and limit the availability of bed material for transport (Brayshaw, 1984). Preservation of this bedform in vertical section, where sedimentation rates were rapid, may also have been responsible for producing the vaguely lenticular organization of the heterogeneous, unstratified gravel facies. Longitudinal sediment sorting, and sorting attributed to the development of cluster bedforms, are the most probable mechanisms responsible for the formation of the heterogeneous, unstratified gravel facies.

The coarse grain size of heterogeneous, unstratified gravel suggests relatively powerful flows. Entrainment velocities \( (U) \) on the order of 2.1–14.5 m s\(^{-1}\) are estimated for the largest clasts in this facies \((-10\phi \text{ diameter}; \text{cf. Williams, 1983})\). These velocities are comparable to those reported for the drainage of ice-dammed lakes (cf. Elfström, 1987; Lord and Kelew, 1987). Assuming closed-conduit conditions, an estimated conduit cross-sectional area \( (A_{\text{con}}) \) of 8849 m\(^2\) (where height \( \sim 30 \text{ m} \) and half width \( \sim 200 \text{ m} \), estimated from the maximum width of Norwood esker) for a conduit with the geometry of a segment defined by the space between the arc of a circle and its chord (cf. Hooke et al., 1990), and the continuity equation \( Q_{w} = U \times A_{\text{con}} \); discharge \( (Q_{w}) \) is calculated at \( 1.3 \times 10^{5} \text{ m}^{3} \text{ s}^{-1} \), if \( U = 14.5 \text{ m s}^{-1} \), during the events responsible for depositing the coarsest members of this facies. The polymodal, framework-dominant character of the facies suggests that although all grain sizes were available in transport, most fines were transported away from the sites of heterogeneous, unstratified gravel deposition. Banerjee and McDonald (1975) reported an esker ridge gravel with a disrupted framework, which may be similar to the heterogeneous, unstratified gravel facies described here.

### 7.1.2. Massive, imbricate, clast-supported gravel

This facies is texturally and structurally gradational with the heterogeneous, unstratified gravel.

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Fig. 12. (a) Massive, imbricate, clast-supported gravel alternating with sand (truncated at arrow), Tweed esker (pit T12; Fig. 2). Metre rod for scale. (b) Imbricate, polymodal, matrix-rich gravel within an in-phase wave structure, Norwood esker (unit N36/7–6 (1989); Table 7). Metre grid for scale. (c) Horizontally bedded gravel, Campbellford esker (pit C31; Fig. 2). Tape extended to 0.5 m.
facies. It is dominated by boulders (< 10\(\phi\) diameter; Fig. 12a) and cobbles or cobbles, with a matrix of small pebbles, granules and sand, and is largely bimodal and clast-supported (Fig. 12a). Unit thicknesses are generally less than 2 m. Imbrication is pervasive rather than clustered (Fig. 12a). There is a mixture of \(a(t)\) and \(a(p)\) clast imbrication, with \(a(t)\) being dominant (Table 7). Mean plunge angles are high (> 40\(^\circ\); Table 7), and imbrication is strong (Table 7). Units are generally ungraded, although one exhibits inverse grading (unit C31/6–2, Table 7).

This facies is similar to the heterogeneous, unstratified gravel facies, and is also interpreted as the product of deposition from fluidal flows. However, there is less matrix, stronger and more pervasive imbrication, and a more massive structure than for the heterogeneous, unstratified gravel. These differences suggest that transport was primarily by traction, and that deposition was perhaps more gradual for massive, imbricate, clast-supported gravel than for heterogeneous, unstratified gravel. Traction transport is consistent with the observation that both cobble and pebble \(a\)-axis orientations are transverse to flow direction. Some transport by suspension and saltation is recorded by \(a(p)\) clasts. Dispersive pressure within a suspension may account for the inverse grading observed in unit C31/6–2 (Table 7). Alternatively, and more probably, this coarsening-upwards sequence may represent the rising limb of a flood hydrograph. Entrainment velocities of 2.1–14.5 m s\(^{-1}\) are estimated for clasts of −10\(\phi\) diameter (cf. Williams, 1983).

### 7.1.3. Imbricate, polymodal, matrix-rich gravel

Visually, this facies is structurally and texturally gradational with the previous two facies (Figs. 10, 12a, 12b). In general, facies architecture is tabular in lateral, upflow portions of the ridge, while it is more lenticular in downflow locations, and has a pronounced undulatory surface (in-phase wave structure) in proximity to the major downflow fans (pit N36; Fig. 2). Unit thicknesses rarely exceed 1.5 m, except in the major fans. The facies may be ungraded or weakly inverse-to-normally graded (Fig. 12b). An increased proportion of sand effects a matrix-supported character, although the facies is texturally polymodal (Fig. 12b). Maximum clast size does not usually exceed −6.5\(\phi\) diameter.

This facies is inferred to have been rapidly deposited from a highly concentrated dispersion (cf. Smith, 1986; Costa, 1988). Entrainment velocities of 0.6–4.4 m s\(^{-1}\) are estimated for clasts of −6.5\(\phi\) diameter (cf. Williams, 1983). Further observations and interpretations are presented in the section on in-phase wave structures.

### 7.1.4. Horizontally bedded gravel

This is the least frequently observed facies in the main esker ridges. Observed unit thicknesses vary from 1.5–4 m. Beds are variously polymodal to graded, with horizontal stratification (Fig. 12c). Stratification is less well developed (Walker, 1975; Smith, 1990) in the coarser units. This facies is dominated by small boulders (< 8.2\(\phi\) diameter) with cobbles and pebbles, cobbles with pebbles (Fig. 12c), or pebbles with granules. Pebbles, granules and sand fill the void spaces (Fig. 12c). Clast imbrication is a mixture of \(a(t)\) and \(a(p)\) (Table 7), with mid-range vector strengths (Table 7), and mean plunge angles comparable to those of heterogeneous, unstratified and massive, imbricate, clast-supported gravels (Table 7).

Stratification within this facies is indicative of deposition from traction transport in fluidal flows. Entrainment velocities of 1.1–8.0 m s\(^{-1}\) are estimated for clasts of −8.2\(\phi\) diameter (cf. Williams, 1983). Horizontal bedding may be attributed to the downflow migration of bed waves which formed by interaction between eddies in the flow and the bed (Allen, 1984a), or to the effects of the burst/sweep process on local rates and modes of sediment transport in a turbulent flow (cf. Cheel and Middleton, 1986). Transport in a turbulent suspension prior to deposition is inferred for unit C31/10–1 (Table 7), where \(a(p)\) orientations are dominant.

### 7.1.5. Cross-bedded gravel

Trough and tabular cross-bedded gravel (Figs. 13a, 13b) is observed in all of the main esker ridges, and more frequently with distance downesker. Individual foreset beds are generally < 50 cm thick, cross-bed sets are generally < 2.5 m
thick, and cosets are generally < 3 m thick. Foresets dip at approximately 26° to 30° downflow. Cross-bedded gravel is dominated by cobbles and pebbles (Fig. 13a), with small boulders (< −8.2 φ diameter) at the base of some rhythmically graded foreset beds (Fig. 13b). A typical rhythmically graded bed exhibits bimodal, clast-supported boulders, cobbles or pebbles with a sandy matrix, which often exhibits convolute laminations, at the base (part A; Figs. 13c, 14). Part A usually accounts for 50–60% of the bed thickness, and passes up-sequence into relatively openwork cobbles or pebbles (part B; Figs. 13c, 14), then into openwork granules (part C; Fig. 13c). Occasionally, foreset beds are polymodal with poorly defined cross-bedding. However, larger clasts often occur in imbricate clusters towards the base of such foreset beds. Similar rhythmically graded

Fig. 13. (a) Trough cross-bedded gravel, Campbellford esker (pit C43; Fig. 2). Flow out of face. (b) Rhythmically graded foresets of tabular cross-bedded gravel, Tweed esker (pit T16; Fig. 2). Flow to the right. Metre grid for scale. (c) Rhythmically graded gravel triplet, Campbellford esker (pit C30; Fig. 2). Note convoluted laminations in sand, part A (bottom). Scale in cm. (d) Rhythmically graded gravel with tabular architecture on the scale of the esker ridge, Tweed esker (pit T26; Fig. 2). Tape extended to 0.5 m.
gravel sequences, but in vaguely delineated lenticular packages, are observed within single gravel units on the scale of the esker ridges. Such units are up to 10 m thick and continue laterally over the width of the esker, while rhythmic packages are 0.5–1.0 m thick and 1.0–2.0 m wide. Some packages have downflow-inclined lower contacts (cf. Shulmeister, 1989).

The cross-bedded gravels are products of bedform migration (cf. McDonald and Vincent, 1972). Traction was the dominant transport mechanism. Longitudinal sediment sorting during transport of heterogeneous gravels, and lee-side deposition of suspended load in return flow beneath a separation eddy, contributed to the characteristic foreset grading (cf. Shaw and Gorrell, 1991). Shaw and Gorrell (1991) described a basal matrix-supported gravel and reported a concomitant fining of clasts and a coarsening, or decrease in the volume of matrix, up-sequence in each foreset bed within large subglacially formed dunes. Such gradual up-sequence changes are not observed within the foreset beds in south-central Ontario eskers; rather, a basal clast-supported gravel passes up-sequence into openwork gravel. Overall, the cross-bedded gravel facies (in addition to heterogeneous, unstratified and massive, imbricate, clast-supported gravel facies) in south-central Ontario eskers is relatively matrix-poor. The constrained nature of flow within conduits may be responsible for the lack of fines, as fines remained in suspension and were transported downflow. The occasional occurrence of a polymodal gravel foreset bed may indicate that the heterogeneous gravel mixture feeding the foreset had only poorly developed longitudinal sorting. In addition, continuous deposition from bedload and suspended load may have produced the polymodal texture (Shaw and Gorrell, 1991).

Continuous upslope flow paths for the eskers of south-central Ontario suggest closed-conduit conditions during esker formation. Flow depths of 5–25 m at the time of dune formation are estimated from dune heights of 2.5 m (cf. McDonald and Vincent, 1972). Entrainment velocities are estimated to have ranged from 1.1 to 8.0 m s⁻¹, for clasts of −8.2 φ diameter (cf. Williams, 1983).

Traditionally, bimodal gravel has been interpreted as a sliding bed facies, associated with high-Qw events in eskers (e.g. Saunderson, 1977, 1982; Ringrose, 1982; Lindström, 1985; Shulmeister, 1989), or as the product of a change in flow power, such that there was initial deposition of gravel and later deposition of sand in open-channels (e.g. Baker, 1973; Smith, 1974). Openwork gravel has been interpreted as a winnowed (e.g. Lundqvist, 1979) or waning flood (e.g. Shulmeister, 1989) deposit. However, rhythmically graded, vaguely lenticular gravel packages with downflow-inclined lower contacts, within units on the scale of an esker, may be more appropriately attributed to deposition in the lee of a bedform (cf. Anketell and Rust, 1990; Shaw and Gorrell, 1991), a nega-
Fig. 16. Oblique accretion avalanche bed macroform, Marlbank esker (pit M42; Fig. 2). (a) High-angled inclined avalanche beds truncated by lateral scour-and-fill structures. Primary flow was into the photograph. Metre rods for scale. (b) Graded gravel within inclined avalanche beds. Metre grid for scale.
tive step (cf. Carling and Glaister, 1987), or a macroform, and related to longitudinal sediment sorting during transport of heterogeneous gravel to the brink of the form (cf. Shaw and Gorrell, 1991). Alternatively, rhythmic grading within these downflow-inclined packages may have been a product of the migration of smaller gravel bedforms, themselves possessing rhythmically graded foreset beds, over the brink of a larger bedform or macroform (cf. Anketell and Rust, 1990).

Similar grading in tabular beds arranged in sets on the scale of an esker, and observed in sections cut perpendicular to flow direction (Fig. 13d), may also be interpreted as cross-beds related to bedform migration or macroform progradation. However, without observation of downflow-inclined lower contacts between beds, it is equally possible that such units could be graded, horizontally bedded gravels (cf. Cheel and Middleton, 1986; Smith, 1990).

7.2. Pseudoanticlinal macroforms

Pseudoanticlinal macroforms (or coarse gravel anticlinal macroforms, Garbutt, 1990) are composed of low-angled, arched or anticlinal bedding in gravel up to $-10^\phi$ diameter, on the scale of the main esker ridges (Fig. 15, p. 26). They are generally observed where a ridge is narrow and geometrically uniform. Facies within this macroform include heterogeneous, unstratified gravel (Fig. 15) and massive, imbricate, clast-supported gravel. Occasionally, continuous or discontinuous sand units alternate with gravel facies to form this macroform. Palaeoflow azimuths, inferred from gravel fabrics, are downflow and convergent on the crest of the anticline (Fig. 15; Table 7). Clast orientations are dominantly $a(t)$ (Table 7).

Powerful flows were required to move the boulders and cobbles which make up this macroform. Entrainment velocities of 2.1–14.5 m s$^{-1}$ are estimated for clasts of $-10^\phi$ diameter (cf. Williams, 1983). The formation of a primary pseudoanticlinal macroform with crest-convergent fabrics is inferred to have been the product of secondary currents or vortices. Secondary currents have been reported for flow in closed-conduits (Rouse, 1961). Here, circulation was set up in the plane of a cross-section and superimposed on primary longitudinal flow; mean motion formed a spiral or vortex moving downflow. While double helicoidal vortices can form in long, straight, deep, narrow, open-channel flow (Mathies, 1947), closed-conduit conditions have been inferred during the formation of the main esker ridges from continuous, upslope crest long-profiles (Fig. 3). Pseudoanticlinal macroforms are observed where the esker ridges are fairly narrow and geometrically uniform, and not where they pinch and swell, laterally and vertically. By inference, narrow subglacial closed conduits of uniform geometry may have been a prerequisite for the maintenance of strong secondary vortices (cf. Shreve, 1972, 1985; Garbutt, 1990).

Dominance of $a(t)$ clast orientations is consistent with observations from both heterogeneous, unstratified gravel facies, and massive, imbricate, clast-supported gravel facies which form the bulk of this macroform. Tractive forces generated by fluidal secondary flows, as they converged on the crest of the anticline, are inferred to be responsible for this orientation.

7.3. Oblique accretion avalanche bed macroforms

Oblique accretion avalanche bed (OAAB) macroforms exhibit high-angled inclined surfaces ($\sim 30^\circ$), or avalanche beds, dominated by pebbles...
(< -6\phi diameter; Fig. 16). They were observed only in the Marlbank esker. Morphologically these macroforms occur at zones of slight expansion along an otherwise relatively continuous, narrow esker ridge.

At pit M6\textsuperscript{*} (~6 m vertical exposure; Fig. 2), the angle of inclination of the beds decreases upward (32–14\degree dip towards ~280\degree orientation) and reactivation surfaces are observed. Fabrics from different locations within the structure exhibited both weak \(a(t)\) and \(a(p)\) imbrication, downflow and oblique to the esker ridge crest (Table 7). Facies within this structure include polymodal gravely sand (Fig. 17), openwork cobbles and openwork pebbles. Immediately lateral to the gravel avalanche beds, and at a lower elevation, the sedimentary sequence (~2 m thickness) is composed of fining-upward sets (0.25–0.30 m thick) of massive, cross-laminated, parallel-laminated and penecontemporaneously deformed, coarse or medium sand to silt or clay.

At pit M42 (~8 m vertical exposure; Fig. 2), steeply inclined avalanche beds (~30\degree) are observed directly on bedrock (Fig. 16a). Each bed is 0.5–1.5 m thick and composed of graded pebbles and granules (Fig. 16b) with a maximum grain size of ~5.6\phi diameter. The grading in each bed is relatively uniform, making differentiation of the top and bottom difficult (Fig. 16b). Three gravel structures are observed within most avalanche beds: polymodal, matrix-rich pebbly sand; polymodal, framework-supported pebbly sand; and openwork pebbles (Fig. 16b). If this sequence is correctly ordered, the proportion of matrix decreases up-sequence in each avalanche bed. The \(ab\)-planes of the larger clasts tend to parallel the avalanche beds. Scour-and-fill structures truncate these beds on the lateral flank of the ridge (Fig. 16a). Fills have a maximum grain size of small boulders (< -8.2\phi diameter).

Although structures similar to OAAB macroforms have been interpreted as esker-deltas (cf. Banerjee and McDonald, 1975; Thomas, 1984), some of their characteristics distinguish them from previously described esker-deltas. First, no obvious topset/foreset/bottomset relationships (cf. Clemmensen and Houmark-Nielsen, 1981) are observed, and fines lateral to the avalanche beds at pit M6\textsuperscript{*} (Fig. 2) may be later glaciolacustrine deposits associated with Glacial Lake Iroquois (cf. Mirynech, 1962). The avalanche beds at pit M42 (Figs. 2, 16) lie directly on bedrock; no bottomsets exist. Second, OAAB macroforms do not punctuate segments of the Marlbank esker, but rather occur at zones of morphologic expansion within a continuous ridge. Third, lateral scour-and-fill structures truncate the earlier avalanche beds and run parallel to the esker ridge (Fig. 16a).

Flow separation and expansion of a secondary current vortex, from the main conduit into a lateral cavity, is inferred for the formation of OAAB macroforms. Avalanche beds are interpreted as prograding avalanche fronts into a subglacial cavity connected to the main conduit. Similar large avalanche beds have been reported within flood related expansion bars (Baker, 1973). Sliding and rolling of clasts down the avalanche surfaces may have contributed to weak \(a(t)\) and \(a(p)\) clast orientations parallel to the inclined beds observed at pit M6\textsuperscript{*} (Fig. 2; cf. Johansson, 1976). Fluctuation of discharge over time may have accounted for the crudely developed inverse grading of avalanche beds at pit M42 (Figs. 2, 16). Alternatively, the grading may be attributed to the delivery of longitudinally sorted sediment to the brink of the macroform as described previously for gravel cross-beds (cf. Shaw and Gorrell, 1991).

Lack of bottomset beds at pit M42 (Figs. 2, 16) is interpreted to be the result of a relatively rapid rate of cavity opening or exposure, and rapid sedimentation and progradation in a subglacial environment. Fine sediment would have been carried off downflow within a continuous conduit. The lateral scours may have been eroded during higher flow velocities caused either by a decrease in conduit cross-sectional area during sedimentation, or by a reduction in the sediment load carried by the flow. The larger clast size in scour fills may be attributed to a higher flow velocity during a later event, or to a change in the clast size available to the flow.

Reduction in the angle of inclination of foreset beds towards the ridge crest, at pit M6\textsuperscript{*} (Fig. 2), suggests aggradiation and indicates that there may
have been a process link between pseudoanticlinal and OAAB macroforms; that is, both may
have required the operation of secondary vortices for their formation.

7.4. Rhythmically alternating sand and gravel units

A secondary signal of the hydrodynamic conditions within a subglacial conduit may be inferred
from rhythmically alternating sand and gravel units (Figs. 8, 9, 12a, 18), within the macroforms.
Gravel units may be members of any of the facies previously discussed. Although sand units are
predominantly cross-laminated, they also may be plane-bedded, cross-bedded, parallel-laminated/
draped, or massive. The contact between gravel and sand units is often very sharp (Fig. 18).
Occasionally, truncation of a sand unit (Fig. 12a) highlights the probability of amalgamated gravel
units (events) which may not be obvious from vertical sections of heterogeneous, unstratified
gravel, or massive, imbricate, clast-supported gravel. Between 1 and 7 rhythmic sequences are
observed at most exposures in the esker ridges.

Rhythmicity similar to that described above is common to many eskers (e.g. Jewtuchowicz, 1965;
Lobanov, 1967; Allen, 1971; Shaw, 1972; Banerjee and McDonald, 1975; Ringrose, 1982). It has
been attributed to changes in depositional conditions on seasonal or annual time-scales, suggesting
a supraglacial to subglacial connection in the glacial hydrologic system (cf. Banerjee and Mc-
Donald, 1975; Ringrose, 1982). However, rhythms may equally represent episodic flood deposition
caused by lake (cf. Whalley, 1971; Nye, 1976; Haeblerli, 1983) or subglacial cavity drainage (cf.
Iken et al., 1983; Kamb et al., 1985; Walder, 1986; Kamb, 1987). What is certain, is that rhyth-
micity may involve: (i) a temporal or spatial change in sediment supply (cf. Shaw, 1972); (ii) a
spatial change in flow conditions, that is, the headward growth of the main conduits and cap-
ture of smaller conduits and cavities (cf. Willis, 1990); or (iii) a temporal change in flow compe-

Fig. 18. Rhythmically alternating sand and gravel units with sharp contacts, Norwood esker (pit N36; Fig. 2). Metre rod for scale.
tence. The latter may be directly related to seasonal melting, producing a direct connection between the supraglacial and subglacial meltwater subsystems or, indirectly related by the influence of seasonal meltwater on subglacial water pressure, effecting periodic capture of subglacial, water-filled cavities. In this sense, possibilities (ii) and (iii) may be related. Alternatively, temporal change may be episodic (jökulhlaups), related to the drainage of large supraglacial lakes or subglacial water bodies.

While the migration of large bedforms and progradation of macroforms within a conduit may explain some rhythmically alternating sand and

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**Fig. 19.** (a) Schematic jökulhlaup hydrographs. 1 = sudden meltwater release associated with the drainage of a large water body; 2 = progressive enlargement of the catchment area of a main conduit associated with capture of minor conduits and cavities. Both show a rapid reduction in meltwater discharge on the falling limb (modified from Haeberli, 1983). (b) Critical shear velocities \( U_s \) for traction and suspension transport. Data for critical suspension criterion \( U_s / w > 1.0 \), where \( w \) = settling velocity and flow regime from Mamak (1964, cited in Graf and Acaroglu, 1966). Data for critical traction criterion (Shields’ criterion) from Blatt et al. (1980) and Walker (1975), and for diameter \( d > 100 \) mm, \( U_s \) calculated from \( U = 0.46d^{0.30} \) (cf. Williams, 1983) and \( 15U_s = \bar{U} \) (Walker, 1975), assuming \( U = \bar{U} \).
gravel sequences (e.g. related to flow separation zones in the lee of these forms), the pervasive occurrence of these rhythms argues for a flow dynamic control. Peaked or flood-type (jökulhlaup) hydrographs (Fig. 19a) may be invoked to explain both rhythmicity and the sharp contacts between sand and gravel units. Most deposition occurs on the falling limb of a flood hydrograph. Whalley (1971) reported a 93% reduction in meltwater discharge \( (Q_w) \) over a 10 minute period, associated with a jökulhlaup shut-off. Assuming a conduit of uniform area, such a reduction in \( Q_w \) would effect a concomitant reduction in mean flow velocity \( (\bar{U}) \), and consequently shear velocity \( (U_s) \) (Fig. 19b). A 93% reduction in \( \bar{U} \) from 11.25 m s\(^{-1}\) to 0.75 m s\(^{-1}\) would result in a reduction in \( U_s \) from 75 cm s\(^{-1}\) to 5 cm s\(^{-1}\). At \( U_s \sim 75 \) cm s\(^{-1}\), all clasts smaller than medium pebbles \((<30 \text{ mm diameter})\) would have been in suspension, large pebbles to small boulders \((30-600 \text{ mm diameter})\) would have been transported as bedload, and medium to large boulders \((>600 \text{ mm diameter})\) would have been stationary. Within minutes \( U_s \) may have fallen to \( \sim 5 \) cm s\(^{-1}\), such that material finer than medium sand \((<0.4 \text{ mm diameter})\) was in suspension, medium to very coarse sand \((0.4-2 \text{ mm diameter})\) was transported as bedload, and all gravel clasts \((>2 \text{ mm diameter})\) were stationary. Therefore, on the falling limb of a flood discharge within a subglacial conduit, a rapid reduction in \( U_s \) may have been responsible for the sharp contact between the sand and gravel units. Consequently, a sand and gravel couplet may be interpreted as the product of a single flood event.

Caution should be exercised when interpreting all apparent couplets as the product of a seasonally induced flood (cf. Banerjee and McDonald, 1975) as gravel units may have been amalgamated. In addition, it is possible that either: more than one flood event may have occurred in any one season, if the occurrence was controlled by the capture of conduits, water-filled cavities or large subglacial water bodies, due to changes in subglacial water pressure; or, no flood of appreciable magnitude may have occurred in a particular season. This emphasises the notion that esker ridge sediments are a very coarse record of sediment and meltwater discharge fluctuations within a conduit, and possibly only record the most powerful events within that conduit (Gorrell and Shaw, 1991).

7.5. Discussion: esker ridge macroforms

Subglacial conduits are unlikely to have uniform geometry; they may bend (cf. Hagen et al., 1983) and/or pinch and swell, laterally and vertically (cf. Walder and Hallet, 1979; Hallet and Anderson, 1980). Studies on slow-flowing contemporary glaciers have suggested that with an increase in meltwater discharge, the drainage system evolves from a linked-cavity network into an integrated conduit system (cf. Willis et al., 1990). Some of the bulges in pre-esker conduits may have been cavity remnants (cf. Hooke, 1989, fig. 4). Alternatively, with increased meltwater discharge, conduits may have enlarged by melting or localized flotation (cf. Gorrell and Shaw, 1991) to capture adjacent water-filled cavities with which they co-existed (cf. Iken and Bindshadler, 1986; Walder, 1986; Fowler, 1987). In this manner, the location and spacing of expansion zones may have changed over time within conduits.

Esker macroforms should not be viewed as normal fluvial bedforms (e.g. large dunes or mesoforms, cf. Hoey, 1992) migrating down a conduit of uniform geometry and responding solely to flow power (variation in the fluid dynamic regime of the boundary layer, Hoey, 1992). Rather, they are envisaged as being initially spatially controlled by the pre-existing geometry of the conduit. Deposition and augmentation of composite and OAA macroforms are inferred to have occurred in zones of conduit expansion and concomitant flow expansion; the style of the macroform having been controlled by local conduit geometry downflow from the point of expansion. Macroforms are believed to scale to flow width (cf. Hoey, 1992). Later, as the macroform developed, feedback between macroform and conduit geometries may have ensued. On occasion, progradation of macroforms may have temporarily blocked constricted portions of conduits and acted as possible internal (autogenic) controls on sediment availability and flow resistance.
(cf. Ashmore, 1991). More constricted zones of the conduits would have experienced relatively higher flow velocities than expanded zones. Pseudooanticlinal macroforms are inferred to be products of relatively constricted, geometrically uniform conduit segments. Very constricted conduit segments may have experienced erosion, or simply transportation, along their length during high discharges. Some minimal deposition may have occurred along these segments during lower meltwater discharges. Consequently, conditions along a continuous conduit are inferred to have changed from erosional to transportational to depositional downflow, controlled by conduit geometry.

Although exposure was not extensive enough to reveal the spatial dimensions of the macroforms, vertical exposures suggest they attain heights of at least 10 m; crestline undulations (Fig. 3) are on the order of 20 m maximum relief. Longitudinally, the macroforms are inferred to conform to the wavelength of crestline undulations (1–5 km; Fig. 3). However, the construction of the long-profiles (Fig. 3) was constrained by the contour interval of the topographic maps (10

Fig. 20. (a) Ball and pillow structures in the upper part of a minor fan along the Tweed esker (pit T17; Fig. 2). Scrapper is 20 cm long. (b) Tabular, gently inclined architecture of a minor fan along the Tweed esker (pit T17; Fig. 2). Arrow marks the location of (a). Shovel handle is ~ 1 m long.
m on 1:50,000 NTS sheets), so it is possible that shorter wavelengths may be differentiated on larger-scale maps.

The presence of macroforms within eskers may not be restricted to the examples from south-central Ontario reported here. Hebrand and Åmark (1989, p. 77) inferred for some Swedish eskers that "deposition took place only in restricted parts of the conduit at one and the same time". They proposed a large and orderly reduction in meltwater competence at the ice margin, with this sequence repeated time-transgressively in an up-glacier direction as the ice margin retreated. Spatial constraints on erosion, transportation and deposition within a continuous conduit, imposed by conduit geometry, may provide an alternative explanation. Hebrand and Åmark (1989, pp. 73–74, fig. 9D) also noted undisturbed, undulating bedding in longitudinal section, which can be related to esker morphology. In addition, Saunderson (1977) alluded to the presence of macroforms in the Guelph esker, Ontario, when he speculated that cross-beds may have been deposited in the lee of a core of sliding bed deposits. Bedforms related to flow expansion in a conduit have been investigated experimentally by Johansson (1976). He describes simple delta-like forms with backsets, topsets, foresets and bottomsets. Large backset beds were not observed in the esker ridges of this study, but they have been observed in the Peterborough esker (Banerjee and McDonald, 1975) and in expansion zones in the Harricana glacifluvial complex, Québec (Brennand, 1991a, 1991b).

8. Sedimentology of major and minor fans, beads with minor ridges, and laterally fining deposits

In major and minor fans, massive, plane-bedded, cross-bedded, cross-laminated (ripple-drift cross-lamination types A, B, C and S, cf. Jopling and Walker, 1968) and draped, sand, silt and clay are observed in proximal to distal, and upward-finering sequences. Cross-laminated and massive sand and silt dominate these sequences. Fining-upward rhythms decrease in thickness upward from metres to millimetres. Coarse gravels are limited to basal and proximal locations, while clay is, in general, limited to upper rhythmites, some of which may be related to later sedimentation in Glacial Lake Iroquois (cf. Mirynech, 1962). However, thin clay drapes (1–2 mm thick) are also observed in some fining-upward rhythms within fans. Towards the top of some minor fan deposits, fine sand and silt units exhibit ball and pillow structures (Fig. 20a). Microfaults, convolute laminations, flame structures, drag structures, and rip-ups are common throughout the fan deposits. Sedimentary architecture is tabular, gently inclined (Fig. 20b) or undulatory.

Similar sequences to those described above are observed in beads and laterally fining deposits. Beads tend to exhibit more rapid downflow fining, and normal and thrust faulting (pit N35; Fig. 2; cf. Banerjee and McDonald, 1975; Gorrell and Shaw, 1991). Minor ridges, distributary from the main esker ridge to the beads, are composed of rhythmically alternating sand and gravel facies which have been faulted and overfolded (pit N35; Fig. 2). Laterally fining deposits display more pervasive normal faulting than minor fans. All of these observations have been made by previous authors (e.g. Banerjee and McDonald, 1975; Diemer, 1988; Henderson, 1988; Gorrell and Shaw, 1991). Consequently, further description is limited to a new observation which has significance for the interpretation of these deposits as depositional components of continuous subglacial conduits.

8.1. Hummocky or in-phase wave structures

Hummocky or in-phase wave structures are observed: (1) at zones of morphologic expansion along the main esker ridges; (2) near the base of some minor fans and laterally fining deposits; and (3) toward the downflow ends of the main esker ridges in major fans (Fig. 21). These structures are composed of diffusely graded sand or granules (Fig. 21a) or imbricate, polymodal, matrix-rich gravel (Fig. 21b), exhibit undulatory surfaces between beds (Fig. 21), and are often draped by fine sand and silt (Fig. 21). In some cases, a single in-phase wave structure is observed; in others, in-phase wave structures are stacked and offset,
Fig. 21. In-phase wave structures within a major fan or fan complex at pit N36 (Fig. 2). (a) Differently graded sandy in-phase wave structures draped by fine sand and silt, with a scour-and-fill structure (arrow) and dispersed clasts within hummocks. (b) Gravel-in-phase wave structures draped by fine sand and silt. Note water escape/lead structure (arrow), and drag and load structures at base of gravel hummocks. Metre rods for scale.
effecting an overall repetitive lenticular geometry in vertical section (Fig. 21). These structures have similar morphologies in sections perpendicular to one another, consequently, they are inferred to be three-dimensional forms.

Sandy in-phase wave structures are 5–20 m in wavelength and 0.5–2.5 m in height (Fig. 21a). Internally, sediments are well sorted, and may be massive or diffusely graded. Diffusely graded sand exhibits diffuse or wispy laminae (Fig. 21a). In places these laminae are concordant with the upper surface, in others they are truncated by that surface. That is, in-phase wave surfaces may be erosional or aggradational. Pebbles and cobbles are dispersed, or “suspended”, within the diffusely graded deposit (Fig. 21a), often with their a-axes parallel to flow (Table 7); vector strength is weak (Table 7). Clasts are commonly concentrated along surfaces internal to the hummock (Fig. 21a). Occasionally, scours (up to 0.5 m in width) truncate an in-phase wave surface and are filled with bimodal cobbles or pebbles and sand (Fig. 21a). Fine sand and silt rafts, or soft sediment clasts, are also observed within the hummocks. In-phase wave structures are separated by fine sand and silt drapes 0.1–0.6 m thick (Fig. 21a), which may be cross-laminated, parallel-laminated, or massive, and exhibit microfaulting, flames and other penecontemporaneous deformation structures.

Gravel in-phase wave structures are up to 12 m in wavelength and 1 m in height (Fig. 21b). Truncation of these structures has produced an elongate, low-angled, lenticular architecture in vertical section (Fig. 21b). Gravel in-phase wave structures are bounded above and below by sandy in-phase wave structures at pit N36 (Fig. 2), and are composed of imbricate, polymodal, matrix-rich pebble and cobble gravel (Fig. 12b). In places low-angled beds are observed to dip upflow or downflow within the hummocks. Some hummocks exhibit inverse-to-normal or normal grading, others have clast concentrations along bedding surfaces within the hummock. Occasionally, larger clasts form weakly imbricate, a(Δ) clusters (Table 7). Soft sediment clasts and discrete openwork pebble pods (possibly representing frozen rafts of openwork pebbles) are observed within the hummocks. Again, gravel in-phase wave structures are draped by coarse to fine sand and silt in fining-upward rhythms (Fig. 21b), with cross-lamination, parallel-lamination, massive and penecontemporaneous deformation structures. In one case, clay (1 mm thick) drapes a fining-upward rhythm. Draped rhythms are often truncated and exhibit load structures associated with the formation of subsequent gravel in-phase wave structures (Fig. 21b). In another case, a water escape structure in gravel (Fig. 21b) disrupts the overlying sandy drape; clasts within it are aligned parallel to the direction of water escape, with mean plunge angles equivalent to the angle of drape disruption (Table 7).

The three-dimensional, lenticular geometry of these structures resembles that of smaller in-phase wave structures (often termed antidunes) reported for open-channels (e.g. Middleton, 1965; Shaw and Kellerhals, 1977; Rust and Gibson, 1990), coasts (e.g. Allen, 1984b; Barwise and Hayes, 1985) and turbidity currents (e.g. Skipper, 1971; Hand, 1974; Prave and Duke, 1990), and larger in-phase wave structures reported for volcanlastic sediments (e.g. Pierson and Scott, 1985; Boudon and Lajoie, 1989; Charland and Lajoie, 1989; Fisher, 1990).

With increasing flow velocity, bedforms in open-conduit flow develop from ripples to dunes, to plane beds, to in-phase waves (cf. Ashley, 1990; Cheel, 1990). It has been suggested that in-phase waves would not form under closed-conduit conditions, since surface waves are suppressed by the conduit roof (e.g. Banerjee and Mcdonald, 1975; Saunderson, 1977, 1982; Ringrose, 1982). However, an interfacial in-phase wave may form at any density interface (cf. Hand, 1974). In addition, in closed-conduit flume experiments, McDonald and Vincent (1972) reported undulations on the bed similar to the hummocky surfaces described here. They hesitantly suggested that these may have been related to standing waves.

Necessary conditions for the formation of in-phase wave structures may occur in subglacial environments where meltwater with high sediment concentrations flows from a constricted to an expanded reach within a subglacial conduit.
Hummocky surfaces may be concordant with internal laminae (drape laminae, Cheel, 1990), or truncate backset or foreset cross-laminae. In these cases, the complete lenticular deposit may be explained as the product of upper flow regime, in-phase wave deposition (and migration). For such structures to be preserved in the sediment record would have required net sediment deposition (cf. Skipper, 1971). Diffusely graded sediments, penecontemporaneous deformation structures (cf. Aario, 1971b), dewatering structures (cf. Lowe, 1975) and preservation of soft sediment rafts, within in-phase wave structures, attest to high sedimentation rates. In other cases, massive sand or gravel with little internal lamination or bedding is topped by a hummocky surface. While it may be possible that sedimentation rates were so rapid during in-phase wave formation that internal laminae or beds were not preserved, it is also possible that massive sediments were first deposited, and later eroded by in-phase waves.

The sediments within in-phase wave structures are inferred to have been deposited from hyperconcentrated flood flows (e.g. McCutcheon and Bradley, 1984; Smith, 1986; Costa, 1988). Such flows are non-Newtonian and have sediment concentrations between 40 and 80% by weight (20 and 47% by volume) (cf. Smith, 1986; Costa, 1988). In-phase wave structures have previously been reported in hyperconcentrated flood flow deposits (e.g. Pierson and Scott, 1985). Weak stratification and imbrication, framework-supported, poorly sorted gravel, and normal grading are characteristic of hyperconcentrated flood flow deposits (cf. Smith, 1986). Similarly, the presence of both weak \( a(p) \) and \( a(t) \) imbrication has been reported and attributed to grain-by-grain deposition from traction and suspension (cf. Smith, 1986). The anomalous observation of \( a(p) \) cobbles and \( a(t) \) pebbles within one gravel hummock (unit N36/17–8, Table 7) suggests that flow may have been unsteady during the deposition of this facies. Weak inverse grading towards the base of some hummocks may be explained by the sediment support mechanisms within hyperconcentrated flood flows; primary sediment support was from fluid turbulence, with secondary support from dispersive pressure, buoyancy and hindered settling (cf. Smith, 1986).

Sand and silt drapes over single in-phase wave trains or vertically stacked in-phase wave structures in sand or gravel are inferred to be the result of waning flow events, some of which may have been associated with rhythmically alternating sand and gravel units, or flood deposits, in the main esker ridges. In general, conduits enlarge by melting to accommodate increased meltwater discharge \( Q_w \) (cf. Röthlisberger, 1972; Shreve, 1972). However, during a flood event, conduits may have been unable to accommodate the rapidly increasing \( Q_w \). High water pressures, particularly at bends, would have resulted, locally exceeding the high ice overburden pressures along the margins of subglacial conduits and connecting the main conduits to adjacent water-filled cavities, with or without minor connecting conduits. As meltwater with a high sediment concentration expanded into a lateral cavity, it may have also undergone transition from supercritical to subcritical flow. This is inferred to have created the hydraulic jump \( F_{d} \approx 1 \) necessary to form in-phase wave trains. Gorrell and Shaw (1991) have inferred similar denser wall jets with hydraulic jumps to explain the morphology and sedimentary characteristics of subaqueous (subglacial) fan and bead deposits. Indeed, they showed diffusely graded sand with in-phase wave surfaces filling a scour (Gorrell and Shaw, 1991, fig. 17.A.). Jensen (1988, fig. 3a) also showed in-phase wave structures within scours in a Danish esker.

Approximations for inflow velocity \( (U) \) and depth \( (h) \) are made from measurements of the wavelength \( (L) \) of in-phase waves and estimates of the density of hyperconcentrated inflow \( (\rho_1) \) and ambient fluid \( (\rho_2) \), using equations from Allen (1984b, p. 407, eqs. 10.29 and 10.30) (Table
Table 8
Estimates of flow depth and velocity for in-phase wave structures

<table>
<thead>
<tr>
<th>L (m)</th>
<th>$\rho_1 = 1.33$ g cm$^{-3}$</th>
<th>$\rho_1 = 1.80$ g cm$^{-3}$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$h$ (m)</td>
<td>$U$ (m s$^{-1}$)</td>
</tr>
<tr>
<td>20</td>
<td>1.82</td>
<td>2.10</td>
</tr>
<tr>
<td>5</td>
<td>0.45</td>
<td>1.05</td>
</tr>
</tbody>
</table>

(8). This method provides similar results to the equations of Hand et al. (1972) and Hand (1974). In-phase wave surfaces have wavelengths ($L$) of 5–20 m. Costa (1988) suggested a bulk density ($\rho_1$) range of 1.33–1.80 g cm$^{-3}$ for hyperconcentrated flood flows. The density of the ambient fluid ($\rho_2$) is assumed to be $\sim 1$ g cm$^{-3}$. Inflow velocity estimates ($U$, Table 8) fall within the range (0.6–4.4 m s$^{-1}$; cf. Williams, 1983) estimated from the maximum clast size (−6.5φ diameter) in gravel in-phase wave structures.

8.2. Interpretations / discussion

Major fan complexes are inferred to have been deposited at a subglacial grounding line (cf. Gorrell and Shaw, 1991), while a subglacial conduit/cavity environment is inferred for minor fans, beads and laterally fining deposits; all lack the pervasive deformation characteristic of sediments which have been let down from supraglacial or englacial positions (cf. Charlesworth, 1957). Where fans and beads punctuate esker segments, they have been traditionally interpreted as products of ice-marginal, time-transgressive sedimentation (cf. Banerjee and McDonald, 1975). In south-central Ontario, the main esker ridges were deposited in continuous, closed conduits, and minor fans and beads do not punctuate depositional esker segments. Downstream continuation of distributary ridges and beads as contributory ridges to the main esker, and smooth continuation of minor fans upflow and downflow with the esker ridges, argue for synchronous subglacial deposition of the main esker ridges, beads and fans. Removal of lateral ice support after deposition in a subglacial conduit may be invoked to explain faulting within beads and laterally fining deposits (cf. McDonald and Shilts, 1975).

Rapid fining of sediment in the minor fans and beads is attributed to a reduction in flow competence during flow expansion, or differential transport of coarse and fine sediment caused by rapid aggradation (cf. Gorrell and Shaw, 1991). It is suggested that as meltwater discharge declined on the falling limb of a flood hydrograph, the normal high-pressure zones at the conduit margins were re-established. These zones would have acted as the controlling valves (cf. Willis et al., 1990) which sealed water-filled cavities and minor conduits (cf. Gorrell and Shaw, 1991). Overfolding and faulting in distributary ridges at pit N35 (Fig. 2) may be explained by deformation as ice locally reattached to the bed, causing cavities and minor conduits to be pinched off from the main esker ridge at the cessation of a high-$Q_w$ event. Cessation of flow in cavities would have produced standing water bodies which may have resulted in deposition of thin clay drapes in minor fans. The fact that cavities and minor conduits remained water-filled, provided a zone of weakness such that the same flow path was reutilised during succeeding high-$Q_w$ events (cf. Gorrell and Shaw, 1991). This repetition is inferred to have produced repeated fining-upward sequences in beads and minor fans. A similar $Q_w$-controlled grounding line environment may account for the presence of, and rhythmicity within, sand and silt drapes over sand and gravel in-phase wave structures in major downflow fans (pit N36; Figs. 2, 21).

While the valve or conduit avulsion mechanism (primarily changes in meltwater discharge over time) may account for most of the pulsations observed in the sedimentary record of fans and beads (cf. Gorrell and Shaw, 1991), some may be attributed to a second mechanism. It has been suggested that within-conduit macroforms may have temporarily stored sediment, providing a spatial control on sediment availability. If macroforms indeed acted as temporary sediment storage features, some of the rhythmic sedimentary sequences in the minor, lateral and major, downflow fans, may be related to internally generated pulses (autopulses, Ashmore, 1991) in the bed-
load transportation rate (e.g. Ashmore, 1988, 1991; Gomez et al., 1989; Hoey and Sutherland, 1991; Young and Davies, 1991; Hoey, 1992).

While laterally fining deposits may simply represent deposition during the waning stages of conduit operation, in reality, similar sedimentary sequences in minor fans and laterally fining deposits make distinction between them difficult. In this paper, the distinction has been made primarily on morphologic grounds. Minor fans and beads are both intimately associated with the main ridge. It is likely that they represent a morphologic and sedimentologic continuum related to the tunnel avulsion mechanism proposed.

9. Sedimentology of anabranching reaches of the main esker ridges and hummocky, extended deposits

Few exposures are available in these morphologic elements. A section exposed in the Marlbank esker (pit M10, Fig. 2) is the only example of sediments within an anabranching reach in this study. Rhythmically alternating sand and gravel units are intensely faulted (normal and reverse). Pebbles and cobbles exhibit heterogeneous, unstratified, massive and openwork structures. Sand units are primarily massive or diffusely graded and range texturally from coarse to fine sand.

Observations from pits N53 and N52 (Fig. 2) provide some insight into the characteristics and context of extended deposits. Faulted and folded, rhythmically bedded sand and gravel units occur adjacent to the esker ridge (pit N52, Fig. 2). Facies include heterogeneous, unstratified and cross-bedded gravel, and cross-laminated and massive sand and silt. Further from the ridge, rhythmicity is still observed in the sand and gravel, but sand units are thicker (pit N53, Fig. 2).

Multiple-crested eskers and esker nets have long been reported in the literature (e.g. Crosby, 1902; Price, 1966, 1969; Shaw et al., 1989). Three main theories have been advanced for their formation. First, it has been suggested that such eskers are the product of glaciofluvial deposition in supraglacial channels (e.g. Crosby, 1902; Price, 1966). It has been proposed that topographic inversion caused by differential ice melting resulted in sediments capping an ice ridge. With time these deposits are inferred to have slid down the ice ridge flanks to form double-crested eskers. Repetition of this process is believed to result in multiple-crested eskers (e.g. Crosby, 1902; Price, 1966). This suggestion is difficult to rationalize for the eskers of south-central Ontario, as the sediments exhibit little post-formational disturbance. Second, development of kettle-holes, due to ice-block disintegration within a ridge, may have also resulted in a multiple-crested appearance (e.g. Howarth, 1971). The intricate pattern of anabranching reaches (Fig. 2) argues against such a stochastic control in this study. Third, under steady-state, closed-conduit conditions and along gently ascending esker paths, Shreve (1985) suggested that a relatively low rate of conduit roof melting during ridge aggradation may have caused flow to be concentrated at low points lateral to the main ridge. Pseudo-separation of the flow into two parallel channels, subsequent deposition in these channels and repetition of this process, may have produced multiple-crested eskers. Hooke (1984) suggested a similar mechanism for subglacial open-channel flow. In Pleistocene eskers, the flow separation mechanism has the advantage of exploiting subglacial sediment sources, rather than requiring large volumes of supraglacial or englacial sediment. While this process may be appropriate for anabranching reaches in this study, where subglacial, closed-conduit, ascending conditions have been inferred, the mechanism falls short of explaining the complex anabranching pattern.

An alternate mechanism, consistent with subglacial, continuous, closed-conduit conditions inferred for the main esker ridges, is presented here. Hummocky, extended deposits and anabranching reaches are inferred to be further responses to rapid changes in meltwater discharge ($Q_w$) and water pressure within a conduit. Shaw et al. (1989) suggested that anabranching eskers may indicate subglacial floods. If melt rates within a conduit were inadequate to accommodate a rapid increase in $Q_w$, high-pressure zones at the conduit margins may have been locally breached and a broad zone of minor conduits established.
Close proximity of these conduits may have aided further breaches between them, producing complex interconnections. Multiple conduits would have effectively increased the cross-sectional area and reduced the flow velocity. A concomitant decrease in flow competence would have resulted in rapid deposition. Some minor conduits may have become choked with sediment, while others may have opened to accommodate diverted meltwater. With a decrease in $Q_w$ during the waning stages of a high-discharge event, the normal-pressure field at the margins of the main conduit would have re-established and this conduit would have regained dominance (cf. Röthlisberger, 1972).

In some cases, the increased cross-sectional area created by a multiple conduit system may have been inadequate. With a very rapid increase in $Q_w$, larger breaches in the marginal high-pressure seal to the main conduit may have ensued. Here, localised flotation or hydraulic lifting, over a broad zone either side of the main conduit, may have occurred. This plane of separation may have been at a frozen bed. In addition, the associated high-$Q_w$ and conduit rupture may have contributed ice blocks to the flow. Meltwater flow within this broader zone may have approximated a narrow sheet. Flow expansion would have resulted in loss of competence and sedimentation. As $Q_w$ declined, the main conduit would have been rapidly re-established (cf. Nye, 1976).

In both cases, ice is inferred to have settled back down onto the bed adjacent to the main conduit and may have experienced minor forward motion. This recoupling may account for the shearing and folding deformation observed within the deposits. Subsequent melting of buried ice blocks (or frozen ground) and removal of lateral ice support on ice sheet disintegration explains the intense faulting and hummocky topography of the extended deposits. Rhythmicity in the sand and gravel units necessitates repetition of this process at the same location. The reasons for the location and repetition of channel avulsion responsible for the anabranching reaches of the main eskers, and the extended, hummocky deposits, are enigmatic at present. While the mechanism described here is not directly equivalent to that proposed for anastomosed rivers (cf. Smith and Putnam, 1980; Smith, 1983), both invoke channel avulsion.

10. Discussion: implications for esker genesis and meltwater regime

10.1. Subglacial closed-conduit flow and macroforms

By concentrating on the morphologic and sedimentologic characteristics of the main esker ridges, it has been possible to infer the mode of esker genesis. Synchronous erosion, transportation and deposition by pulsed meltwater discharges in up-sloping subglacial closed conduits has been inferred. The depositional products of these processes are the large bedforms, macroforms, and rhythmic sedimentary sequences. By extension, minor fans, beads with minor ridges, anabranching reaches of the main esker ridges, and hummocky, extended deposits have been interpreted primarily as consequences of subglacial deposition in cavities or minor conduits connected to the main conduit, in response to elevated subglacial water pressures and flood discharges through the main conduit. Sediment supply for esker formation was primarily from squeezing in of adjacent sediment, either during low meltwater discharges in the winter (cf. Shaw and Gorrell, 1991), or during a high-$Q_w$ event with high melt rates and low local pressures.

It is, perhaps, interesting to speculate on the effects of the progradation of macroforms and the migration of large bedforms to conduit margins which feed into standing water bodies. It is likely that the arrival of macroforms and bedforms at the end of a conduit would have produced a high-magnitude, episodic sediment delivery to the margin. Flow expansion into a standing water body would have resulted in a reduction in flow power and would have favoured the preservation of macroforms and bedforms at the margin. Jet inertia may have caused these forms to be transported a short distance away from the end of the conduit. Stacking of such macroforms and bedforms at a stable conduit margin may have
contributed to grounding line or ice-marginal subaqueous fan sequences. South-central Ontario eskers appear to feed into the Oak Ridges glaciofluvial complex (cf. Barnett et al., 1991). Original interpretations of sedimentary packages (Duckworth, 1979) within the complex include reference to small deltas (≈ 3 m in height); these look (cf. Duckworth, 1979, fig. 7) remarkably similar to large gravel bedforms. Interpretations presented here suggest that in cases where foresets are observed within, lateral to, or downflow from, large subglacial eskers, original interpretations of these foresets as deltas (e.g. Shaw, 1972) may need to be reconsidered.

Subglacial, closed-conduit conditions are primarily inferred here from continuous upslope paths and lack of post-depositional disturbance, particularly at the centre of the main esker ridges. No obvious sliding bed facies was observed, despite the common use of the presence of this facies as the main criterion for inferring closed-conduit conditions (e.g. Saunderson, 1977, 1982; Ringrose, 1982; Lindström, 1985; Henderson, 1988). This apparent omission requires further comment.

10.2. The sliding bed facies

The sliding bed facies has been described as a poorly sorted, massive, matrix-supported sand and gravel (Saunderson, 1977, 1982; Ringrose, 1982). It has been attributed to flows with shear stresses between those required for plane beds and heterogeneous suspensions in pipes; the result being an en bloc moving bed, during full-conduit flow (cf. Newitt et al., 1955; Saunderson, 1977, 1982). Proposed conditions necessary for sliding include excess hydrostatic pressure from proximal to distal ends of the conduit, and seepage flow through the gravel due to surface waves being suppressed by the conduit roof (cf. Saunderson, 1977, 1982). The buoyant effect of sand on gravel and dispersive pressure in a heterogenous (poorly sorted) sediment are invoked as sediment support mechanisms (cf. Saunderson, 1977). The sliding bed facies, then, has been associated with high-discharge events, probably of flood magnitude, in closed conduits (cf. Saunderson, 1977, 1982).

Unfortunately, similar massive, matrix-supported, poorly sorted facies have been reported in open-channel deposits (cf. Ringrose, 1982), in proximal outwash deposits (e.g. Boulton and Eyles, 1979), and in hyperconcentrated flood flow deposits (e.g. Lord and Kehew, 1987), posing a problem of equifinality. Saunderson (1977, 1982) suggested that the sliding bed facies may replace in-phase wave structures in closed-conduit flows. Clearly, in-phase wave structures can occur where interfacial in-phase waves are established in closed-conduit/cavity environments. Similar gravels to those described as sliding beds, and with in-phase wave surfaces, are described here and explained by deposition from hyperconcentrated dispersions, the surfaces resulting from interfacial in-phase waves. Further problems exist in the description of the sliding bed facies as matrix-supported. First, by definition, this infers a bimodal grain size. In reality, photographs of this facies display polymodal sediments (cf. Saunderson, 1977, fig. 4; Saunderson, 1982, fig. 6). This problem of semantics has had further implications, namely that bimodal, matrix-supported gravels have been, perhaps incorrectly, attributed to the sliding bed mechanism (e.g. Lindström, 1985). It is necessary to question how this facies becomes bimodal, when all sediment sizes were probably available in transport (cf. Saunderson, 1977). As discussed previously, bimodality would appear to necessitate conditions of flow separation over bedforms or macroforms, and longitudinal sediment sorting during transport. Further, with dispersive pressure as the primary sediment support mechanism, inverse grading may be expected in a sliding bed facies. Presumably, as sand shears past gravel clasts in a sliding bed (cf. Saunderson, 1977) a(p) clast orientations should result. This type of information has not been recorded for so-called sliding bed deposits.

The heterogeneous, unstratified gravel facies present in the main esker ridges in south-central Ontario is a close visual equivalent to Saunderson's (1977, 1982) sliding bed facies. Predominant a(t) clast orientations and its characteristic structure are explained by deposition from fluidal flows, where sediment support was provided by fluid turbulence and the bed.
Engineering literature indicates that sliding beds may form in closed conduits. This has been demonstrated experimentally for slurry flow through narrow pipes (e.g. Newitt et al., 1955). Its absence in the esker ridges of south-central Ontario indicates that the necessary conditions for its formation were not attained. That is, water was not forced into the bed upflow and a pressure head of sufficient magnitude necessary to drive the coarse bed was not realized.

10.3. Conduit maintenance between seasons / events

Rhythmcity in sand and gravel units suggests that multiple flood events are recorded in the main esker ridges. In addition, the ridges are relatively undeformed. If each sand and gravel couplet represents an annual flood, and more than one rhythm is observed without deformation, the conduits must have remained open between events. Further, esker deposits are often transitional to lacustrine rhythmites (cf. Hebrand and Ámark, 1989). This suggests a strong seasonal control on meltwater supply to the conduits and to supraglacial to subglacial connection in the meltwater system. In contemporary glaciers, conduits appear to close down in the winter; the system redevelops in the following spring (cf. Willis et al., 1990). Conduit closure would have resulted in deformation of esker sediments. Intense deformation was not observed; several explanations are possible. First, if conduits fed into a standing water body, as appears to have been the case for those draining the Laurentide ice sheet, they may have remained open when the meltwater supply was shut down in the winter, at least to the height of the piezometric surface formed by the water body (cf. Powell, 1990). Second, closure rate by plastic deformation of ice may have been insufficient to cause the conduit walls to impinge upon esker gravels. Third, the inferred non-uniform geometry of the conduits may have been the key. Constricted segments of the conduit may have pinched off as pressure within the conduits fell, thereby trapping water within expanded segments and creating linearly arranged water-filled cavities. Re-establishment of conduit flow between cavities in the next melt season may have been facilitated by the location of these repeatedly occupied conduits in topographically low tunnel channels (cf. Brennand and Shaw, 1994).

The explanations above assume that the rhythms approximate to single seasonal events. A divergent hypothesis is that eskers were deposited in a single year, the rhythms representing multiple flood events within a single melt season. However, the complexity and number of finer-grained rhythmic sequences in adjacent fans and beads, and downflow fans and glaciolacustrine rhythmites argue against this. In this paper, multiple discharge events have been preferred as explanation of morphologic and sedimentologic observations. Esker ridges are coarse records of the most powerful pulses, whereas fans and beads record flow events in finer detail; they are primarily depositional reservoirs.

10.4. Caveat

While the sedimentary and macroform descriptions and interpretations presented here are rigorous, as in all realist science, the necessary link to processes and controls is tenuous. The main problem lies in the dearth of research on the mechanics of sediment movement of extreme grain sizes in unsteady flows (cf. Allen, 1983) and the flow dynamics, sedimentation mechanisms and potential sedimentary structures expected in geometrically non-uniform, closed conduits. Engineering literature has furnished information on slurry flow through pipelines (e.g. Newitt et al., 1955). However, concern has been with maintaining conduit movement rather than with depositional consequences, and with uniform rather than non-uniform conduit geometries (excluding bends, Rouse, 1961). My interpretations are of necessity theoretical, although grounded in observation; we have no modern analogues.

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