8.11 Depositional Features

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Glossary

De Geer moraine A term derived from Sweden that can also be used to describe cross-valley, Rogen, and ribbed moraines that are typically subglacially squeezed forms.

De Kalb mounds These are found in the mid-west USA, a term to describe a series of subglacially squeezed moraines.

Equifinality A term used to convey the fact that many landforms or bedforms, although of different origins and with differing sediment contents, may end up looking remarkably similar in the final form.

Equilibrium It is the altitude on an ice mass that marks the point below which all previous year's snow has melted. This lower zone, of course, marks the Zone of Ablation, and the upper zone is the Zone of Accumulation on an ice mass.

Erosivity A term used to indicate the susceptibility of bedrock or sediment to erosion.

Grounding-line As ice masses approach open water, whether an ocean or a large lake, the ice due to buoyancy will begin to lift off its solid bed and float. The point in the terrain where this occurs is the grounding-line.

Kalixpinmo, Blattnik, Viekki and Pulju Moraines These are terms used in Sweden and Finland to describe a series of subglacially squeezed moraine ridge types.

Mélange A term used to express the generally vast variations in sediment content found within landforms and bedforms within glacial environments, especially those derivative of the subglacial and proglacial environments.

Polar ice mass It is commonly termed a cold-based or a dry-based ice mass – this term is used to refer to those ice masses that exist below the pressure melting point, in other words, ice masses that are frozen to their beds and in which there is no free meltwater at their base.

Rogen moraine These subglacial or submarginal transverse moraines were first named in Sweden. They are also termed cross-valley, ribbed, transverse, or washboard moraines.

Sediment rheology This refers to the deformability of a sediment in terms of its plasticity or otherwise generally as a function of particle size, porewater content, and impact of the stress applied.

Stauchmoräne A German term used to describe frontal squeeze or push moraines.

Temperate ice masses This is commonly termed warm-based or wet-based ice mass – this term is used to refer to those ice masses that exist at or above the pressure melting point, in other words, ice masses that are not frozen to their beds and in which there is free meltwater at their base.

Menzies, J., Hess, D.P., 2013. Depositional features. In: Shroder, J. (Editor in Chief), Giardino, R., Harbor, J. (Eds.), Treatise on Geomorphology. Academic Press, San Diego, CA, vol. 8, Glacial and Periglacial Geomorphology, pp. 127–140.

Abstract

Processes of transport and deposition are discussed with reference to glaciers and ice sheets. Sedimentary delivery systems (SDS) and the complex, interlinked sediment transport systems within glacial environments are discussed. Glacial landforms (drumlins, fluted moraines, MSGL, Rogen moraines, end and lateral moraines, eskers, kame terraces, and outwash fans) are described and their possible origins discussed. Many of these landforms are part of a spectrum of related bedforms. These depositional forms are reviewed within the context of soft sediment deforming bed conditions. Many subglacial bedforms result from instabilities in subglacial glaciodynamic conditions. Hydrological states within glacial environments are discussed.

Introduction

The transport of sediments within the glacial system requires an appreciation of cascading sediment delivery systems (SDS) in supraglacial, proglacial, and subglacial subenvironments. It is axiomatic that glacial deposits and landforms are end products of these complex SDS (cf. Alley et al., 1997). Sediment delivery occurs as a function of ice dynamics, the type of ice mass, basement and subjacent geology and sedimentology, the temporal and spatial variability of sediment discharge (flux), associated hydrological regimens, and topography. **Figure 1** illustrates the many and complex pathways of these transport systems.

In terms of ice dynamics and its influence on sediment transport, each local SDS depends on ice internal and basal thermal conditions whether polar or temperate. Under polar frozen bed conditions, it was once thought that little or no transport of sediment occurred other than in the supraglacial SDS. However, evidence now shows that even under polar or subpolar bed conditions, some limited transportation occurs as part of a slow deforming bed (cf. Shreve, 1984; Echelmeyer and Zhongxiang, 1987; Hallet et al., 1996; Alley et al., 1997). Immense amounts of sediment are transported in all SDS subsystems under temperate wet bed conditions (cf. Kirkbride, 2002; Ottesen et al., 2005; Benn and Evans, 2010) (**Figure 1**). With changes in ice conditions within an ice mass over diurnal, seasonal, and annual cycles, there occur changes in the volume and rate of sediment transfer. On a daily basis, as the volume of meltwater increases over the day as a function of solar radiation, transport of both bedload and suspended sediment increases within the supraglacial and proglacial subenvironments. Daily there may be a slight increase in subglacial and englacial transport rates as meltwater penetrates a thin ice mass, such as a cirque or valley glacier or the marginal edge of an ice sheet. In general, however, sediment transport in subglacial systems are likely to remain relatively steady, even over an annual cycle (cf. Truffer et al., 2000). Seasonal changes in sediment transport are influenced by local weather conditions such that when freezing temperatures prevail, supraglacial and subsequently englacial meltwater transport ceases. Under conditions of freezing temperatures, debris flows and other forms of mass movement transport are likely to continue. Debris flows may slow down and be less active and only when frost penetration to depth occurs will mass movement transport cease. As winter approaches, likewise, proglacial meltwater transport and later mass movement transport in the proglacial subsystem will terminate until temperatures again rise with the onset of spring. Annual variations in transport activity are likely to be at a much broader scope than either seasonal or diurnal changes and are a reflection of the overall activity of each individual ice mass or region of an ice mass such as an ice stream or an ice shelf (cf. Ottesen et al., 2005; Nygård et al., 2007; Quincey and Luckman, 2009). Changes in ice mass balance will affect sediment transport pathways and the rates of transport in addition to altering the dominance and importance of specific SDS, for example, as areas of proglacial environments are exposed or overrun.

Differing ice mass types contain different SDS, as shown in **Figure 1**. Valley glaciers and other confined ice masses such as cirque glaciers have distinct and well-developed supraglacial and marginal SDS, whereas ice sheets and associated ice streams commonly have limited supraglacial SDS other than at the frontal margins. Ice shelves, typically, have no supraglacial SDS and dominantly transport sediment as either frozen-on subglacial debris or englacial debris (cf. Anderson et al., 1991).

It is evident that sediment transported by ice masses reflects the underlying and surrounding bedrock geology as well as the subjacent and marginal sediments. Although ice masses may transport sediment extremely long distances of several hundreds of kilometers, much of the sediment load, for example, in the subglacial system, is derived from little more than a few kilometers up-ice perhaps as little as 10–15 km (cf. Hallet et al., 1996; Menzies and Shilts, 2002; Benn and Evans, 2010). In areas of previous glaciation, sediment is commonly sourced directly from relict deposits. Sediment from confined valley systems provides overwhelming supplies of transportable debris from the surrounding steep terrains, where debris flows, landslides, rock avalanches, and surface runoff in the summer supply considerable volumes of sediment (cf. Hewitt, 2009) (**Figure 2**). Hallet et al. (1996) have demonstrated that as basement bedrock types vary in terms of erosivity, the sediment yields for transport change drastically. Slow-moving polar ice masses sliding over hard crystalline bedrock yield many magnitudes less eroded material than fast-moving temperate ice crossing soft sedimentary rock types (cf. Colgan et al., 2002; Dühnforth et al., 2010).

It is very clear that sediment transport pathways within glacier/ice sheet systems are exceedingly complex and much remains to be understood (cf. Gurnell and Clark, 1987; Evans, 2003). Huge temporal and spatial variations in transport processes and their effectiveness and flux rates occur in all glacial subenvironments. Variations in both temporal and spatial transport result in large fluctuations in sediment flux rates. Although flux rates have been studied quite extensively in valley glacier systems (cf. Hallet et al., 1996; Kirkbride, 2002;

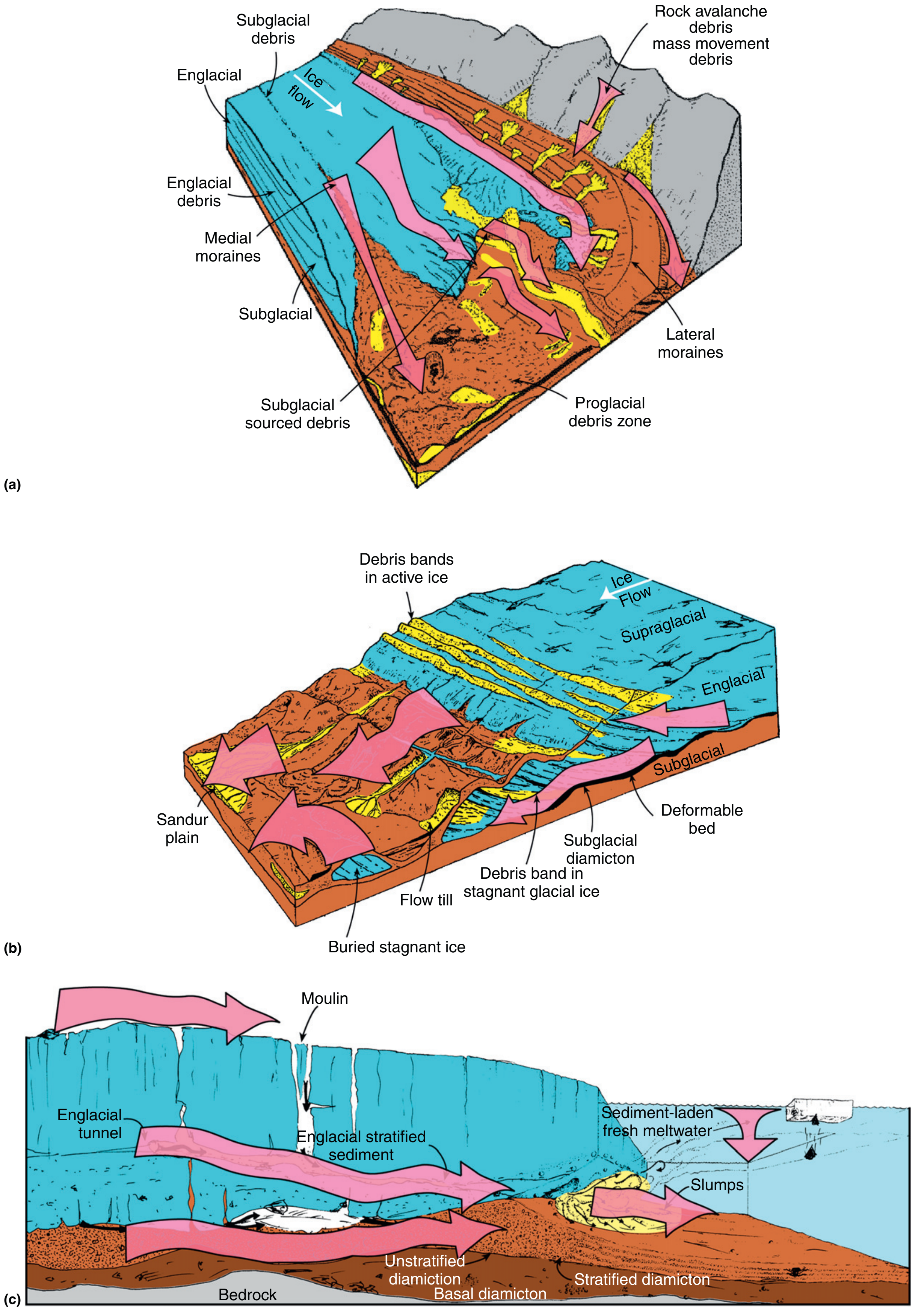
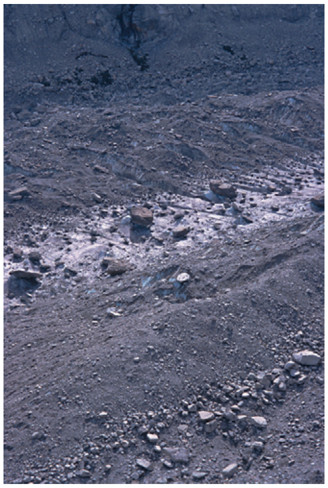


Figure 1 Sediment delivery systems (SDS) within glacial environments (a) SDS within valley glacial systems, (b) SDS within marginal ice sheet systems, and (c) SDS within subaqueous glacial environments. (a) Modified from Boulton, G.S., Eyles, N., 1979. Sedimentation by valley glaciers; a model and genetic classification. In: Schlüchter, C. (Ed.), Moraines and Varves: Origin/Genesis/Classification. A.A. Balkema, Rotterdam, pp. 11–23.

(a)



(b)

Figure 2 (a) Debris covered Bualtar Glacier in the Karakoram within the Gilgit District, Pakistan. Photo courtesy of Ken Hewitt. (b) Supraglacial debris on the Mer de Glace, France.

Benn et al., 2003) there has been limited discussion of sediment flux rates below present-day ice sheets, ice streams, and ice shelves (cf. Alley et al., 1989; Dowdeswell and Siegert, 1999; Ottesen et al., 2005; Dowdeswell et al., 2006, 2010). From modeling, it has been predicted that ice streams transporting sediments to the margin of the northern portion of the Barents Kara Sea ice sheet northern margin delivered sediment at a rate of $\sim$4 cm a^{-1} (0.13 cm a^{-1} averaged over the fan) over a 200-km-wide mouth of the Bear Island trough (cf. Dowdeswell and Siegert, 1999). It is clear, however, that fast-flowing ice streams are responsible for the bulk of sediment transfer by ice sheets (Alley and Macayeal, 1994; Dowdeswell and O Cofaigh, 2002; Ottesen et al., 2005; Dowdeswell et al., 2006, 2010; Bingham et al., 2010). Anandakrishnan et al. (2007) suggest a sediment flux rate in the order of 150 m^3 m^{-1} a^{-1} beneath the Whillans Ice Stream in Antarctica. However, great care needs to be exercised in attempting to estimate such sediment flux rates as only limited "snapshots" of subglacial conditions exist at this time. It is interesting to speculate that a relationship would seem to exist between drumlins, ribbed moraine, and rates of sediment discharge at the ice/bed interface (cf. Dunlop et al., 2008; Menzies and Hess, in press).

8.11.1 Transport

In terms of sediment transport within SDS, different pathways can be examined viz. (1) within glacier ice, (2) meltwater, (3) basally below glacier ice, and (4) by gravity (cf. **Figure 1**) (Chapter 8.10).

(1) Ice mass transport of sediment can be on the ice supraglacially, within an ice mass, englacially, or at the base of the ice and frozen on. On confined ice masses such as valley glaciers or on ice sheets along the edges of nunataks, supraglacial debris can accumulate and move down ice (cf. **Figure 1(a)**). Typically, such debris only appears on ice mass surfaces below the Equilibrium Line. This debris is generally avalanched onto the ice surface or from rock falls or other forms of mass movement and commonly reflects the processes of erosion that result in the debris arriving at the ice surface (cf. **Figure 2**). This frost-driven debris or rock fall materials are typically angular to subangular, with limited evidence of transport of any significant distance (Chapters 8.8 and 8.9).

Englacial debris, on the other hand, may enter the SDS of an ice mass at any point where ice masses move along valley sides. In valley glaciers englacial debris may be acquired from back and sidewall erosion and again, often, the debris is angular to subangular and of very short transport distances. Within ice sheets, unless nunataks are present, there tends to be a very limited if any englacial debris transport (cf. **Figure 1(b)**). Only in those marginal areas of the ice where subglacial debris flow along upward-moving glide planes do debris first move into the englacial position and at times into the supraglacial environment close to the very margin of the ice mass. Most englacial debris, other than that uplifted onto the basal layer of the ice, carries little or no evidence of glacial attrition. Where basal debris has frozen onto the basal layers of the ice and has, in some instances, then been transported to some height within the ice (as much as 20–50 m within the basal Greenland Ice Sheet, e.g.), does this debris carry very clear signatures of basal glacial erosion and comminution (cf. **Figure 1(b)**) (Chapter 8.7).

(2) Most sediments transported by meltwater within all SDS are subject to fluvial erosion in transport, resulting in most particles losing evidence of glacial surface wear. The main characteristics of meltwater transport tend to be the impact of rapid changes in meltwater discharge, and hydrostatic pressure either in englacial or in subglacial channel systems.

(3) Under many ice masses, substantial transport of mobilized sediment occurs as a result of basal ice shear stresses, causing saturated sediment to move as a layer. This layer, often referred to as a deforming or a soft bed layer of sediment, acts to some extent as a lubricating zone between the ice mass and its bed (cf. **Figures 1(b)** and **1(c)**). Transport of sediment in this layer is, as yet, only partly understood (Cuffey and Alley, 1996; Boulton et al., 2001a; Hart and Rose, 2001; Hart et al., 2011). The effect of the transport processes on such deformed sediments would appear to be relatively minor other than possible edge-to-edge grain fracture and some surface wear impacts. Transport of sediment is most likely sporadic and episodic, with sediments subjected to varying levels of stress, porewater saturation, and temperature fluctuations (van der

Meer et al., 2003; Menzies et al., 2006; Phillips et al., 2000, 2002; Bartholomaus et al., 2008).

(4) In all SDS, there are sediments transported by mass movement. Such gravitational transport occurs in many instances after initial deposition, especially on steep and unstable slopes. Mass movement in englacial tunnels is likely to be minimal. In proglacial areas, where sediments may be initially deposited at steep angles, or where buried ice causes areas to be subject to local instability on ice melting, or rapid changes in proglacial streams lead to active slope undercutting, active mass movement is pervasive other than during winter months (cf. **Figures 1(a)** and **1(b)**). The impact of transport during mass movement is relatively limited but changes in sediment clast fabrics and internal stratigraphy tend to occur.

8.11.2 Deposition

The term 'deposition' within glacial environments is fraught with issues concerning place and time. In many glacial environments, deposition occurs for brief moments to weeks to years of deposition at a time and then reworking occurs and the sediment is in transit once again before being finally deposited. The concept of primary and secondary deposition attempts to address this issue (cf. Boulton and Deynoux, 1981), which may be quite successfully used in areas of modern glacial deposition but in those areas covered by Quaternary and pre-Quaternary glaciations, such a distinction is probably of little value. When sediment comes to rest at a moment and place, one can say that deposition has occurred, however temporary that might be. In terms of 'glacial deposition,' there is another distinction that needs to be discussed, namely when is it glacial or nonglacial? When deposition occurs in direct contact with an ice mass, the sediment in question has all the characteristics of glacial erosion and transport and the term 'glacial deposition' can be correctly applied. However, since sediments deposited far from the margins of an ice mass such as on or beyond the continental shelf carry a 'glacial' signature, are these sediments also glacial deposits? At some distance from an ice source, the term glacial deposition can no longer apply but such a distinction remains somewhat elusive.

Concerning glacial depositional forms (landforms/bedforms) it seems reasonable to assert that such forms need to be formed or evolved in direct contact (more or less) with either active or passive ice masses. Many of these forms combine both glacial erosional and depositional aspects such as the streamlined form that is attained by drumlins or the commonly noted asymmetrical forms of fluted moraine where postdepositional meltwater erosion or mass movement has sculpted and altered the form.

The enormous range and types of glacial depositional landforms are well known and their typical forms and styles of development are discussed in many textbooks (cf. Sugden and John, 1976; Hambrey, 1994; Bennett and Glasser, 1996; Menzies, 1995, 1996; Benn and Evans, 2010). In past decades, these landforms have been considered unique individual landforms that require unique explanations of origin. Today, more typically, the numerous forms are perceived as a series or a subset of bedforms that have many broad similarities not only in sediment content but also forming in comparable environments, albeit, at times under different conditions of stress, temperature, sediment rheology, and ice mass conditions and glaciodynamics (cf. Aario, 1977; Rose, 1989; Menzies, 2004). It is apparent that many landforms should be considered as part of a group of associated bedforms (continuum) that have developed in response to slightly differing conditions but have remarkable similarities. Thus, when considering streamlined subglacial forms developed at the ice/bed interface, it can be construed and expected that, for example, drumlins, fluted moraines, and Rogen moraines have analogous internal sedimentology, stress histories, and some commonly shared developmental aspects. In the past, it was common to differentiate glacial landforms based on internal sedimentology, for example, stratified and unstratified sediments, or because of their formative location within glacial environments. Today, it is perhaps more accurate to subdivide glacial landforms on the basis of: (1) those forms directly attributable to the interplay of glaciodynamics and sediment availability and rheology; and (2) those more directly influenced by glaciodynamics and meltwater activity. In the former subdivision, it is also relevant to differentiate as to whether the forms have developed as ice/bed interface bedforms, parallel or transverse to ice movement, or are developed unoriented in relation to ice streaming, or are topographically ice mass marginal forms (**Table 1**).

8.11.2.1 Landforms/Bedforms Directly Attributable to Active/Passive Ice Activity

At the interface between an ice mass and its bed, fluctuations or perturbations can occur that translate into landforms/bedforms. Such perturbations are the result of interactions across this basal interface due to changes, for example, in basal ice stress conditions, basal ice velocities, thermal variables, sediment rheology, bedrock lithology and bulk strength, sediment flux rates, preexisting topographies, and interface 'roughness.' Such perturbations that can and do occur translate in some instances into bedforms that align parallel or transverse to the main flow direction of the overlying ice mass or are roughly topographically planar forms. A fundamental question that still requires an adequate solution is why do such interface perturbations occur where and when they do? Is there an inherent instability in basal ice interface conditions that results in forms being developed or evolved at various scales and orientations and why? To answer these critical questions, a better understanding of basal ice dynamics and interface ephemeral conditions and sediment rheologies is necessary (cf. Menzies and Shilts, 2002; Hart, 1997; Hindmarsh, 1999; Schoof, 2002, 2007; Dunlop et al., 2008).

Whether drumlins, fluted moraines, mega-scale glacial lineations (MSGLs), Rogen moraines, or other basal ice interface forms, there are several crucial, yet enigmatic aspects of their origins and development to be considered.

1. The patterns formed are distinctive and would appear related to the position and location beneath the ice.
2. Are the forms developed as a group over a short period of time or over longer and repeated episodes of formation?

Table 1 Landforms and bedforms associated with glacial environments

Direct glacial landforms					
Subglacial forms parrallel to ice motion	Subglacial forms transverse to ice motion	Subglacial forms non-linear – unoriented to ice motion	Ice-marginal forms (ice proximal)		
Streamlined sediments	Moraine ridges	Ground Moraine	End Moraines	Lateral Moraines	Other Moraines
Drumlins drumlinoids fluted moraines radial moraines lee-side cones	Washboard moraines De Geer moraines Rogen moraines ribbed moraines thrust moraines Kalixpinmo Hills shear moraines squeeze moraines	Corrugated moraines cover moraines till veneers hummocky moraines Pulju moraines Veiki moraines Blattnick moraines till plains squeeze moraines ice-pressed moraines (ridges)	End moraines terminal moraines recessional moraines push moraines – (Stauchmoräne) boulder belts aprons kame moraines	Perched moraines valley-sided moraines looped moraines interlobate moraines radial moraines trimline moraines	Subaqueous moraines grounding-line ridges De Geer moraines cross-valley moraines till plateau disintegration ridges De Kalb mounds

Indirect glacially induced landforms	
Subglacial meltwater forms	Submarginal/ice-marginal forms
Esker systems	Kame deposits and forms
Eskers beaded eskers engorged eskers squeeze-up eskers kame ridges esker chains crevasse fillings kettles esker fans subaqueous fans subaqueous eskers	Kames moulin kames kame plateau kame terraces kame deltas

Note: Based upon Prest (1968), Sugden and John (1976), Goldthwait (1988), Menzies and Shilts (2002).

3. Most forms are composed of a variety of sediments rather than simply till although the concept of mélange probably incorporates most sediment types available for incorporation into such forms (cf. Menzies, 1990; Hoffmann and Piotrowski, 2001).
4. Most forms developed parallel to ice motion would seem to be more or less elongated as a result of either higher ice velocities or more ductile sediment rheologies or a combination of both factors, also perhaps in relation to the bedforms' positions close to or beneath ice streams.
5. It is likely that all forms developed at the ice/bed interface are not formed or subsequently developed by the same set or a combination of processes (equifinality).
6. In many instances, overprinting or reorientation of some or all forms may occur.
7. There may well be interrelationships between these forms and other apparently unrelated forms such as drumlins and end moraines or between the interface forms and topographic slope or proximity to lakes or other large bodies of water where basal ice dynamics suddenly change.
8. Finally, is it possible that there is a relationship between the size, morphology, and shape of, at least, the parallel forms (drumlins, fluted moraines, etc.) and the sediment flux rates at the ice/bed interface?

8.11.2.1.1 Drumlins

Drumlin morphology is commonly varied and can deviate considerably from the classical tear-shaped form so commonly portrayed in textbooks (**Figure 3**). Drumlins can vary in size from a few meters in height to over 200 m in height and can stretch a few meters long to over a kilometer. Typically, they commonly form in large 'swarms' or fields many thousand in number (e.g., western New York Drumlin field >6000) (cf. Hess and Briner, 2009; Hess and Menzies, in press), but may also occur singly or in small groups (**Figures 3(a)** and **3(b)**) (cf. Menzies, 1986). Recent satellite images clearly demonstrate the apparent close relationship between drumlin fields and ice stream locations (cf. Clark, 1993; Stokes and Clark, 2001; Clark and Stokes, 2003, Fig. 9.12; Clark et al., 2003, 2009).

As noted above, in all cases of flow developed at the ice/bed interface, several current hypotheses exist as to the formation of drumlins and drumlin fields. In all cases, an 'event' or a 'trigger' appears to be necessary for their formation and development. Once an initial nucleation occurs it can be demonstrated that in some places the form will persist, grow, and possibly migrate (cf. Menzies, 1982). The problem with all drumlin formative hypotheses is what is the 'trigger' (cf. Aronow, 1959)? Currently, three broadly acceptable hypotheses exist that attempt to account for the formation and subsequent development of drumlins:

1. Deforming sediment bed (Boulton, 1987; Menzies, 1989; Smith et al., 2007).
2. Groove 'ploughing' (Tulaczyk et al., 2001; Clark et al., 2003).
3. Interface instability (Hindmarsh, 1989, 1999; Bowler, 2010a, b).

(a)

(b)

Figure 3 (a) Drumlin within the New York Drumlin field, ice direction is right to left. (b) An example of a single drumlin in front of the Biferten Glacier, eastern Switzerland. (Drumlin in center of photograph is approximately 10 m in height.)

8.11.2.1.2 Fluted moraines, and mega-scale glacial lineations (MSGLs)

As in the case of drumlins, there would appear to be a bedform association between fluted moraines, MSGLs, drumlins, and Rogen moraines (cf. Rose, 1989; Clark, 1993, 1994). Hypotheses similar to those evoked to explain drumlin origin can be advocated, except that, unlike drumlins, these other forms can extend for many kilometers, are much narrower, in general, and typically are of lower height.

Fluted moraine typically range in morphological dimensions between a few centimeters to 1 or 2 m in height to up to 50–70 m and can range from a very short distance of a few meters to several kilometers (**Figure 4**). MSGLs, on the other hand, can be several meters to over 10 m in height and extend for many kilometers in length. Both forms are generally composed of a mélange of sediments that have apparently been collectively scavenged in the process of formation.

The old hypothesis that flutes grow in the lee of boulders is generally correct (**Figure 5**) but in many cases and especially so with MSGLs, the cause of nucleation and streamlined elongation is often missing. It seems likely that the formative processes involved in both drumlins and MSGLs are very similar and any differentiation may be, in the case of flutes and MSGLs, more the result of a relatively narrowly confined, high-sediment flux rate at the ice/bed interface and the relatively high basal ice velocities as is typical for ice stream

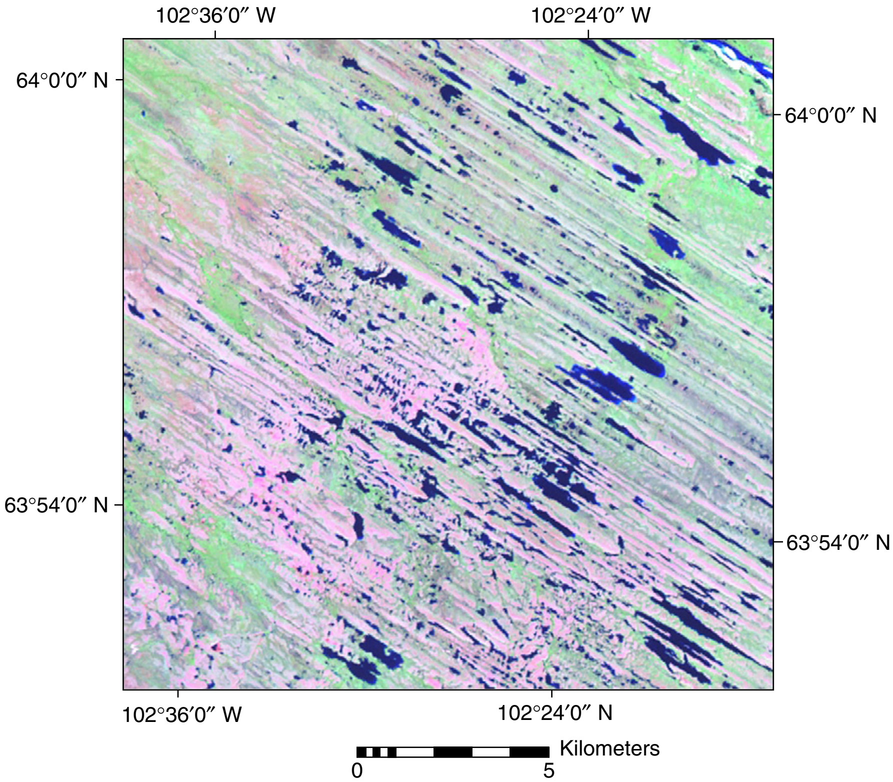


Figure 4 Landsat ETM+ satellite image of MSGL from central Nunavut, Canada. The spatial resolution of the image is 25 m. Copyright Department of Natural Resources Canada. All rights reserved.

locations (cf. Dunlop and Clark, 2006; Dunlop et al., 2008; Stokes et al., 2008; Winsborrow et al., 2010).

8.11.2.1.3 Rogen moraines

These morainal forms, in the past termed ribbed, washboard, and cross-valley moraines, are formed transverse to the dominant ice flow directions. It was probably Lundqvist (1989) who first suggested that rather than considering Rogen moraines in isolation or as unique landforms, the moraines were part of a continuum of forms either emanating from or passing into parallel fluted moraines and drumlins (**Figure 6**).

In morphology, Rogen moraines occur as ridges 10–100 m in height stretching transverse to ice flow for hundreds of meters to several kilometers. Like fluted moraine, MSGLs, and drumlins, the sediment content of these ridges is equally varied and essentially a mélange of basal available sediment.

Hypotheses of Rogen moraine origin are similar to drumlins in that both are a subglacial interface 'problem.' However, these moraines can be viewed as wave-like forms that may result from rapid transverse glaciodynamic responses such as proximal grounding-line lift-off events or instabilities inherent in the sediment flux or basal ice stress patterns (cf. Bouchard, 1989; Lundqvist, 1989; Fisher and Shaw, 1992; Hättestrand, 1997; Knight and McCabe, 1997; Dunlop and Clark, 2006; Dunlop et al., 2008).

8.11.2.1.4 Marginal moraines

Other moraines that form transverse to the ice flow direction, but not formed subglacially, may be the result of push from advancing ice or the upward squeezing of sediments at ice margins or accumulate at the ice frontal margin as end, recessional, or terminal moraines (cf. Bennett and Boulton, 1993; Krüger, 1996; Bennett, 2001; Evans and Hiemstra, 2005). Such moraines can vary in height from a few meters to several tens of meters and commonly have an asymmetric transverse profile (**Figure 7**). In some cases where the clay content is sufficiently high, the moraines may attain an almost vertical slope profile. The volume of any marginal moraine is very much a function of the residency time the ice margin is at or close to a specific location. In many instances, the ice margin may return to a particular location within the topography, thus continuing to build up the moraine over time repeatedly (cf. Krüger, 1995; Vacco et al., 2009). The sediment content of most marginal moraines reflects a wide diversity of sediment supraglacial, englacial, and subglacial. In addition, lateral and medial moraines commonly contain a large percentage of mass movement sediments that, in the case of valley glaciers, mirrors the surrounding geology of the mountainous terrain in which the particular valley glacier resides.

Other ice marginal landforms that occur within the proglacial zone may reflect the effects of episodic meltwater discharge traversing these environments, terrain collapse due to buried ice melting, glaciotectonic deformation from fontal marginal ice deformation of proglacial sediments, mass movements, or where meltwater channels undercut slopes (cf. Hart, 1990; Maizels, 2002; Evans, 2003).

Hummocky moraine is a distinctive form of moraine that occurs over large areas of submarginal and proximal proglacial


(a)


(b)

Figure 5 (a) Flute developed in the lee as a boulder on the forefield of Storbreen Glacier, Jostedal, Norway. The boulder is approximately 1.5 m in height, (b) long flutes within the New York Drumlin field.

Figure 6 Rogen moraine from near Whitbourne, the Avalon Peninsula, Newfoundland, Canada. Photograph courtesy of Tom Fisher.

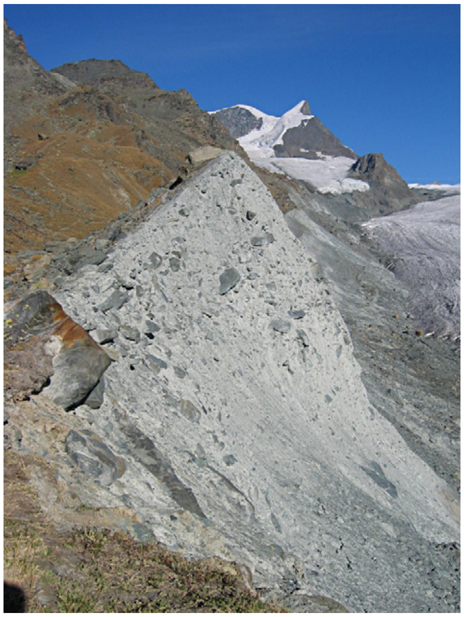


Figure 7 Asymmetrical lateral moraine of the Findelen glacier, near Zermatt, Switzerland. The boulder in the forefront of the photograph is approximately 3 m in height.

areas. Current depositional models argue that hummocky moraine was deposited supraglacially from stagnant debris-rich ice ('disintegration moraine') (cf. Bennett and Boulton, 1993; Eyles et al., 1999; Ham and Attig, 1996; Munro-Stasiuk, 1999); (**Figures 8(a)** and 8(b)). Boone and Eyles (2001), in contrast, suggest that hummocky moraine may be a product of subglacial erosion rather than supraglacial letdown during ice disintegration. Eyles et al. (1999) note that across southern Alberta hummocky moraine is composed of fine-grained till as much as 25 m thick containing rafts of soft, glaciotectonized bedrock and sediment. Much hummocky moraine is chaotic, nonoriented that appears, in places, to pass downslope into weakly oriented hummocks ('washboard moraine') that are transitional to drumlins in topographic lows. In northwest Scotland, hummocky moraine is viewed as evidence of ice-marginal disintegration possibly linked to marginal englacial stacking of debris-laden englacial shear zones that, on disintegration, collapse chaotically to form hummocky moraine (cf. Lukas, 2005). It is intriguing to speculate that many forms of hummocky moraine exist that in topographic appearance and in some cases sediment content bear remarkable similarities to each other, but it seems likely that different processes may have occurred, resulting in an equifinality of form. Ice disintegration, basal ice 'pressing,' and possibly subsequent chaotic overprinting on drumlins, fluted moraines, and Rogen moraines may all help explain hummocky moraines in different parts of the world.

8.11.2.2 Landforms/Bedforms Indirectly Attributable to Active/Passive Ice Activity

Landforms/bedforms within this category of glacial landforms are largely attributable to the influence or the impact of


(a)


(b)

Figure 8 (a) Landsat ETM + satellite image of upper Glen Turret (red dot and arrow marks the location of photograph in (b)), (b) Hummocky moraine in Glen Turret, Scotland, of Loch Lomond Readvance age. Moraine in foreground is approximately 8–9 m in height.

meltwater activity (cf. Table 1) (Chapter 8.13). A division of these forms can be made based on those formed subglacially, and others developed either submarginally or marginally to an ice mass. Although these forms may grade into other topographic features, they tend not to be part of a continuum of forms but typically occur often together. Where one finds eskers, it is not unusual to find kames and kame terraces.

8.11.2.2.1 Esker systems

Esker systems form as a function of the location and form of a meltwater channel or drainage system either subglacially or at least submarginal along the outer edge of an ice mass (cf. Veillette, 1986; Clark and Walder, 1994; Kleman and Hättestrand, 1999; Boulton et al., 2001b, 2009). Essentially an esker is the result of a sediment-choked meltwater channel (Bannerjee and McDonald, 1975; Clark and Walder, 1994; Boulton et al., 2007a, b; Hooke and Fastook, 2007; Boulton et al., 2009). Some eskers would appear to form within part or completely within an englacial conduit that, on ice melting, becomes overlaid on subglacial sediments. In some instances, eskers can occur draping drumlins in some places at an oblique angle to the drumlin long-axis orientation. Since eskers mirror the pathways taken by meltwater channels and drainage systems, they generally do not follow directly the same azimuth as the ice flow lines within an ice mass. Esker ridges (Figure 9) can be a few meters to several tens of meters in height and may run across terrain for a few tens of meters to many hundreds of kilometers. Examples from the North West Territories and Nunavut, Canada, illustrate the location and distribution of eskers on the Canadian Shield and below the central area of the Laurentide Ice Sheet during the Late Wisconsinan (cf. Menzies and Shilts, 2002, Figure 8.44). It seems likely that the location and size of eskers and entire esker patterns may be related to groundwater sources below ice sheets in which the transmissivity of the bed exerts control (cf. Boulton et al., 2007a, b, 2009). Boulton et al. (2009) have suggested that esker patterns can be deduced from basal meltwater recharge rates coupled with patterns of paleo-groundwater flow and the seasonally varying magnitude of discharge from stream tunnels at the retreating ice sheet margin. Major channel/esker systems appear to form under quasistable conditions close to the ice margin, at least over several centuries, during the retreat of an ice sheet (cf. Hooke and Fastook, 2007). The development of esker systems would appear to be interlinked with hydraulic systems supraglacially, englacially, subglacially, and crucially, within the coupled underlying groundwater systems (cf. Stenborg, 1970; Collins, 1982; Röthlisberger and Lang, 1987; Boulton and Caban, 1995; Boulton and Zatsepin, 2006).

Figure 9 Landsat ETM + satellite image of Carstairs Esker system, Scotland. Photograph courtesy of C. Zadowicz.

8.11.2.2.2 Kames and kame terraces

In most areas of ice mass melting where massive amounts of glaciofluvial sediments have been transported, kames and

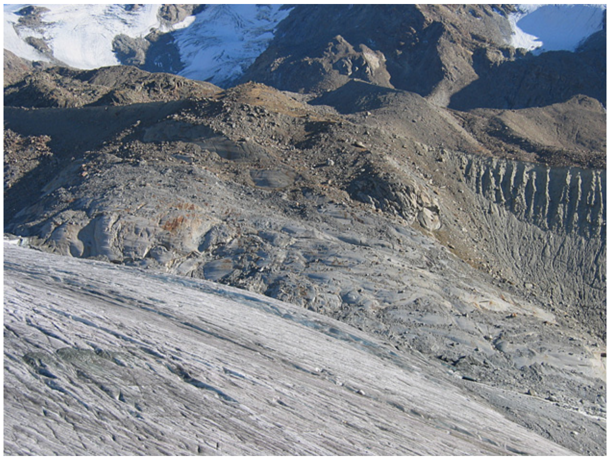


Figure 10 A series of kame terraces along the southern flank of the Findelen Glacier near Zermatt, Switzerland.

marginal kame terraces occur where terrain or slope conditions permit. Kame terraces are normally associated with valley glaciers, where the confining slopes act as a marginal route way for meltwater and transport of sediment. Commonly the slope of a valley glacier trim line is traced by a kame terrace (cf. Gray, 1995; Huddart and Bennett, 1997; Paterson and Cheel, 1997; Terpilowski, 2007; Bennett et al., 2009); (Figure 10). In some cases where crevasse filling collapses on the melting of the ice mass, kames form as roughly circular dumps of glaciofluvial sediment that exhibit marked faulting and slumping on their sides. In other instances, kames develop as unoriented accumulations of glaciofluvial sediment within subglacial and englacial cavities or abandoned meltwater channels (cf. Houmark-Nielsen et al.,1994; Ham and Attig, 1996; Huddart and Bennett, 1997). Where subglacial meltwater channels discharge into lakes, deltas of glaciofluvial sediment can build up and are termed kame deltas (nb. the Salpausselka kame deltas in Finland, Glückert, 1977).

8.11.2.2.3 Outwash fans and deltas

In most proglacial environments where large meltwater or multiple streams emanate from the frontal and lateral margins of ice masses and glaciofluvial sediments enter the proglacial zone and bed load competency declines, then characteristically outwash fans of various dimensions develop (cf. Maizels, 1989, 1991, 2002; Cutler et al., 2002). Such fans ('sandur' in Icelandic) then spread out across the proglacial zone and may extend for many tens of meters to several kilometers away from the ice mass (cf. Russell and Marren, 1999; Maizels, 2002; Russell and Knudsen, 2009); (Figure 11). Outwash fan gradients develop largely as a function of grain size such that the further down ice within the fan, the finer the sediment (cf. Maizels, 2002, Fig. 9.7). Commonly fans are pock marked with abandoned ice masses that subsequently develop into kettle holes.

8.11.2.2.4 Till deltas/tongues and grounding-lines

Where ice masses enter large bodies of water and begin to float at a grounding-line, a large tongue of till may develop that has been variously termed a till delta or a till tongue (cf. King

Figure 11 The proglacial zone of the Mont Miné Glacier, Valais, Switzerland.

et al., 1991; Larter and Vanneste, 1995; Boulton et al., 1996; King, 1996; Anandakrishnan et al., 2007; Smith and Anderson, 2010). Under conditions of subglacial soft sediment deformation, it seems likely based on research in Antarctica and on Pleistocene ice sheets that the till may emerge as a deforming unit into the water body, thus developing a wedge of till out into the bed of the water body. Likewise, under these soft bed conditions, as an ice mass retreats, a till tongue may slowly begin to from beneath the ice subglacial margin at or very close to the ground-line. At a grounding-line, there are significant changes in subglacial stress and hydraulic conditions that will lead to rapid changes in sediment rheology at that point. In many instances, major meltwater portals emerge at the grounding-line producing large marginal glaciofluvial deposits as subaqueous fans and deltas. These major changes at the grounding-line in terms of stress and hydraulics are transmitted back up-ice and affect subglacial conditions for some considerable distance back under the ice. The impact of such grounding-lines is only now being investigated but already ideas with regard to subglacial soft sediment deformation, and for example, how this bears upon drumlins and MSGL development, require investigation in the field (Le Meur and Hindmarsh, 2001).

8.11.3 Future Perspectives

The SDS in any ice mass is a complex set of interlinked and interrelated processes that have the glacial hydraulic system as the single underlying and controlling variable. Where, in the past, individual subsystems, sediment cascading processes, glaciodynamics, and stress fields have been viewed somewhat in isolation, it is clear that the hydraulic system is the unifying factor in almost, if not all, glacial processes of transport and deposition. The structuring, seasonality, spatial, and temporal episodic fluctuations in the glacial hydraulic system tend to produce characteristic structuring of other processes and properties that depend on the overall glacial hydraulic regime. In the supraglacial, englacial subglacial, or marginal proglacial environments, the integration of the hydraulic systems is crucial to understanding the full glacial system as a single, integrated, process–system entity. This integration must be investigated in terms of meltwater channel placement, size, survivability, and discharge; sediment rheology through porewater-controlled, effective stresses, sediment fluxes, the evolution of subjacent groundwater patterns; sediment shear failure, and for example, the evolution of deformational drumlins and other streamlined terrain bedforms. This integrated approach should allow the investigation of inherent instabilities or perturbations within glacial systems that may aid in explaining drumlin development and at the same time soft sediment deformation and transport at the subglacial interface. Where the topographic placement of a subglacial drainage system will influence groundwater and subglacial meltwater drawdown as well as till rheologies, it may also help explain specific locations of preferential subglacial streamlining, ice streaming, and ice mass marginal retreat or advance, the latter influencing marginal moraine and other landform development.

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