Glacial Processes and Landforms—Transport and Deposition☆

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Glossary

De Geer moraine Named after Swedish geologist G.J. De Geer (1858–1943), these moraines are low amplitude ridges that developed subaqueously by a combination of sediment deposition and squeezing and pushing of sediment along the grounding-line of a water-terminating ice margin. They typically occur as a series of closely-spaced ridges presumably recording annual retreat-push cycles under limited sediment supply.
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 line 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.
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.
Nunatak (from Inuktitut (Inuit language); *lonely peak*)) This is an exposed bedrock protrusion often a mountain ridge or peak above an ice mass (glacier, ice sheet or ice shelf).
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 whose ice-bed interface is 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.
Regelation A process that occurs when ice melts under pressure and refreezing when the pressure is reduced. This results in the production of meltwater that may enhance ice movement at that interface and also melting and re-freezing of ice around debris lead to their incorporation into basal ice.
Rogen moraine (ribbed moraine) These subglacial or submarginal transverse moraines were first named in Sweden. They are also termed cross-valley, ribbed, transverse, or washboard moraines.
Sediment flux This refers to the movement of sediment transported within any glacial system (in a sense the discharge of sediment at any locale).
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.
Temperate ice masses This is commonly termed warm-based or wet-based ice mass—this term is used to refer to those ice masses whose ice-bed interface is 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.

☆*Change History*: October 2020. J Menzies and M Ross this chapter has been updated by both John Menzies and Martin Ross in a totally shared capacity such that each of us added and edited each section of the text of the complete chapter. Fig. 4 was added by John Menzies and Fig. 9c was added by Martin Ross.

 https://doi.org/10.1016/B978-0-12-818234-5.00027-4

1 Introduction

Glacial deposits and landforms are end products of complex sediment delivery systems (SDS) (Alley et al., 1997; Evans, 2014). Net sediment deposition occurs as a function of ice dynamics, the type of ice mass, basement and subjacent geology and sedimentology, the temporal and spatial variability of erosion and sediment discharge (flux), associated thermal and hydrological regimens, and topography. Once sediments are produced by erosion, their transport and subsequent deposition within the glacial system requires an appreciation of cascading SDS in supraglacial, englacial, subglacial, and proglacial subenvironments. Fig. 1 illustrates the many

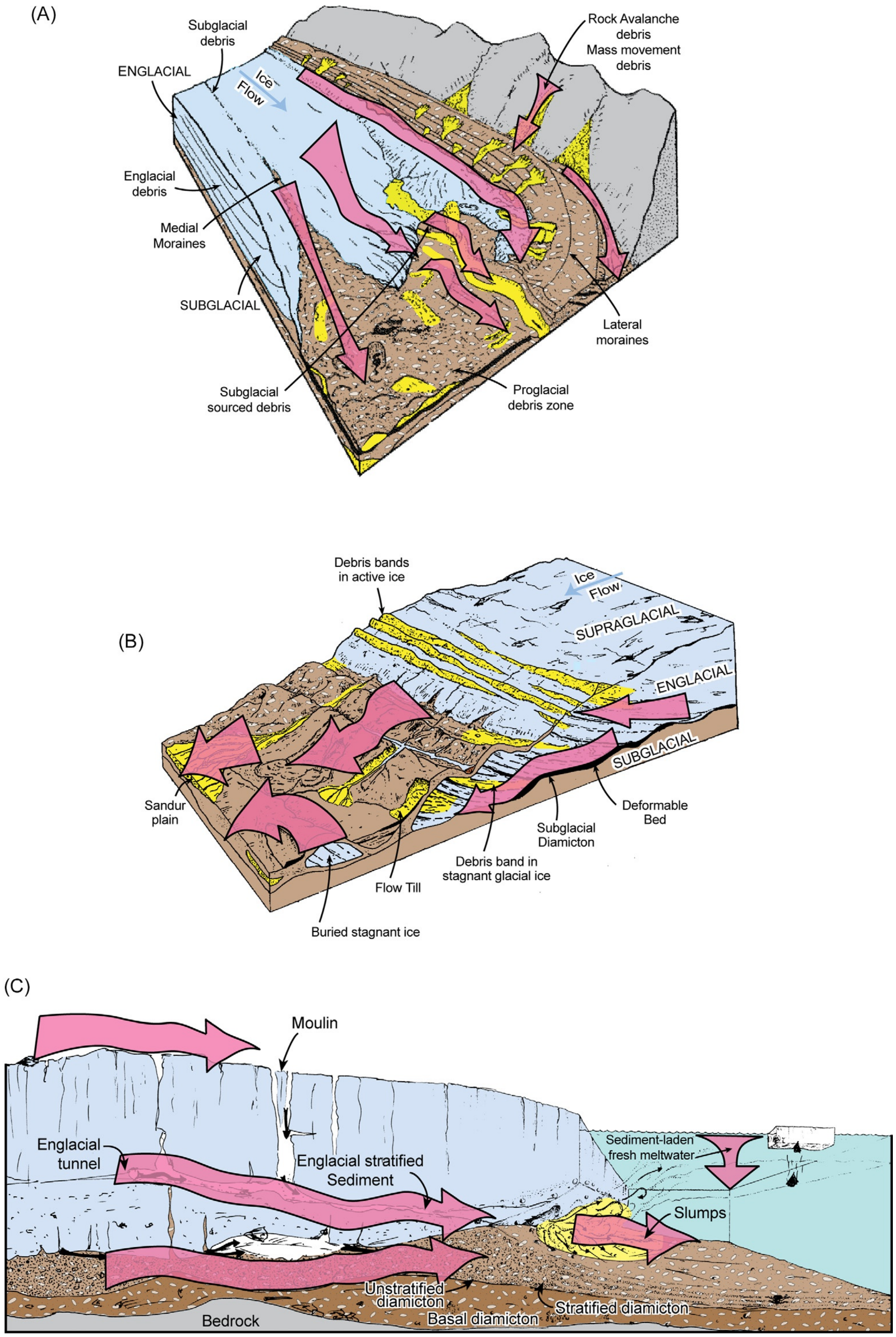

Fig. 1 Sediment delivery systems (SDS) within glacial environments (A) SDS within valley glacial systems, (B) SDS within marginal ice sheet systems, and (C) SDS within subaqueous glacial environments. (A) Modified from Boulton GS and Eyles N (1979) Sedimentation by valley glaciers; A model and genetic classification. In: Schlüchter C *Moraines and Varves: Origin/Genesis/Classification*, pp. 11-23. Rotterdam: A.A. Balkema Publishing.

and complex pathways of these delivery systems. The purpose and scope of this chapter is to link sedimentary transport processes to depositional mechanics of glacial sediments and the landforms/bedforms produced as a consequence.

In terms of ice dynamics and its influence on sediment deposition, 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 (Echelmeyer and Zhongxiang, 1987; Hallet et al., 1996; Alley et al., 1997; Nygård et al., 2007; Gulley et al., 2009; Batchelor and Dowdeswell, 2014; Dowdeswell et al., 2015; Livingstone et al., 2016; Stokes, 2018; Hogan et al., 2019; Reinardy et al., 2019).

Large volumes of sediment are transported in all SDS subsystems under temperate wet bed conditions (Ottesen et al., 2005; Benn and Evans, 2010; Bradwell and Stoker, 2015; Peters et al., 2016; Prothro et al., 2018; Rempel, 2008; Swift et al., 2018) (Fig. 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. As the volume of glacial 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 glaciofluvial sediment 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 (Truffer et al., 2000; Jaeger and Koppes, 2016; Gardner, 2019). 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, mass movement transport ceases. 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. Large annual variations in sediment fluxes are expected across different glacial subsystems (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 and deposition 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 (Fig. 1). Valley glaciers and other confined ice masses such as cirque glaciers have distinct and well-developed supraglacial and marginal SDS, whereas ice sheets 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 (Anderson et al., 1991; Domack and Powell, 2018).

Because glaciers can erode bedrock directly, mainly through abrasion and quarrying (cf. Alley et al., 2019), sediment transported and deposited by ice masses largely reflects the bedrock geology encountered by flowing ice along its path (Trommelen et al., 2013; McClenaghan et al., 2018). However, glaciers can also re-entrain pre-existing sediment, which can complicate the transport history of glacial sediment due to the polycyclic origin of a portion of its constituents. For instance, the distribution of clasts of known provenance in poorly-sorted glacial sediment (till) is interpreted to reflect more than one glaciation (Prest et al., 2000) because it requires multiple and prolonged ice flow phases to transport sediment over several hundreds of kilometers in different directions. These continental-scale compositional distributions are thus more likely to reflect several cycles of entrainment and deposition than a single cycle. Although ice sheets and long valley glaciers may transport sediment over long distances of several hundreds of kilometers, the bulk 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 (Paulen and McMartin, 2009; Stanley, 2009; Benn and Evans, 2010; Menzies et al., 2018). There are exceptions to this, such as along ice streams, where evidence suggests sediment can be transported over hundreds of kilometers in a single glaciation (e.g., Ross et al., 2009). 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 (Hewitt, 2009) (Fig. 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 orders of magnitude less eroded material (Woodard et al., 2019) than fast-moving temperate ice crossing soft sedimentary rocks (Colgan et al., 2002; Dühnforth et al., 2010; Alley et al., 2019).

Huge temporal and spatial variations in transport and depositional processes and their effectiveness and flux rates occur in all glacial subenvironments. Large temporal and spatial variations result in large fluctuations in sediment flux rates that result in differences in depositional rates. Although flux rates have been studied quite extensively in valley glacier systems (Hallet et al., 1996; Kirkbride, 2002; Kjaer et al., 2003; Riihimaki et al., 2005; Alley et al., 2019), there has been limited discussion of sediment flux rates below present-day ice sheets, ice streams, and ice shelves (Alley et al., 1989; Dowdeswell and Siegert, 1999; Ottesen et al., 2005; Dowdeswell et al., 2006, 2010; Hogan et al., 2019). 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 4 cm a^{-1} (0.13 cm a^{-1} averaged over the fan) over a 200-km-wide mouth of the Bear Island trough (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 Ó 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 estimating such sediment flux rates as only limited "snapshots" of subglacial conditions exist at this time. The variations in transport of sediment volumes across the bed of an ice mass and in margins lead to considerable volumetric variations in sediment delivery pathways and "end-points." It is at these "end-points" that deposition occurs, leading to large ranges and fluctuations in landform/bedform evolution and development over time.

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

2 Towards deposition—Sediment transport

Different sediment transport pathways within SDS that lead to deposition/emplacement can be examined viz. (1) at the top of and within glacier ice (supraglacial and englacial transport, excluding basal debris-rich ice), (2) meltwater, (3) basal debris-rich ice and below glacier ice (subglacial transport), and (4) by gravity (Fig. 1).

Ice mass transport of sediment can be on the ice supraglacially, or within an ice mass, englacially, or at the base of the ice and frozen on, or as deforming subglacial sediment. On confined ice masses such as valley glaciers or ice sheets along the edges of nunataks, supraglacial debris can accumulate and move down ice (Fig. 1A) (Schomacker and Benediktsson, 2018. 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 (Fig. 2). This frost-driven debris or rock fall materials are typically angular to subangular, with limited evidence of transport of any significant distance. Supraglacial debris, if based on the mass movement, would be classified as flow till and where meltwater transport has been the dominant transport and depositional process, the sediment would be stratified and classified as a glaciofluvial sediment.

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, generally, the debris is angular to subangular and was thus transported over short distances (Schomacker and Benediktsson, 2018; Swift et al., 2018). Within ice sheets, unless nunataks are present, there tends to be very limited, if any, englacial debris transport (Fig. 1B). 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 (Hindmarsh and Stokes, 2008; Winter et al., 2019). Basal debris can be incorporated within basal ice by basal freeze-on or regelation intrusion processes (Iverson and Semmens, 1995), forming debris-rich basal ice layers typically just a few meters thick, but that can exceptionally reach greater thicknesses due to a variety of mixing processes such as basal folding (e.g., as much as 20–50 m within the basal Greenland Ice Sheet). The subglacial origin of this material is supported by clear evidence of basal glacial erosion and comminution (Fig. 1B).

Most sediments transported by meltwater within all SDS are subject to reshaping and polishing by fluvial processes, 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 (Alley et al., 1997, 2019; Gulley et al., 2009; Delaney et al., 2018). These sediments are all forms of glaciofluvial deposits and making a distinction between supraglacial or englacial origin can occasionally be difficult and highly uncertain.

Under many ice masses, substantial transport of mobilized sediment occurs as a result of basal ice shear stresses, causing saturated and poorly sorted sediment to move as a layer. This layer, referred to as deformation till, contributes to basal ice flow velocity and sediment flux to the margin (Figs. 1B and C). Transport of sediment in this layer is, as yet, only moderately understood (Boulton et al., 2001a; Hart et al., 2011, 2018; Swift et al., 2018; Narloch et al., 2020), but low effective pressures (weight of the ice minus porewater pressure) at the ice bed interface plays an important role in lowering till strength allowing for soft deformation to take place even under relatively low basal shear stress (e.g., Alley et al., 1987). However, at even higher porewater pressures, the ice-bed interface may decouple reducing till deformation and promoting sliding of ice on soft bed. However, several other factors need to be considered to explain temporal variations in both the spatial extent and thickness of the deforming layer (cf. Dowdeswell et al., 2004; Livingstone et al., 2012, 2016). 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. Notably, deformation tills in some instances include fragile, yet well-preserved, transported particles such as marine shells (e.g., McMartin et al., 2019) attesting to the dilated state of the deforming layer under high porewater pressures. 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 and Auton, 2000; Phillips et al., 2002; Bartholomaus et al., 2008). Subglacial sediments can also be smeared onto the underlying surfaces by bedrock or frozen sediments or simply by adhesion forming lodgement tills. Likewise, in some cases within subglacial cavities or parts of subglacial meltwater tunnels, it is not unusual to find mass movement occurring, thus producing flow tills. Where glacier ice meltout or sublimation occurs, various forms of meltout tills are produced, exhibiting limited shear stress (cf. Larson et al., 2016).

In all SDS, there are sediments transported by mass movement (as noted above). Sediment gravity flows occur 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 (Fig. 1A and B). The impact of transport during mass movement is relatively limited, but changes in sediment clast fabrics and internal stratigraphy tend to occur.

3 Sediment deposition

Sedimentary processes in some depositional environments, such as glacial environments, are controlled by short period (high frequency) cycles, whereby erosion, transport and deposition occur rapidly and repeatedly before a depositional sequence is formed and preserved in the long-term geological record. 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; the term "glacial deposition" can be correctly applied. At some distance from an ice source, the term glacial deposition can no longer apply, but sediments may have a glacial origin and the processes of deposition may still be influenced by the distal glacier (e.g., sand and gravel on the seafloor that were dropped by melting icebergs). Different depositional environments thus typically co-exist laterally and influence each other (e.g., glacial, fluvial, aeolian, marine). In the proglacial environment, the glacial influence on sediment deposition decreases with distance to the glacial margin. For instance, the dynamics of a braided river may be controlled by the melting rate of the upstream glacier, but fluvial sediments may be redistributed at the river mouth by coastal processes and shelf processes. The particles deposited on the beach or the shelf were originally produced by glacial processes but have been extensively reworked and redeposited by fluvial and marine (nonglacial) processes.

It seems reasonable to assert that glacial forms, landforms/bedforms, develop 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 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 (Eyles et al., 2016).

The enormous range and types of glacial landforms are well known, and their typical forms and styles of development are discussed in detail in several textbooks (Bennett and Glasser, 2011; Benn and Evans, 2014; Menzies and van der Meer, 2018). 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 (Aario, 1977; Rose, 1987; Stokes et al., 2013a; Ely et al., 2018; Hart et al., 2018). 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 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 based on: (1) those forms directly attributable to the interplay of basal glaciodynamics and sediment availability and rheology; and (2) those that dominantly reflect meltwater and/or sediment-gravity processes. In the former subdivision, it is also relevant to differentiate as to whether the forms have developed at the ice/bed interface, parallel or transverse to ice movement, or have developed unoriented in relation to ice flow, or as ice marginal forms (Stokes et al., 2008, 2013a,b; Menzies et al., 2016) (Table 1).

3.1 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 basal interface due to changes, for example, in basal ice stress conditions, basal ice velocities, thermal regime, sediment rheology, bedrock lithology and bulk strength, sediment flux rates, preexisting topographies, and interface "roughness." Such perturbations translate in some instances into bedforms that align parallel (e.g., drumlins) or transverse (e.g., Rogen moraine) to the main flow direction of the overlying ice mass or are roughly topographically planar forms. A fundamental question that still requires an adequate answer is why such interface perturbations occur where and when they do? Is there an inherent instability in basal ice interface conditions (Hindmarsh, 1999; Fowler and Chapwanya, 2014) 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 (Hart, 1997; Hindmarsh, 1999; Schoof, 2007; Dunlop et al., 2008; Menzies et al., 2018; Narloch et al., 2020). 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.

- The patterns formed by these bedforms are distinctive and would appear related to the position and location beneath the ice (Wagner, 2018).
- Are the forms developed as a group over a short period of time or prolonged and repeated episodes?
- 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 (Stea and Brown, 1989; Hoffmann and Piotrowski, 2001; Sookhan et al., 2018).
- Most forms developed parallel to ice motion would seem to be elongated as a result of either higher ice velocities, such as below ice streams (Clark et al., 2003a), or more ductile sediment rheologies or a combination of both factors.
- 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).
- In many instances, overprinting or reorientation of some or all forms may occur.
- There may well be interrelationships between these forms and other 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.
- Finally, it is 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 (Dunlop et al., 2008; King et al., 2009; Ross et al., 2011; Stokes et al., 2013a,b; Barchyn et al., 2016; Spagnolo et al., 2017; Sookhan et al., 2016; Hart et al., 2018).

3.1.1 Drumlins

Elongated landforms aligned parallel to ice flow are common on subglacial beds. They have been classified in different ways based on their morphology. Drumlins are intermediate forms, varying 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 (cf. Spagnolo et al., 2014; Hillier et al., 2016; Ely et al., 2018). Their varied morphology can deviate considerably from the classical tear-shaped form commonly portrayed in textbooks (Fig. 3). Typically, they form in large "swarms" or fields many thousand in number (e.g., western New York Drumlin field ~6000) (Hess and Briner, 2009; Menzies et al., 2016; Sookhan et al., 2018), but may also occur individually or in small groups (Fig. 3A and B). Recent satellite images clearly demonstrate the apparent close relationship between drumlin fields and ice stream locations (Clark, 1993; Stokes and Clark, 2001; Clark and Stokes, 2003, Fig. 9.12; Clark et al., 2003b, 2009).

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.

(*Modified from Menzies, J. and Shilts, W.W. (2002). Subglacial environments. In: Menzies, J. (ed.) Modern and past glacial environments, pp 183–278. Oxford: Butterworth-Heineman.*)

Fig. 3 (A) Drumlins in Clew Bay, Ireland (image from Google). (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).

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 initial nucleation occurs, it can be demonstrated that in some places, the form will persist, grow, and possibly migrate. The problem with all drumlin formative hypotheses is identifying the "trigger" (Aronow, 1959). Currently, four broadly acceptable hypotheses exist that attempt to account for the formation and subsequent development of drumlins (Fig. 4):

- Deforming sediment bed (Boulton, 1987; Menzies, 1989; Smith et al., 2007).
- Groove "ploughing" (Tulaczyk et al., 2001; Clark et al., 2003b).
- Interface instability (Hindmarsh, 1999; Fowler, 2009; Fowler, 2010a,b; Stokes et al., 2013a; Fowler and Chapwanya, 2014).
- Erosion of pre-existing sediments into streamlined forms (Eyles and Doughty, 2016; Eyles et al., 2016; Möller and Dowling, 2016; Iverson et al., 2017; Hart et al., 2018).

3.1.2 Flutes moraines and mega scale glacial lineations (MSGLs)

Subglacial streamlined landforms that are either shorter or longer than typical drumlins are generally classified separately, although all these landforms may constitute a continuum (e.g., Ely et al., 2016). 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

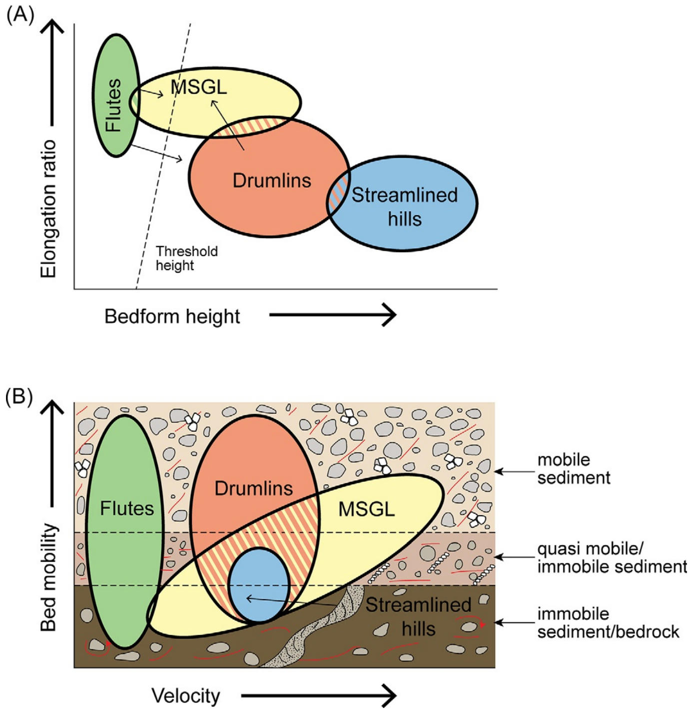

Fig. 4 The bedform continuum: (A) schematic relationship between bedform height and elongation ratio; (B) schematic relationship between ice velocity and bed mobility. (B) Modified from Hart JK, Clayton AI, Martinez K and Robson BA (2018) Erosional and depositional subglacial streamlining processes at Skálafellsjökull, Iceland: An analogue for a new bedform continuum model. *GFF* 140: 153–169. doi:10.1080/11035897.2018.1477830.

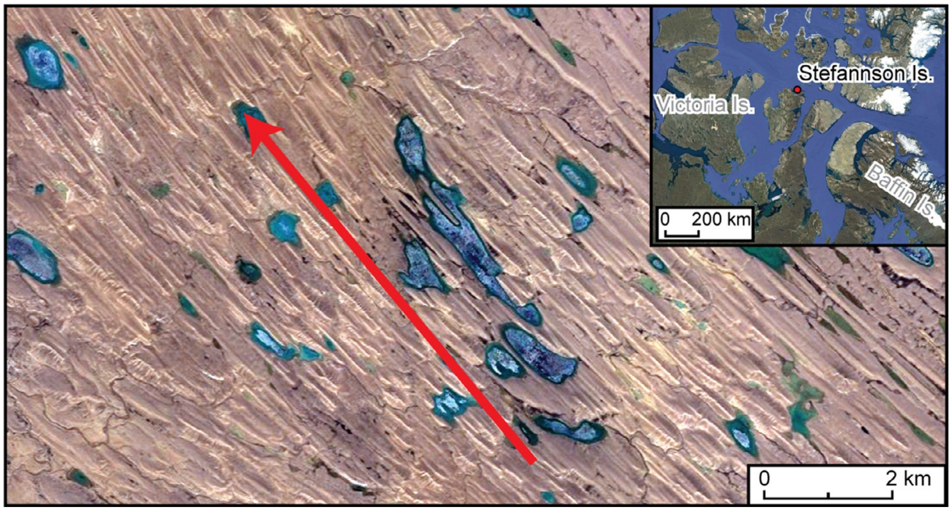

Fig. 5 Image of drumlins on west side of Stefansson Island, Nunavut. Note the sweep of the curving ice stream (Innuitian Ice Sheet (IIS) and the ensuing drumlin morphometry (image from Google maps) (cf. England et al., 2006).

kilometers (Fig. 5). Mega-scale glacial lineations (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 mélange sediments that have been collectively scavenged in the process of formation, although MSGLs, like drumlins, in some instances, have internal stratigraphy.

As in the case of drumlins, there would appear to be a bedform association between fluted moraines, MSGLs, drumlins, and Rogen moraines (Rose, 1987; Clark, 1993, 1994; Menzies et al., 2016; Stokes, 2017; Hart et al., 2018). Hypotheses like 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. Hart et al. (2018) suggest that there are two possible pathways through this continuum. At lower basal ice velocities or, more likely, lower sediment flux rates, flutes and drumlins form at differing scales and dimensions, possibly depending on the threshold height of obstacles (Boulton, 1982; Menzies, 1982; Phillips et al., 2018a). It seems generally accepted that as basal ice velocities and/or sediment flux rates increase, drumlins will become elongated (Stokes and Clark, 2002; Stokes et al., 2013a; Barchyn et al., 2016; Menzies et al., 2016). In some cases, as drumlins become elongated, they essentially "morph" into MSGL, making the distinction between the two landforms might be arbitrary (Graham et al., 2009; Stokes et al., 2013b; Spagnolo et al., 2014; Stokes, 2017) (Fig. 4).

The old hypothesis that flutes grow in the lee of boulders is generally correct (Fig. 6), but in many cases and especially so with drumlins and MSGLs, the cause of nucleation and streamlined elongation is commonly missing. It seems likely that the formative processes involved in both drumlins and MSGLs are very similar. Any differentiation maybe, in the case of flutes and MSGLs, the result of a relatively confined, high-sediment flux rate at the ice/bed interface and the associated relatively high basal ice velocities typical of ice streams (Dunlop et al., 2008; Stokes et al., 2008; Winsborrow et al., 2010; Barchyn et al., 2016; Hart et al., 2018; Sookhan et al., 2018).

3.1.3 Ribbed (Rogen) moraines

In the past, these morainal forms, 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

Fig. 6 (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.

Fig. 7 Rogen moraine from near Whitbourne, the Avalon Peninsula, Newfoundland, Canada. Photograph courtesy of Tom Fisher.

unique landforms, the moraines were part of a continuum of forms either emanating from or passing into parallel fluted moraines and drumlins (Möller, 2006; Barchyn et al., 2016; Möller and Dowling, 2016, 2018) (Fig. 7).

In morphology, Rogen moraines occur as slightly sinuous ridges 10–100 m in height stretching transverse to ice flow for hundreds of meters to several kilometers. The ends of many ridges bend down-ice and the ice-flow parallel profile is typically asymmetric with a gentle up-ice (stoss) side and a steeper down-ice (lee) side (Fig. 6). Like fluted moraine, MSGLs, and drumlins, the sediment content of these ridges is equally varied and essentially a mélange of available basal sediment. Notably, the higher amplitude moraines that have not been drumlinized are commonly covered with boulders (Sarala, 2006; Trommelen et al., 2014). Hypotheses of Rogen moraine origin are like drumlins in that both form at the ice-bed interface from complex thermomechanical conditions. These moraines can be viewed as wave-like forms that may result from rapid glaciodynamic changes such as proximal grounding-line lift-off events or instabilities (Fowler and Chapwanya, 2014) inherent in the sediment and water flux and basal ice stress and effective pressure (Bouchard, 1989; Lundqvist, 1989; Fisher and Shaw, 1992; Hättestrand, 1997; Knight and McCabe, 1997; Sarala, 2006; Dunlop et al., 2008; Ross et al., 2009; Chapwanya et al., 2011; Trommelen et al., 2014; Barchyn et al., 2016).

3.1.4 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 (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 (Fig. 8). 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 location within the topography, thus continuing to build up the moraine over time repeatedly (Krüger, 1995; Vacco et al., 2009; Barr and Lovell, 2014). The sediment content of most marginal moraines reflects a wide diversity of sediment facies characteristic of supraglacial, englacial, and subglacial environments. 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 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 (Maizels, 2002; Evans, 2009, 2014; Schomacker and Benediktsson, 2018).

Hummocky moraine is a distinctive form of moraine that occurs over large areas of submarginal and proximal proglacial areas. Current depositional models argue that hummocky moraine was deposited supraglacially from stagnant debris-rich ice ("disintegration moraine") (Bennett and Boulton, 1993; Eyles et al., 1999; Ham and Attig, 1996; Munro-Stasiuk, 1999; Lukas et al., 2005) (Fig. 9A and B). In contrast, Boone and Eyles (2001), 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 of the hummocky moraine is chaotic, non-oriented that appears, in places, to pass down the adverse slope (up-ice direction?) into weakly oriented hummocks ("washboard moraine") that are transitional to drumlins in topographic lows further into the subglacial landsystem. There are also fields of hummocky moraine that have sharp lateral boundaries delimited by ice-stream shear margin

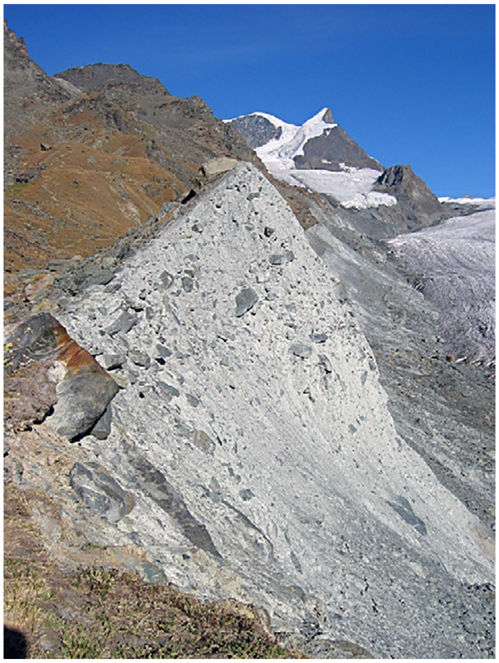

Fig. 8 Asymmetrical lateral moraine of the Findelen glacier, near Zermatt, Switzerland. The boulder in the forefront of the photograph is approximately 3 m in height.

moraines, which suggest these fields of hummocky moraine have a subglacial origin related to processes taking place in inter-ice stream zones (Ross et al., 2009).

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 (Lukas, 2005; Lukas et al., 2005). It is intriguing to speculate that many forms of hummocky moraine exist in topographic appearance, and in some cases, sediment content shows 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 help explain hummocky moraines in different parts of the world.

3.2 Landforms/bedforms indirectly attributable to active/passive ice activity

Landforms/bedforms within this category of glacial landforms involve sediment transport and deposition by meltwater-related processes (Table 1). 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 together, forming sediment-landform associations. For instance, where one finds eskers, it is not unusual to also find kames and kame terraces. However, a distinction between all the different groups of meltwater-generated landforms is not always clear.

3.2.1 Esker systems and meltwater corridors

Eskers are sinuous ridges, typically consisting of coarse-grained and heterogeneous sediment packages (Brennand, 2000). Esker ridges (Fig. 10) 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 Northwest 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 Late Wisconsin (Menzies and Shilts, 2002, Fig. 8.44). Esker systems form as a function of the location and type of meltwater channel or drainage system either subglacially or at least submarginal along the outer edge of an ice mass (Veillette, 1986; Clark and Walder, 1994; Kleman and Hättestrand, 1999; Boulton et al., 2001b, 2009; Bennett et al., 2007; Menzies et al., 2018; Knight, 2019).

Essentially, an esker is the result of a sediment-choked meltwater conduit that developed within the ice (Boulton et al., 2007a,b; Hooke and Fastook, 2007; Boulton et al., 2009; Ahokangas and Mäkinen, 2014; Dowdeswell and Ottesen, 2016; Storrar et al., 2013, 2014, 2019). Some eskers appear to conformably overly subglacial sediments suggesting they formed within an englacial conduit and sediments within the conduit later draped on subglacial sediments due to loss of ice support by melting (Warren and Ashley, 1994; Huddart and Bennett, 1997). In most cases, however, eskers appear to have formed at the base of the glacier within basal ice

Fig. 9 (A) Landsat ETMþ 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. (C) Hummocky controlled moraine ridges on the flank of a latero-frontal moraine at Kvíárjökull in southern Iceland; some hummocks may still be ice-cored.

Continued

Fig. 9, Cont'd

Fig. 10 Landsat ETMþ satellite image of Carstairs Esker system, Scotland. Photograph courtesy of C. Zadowicz.

conduits directly over a rigid subglacial substrate (typically bedrock). It is also common to find eskers along larger meltwater corridors (e.g., Rampton, 2000; Peterson et al., 2018), suggesting eskers record the waning stage of meltwater drainage along these corridors. The larger routeways corridors appear to be mainly erosional. However, discontinuous gravel deposits and associated depositional bedforms have been documented within these meltwater routeways corridors (Rampton, 2000). Eskers have been documented draping drumlins in some places at an oblique angle to the drumlin long-axis orientation and along the flank of valleys rather than along their bottom. Eskers mirror the pathways taken by meltwater channels and drainage systems, which reflect the general direction of hydraulic gradient, controlled by ice thickness and surface slope, within a glacier at the time of esker formation. Esker tends to form dendritic patterns, which means that locally, esker segments can differ in orientation from the regional ice flow directions. The location and size of eskers and entire esker patterns may also be influenced or controlled by groundwater flow within the subglacial aquifer (Boulton et al., 2007a,b). 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 quasi-stable conditions close to the ice margin, at least over several centuries, during the retreat of an ice sheet (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 (Boulton and Caban, 1995; Boulton and Zatsepin, 2006).

3.2.2 Kames and kame terraces

In most areas of ice mass melting where massive amounts of glaciofluvial sediments have been transported, kames and 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 for meltwater and transport of sediment. Commonly the slope of a valley glacier trim line is traced by a kame terrace (Gray, 1995; Huddart and Bennett, 1997; Terpilowski, 2007; Bennett et al., 2007); (Fig. 11). In some cases where crevasse filling collapses on the ice mass melting, kames form as roughly circular accumulations of glaciofluvial sediment that exhibit marked faulting and slumping on their sides.

In other instances, kames develop as unoriented deposits within subglacial and englacial cavities or abandoned meltwater channels (Houmark-Nielsen et al., 1994; Ham and Attig, 1996; Huddart and Bennett, 1997) or kame ice-contact deltas (Schaetzl et al., 2017; Włodarski and Orłowska, 2019). Where subglacial meltwater channels discharge into lakes, deltas of glaciofluvial sediment can build up and are termed kame deltas (nb. the Salpausselkä kame deltas in Finland, Glückert, 1977; Ahokangas and Mäkinen, 2014; Ojala et al., 2019; Winsemann et al., 2018).

3.2.3 Outwash fans and deltas

In most proglacial environments, outwash fans of various dimensions develop when large meltwater or multiple streams emanate from the frontal and lateral margins of ice masses, or glaciofluvial sediments enter the proglacial zone and bed load competency declines (Maizels, 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 (Russell and Knudsen, 2009; Carrivick and Tweed, 2019) (Fig. 12). Outwash fan surface gradients develop largely as a function of grain size such that steep gradients develop proximal to the margin where sediments are coarser and then decrease further down ice along with the sediment fining trend (Maizels, 2002, Fig. 9.7). Small abandoned ice masses are common on their surface and they subsequently develop into kettle holes upon melting. Sedimentary deposits associated to outwash fans or plains tend to consist of laterally extensive tabular bodies, or sheets, of coarse-grained assemblages characteristically reflecting rapid (debris-flow type) deposition (poorly to moderately sorted sediment), as well as well-sorted imbricated packages. Lenses of fine-grained sediments may occur but are generally thin and of limited extent.

3.2.4 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 crudely stratified till may develop that has been variously termed a till delta or a till tongue (King et al., 1991; Larter and Vanneste, 1995; King, 1996; Powell and Cooper, 2002; Anandakrishnan et al., 2007; Smith and Anderson, 2010; Reinardy et al., 2011; Phillips et al., 2018b). 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 form beneath the subglacial ice

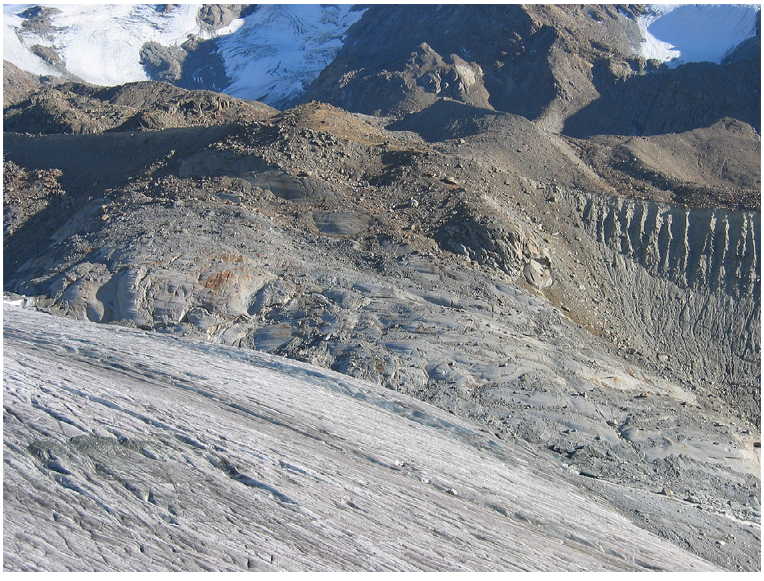

Fig. 11 A series of kame terraces along the southern flank of the Findelen Glacier near Zermatt, Switzerland.

Fig. 12 The proglacial zone of the Fidelen Glacier, near Zermatt, Valais, Switzerland. Photo taken looking from the glacier snout out across the proglacial zone.

margin at or very close to the grounding-line (Domack and Powell, 2018; Demet et al., 2019). 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. For example, how subglacial soft-sediment deformation exhibits upon drumlins and MSGL development requires field investigation (Le Meur and Hindmarsh, 2001; Christianson et al., 2016).

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, 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, integrating the hydraulic systems is crucial to understanding the full glacial system as a single, integrated, and 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 approach should also 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. The utilization of Lidar in conjunction with ground mapping (cf. Yu et al., 2015; Mayoral et al., 2017; Sookhan et al., 2018; Wagner, 2018; Lewington et al., 2019) is rapidly showing enormous success in explaining glacial depositional processes and holds great potential for the future. Our overall understanding of many glacial depositional processes still leaves much research to be achieved. As new advances in technology and our comprehension of glaciodynamics in many glacial subenvironments advances, the rate of knowledge acquisition is rapid and holds great promise for the immediate future. With global warming and climate change upon us, such advances are critical if we are to grasp any hope of controlling climate changes and their potential dire consequences.

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