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Subglacial basins: their origin and importance in glacial systems and landscapes

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ABSTRACT

Closed topographic basins are found beneath contemporary ice masses and within the footprint of former ice masses in all glaciated regions. We present the first integrated review of subglacial basin occurrence and formation and the implications of such basins for glaciological processes and the evolution of landscape. Our purpose is to motivate research in areas where understanding of basin origin and process significance is weak. Basins on the order of $10\text{-}10^2$ m deep and $10^2\text{-}10^3$ m long are produced by glacial erosion of subglacial rock and/or sediment and are known as 'overdeepenings'. Outlet and valley glaciers can 'overdeepen' their beds far below sea level or local fluviatile base level. Larger basins, typically in ice sheet contexts, may have a pre-glacial (usually tectonic) origin. Subglacial basins are important glaciologically because they require ice, water and sediment to ascend an adverse subglacial slope in order to exit the glacial system, the efficiency of which is dependent upon the gradient of the adverse slope and that of the ice surface. Basins thus influence subglacial drainage system morphology and transmissivity, the thickness and distribution of basal ice and sediment layers, and the mechanisms and dynamics of ice flow. Adverse gradients that exceed 11 times that of the ice surface may even permit the formation of subglacial lakes. We speculate that, in comparison to ice masses with few or no subglacial basins, those with numerous or very large basins may respond to climatic changes with unexpected vigour. In addition, erosion rates and transport pathways of water and sediment through the glacial system, and the expression of these processes in the sediment and landform record, may be unexpectedly complex. Further, our review shows that, in a warming climate, ice masses resting on adverse slopes will be vulnerable to rapid and potentially catastrophic retreat; new lakes in subglacial basins exposed by mountain glacier retreat will present

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an increasing hazard; and subglacial lakes may drain catastrophically. On even longer time scales, we speculate that the glacial excavation and post-glacial filling of basins in mountainous regions should contribute importantly to climate-related changes in isostasy and relief. Although the controls on overdeepening and their influence on other glacial and landscape processes remain uncertain, we hypothesise that overdeepened glacial systems reflect an equilibrium ice-bed geometry that maximises the efficiency of ice discharge. Improved understanding of overdeepening processes, especially overdeepened-bed hydrology, is therefore necessary to understand fully the dynamic behaviour of valley and outlet glaciers, and thus the fate of Earth's largest ice masses.

Keywords: glacier dynamics; glacier hydrology; overdeepening; sediment transfer; glacial geomorphology; landsystems; landscape evolution; hazards; climate sensitivity

1. INTRODUCTION

Subglacial basins (i.e. closed topographic depressions in the beds of present and former ice masses) are a ubiquitous feature of glaciated environments. In this paper, we demonstrate that many subglacial basins are produced by glacial erosion. This type of basin is known commonly as an 'overdeepening'. However, many basins are likely to be non-glacial in origin, and may have undergone only limited modification by glacial processes. Our intention is to highlight the process significance of subglacial basins in general; hence, we do not restrict our attention to those basins of exclusively glacial origin. Nevertheless, the appetite of ice masses to 'bite deeply into sound rock' (Linton, 1963) means we focus primarily on the origin and process significance of glacial basins, and consider the process significance of presumed non-glacial basins only when it is pertinent.

This paper is primarily a review that aims to motivate research where understanding of the origin and process significance of overdeepening is weak. Nevertheless, wherever possible, we have attempted to draw out and present new ideas and hypotheses in order to illustrate the diverse and potentially important implications of overdeepening for glacial systems and geomorphic processes.

The first three sections review the context, occurrence, and origin of subglacial basins, focussing on the distribution and formation of glacial overdeepenings. We then review published work pertinent to the glaciological importance of subglacial basins in two parts: firstly, the importance for glacier hydrology and seasonal dynamics, ice flow and the stability of ice masses, and related phenomena; and secondly, the importance for glacial geomorphic processes and long-term landscape and ice sheet evolution. This review and analysis of published literature does not intend to be exhaustive; rather, we seek to identify and support essential concepts. For brevity, we do not include governing equations that can be found in cited literature. The sixth section is a discussion that draws on the evidence provided by published work to highlight areas of uncertainty and future research questions. It is recommended that readers who are familiar with the concept of overdeepening and have an expert understanding of glacier erosional, hydrological and ice dynamic processes skip to this section. The final section concludes the review and discussion.

2. GLACIOLOGICAL CONTEXT AND DEFINITIONS

Glacial erosion during the Quaternary period has contributed enormously to the sculpting of large regions of the globe, carving spectacular alpine landscapes, scouring vast areas of continental crust, and generating prodigious quantities of sediment for transport and deposition by glacial, aeolian and fluvial processes. The evolution and glaciological significance of many large-scale glacial erosional forms common to such regions, including U-shaped fjords and valleys, expanses of aerial scouring, and cirques and arêtes, is now well understood. In formerly glaciated regions, the distribution and morphometry of these landforms has been exploited widely for the purposes of palaeoglaciological

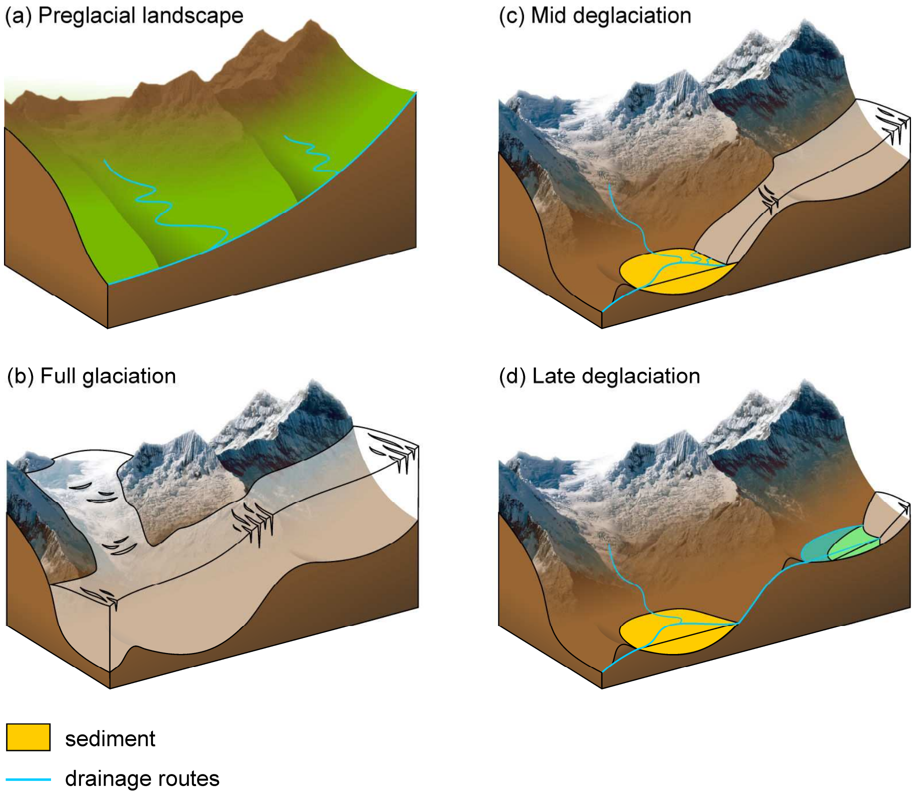

Figure 1. Schematic diagrams illustrating the typical glacial and post-glacial context of over-deepened basins under temperate valley glaciation.

and process reconstruction (Glasser and Bennett, 2004). Advances in remote sensing, numerical modelling and geochronology continue to enable investigation of landscape evolution at ever greater spatial and temporal scales, providing new insight into the glaciological significance of landscapes and their component forms, even beneath contemporary ice sheets (Bingham et al., 2010).

Glacial erosion is closely linked to spatial patterns of ice flow and sediment transport and deposition (e.g. Alley et al., 1997), and, over million-year timescales, glacial landscape evolution reflects and influences the dynamics of tectonics and climate (e.g. Stern et al., 2005; Nielsen et al., 2009; Whipple, 2009; Norton et al., 2010). Landforms record the growth and decay of past ice masses and the processes and patterns of sediment production and transport, such that analyses of formerly glaciated landscapes can inform understanding of contemporary ice mass dynamics and response. Further, landforms themselves, large-scale erosional landforms in particular, exert a strong control on basal properties and the mechanisms and patterns of ice flow (e.g. Marshall et al., 1996; Raymond et al., 2001; Jamieson et al., 2010), such that accurate knowledge of large-scale bed geometry is essential to fully understand ice-mass behaviour. Importantly, the evolution of landscape over successive glacial cycles will influence patterns of ice-mass inception, evolution, dynamics, and maximal extent (e.g. Taylor et al., 2004; Kessler et al., 2008; Schoof, 2007; Kaplan et al., 2010), and the widening and deepening of pre-glacial valleys will increase relief and stimulate isostatic adjustment, including the uplift of adjacent peaks (e.g. Stern et al., 2005; Champagnac et al., 2007; Norton et al., 2010). The latter process has been invoked as a key driver of late Cenozoic global cooling (e.g. Molnar and England, 1990). On a similar time scale, rates and patterns of glacial erosion during forthcoming glacial cycles are of particular concern for countries endeavouring to build underground repositories for the safe long-term storage of nuclear waste (e.g. Fischer and Haeberli, 2010).

Overdeepenings (Figures 1 and 2) are large-scale glacial erosional landforms that have long been recognised to be key elements of glaciated landscapes and to contribute importantly to the development of glacially carved relief. Despite the wealth of related research on glacial landscape

Figure 2. (a) An overdeepened cirque basin (Blea Water, UK). (b) A trunk-valley overdeepening bisected by post-glacial alluvial deposition (Buttermere, UK) (photo: Rob Larkamb); (c) A formerly ‘terminal’ overdeepening exposed by glacier retreat; the prominent lateral moraine (left of lake) marks the Little Ice Age glacier extent (Steingletscher, Switzerland) (photo: Tino Moeller); (d) A ‘terminal’ overdeepening partially exposed by glacier retreat; the overdeepening is partly confined by an arcuate moraine ridge (Bagley Icefield, Alaska) (photo: Don McCully).

evolution, the concept of overdeepening has received little attention and is often absent from reviews of these subjects (e.g. Glasser and Bennett, 2004).

Originally described as ‘rock basins’, overdeepenings were first attributed to glacial erosion by A.C. Ramsay in the late 1850s, and were described as a distinctive characteristic of glaciation by W.J. McGee in the early 1880s (Harbor, 1989). The term ‘overdeepening’ did not appear until the 1890s (Linton, 1963; Evans, 2008), when Albrecht Penck, having been “seized of the abnormality of the depth of glacial troughs” (Linton, 1963, p.14), expressed to W.M. Davis his view that “[alpine] valleys are certainly deepened and ‘overdeepened’ in the direction of ice flow, and [that] alpine lakes are merely overdeepened valley floors” (Evans, 2008, p. 422). Use of this term evolved later to apply specifically to “those cases where a trough floor had been deepened below the profile of equilibrium of the pre-glacial stream...[implying] mass quarrying of rock far below the local fluviatile base-level and the evacuation of the excavated material down low overall, and often reversed, gradients” (Linton 1963, pp. 15–17). Today, the term is used almost exclusively to describe a closed basin in a contemporary or former subglacial context that, under non-glacial conditions, would form a lake or allocthonous sedimentary basin (e.g. Fountain and Walder, 1998) (Figure 2).

In addition to being used as a noun, it is now common to use ‘overdeepening’ as a verb to describe the suite of glaciological processes that produce closed topographic basins in the glacier bed (cf. Hooke, 1991; Alley et al., 2003a, 2003b). In this paper, we accept this usage, but also use the term ‘subglacial basin’ to describe closed basins in general and in particular those that are known or presumed to be non-glacial in origin. Glacial valleys that have been widened and deepened by glacial erosion to a level below the equilibrium level of a pre-glacial valley (e.g. Figure 1a), but not sufficiently deeply to produce a closed basin, are described as ‘glacially deepened’, the important distinction being that ‘true’ overdeepening requires ice and water at the glacier bed to traverse a locally-reversed (or adverse) subglacial slope (e.g. Figure 1b), along which the products of glacial erosion must also be transported.

3. DISTRIBUTION OF SUBGLACIAL BASINS AND OVERDEEPENINGS

Subglacial basins occur in all glaciated contexts. Cirque floors, trunk valleys and fjords are commonly overdeepened or have overdeepened sections (Figure 2a and b), and overdeepening appears to be especially common beneath former glacier termini (Figure 2c). Overdeepening is common in bedrock but overdeepenings are not always entirely erosional in origin: some are confined, either partly or wholly, by sediment deposited at the terminus in the form of a moraine or, in lacustrine and marine contexts, a grounding-line 'shoal' (Figure 2d and Figure 3). Neither are overdeepenings necessarily glacial in origin: some are partly or wholly pre-existing tectonic bedrock basins that have been overridden by ice (e.g. Figure 4a and b).

In contemporary environments, basins have been located during borehole and geophysical (e.g. radar) investigations at mountain and outlet glaciers in the European Alps (e.g. Hantz and Lliboutry, 1983; Hock et al., 1999; Iken et al., 1996), Scandinavia (e.g. Hooke et al., 1988) (notably Storglaciären), Iceland (e.g. Roberts et al., 2002; Knudsen et al., 2001; Spedding and Evans, 2002), North America (e.g. Arcone et al., 1995; Fleisher et al., 1998; Fountain, 1994) and Greenland (e.g. Howat et al., 2008). Similarly, basins of glacial and non-glacial origin have been located beneath ice masses in East and West Antarctica (Del Valle et al., 1998; Fricker et al., 2001; Siegert et al., 2004; Hubbard et al., 2004; Bo et al., 2009; Vaughan et al., 2006, 2007; Ross et al., 2012) (Figure 4a–f) and beneath the Greenland ice sheet (e.g. at Swiss Camp; Price et al., 2008; Mottram et al., 2009). Smaller basins are often revealed by glacier retreat (e.g. Meier and Post, 1987; Frey et al., 2010), and their existence beneath contemporary ice masses can be inferred using associated phenomena (e.g.

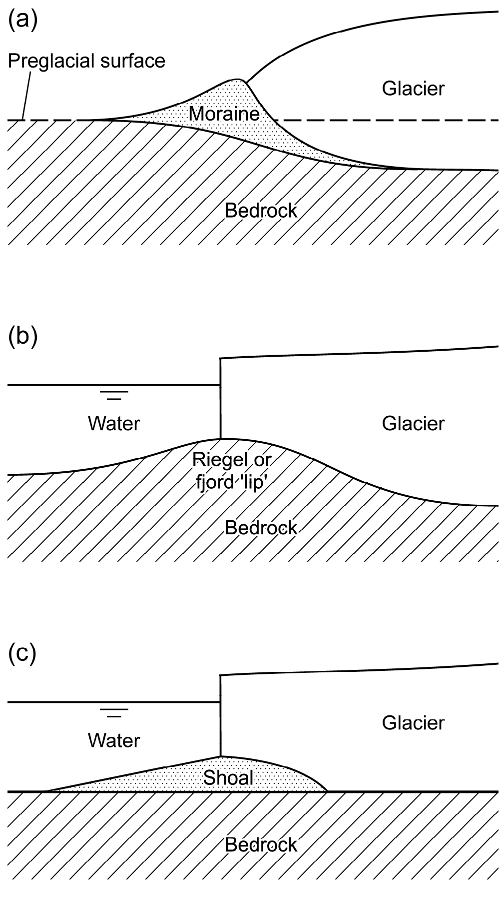

Figure 3. (a) A terrestrial valley or outlet glacier with a partly moraine-confined overdeepened rocky bed (ice flow is right-to-left). (b) A lacustrine- or marine-terminating outlet glacier with an overdeepened rocky bed, producing a pronounced riegel or fjord 'lip'. (c) A lacustrine- or marine-terminating outlet glacier with an overdeepened bed confined by a 'shoal' of glacially-deposited sediment.

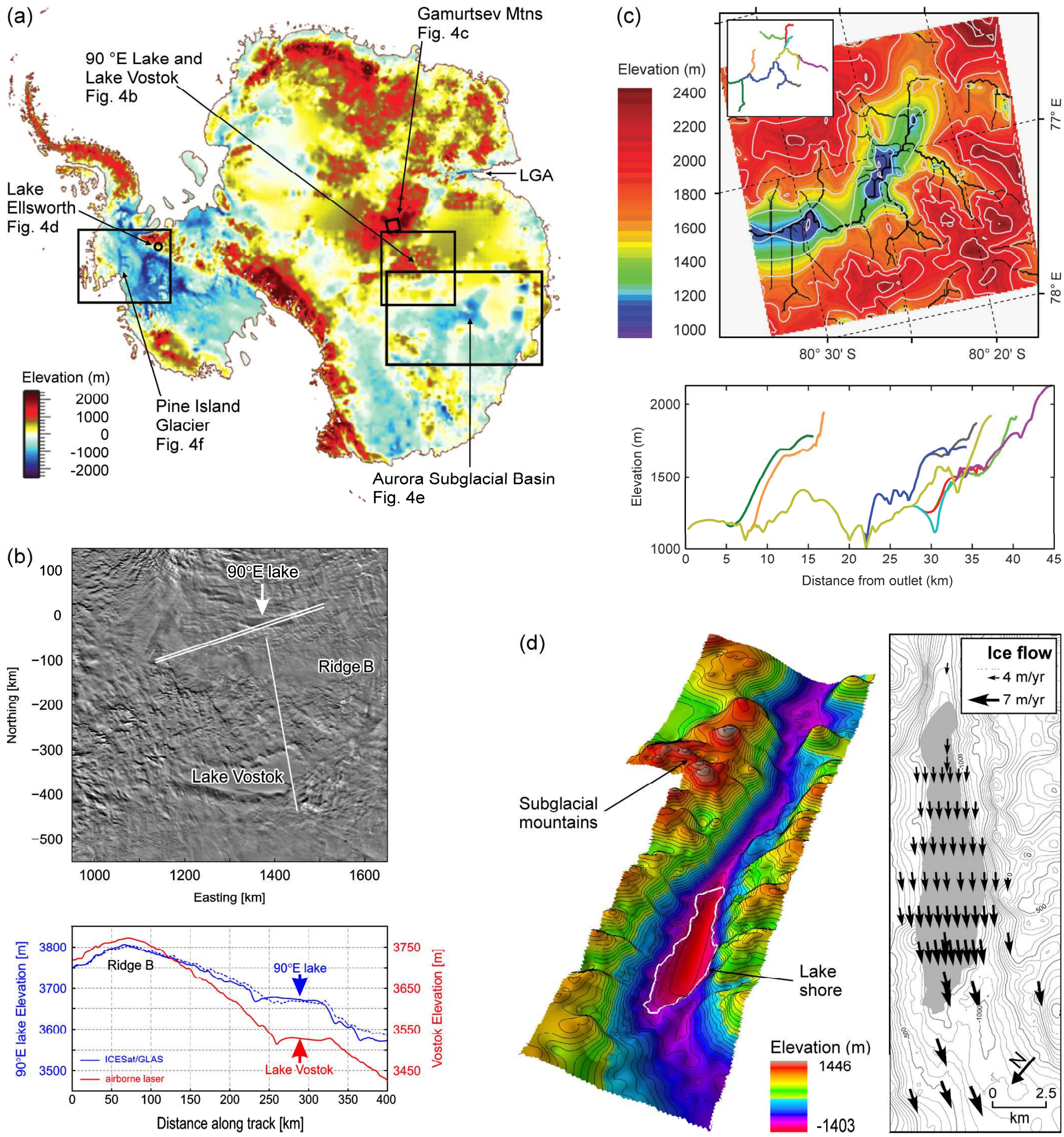

Figure 4. (a) Bed topography of Antarctica showing major marine subglacial basins (blue) and locations of Figures 4b-f (reproduced with permission from Young et al., 2011). (b) MODIS mosaic of the East Antarctic ice sheet surface (upper panel); flatter areas of the ice-surface (measured along the solid-white transect-lines) reveal the presence of tectonically-controlled fault-bounded basins containing subglacial lakes (lower panel) (reproduced with permission from Bell et al., 2006). (c) Contoured DEM of bed topography (upper panel) in the Gamburtsev Mountains, showing classic alpine topography and overdeepened basins occupying a pre-glacial fluvial valley network; profiles along major valley-axes (inset, upper panel) reveal the depth and extent of overdeepening (lower panel) (modified with permission from Bo et al., 2009). (d) Contoured map (left) and DEM (right) of bed topography showing the glaciated-valley context of the overdeepened basin occupied by Ellsworth subglacial lake; map also shows surface ice velocities (contours are 100 m intervals; figures provided by Neil Ross) (Ross et al., 2011).

debris-rich basal ice; see Section 4.2.1). Within the footprints of former ice masses, former subglacial basins are often evident as bedrock- or sediment-confined lakes (e.g. Fountain and Walder, 1998), though it is common for postglacial sedimentation to obscure their true extent. For example, the true

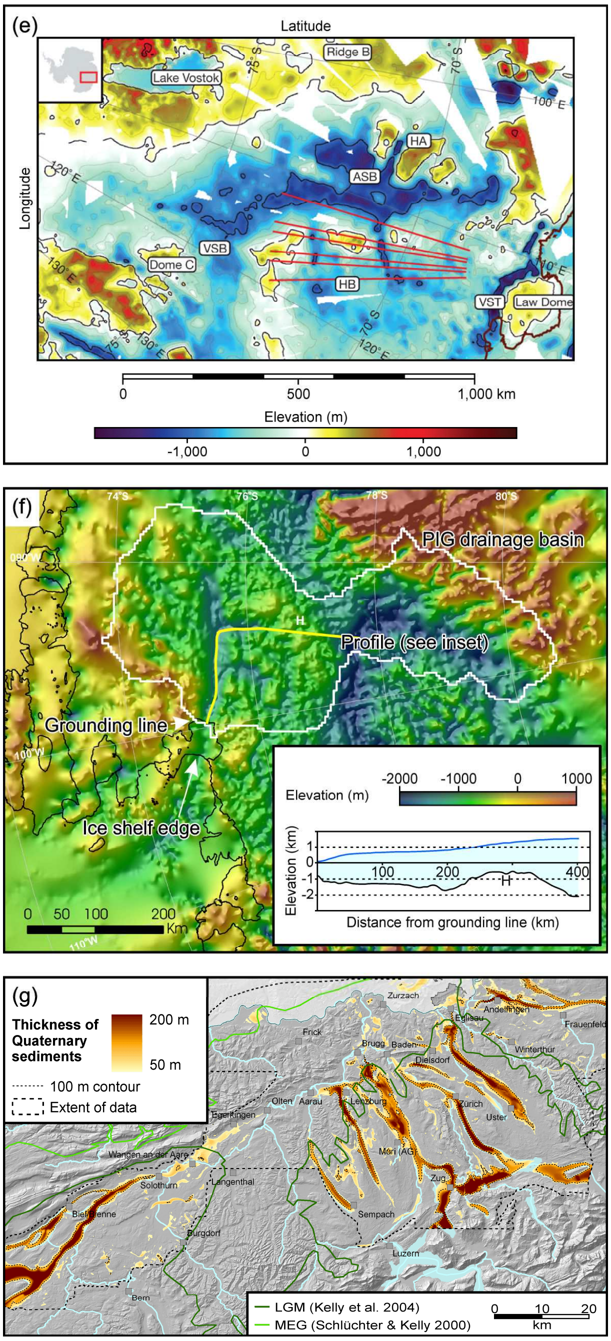

Figure 4. continued: (e) Bed topography of the Aurora subglacial basin showing smaller overdeepened basins between areas of higher topography; red lines are transects (see Young et al. 2011; reproduced with permission). (g) DEM of the bed topography of Pine Island Glacier; inset shows the ice and bed surfaces along the solid-yellow transect-line (reproduced with permission from Vaughan et al., 2006). (f) Thickness of Quaternary deposits (brown shading) showing extensive overdeepening of the Swiss Alpine Foreland; green lines indicate Quaternary and Last Glacial maximum ice extent (reproduced with permission from Fischer and Haeberli, 2010).

extent of basins that occupy the Swiss Alpine foreland has been largely obscured by postglacial sedimentation (Preusser et al., 2010) (Figure 4g).

The ubiquity of such basins and their location, typically along the main flow line of glacially deepened valleys, indicates that the majority of basins are ‘true’ overdeepenings produced by glacial erosion of subglacial sediment or rock. This type of basin is exemplified by overdeepenings in the Gamburtsev and Ellsworth Mountains, which occupy glacial valleys formed during earlier stages of

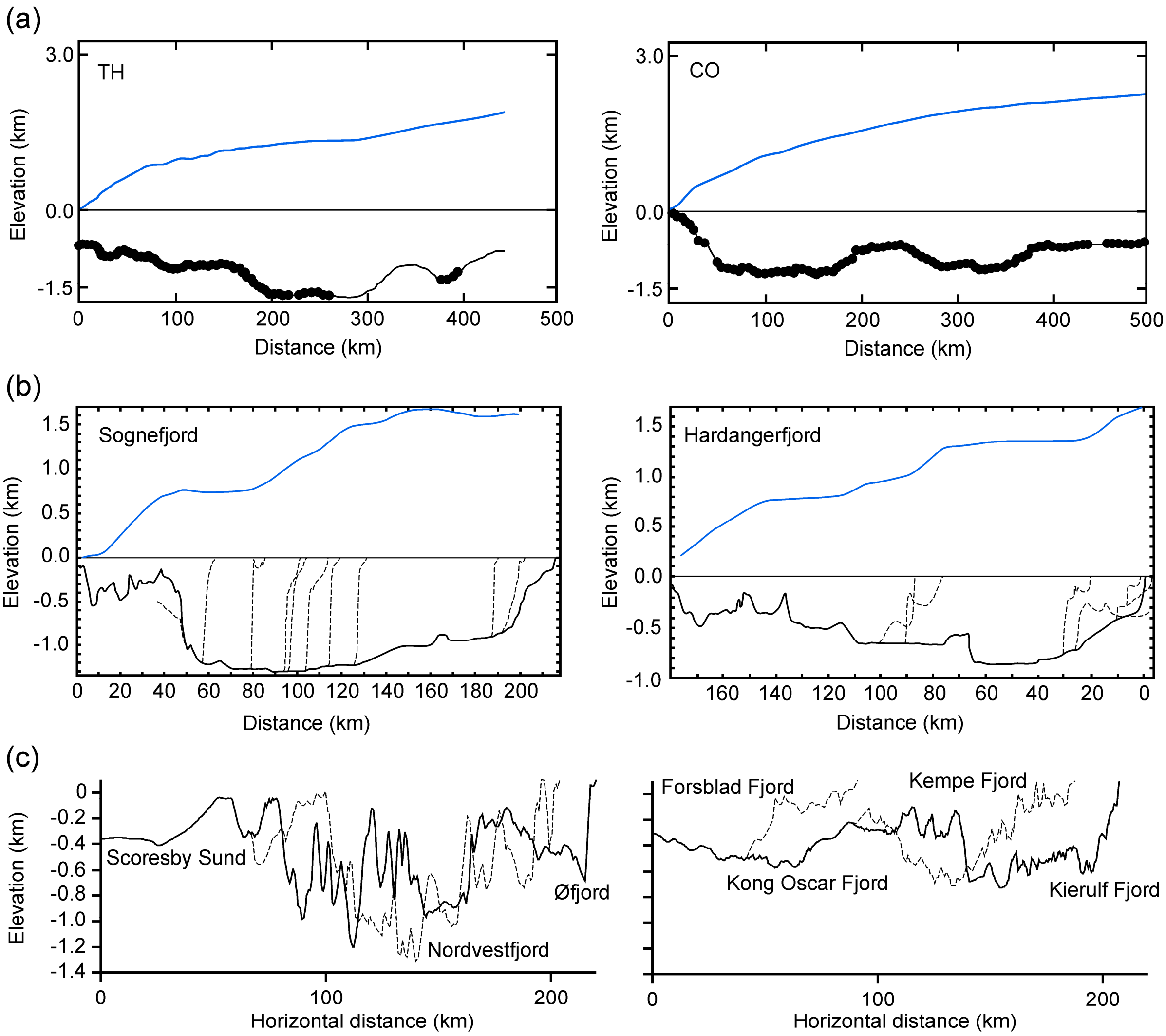

Figure 5. Bed topography of major outlet glacier systems in Antarctica, Norway and Greenland. (a) Thwaites Glacier, West Antarctica and Cook Glacier, East Antarctica; upper line shows the present ice surface (redrawn from Shepherd and Wingham, 2008). (b) Sognefjord and Hardangerfjord, Norway; upper line indicates the general maximum level of adjacent mountainous topography; stippled lines are tributary fjords (redrawn from Holtedahl, 1967). (c) The ice-free regions of the Scoresby Sund and Kong Oscar Fjord fjord systems, East Greenland (redrawn from Swift et al., 2008).

glaciation (e.g. Bo et al., 2009) (Figures 4c and d), and overdeepenings in the Erdalen valley, Western Norway (Hansen et al., 2009), and the Buttermere valley, England (Figure 2b). Even small glaciers can possess several overdeepenings, often comprising both cirque-type overdeepenings beneath (former) accumulation areas and larger trunk-type and terminal overdeepenings beneath the (former) ablation area (as at Storglaciären; Hooke, 1991). Long fjord and valley systems developed under former ice sheets and the troughs of outlet glaciers that drain contemporary ice sheets commonly possess multiple or complex overdeepenings (e.g. Holtedahl, 1967; Swift et al., 2008; Shepherd and Wingham, 2008; Jordan, 2010b) (Figure 5).

A sample of published literature (Table 1) indicates that, though trunk-type overdeepenings are common, overdeepenings also occur where glaciers are (were) confluent and, intriguingly, beneath (former) glacier termini where ice flow is (was) diffluent. Some very large basins (> 1000 km^2) are also included in Table 1 that exist beneath continental ice masses and are therefore presumed wholly or partly tectonic in origin; for example, Lake Vostok (see Bell et al., 1998; Jordan et al., 2010b) and the Wilkes Subglacial Basin (Stern and ten Brink, 1989; Ferracioli et al., 2001).

4. ORIGIN AND EVOLUTION OF BASINS BY OVERDEEPENING

Despite the pervasiveness of overdeepening in glaciated environments, the processes by which ice masses overdeepen their beds have been studied little and are understood poorly. A handful of largely theoretical studies have made significant advances, nevertheless, using insights drawn from the wider, more general body of theoretical and empirical research into the processes and patterns of glacial erosion and sediment transport. These studies indicate that cirque, valley and outlet glacier systems should tend naturally toward overdeepened bed geometries.

4.1. Key concepts and modelling approaches

Overdeepening of glacier beds is to be expected because the pattern of ice flux, which dictates largely the pattern of basal sliding, should cause erosion and sediment transport potential to peak at a location close to the long-term average equilibrium line altitude (ELA) (e.g. Hallet et al., 1996; Boulton, 1996; Anderson et al., 2006). For small cirque-type glaciers, this pattern is controlled by rotational movement about a point at the ice surface that approximates the ELA (e.g. Grove, 1958). For ice sheets, thermal conditions at the bed will also influence the pattern of erosion (Figure 6a). It follows that, whilst erosion should peak close to the ELA, glacial sediment deposition, which is likely to inhibit erosion, will increase toward the glacier terminus (Boulton, 1996; Figure 6a), and this pattern of deposition will be reinforced by melt out from basal ice layers (e.g. Hart, 1998; Boulton, 1996; Figure 6a). Maximum deepening of the bed therefore should occur near the ELA, with sedimentation predominating at the ice-margin (Figure 6b) (e.g. Boulton, 1996; Anderson et al., 2006). Because ice can move water and sediment along adverse bed slopes, deepening of an inclined glacier bed should eventually produce an overdeepening (Figure 6b), provided the ice flux remains sufficient to evacuate water and sediment from the deepening basin. Sedimentation at the ice margin during maximal and retreat stages may further enhance the depth of the basin (Figure 6b).

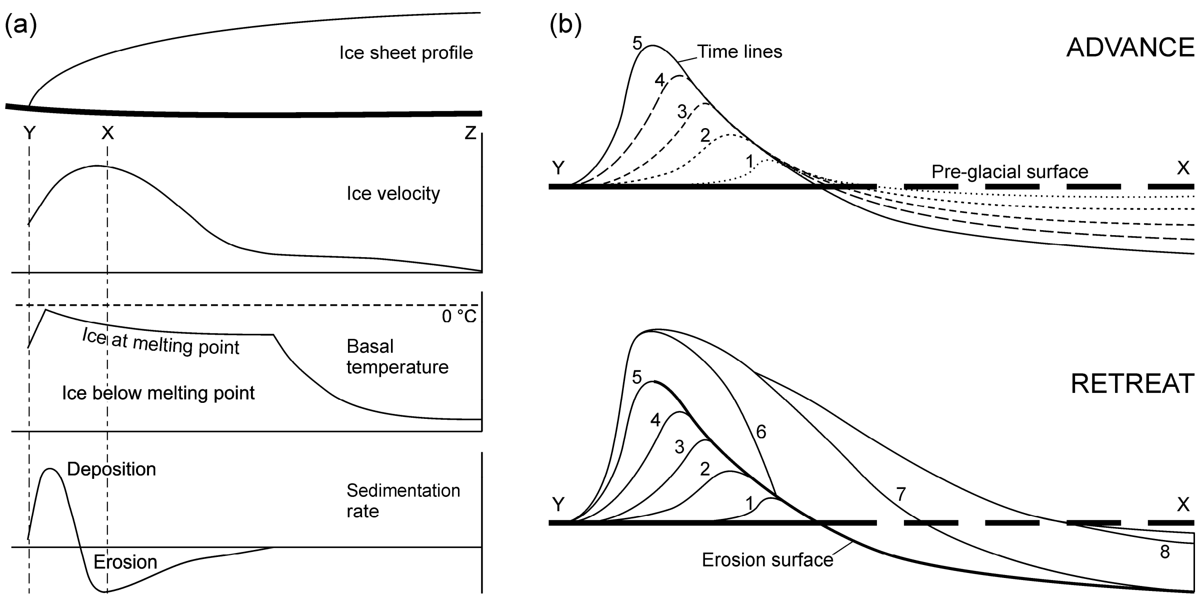

Figure 6. (a) Schematic diagram showing variables that affect glacial erosion and deposition along a flow line for a land-based ice mass, assuming that fluvial erosion and transport is negligible (modified from Boulton et al. 2001). Z and Y mark the ice mass centre and margin, respectively, and X is the approximate location of the equilibrium line, where ice velocity is greatest. Qualitatively similar patterns of velocity and basal temperature exist beneath valley and outlet glaciers. (b) Erosion surfaces and stratigraphy of deposits in the marginal zone of a land-based ice mass during a cycle of advance (above) and retreat (below) (modified from Boulton, 2006). X and Y are as in (a); numbers indicate successive time horizons.

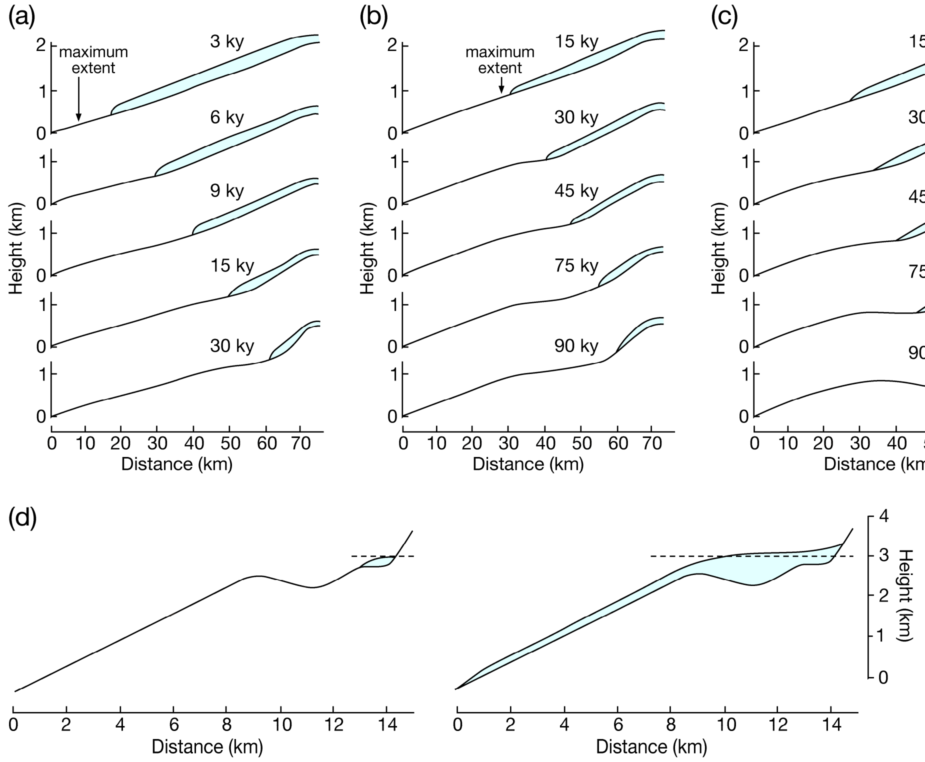

Figure 7. Evolution of ice mass and bed geometry using a simple glacier-erosion model (redrawn from Oerlemans, 1984). (a) to (c): The effects of height-mass balance feedbacks mean on bed erosion by (a) temperate ice and (b) cold-based ice when ELA and accumulation remain fixed, and (c) when the ELA is gradually lowered by 4 m per 1 ka. (d): The presence of an overdeepening creates branching of the equilibrium states, such that two very different ice mass geometries (left and right) are possible under the same climatic conditions. The larger ice mass (right) can only be simulated if it is allowed to grow beyond the overdeepening before raising the ELA to the position shown.

Numerical models that incorporate simple velocity-based erosion rules produce overdeepenings in cirques and trunk valleys with reasonable accuracy (e.g. MacGregor et al., 2000; Anderson et al., 2006). Such models also produce rock-steps and overdeepenings at confluences because ice convergence causes a local peak in velocity, as observed in nature (e.g. Gudmundsson, 1997, 1999). It follows that overdeepenings beneath cold-based or polythermal glaciers with cold beds, where sliding is limited, must develop during periods when ice is thicker (e.g. McCall Glacier; Pattyn et al., 2009). Further, local velocity peaks caused by valley morphology, and long-term changes in the glacier velocity profile caused by changes in the position of the ELA, can account for the presence of multiple individual overdeepenings in a single glacier system. However, the existence of terminal overdeepenings in areas of strongly diffluent ice flow highlights the limitations of these simple ice-erosion models. Notable examples of terminal overdeepening are the basins that occupy the Swiss Alpine foreland (Figure 4g): these must have formed as ice spilled-out across the low-relief foreland basin from the confinement of deep Alpine valleys (e.g. Preusser et al., 2010).

A further problem is that simple ice-erosion models (e.g. MacGregor et al., 2000; Anderson et al., 2006) produce only slightly overdeepened valley profiles. For example, in one of the earliest studies to use a simple ice-erosion model to simulate glacier-bed evolution, Oerlemans (1984) observed only shallow overdeepening (Figure 7a and b) because of a height-mass balance feedback in which erosion of the bed reduces the altitude of the accumulation area, causing the glacier terminus to continually retreat. Oerlemans (1984) showed that significant overdeepening was possible when climate was allowed to cool (Figure 7c) and hence the terminus position could remain stable. Nevertheless, all such models are further limited by the exclusion of subglacial water and sediment transport processes, which may stimulate further important ice-erosion feedbacks.

Three feedbacks that highlight the potential importance of water and sediment transport are discussed by Hooke (1991) and Alley et al. (2003a). Firstly, Alley et al. (1997, 2003a) argued that without the flushing of sediment by water from the ice-bed interface, the products of erosion would accumulate in a till layer that would prohibit further erosion. As a result, erosion should be most efficient not at the ELA but near the glacier terminus, where water is present in the greatest volume and flow often takes the form of hydraulically-efficient channels that have exceptionally high sediment transport capacity (Alley et al., 1997). Secondly, Hooke (1991) emphasised the importance of basal water pressure perturbations for the process of quarrying, arguing that overdeepening should occur where surface crevasses allow seasonally and diurnally peaked surface runoff to access the bed. Hooke further postulated that, because crevasses tend to form over irregularities in the bed, quarrying might initiate overdeepening formation by amplifying an initial bed-irregularity through a positive feedback loop of crevasse development, water supply, and erosion. Finally, both Hooke and Alley et al. argued that adverse slopes approaching or exceeding 1.2–1.7 times the gradient of the ice surface slope will prevent efficient flushing of sediment because water flowing in hydraulically efficient channels begins to freeze (Röthlisberger, 1972; Röthlisberger and Lang, 1987; Hooke, 1991; Creyts and Clarke, 2010) (see section 5.1.1). A slope that is twice that of the ice surface is probably sufficient to close any channels that are present and therefore prevent flushing by water (e.g. Alley et al., 2003a). A sufficiently steep adverse slope may therefore inhibit downglacier evolution of the overdeepening but may not necessarily prevent erosion (especially quarrying) at the overdeepening head.

Similar ice-water-sediment feedbacks are present in the almost visionary work of McGee (e.g. McGee, 1894) who observed that glaciers tend to 'cumulatively intensify' pre-glacial irregularities in the valley long-profile. McGee speculated that this happened because thicker ice occupying minor depressions flowed more slowly but pressed harder on the glacier bed and therefore possessed greater erosional potential. Measurements of ice velocity and subglacial water pressure (see section 5) indicate that McGee's reasoning was wrong, but his instincts regarding the presence of feedbacks were correct. Most insightfully, McGee (1894) was the first to postulate the existence of a negative feedback loop in which the formation of a sufficiently steep adverse slope would limit the depth of overdeepening. Here he asserted correctly that a steepening slope would oppose the transport of sediment out of the deepening basin, whether by glacial or fluvial processes, and that a sufficiently deep basin would result in flotation, causing overdeepening to cease. In recognising this, McGee was probably also the first to anticipate the existence of subglacial lakes.

Recent modelling has attempted to incorporate and explore such feedbacks, with some success. Herman et al. (2011) demonstrated that a simple hydrology-dependent basal sliding and erosion rule produced substantial localised overdeepening because spatially focussed surface runoff contributions enhance significantly the rates of sliding and abrasion at the glacier bed. In addition, Herman et al. observed that overdeepening may be localised further by two important feedback loops: (1) a positive ice-erosion feedback, in which erosion of the overdeepening causes the headwall to steepen and therefore further enhances sliding velocity and therefore headwall erosion; and (2) a negative height-mass balance feedback, which limits headward erosion of the overdeepening because this reduces the altitude of the accumulation area, and thus reduces ice flux (cf. Oerlemans, 1984). Modelling by Egholm et al. (2011), utilising a more sophisticated ice flow model, has further included processes of sediment transport by ice, by basal sediment deformation, and by subglacial water. The morphology of overdeepenings produced using this approach differ according to the presumed importance of the individual systems responsible for erosion and sediment transport, indicating potential to test model output (and hence process significance) against field observations. Most importantly, modelled terminal overdeepenings occur even in regions of diffluent ice flow, mainly because of the importance of sediment flushing by subglacial water, with overdeepening depth and extent being enhanced further when quarrying is made dependent on basal water pressure.

The above modelling approaches, whilst achieving huge advances, nevertheless incorporate process simplifications that may limit their applicability. Notably, for computational reasons, Herman et al. (2011) and Egholm et al. (2011) use simplified models of englacial and subglacial hydrology, and specify only seasonal or long-term average hydrological conditions. As a result, ice-water-quarrying feedbacks (cf. Hooke, 1991, above) cannot be modelled or tested because they are dependent on crevasse-controlled hydrological connections between the ice surface and bed and the strong diurnal variation in basal water pressure that such connections instil (Egholm et al., 2011). Further, in Egholm et al. (2011), net sediment transport by efficient channels is almost certainly underestimated because: (1) the relationship between basal sediment evacuation and discharge is extremely non-linear and subglacial discharge is diurnally highly variable (e.g. Swift et al., 2005); and (2) the temporal and spatial distribution of channel closure by glaciohydraulic supercooling versus enlargement by melting is highly dependent on diurnal hydraulic gradients (Creyts and Clarke, 2010) and only loosely dependent on the ice-surface slope to ice-bed slope ratio (see section 5).

4.2. Alternative and speculative processes

Theoretical and empirical studies indicate a variety of further feedbacks that favour overdeepening and the enlargement of existing basins or glacier bed irregularities. First, the presence of a sufficiently large basin will cause topographic steering and acceleration of ice flow into the basin, causing thicker, faster-flowing ice to result in further deepening (e.g. Kessler et al. 2008). Second, thicker, faster-flowing ice will generate greater quantities of basal melt that will further enhance basal sliding and sediment flushing (e.g. Jamieson et al., 2008). Third, the steering of ice flow and focussed delivery of surface melt via the preferential location of crevasses (cf. Hooke, 1991) will focus heat-transport into basins and thereby further elevate melting and sliding (e.g. Jamieson et al., 2010) (these effects may be especially significant at polythermal ice masses). Fourth, steeper adverse slopes should encourage the elevation of sediment into glacial transport pathways via thrusting (e.g. Swift et al., 2006) and/or glaciohydraulic supercooling (e.g. Cook et al., 2007, 2010; Creyts and Clarke, 2010). Interestingly, modelling work by Creyts and Clarke (2010) has indicated that, where the ice surface and bed slope thresholds for glaciohydraulic supercooling are met, the volume of ice accreted by supercooling will typically be less than that lost by subglacial melting, meaning that glaciers with overdeepenings will tend toward even lower surface slopes and even greater supercooling (see section 5.1.1).

Several alternative explanations for overdeepening exist. An intriguing possibility is that stepped valley profiles consisting of cirques and overdeepenings result from a wave-like instability in ice-flow. For example, Mazo (1989) used a numerical ice-erosion model to show that an initially rough bed evolved to form a repeating pattern of overdeepening-like depressions, the dominant wavelength of which was $2/\tan\theta$ times longer than the ice thickness, where θ is the mean bed-slope. This wavelength was observed to approximate the scale and spacing of cirques and overdeepenings in real glacial valleys. Similarly, Comeau (2009) showed that a more advanced 3-dimensional model, incorporating more realistic sliding and erosion rules, produced physically realistic chains of overdeepenings on modelled timescales of $\sim 10^6$ years that again exhibited depth and spacing characteristic of that observed in nature.

A further mechanism that is viewed as an important limitation of current ice-erosion models (e.g. Egholm et al., 2011) is the source and availability of abrasive tools. This has been discussed by Lliboutry (1994), who argued that quarrying could not be an efficient process of erosion because it requires bedrock to be weakened by frost shattering during interglacial periods. Quarrying is therefore an implausible mechanism of overdeepening of fjords. Instead, Lliboutry (1994) calculated the pattern of erosion that would be produced by the grooving action of clasts in glacial transport, and showed that angular clasts falling onto the glacier near the bergschrund could, as a result of basal melting, come into contact with the glacier bed, where grooving of bedrock would slowly blunt them. Lliboutry calculated that this would result in a peak in grooving efficiency at a characteristic distance from the valley head, and hence the slow formation of an overdeepening without the need for quarrying, although he neglected to consider the influence of basal sediment layers.

Geological controls on the location of overdeepening may also be significant. For example, Swift et al. (2008) observed that overdeepenings in East Greenland fjords (Figure 5c) were generally deepest, and the fjords narrowest, in regions with the most resistant bedrock. Preusser et al. (2010) observed that tectonic structures and weaker lithologies appear to control largely the location of

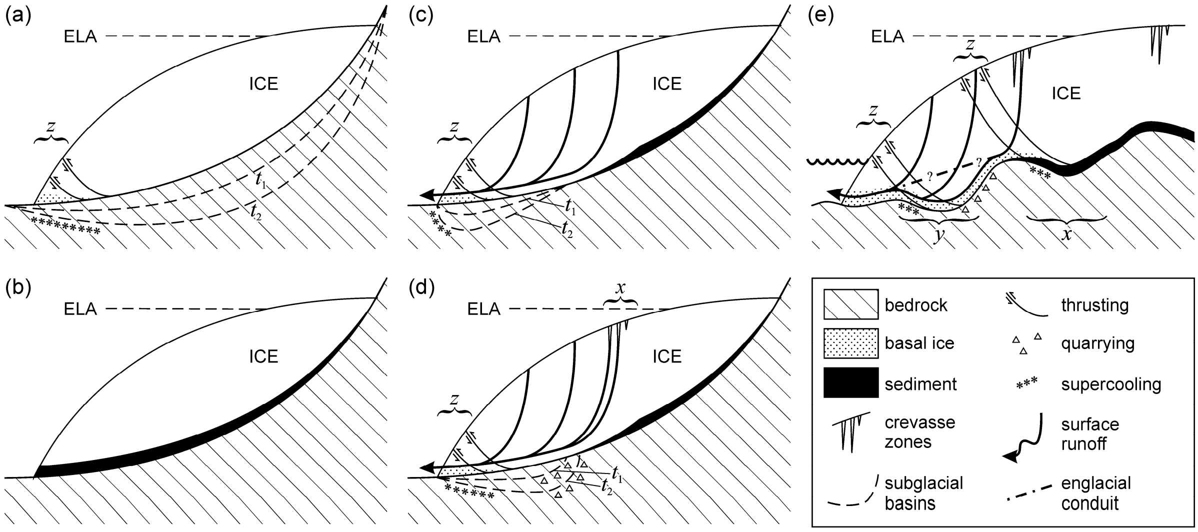

Figure 8. Schematic diagrams summarising the glaciological processes associated with overdeepening. Dashed lines indicate time horizons during the erosion of major subglacial basins. Note: (1) vertical scale and size of the ablation area have been exaggerated; (2) water flow at the ice-bedrock interface is assumed to be hydraulically inefficient, except where conduits carrying surface runoff reach the bed; and (3) basal ice and sediment layers are omitted except where they are predicted to be of significant thickness and extent. (a) Assumed symmetrical pattern of overdeepening that should occur if there is no hydrological forcing of basal sliding and the products of erosion are evacuated efficiently from the glacier bed. (b) No overdeepening as a result of erosion products having accumulated to form a thick basal sediment layer that protects the bed. (c) Strong overdeepening beneath the ablation area, particularly beneath the terminus, as a result of surface runoff enhancing both basal sliding and the efficiency of basal sediment evacuation. (d) Strong headward evolution of an overdeepening because focussed delivery of surface runoff and local steepening of the bed enhances quarrying and abrasion at the overdeepening head. (e) Multiple overdeepenings and complex spatial patterns of erosion resulting from variation in climate and hence ELA position and glacier extent.

overdeepenings within the European Alps, whilst the overdeepenings of the Alpine foreland, which do not appear to have any geological control, end abruptly on reaching the Mesozoic limestone of the Jura (see also Jordan, 2010). Brückl et al. (2010) found that overdeepenings in Austria do not always occur at confluences and suggested that a combination of high basal water pressure and tectonically weakened bedrock, favouring quarrying, might be important, whilst Jansson and Hooke (1989) noted that overdeepenings at Storglaciären appear to be located between outcrops of more resistant bedrock, which may have provided the initial bed irregularities required for surface crevassing to initiate focussed erosion. Apparently strong geological controls on valley/fjord orientation (e.g. Glasser and Ghiglione, 2009) will also influence overdeepening through the location of glacial confluences and the general pattern of ice discharge and valley/fjord evolution.

4.3. Synopsis

Theoretical work provides a framework for understanding the origin and evolution of overdeepening that can be validated using numerical ice-erosion models and comparison of model output with field observations. Maximum deepening of the bed should occur beneath the ELA (Figure 8a), but the accumulation of basal sediment layers will protect the bed from erosion (Figure 8b). The focus of overdeepening should therefore shift away from the ELA, where surface runoff contributions are low and basal sediment forms thick layer, and towards the glacier terminus (Figure 8c), where surface runoff contributions are most substantial, channelised drainage is seasonally-present, and rates of basal sediment flushing are therefore highest (Alley et al., 1997). Where suitable bed irregularities are present, quarrying might also initiate overdeepening, and quarrying and other ice-erosion feedbacks should produce strong headward erosion of overdeepenings (Figure 8d). Together with the diachronous nature of glaciation, these processes are capable of producing complex glacier-bed profiles consisting of multiple overdeepenings of differing origin and relative age (Figure 8e).

Height-mass balance feedback (cf. Oerlemans, 1984), surface runoff availability and the existence of a glaciohydraulic supercooling threshold appear to be strong controls on the extent, depth and geometry of overdeepening. Although the Creyts and Clarke (2010) question the existence of a single glaciohydraulic supercooling threshold, it seems likely that the beds of temperate glaciers should evolve toward an overdeepened geometry in which the gradient of the adverse slope is controlled by the ice-surface slope and the upglacier extent of overdeepening is constrained by ice flux and the availability of surface melt (as shown schematically in Figure 8d). Feedbacks between topography, ice flow, thermal regime and sliding will help to reinforce overdeepening development. Interestingly, although quarrying-related ice-water-erosion feedbacks may initiate multiple overdeepenings in the same glacier bed, efficient headward erosion by quarrying and ice-erosion feedback should cause downglacier overdeepenings to coalesce with their upglacier counterparts (cf. Alley et al., 1999). It remains, however, for the entrainment, transport and deposition of sediment by glacial and fluvial processes to be implemented fully and accurately in glacier erosion models, meaning that testing of model output against field data should be undertaken with caution.

5. THE GLACIOLOGICAL IMPORTANCE OF SUBGLACIAL BASINS

Landforms play an important role in glacial systems and landscapes: at the largest scale, they influence glacier geometry, patterns of ice flow, and basal thermal regime. Basins are likely to be especially significant because they require fundamental components of the glacial system, specifically ice, water and sediment, to be transported along adverse gradients, and may therefore limit the efficiency of key glaciological processes. Further, because the importance of surface melt for erosion indicates that overdeepening will typically be focused beneath the ablation area (section 4.1), adverse slopes will be common near to glacier margins and basins will intercept subglacial drainage systems that transport significant quantities of runoff.

Basins, especially overdeepenings, may therefore be located in dynamic and sensitive regions of the glacial system, and their presence might be expected to have far-reaching consequences for glacier hydrology, ice dynamics, and ice mass stability. Further, associated changes in rates and spatial patterns of subglacial erosion and sediment transport, which in some circumstances act to reinforce and/or limit the process of overdeepening, will influence a wide range of glacial geomorphic processes, including the ice-evacuation efficiency of glacial troughs and the glacial and postglacial evolution of the wider landscape.

5.1. Glacier hydrology and seasonal dynamics

5.1.1. Glacier hydrology and basal water pressure

Hooke (1991) proposed a positive feedback mechanism whereby the formation of crevasses above bedrock irregularities focussed the delivery of diurnally peaked surface melt to the head of incipient overdeepenings (Figure 8d). This mechanism highlights the potentially significant role of glacier-bed geometry in controlling the spatial pattern of diurnal variability in basal water pressure and hence rates of sliding and quarrying. The importance of crevassing for moulin development (Stenborg, 1969; Benn et al., 2009), and the link between crevasse formation and extensional ice-flow (Figure 9), indicates that overdeepenings undoubtedly exert a strong control over both the routing of water at the ice surface and the distribution and density of connections between the surface and bed. Overdeepening geometry might also be important in directing internal glacial water flow, especially at or near the bed (e.g. Hooke et al., 1989; Hubbard et al., 2004). However, the requirement for water to exit an overdeepening along an adverse slope has produced sustained uncertainty regarding the style and spatial pattern of subglacial drainage in overdeepened sections of the bed (e.g. Röthlisberger and Lang, 1987). Such uncertainty is undesirable because mean basal water pressure

Figure 9. Moulins located at the upper-limit of a region of extensional crevassing (foreground) on Svínafellsjökull, Iceland, indicating the approximate upglacier extent of Svínafellsjökull's terminal overdeepening. Photo: D.J. Graham.

and the magnitude of diurnal basal water pressure variability are known to be determined by both the magnitude and peakedness of melt reaching the bed and the nature of the drainage system at the ice-bed interface (e.g. Hubbard and Nienow, 1997; Mair et al., 2003).

Borehole studies at Storglaciären, Sweden have demonstrated that basal water pressure in the main overdeepening is consistently high, averaging ~90% of ice overburden, and that pressure fluctuations are limited, even in the presence of significant variability in surface melt and precipitation (e.g. Hooke et al. 1989; Jansson 1995). In contrast, areas downglacier of the overdeepening at Storglaciären exhibit greater variability in basal water pressure, and transit times of water through the glacial drainage system are shorter (Jansson, 1995). These observations indicate the persistence of hydraulically inefficient drainage within the overdeepening, whilst downglacier areas are free to develop hydraulically efficient channelized drainage system in response to the seasonally evolving contribution of surface-derived melt (Jansson, 1995). Notably, Jansson and Hooke (1989) attributed surface uplift of Storglaciären to slow-moving waves of high water pressure that originated from the crevassed area at the head of the overdeepening; the waves moved down-glacier at speeds of only 20–60 m h^{-1} and only became sufficiently focused to cause uplift at the downglacier end of the overdeepening, 'either due to topography or conduit restrictions'. Moreover, tracer experiments have indicated that the majority of water traversing the main overdeepening at Storglaciären bypasses the overdeepening via an extensive englacial conduit system (e.g. Hooke et al., 1988; Hooke and Pohjola, 1994; Fountain et al., 2005), apparently because englacial flow is more efficient than flow at the bed, which seems to occur 'sporadically' via multiply-interconnected 'storage pockets' at the ice-till interface (Hooke and Pohjola, 1994).

Studies at many other glaciers with overdeepened sections have found similar evidence for inefficient subglacial drainage and relatively efficient englacial drainage (e.g. Hantz and Lliboutry, 1983; Hodge, 1976, 1979; Fountain, 1994). For example, Hock et al. (1999) inferred an extensive 'sub-surface' (presumably therefore englacial) drainage system in the region of a major overdeepening at Grosser Aletschgletscher, Switzerland. Dye injected into a moulin emerged rapidly at the terminus, conceivably having followed an englacial pathway (though the authors are not specific), whereas dye injected at the base of nearby boreholes indicated very poor connectivity to the glacier drainage system. Similarly poor subglacial connectivity was found by Iken et al. (1996) for boreholes drilled near to the centre of an overdeepening at Gornergletscher, Switzerland, despite boreholes drilled near the lateral margin being relatively well connected. In addition, seasonal uplift of the glacier surface was found to be somewhat smaller near the centreline than at the margin, indicating that surface-derived melt was less able to reach the bed of overdeepened areas, conceivably because of capture by englacial flowpaths. At Matanuska Glacier, Alaska, subglacial drainage has been suggested to occur via a shallow canal system that becomes increasingly braided during seasonal increases in surface-derived melt (Ensminger et al. 1999), whilst relict sediment-filled englacial-conduits at Gigjökull and Steinholtsjökull, Iceland (Kirkbride and Spedding, 1996) hint at a significant component of flow via more efficient englacial pathways.

Though subglacial drainage within overdeepenings appears therefore to be hydraulically inefficient, this is true of all areas of the glacier bed where hydraulically efficient channels are not already present. Hydraulically efficient channels convey water rapidly, thereby lowering basal water pressures, but only occur when surface runoff contributions are sufficiently peaked to destabilise pre-existing distributed flowpaths (Kamb, 1987; Nienow et al., 1998). Of critical importance, therefore, is whether seasonal melt can reach the bed of overdeepened areas. Direct access of melt to the bed is perhaps unlikely in light of evidence for pervasive englacial networks in overdeepened areas (e.g. Hooke and Pohjola, 1994; Hock et al., 1999). Indirect access, however, is extremely likely where surface melt that has already reached the bed must traverse an overdeepening. Important questions, therefore, are whether flow continues to follow the bed of the overdeepening or whether it is diverted to follow englacial flowpaths, and, assuming not all flow becomes englacial, whether flow traversing the overdeepening remains in hydraulically efficient flowpaths or 'collapses' to join flow in co-existing high-pressure distributed systems.

Theory indicates that englacial conduits that span overdeepenings should be relatively common. This is because flow within the glacial system occurs mainly at atmospheric pressure, therefore englacial conduits will rapidly descend by preferential melting of the conduit base, and subglacial channels will follow the lowest possible pathway at the glacier bed (Lliboutry, 1983; Hooke, 1984). However, where a descending englacial conduit impinges upon the bedrock lip at the downglacier end of an overdeepening, descent of the conduit upglacier of the lip must cease because the conduit will become water-filled and melting will occur equally on all sides (Hooke, 1984; Fountain and Walder, 1998). Further, if upglacier and downglacier of the overdeepening the conduit descends fully onto bedrock, the result will be a largely subglacial channel with an englacial section that spans the overdeepening. Thus, englacial conduits should be present and should capture and divert at least some surface-derived melt away from or across the bed of overdeepened areas. The englacial section of any such conduit would presumably follow the hydraulic grade line and could therefore be called a "gradient conduit" (Röthlisberger, 1972), which describes the highest possible flooded conduit in which flow remains at atmospheric pressure.

There is greater uncertainty regarding what happens to flow within an existing subglacial channel should it meet an overdeepening. Röthlisberger (1972) showed that water pressure within a steady-state channel that is 'full' would rise toward ice-overburden pressure as it flowed through the overdeepening. Thus, as channel water pressures approach that of the adjacent distributed system, flow in a single, confined channel should become less favourable, and flow will begin to diverge (Kirkbride and Spedding, 1996; Creyts and Clarke, 2010). Equalisation of channel water pressure with ice-overburden pressure occurs when the gradient of adverse slope approaches twice that of the ice surface slope (Röthlisberger, 1972). Above this threshold, flow should favour all types of flowpath equally, and is likely to distribute across the bed to flow as a sheet (Röthlisberger, 1972; Creyts and Clarke, 2010). The same flow might also exploit existing englacial pathways (Röthlisberger and Lang, 1987), especially if the transmissivity of such flowpaths is greater than that of the subglacial system (Flowers and Clarke, 2002). For adverse slopes exceeding 11 times the magnitude of the ice-surface slope, water would be unable to exit the overdeepening and would be forced to pond to form a subglacial lake (Paterson, 1994; Clarke, 2005), in which case water must follow a different route or 'flow across the 'pond'' (Creyts and Clarke, 2010, p. 8).

Whether flow remains subglacial when the adverse slope exceeds twice that of the ice surface slope is particularly unclear. Lliboutry (1983) argued that, in the case of two flooded flowpaths, the highest flowpath should always carry the most water, such that flow divergence should occur that would lead to flow bypassing overdeepenings via englacial conduits or via lateral flow along the flanks (Fountain and Walder, 1998). Although Röthlisberger and Lang (1987) disagreed with Lliboutry's analysis, field evidence for lateral flow exists (Hantz and Lliboutry, 1983; Fountain, 1994), and Röthlisberger and Lang (1987) noted that this conclusion was not *a priori* wrong. Field evidence also exists in support of englacial drainage across overdeepenings, and Hooke & Pohjola (1994) have argued that differences in the shape and roughness of subglacial and englacial channels means that flow in englacial conduits should be preferred. First, because the form of englacial conduits is likely to be circular and subglacial channels are likely to be broad and low, englacial flowpaths should possess lower water pressures (see also Hooke et al., 1990). Second, sediment- or bedrock-floored subglacial channels are likely to have greater roughness than smooth-walled englacial conduits, which again indicates that englacial conduits would possess lower water pressures, even if both subglacial channels and englacial conduits were of similar form.

Englacial flow is also likely to be promoted by the effects of glaciohydraulic supercooling, which can occur within subglacial drainage systems that traverse an overdeepening (Röthlisberger 1968, 1972; Röthlisberger and Lang 1987; Hooke, 1989; Hooke and Pohjola 1994). Here, when an adverse slope exceeds 1.2–1.7 times the ice surface slope, water becomes supercooled because the pressure melting point rises faster than the water can be warmed by viscous dissipation, which causes some of the water to freeze (Röthlisberger 1972; Röthlisberger and Lang 1987; Hooke, 1991). Adverse slopes beneath Matanuska Glacier, Alaska, where the effects of supercooling have been best documented, comfortably exceed this threshold bed-slope to surface-slope ratio (Creyts and Clarke, 2010). The exact threshold is dependent upon the air-saturation state of the flow (Hooke 1991; Lawson et al. 1998; Alley et al. 2003a), but also flowpath size and the variability of surface melt (Creyts and Clarke, 2010). Flowpath size is significant because larger flowpaths (i.e. large channels) conduct more water per unit wetted perimeter than smaller flowpaths (e.g. smaller channels or distributed systems) and hence viscous dissipation is lower (Creyts and Clarke, 2010). Consequently, ice-accretion and frazil-ice formation should occur preferentially within channels, causing the transmissivity of the subglacial drainage system to reduce, basal water pressures to increase, and flow to distribute across the bed. Melt variability is important because hydraulic gradients, which determine flow velocity and hence viscous dissipation, are dependent upon the transmissivity of the subglacial drainage system and the volume of flow. Consequently, growth of channels in response to the seasonal increase in surface melt can initiate supercooling, but supercooling will switch on/off diurnally in response (Creyts and Clarke, 2010). These processes indicate complex seasonal and diurnal changes in subglacial drainage system morphology and basal water pressure (Creyts and Clarke, 2010; see Table 2 and Discussion section 6.2).

Arguments in favour of predominantly englacial flow across overdeepenings are countered by recent field evidence for flood-induced uplift of the surface of Skeiðarárjökull, Iceland (Magnusson et al., 2011). Observations demonstrated that uplift was highly localised and correlated with regions of adverse slope along the subglacial course of the river Skeiðará (Figure 10a and b). Calculation of channel melt-rate ability along the course of the river (Figure 10c and d), which along adverse slopes approximates the energy generated by viscous dissipation that exceeds that required to warm the water to the pressure melting point, demonstrated that uplift occurred where channels possessed little or no excess energy to adapt to the increased discharge through the melting of channel walls. This implies that rapid equalisation of channel water pressures with those of the adjacent distributed system, which may have been enhanced by glaciohydraulic supercooling, encouraged floodwaters to distribute across the bed. Post-flood subsidence further confirmed that uplift was largely by temporary storage of water and not the accretion of supercool-origin basal ice (see section 5.1.2). Whether flood events also promoted englacial flow is unknown.

In summary, observations indicate that crevassing focuses the delivery of surface runoff to the headwalls of overdeepenings, and it has been argued that flow across the overdeepening can occur via low-pressure subglacial, englacial or lateral pathways (Figure 11a). Nevertheless, the morphology

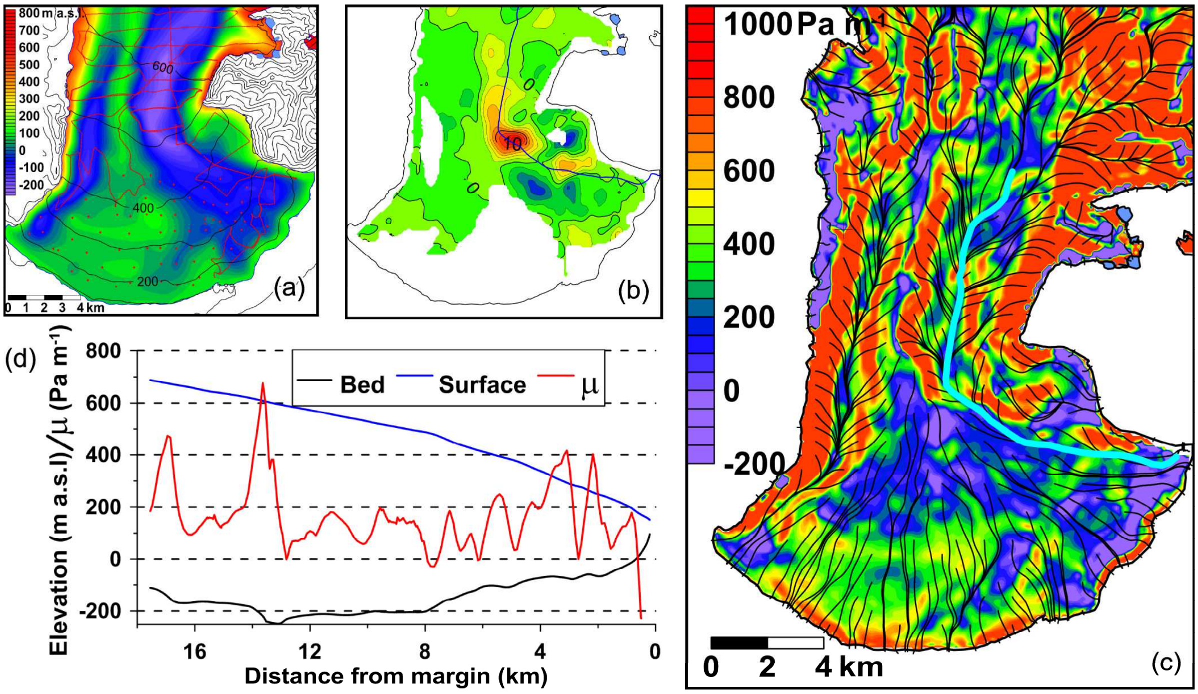

Figure 10. Hydrology and ice dynamics during a flood of the river Skeiðara at Skeiðarárjökull, Iceland (reproduced with permission from Magnusson et al., 2011). (a) Bed topography based on radio-echo sounding (red points and profiles); contours show surface elevation. (b) Vertical velocity (cm d–1) showing local uplift at the beginning of the glacier outburst flood; thin sold line indicates the estimated subglacial course of the river Skeiðara. (c) Colour scale shows melt-rate ability (μ) (see text); black curves show watercourses predicted by subglacial hydraulic potentials. (d) Melt-rate ability (μ) and ice surface and bed elevation along the estimated main course of the river Skeiðara.

of such pathways, and their evolution, is understood very poorly. In particular, subglacial flowpaths that exit overdeepenings should be less efficient, such that adverse slopes exceeding 1.2–1.7 times the ice surface gradient should be characterised by highly inefficient and hence high-pressure distributed flowpaths, and even steeper slopes should result in ponding (Figure 11b). Nevertheless, englacial flowpaths may capture surface runoff before it reaches the glacier bed, and high basal water pressures within overdeepenings may encourage subglacial flowpaths, swollen with surface runoff, to follow lower-pressure englacial ones (Figure 11b). Further, the morphology of flowpaths at the bed may vary seasonally and diurnally in response to changing hydraulic gradients.

5.1.2. Processes and patterns of ice flow

5.1.2.1 Glaciers terminating on land

Efficient basal sliding requires frequent variations in basal water pressure that approach and exceed the ice overburden pressure. Evidence that overdeepenings are characterised by less hydraulically efficient (e.g. distributed) subglacial drainage pathways, whilst remaining susceptible to forcing by seasonal and diurnal variation in surface melt, indicates that overdeepenings could play an important role in glacier dynamics. Further, overdeepenings might be expected to influence glacier flow by encouraging the formation of basal ice and till layers, by introducing steep changes in bed gradient and valley width, and by providing locations favourable to the pooling of subglacial water. The majority of work relating to the processes and patterns ice flow in the context of overdeepening has been undertaken at Storglaciären, where observations support the sensitivity of glacier flow to surface melt, highlight the potential importance of till layers, and illustrate the direct influence of bed geometry.

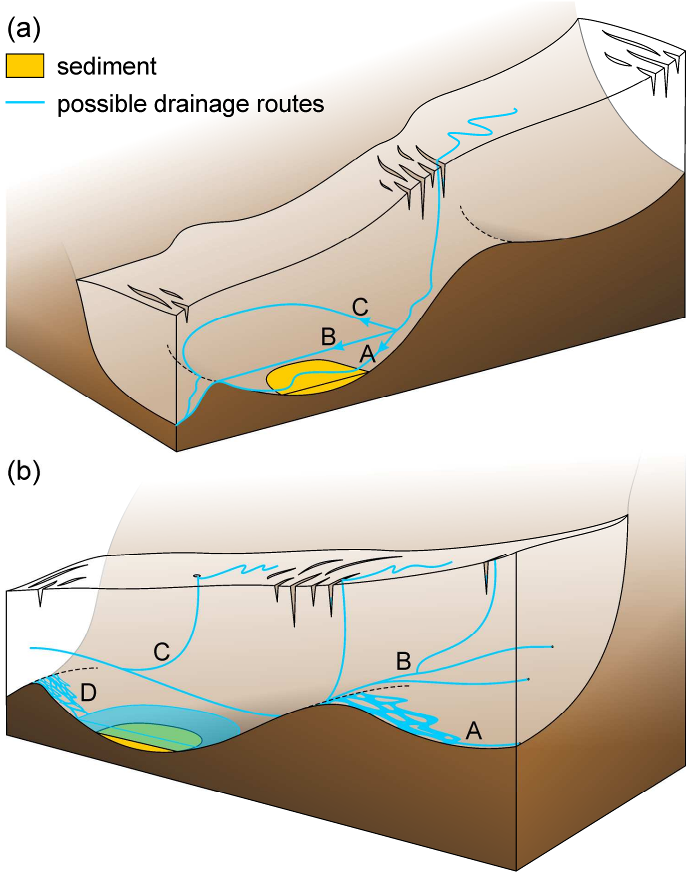

Figure 11. Schematic diagrams showing possible glacial drainage system pathways and morphologies in the presence of subglacial basins and/or overdeepenings. (a) Surface and/or subglacial runoff that reaches a basin or overdeepening may follow subglacial (A), englacial (B) or lateral (C) pathways. The morphology, evolution and stability of such pathways is understood poorly (see text). (b) The morphology and efficiency of subglacial flowpaths that ascend an adverse slope (A) is determined largely by the ratio of the ice surface to ice bed gradients (see text): steep adverse slopes reduce drainage system transmissivity and raise basal water pressures within the overdeepening, and adverse slopes that exceed 11 times the ice surface slope will cause ponding and the formation of a subglacial lake (D). Englacial conduits that traverse an overdeepening may intercept surface runoff before it can reach the bed (B) and may divert subglacial and surface across basins and overdeepenings (C).

Sensitivity of flow to surface melt is supported by numerous observations. Hooke et al. (1989, 1992) found that sustained high basal water pressures in the main overdeepening enabled short-lived increases in water input to increase the rate of basal sliding by 16–40% relative to winter values, whilst Hanson and Hooke (1994) demonstrated that individual rainfall events caused velocity in the region of the uppermost overdeepening (i.e. the ‘cirque’ overdeepening) to increase by up to three times the annual mean. Further, Hooke et al. (1987) and Jansson and Hooke (1989) reported both diurnal uplift and increases in sliding velocity at the downglacier end of the main overdeepening that appeared to result from diurnal pressure-waves originating from the crevassed-area at the overdeepening head. Hooke et al. (1987) further observed that the sliding velocities were five times those at the ice surface, resulting in extrusion flow over the prominent riegel below the main overdeepening, and attributed this to decoupling of ice from the bed over parts of the riegel as a result of cavity-opening on the adverse slope. Interestingly, in terms of seasonal evolution of subglacial drainage and its dynamical effects, Hooke et al. (1992) observed that sliding in the overdeepened section was particularly significant in spring when surface melt was believed to reach an even more restricted ‘winter’ drainage system.

Velocity studies show that non-overdeepened areas at Storglaciären also demonstrate sensitivity to surface melt, but sensitivity diminishes seasonally in response to the development of a hydraulically efficient channel system (Jansson, 1995, 1996). Nevertheless, ice-surface velocities up- and down-glacier of the prominent riegel are often similar, indicating that the effects of basal forcing can be transmitted between overdeepened and non-overdeepened areas by longitudinal stress coupling (Jansson, 1995, 1996). Notably, Hooke et al. (1992) found that ice-bed decoupling sometimes occurred in a very restricted area near the head of the overdeepening and that longitudinal stress coupling was important in transferring enhanced flow velocities both down-glacier and toward the lateral margins. In addition, Jansson and Hooke (1989) observed abrupt movement of tilt-sensors towards the glacier centreline shortly after the beginning of heavy rainstorms, which they attributed to 'longitudinal stretching as the part of the glacier below the riegel accelerates faster than that above'. Extensional strain of this kind across the riegel was also demonstrated by Jansson (1997).

Observations at Storglaciären also support the idea that glacier motion and velocity will be influenced by the formation of a till layer within the overdeepening because till is able to deform and thereby reduce basal drag (Hooke 1991; Alley et al. 1999, 2003 a, b). Notably, Hooke et al. (1989) suggested that the presence of a till layer could help explain enhanced ice flow in the region of the main overdeepening because the shear strength of till will be reduced significantly by the presence of characteristically high basal water pressures. However, later work by Iverson et al. (1995) has indicated that the degree of ice-till coupling can actually decrease when water pressures approach ice-overburden pressures, indicating that basal sliding is likely to remain the dominant flow mechanism. Interestingly, Creyts and Clarke (2010) rejected the idea that supercooling in the region of the adverse slope plays any role in elevating basal water pressures inside the overdeepening, apparently because ice surface and adverse slopes do not provide the conditions necessary, favouring instead the low hydraulic conductivity of flowpaths at the ice-bed interface.

Studies at Storglaciären have also considered the geometry of the overdeepening itself. As well as extensional flow into the overdeepening, Hooke et al. (1989, 1992) observed a peak in vertical velocity near the centreline of Storglaciären that was attributed to longitudinal compression (and thus departure from bed-parallel flow) caused by the prominent riegel and the funnelling effect provided by convergent bed contours at the downglacier end of the overdeepening. Glasser et al. (2003) demonstrated how the structural glaciology of Storglaciären probably reflects these changes in bed geometry (Figure 12a), with primary stratification being deformed into a loop structure as ice accelerates into the main overdeepening, and arcuate bands being markedly more extended in the downglacier direction as the glacier enters the lower overdeepening. Furthermore, crevasses at Storglaciären that are formed as ice accelerates into the overdeepenings close shut as a result of flow compression along adverse slopes, although terminal flow compression at Storglaciären is also enhanced by the cold-based ice margin.

Glasser et al. (2003) also investigated the origin of prominent debris-charged ridges near the northern margin of Storglaciären (Figure 12f), concluding that these were thrust-faults carrying basal ice and sediment that had formed in response to longitudinal compression at the transition of warm- to cold-based ice. However, Moore et al. (2011) rejected evidence for a significant transition in basal thermal regime at this location, finding instead that departure from bed-parallel flow was greatest in the region of a small adverse change in bed-slope. Thrusting arising from a warm- to cold-bed transition is well documented (e.g. Harris and Bothamley, 1984; Alley et al., 1997; Hambrey et al., 1997; Glasser and Hambrey, 2002), but the findings of Moore et al. (2011) indicate that overdeepenings alone can produce thrusting on glacier-thickness scales, at least in the context of valley glaciations where lateral stress-coupling is not significant. Swift et al. (2006) at the temperate Icelandic glacier Kvíárjökull have described glacier-wide thrust-faults (Figure 12c) that, although appearing to exploit existing ogive-parallel foliation, contain water-worked clasts that have presumably been scavenged from a lag-deposit in the overdeepening (Figure 12e). Folding in glacier ice is already commonly associated with irregular bed-geometry (Hambrey and Lawson, 2000; Hooke, 2005), and is therefore likely to be significant in regions of overdeepening.

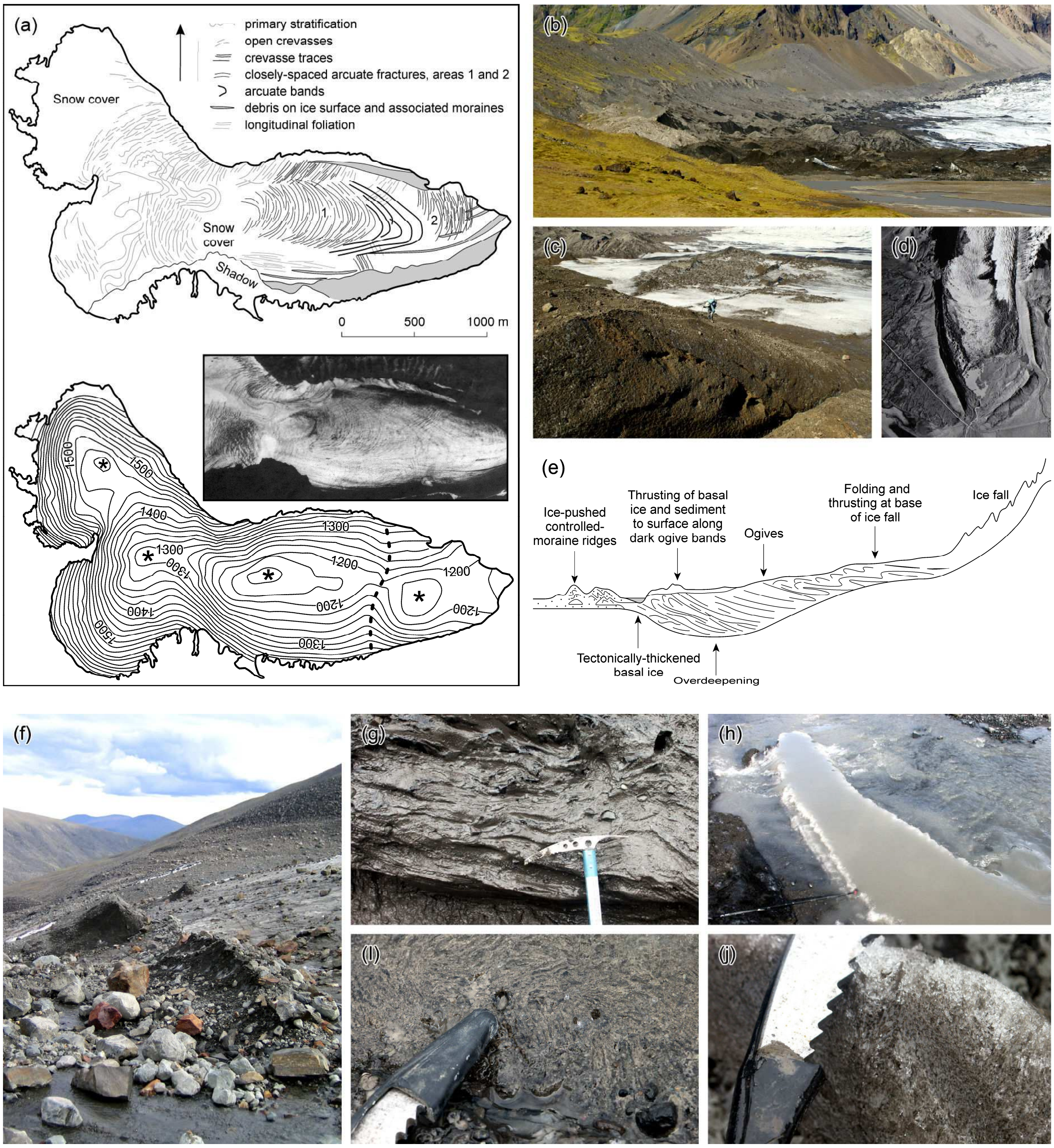

Figure 12. Structural and geomorphological implications of overdeepening. (a) Structural glaciology of Storglaciären, Sweden (above) and bed topography (below), the latter showing four over-deepenings (starred); the lower two overdeepenings are separated by a prominent riegel (thick dashed line) (modified with permission from Glasser et al., 2003). (b) Abundant debris at the margin of Kvíárjökull, Iceland, sourced partly from the melt-out of debris-rich englacial thrusts (c) that contain basal ice and debris of both glacial and fluvioglacial origin (photos: D.A. Swift). (d) The Little Ice Age moraine at Kvíárjökull (aerial photo courtesy D.J.A. Evans). (e) Schematic diagram showing the formation of debris-rich englacial-thrusts and controlled-moraines at Kvíárjökull, based on observations by Spedding and Evans (2002) and Swift et al. (2006) (modified with permission from Evans, 2009). (f) Debris-ridges (thrusts) near the terminus of Storglaciären, Sweden (photo: J. Pomeroy). (g)–(j) Evidence for glaciohydraulic supercooling at Svínafellsjökull: (g) debris-rich 'supercool facies' basal ice (photo: S.J. Cook); (h) an 'anchor-ice' terrace surrounds a vent that is discharging 'supercooled' water (photo: R.I. Waller); (i) an englacial fracture filled with 'herringbone facies' ice (photo: S.J. Cook); (j) frazil ice exhibits a similar 'herringbone' structure to the englacial fracture fill in (i) (photo: S.J. Cook).

Further understanding of ice flow within and across overdeepenings has been aided by numerical modelling. Gudmundsson (1997) demonstrated that moderate overdeepening can cause extrusion flow and, where the amplitude-to-wavelength ratio approaches 0.28, flow separation, such that ice will circulate within the overdeepening and theoretically may never leave it. Gudmundsson thus argued that basal ice stratigraphy should be strongly disrupted and that ice should be expected to pass directly across overdeepenings rather than through them (the 1000 m long and 400 m deep overdeepening beneath the Aletchgletscher is cited as one possible example). On the ice-sheet scale, Hindmarsh et al. (2006) investigated the influence of undulating topography on the internal-layer architecture of ice sheets, demonstrating that for short-wavelength topography (e.g. closely-spaced valleys of wavelength comparable to or less than ice thickness) the layering overrides bed obstacles, whereas for widely-spaced obstacles, the layering drapes over the underlying topography. This appears to result from a similar process to that demonstrated by Gudmunsson (2003), which transmits the form of ice-sheet basal topography to the surface topography of overlying ice, though the implications of this process for ice dynamics at this scale remain unclear.

The pooling of subglacial waters in overdeepenings appears to be another process that influences ice flow. For example, numerous water-filled overdeepenings beneath the East and West Antarctic ice sheets cause pronounced flattening of the ice sheet surface as a result of negligible basal shear stress at the ice-lake interface (e.g. Siegert and Ridley, 1998; Siegert, 2000) (e.g. Figure 4b). It has been further demonstrated that the tributaries of numerous ice streams coincide with the locations of such lakes (e.g. Pattyn et al. 2004; Bell et al. 2007) and that lake drainage is via distributed subglacial drainage pathways (Carter et al. 2009). Notably, Peters et al. (2007) have inferred a major meltwater body beneath a tributary of Bindschadler Ice Stream, West Antarctica, in a region where subglacial water is trapped by a local reversal in ice-air surface slope arising from ice flow over variable basal topography; the same process appears to produce strong variation in the water content of nearby subglacial sediments. Further, Bindschadler and Choi (2007) have shown that acceleration of an ice-stream tributary caused by the pooling of subglacial water produces a positive feedback of further subglacial water pooling, faster-streaming flow, and sustained downstream acceleration along the tributary–ice-stream system. Peters et al. (2007) note that the existence of many such water bodies and the resulting spatial variation in subglacial sediment properties is not captured in current models of subglacial hydrology, lubrication of ice stream motion, and sediment transport.

Another possible role of overdeepenings in the location of fast ice flow features is in their potential influence over the spatial pattern of basal till. Although high basal water pressures promote sliding rather than subglacial sediment deformation (Iverson et al., 1995), till and/or fine-sediment accumulation within overdeepenings, either subglacially or subaerially during periods of more limited glacier extent (Alley et al., 2003b), might sustain till layers downglacier of the overdeepening and thus continue to promote fast flow. The formation of basal ice by glaciohydraulic supercooling could also play an important role in sustaining till layers downglacier of overdeepenings through the gradual release of debris from basal ice as a consequence of basal melting (Alley et al., 1997). Further, it has been suggested that basal ice formed by supercooling should contain high quantities of silt (Lawson et al. 1998; Larson et al. 2006) that may enable ice to deform more easily (Alley et al., 1999), but systematic creep tests of silt-dominated ice have not been undertaken (Alley et al., 1999).

5.1.2.2 Glaciers terminating in lacustrine and marine environments

The implications of overdeepening for the processes of ice flow are most evident for glaciers terminating in water, which applies to most large outlet glaciers of the Greenland and Antarctic ice sheets, but also to terrestrial glaciers that retreat into an overdeepened bed. In such situations, calving processes dominate mass-loss and grounding-line position (e.g. Benn et al., 2007), and retreat of the grounding-line into an overdeepening can accelerate retreat by enlarging the width and depth of the calving-front and/or causing the terminus to become ungrounded.

The key process that accelerates retreat in the presence of a widening and deepening calving-front is the reduction of resistive stresses, mainly lateral- and basal-drag, that prevent the glacial driving-stress from causing catastrophic break-up of the terminus. Notably, basal drag declines as the water-depth increases and the terminus nears flotation, causing lateral- and longitudinal-stress gradients to increase. O'Neel et al. (2005) demonstrated at Columbia Glacier, Alaska that basal and lateral drag were greatest where the valley was shallow and narrow and that lateral drag and longitudinal stress were highest where the valley was deep and wide. Undulations in the glacier bed thus caused longitudinal-stresses to vary along the length of the glacier, indicating zones of compressive and extensional ice flow, with areas of extending flow focussed above overdeepenings being susceptible to rapid calving during retreat. Such rapid calving in these locations is likely to be facilitated by extensional crevasses that form planes of weakness. For example, Sund et al. (2011) observed that extensional flow across overdeepenings at Kronebreen, Svalbard caused reactivation and deepening of transverse crevasses and an enhancement of calving and retreat.

Although the effect remains poorly understood, calving rates are also likely to be influenced by the removal of backstress (Benn et al., 2007), which, like lateral and basal drag, opposes glacier flow. 'Backstress' describes non-local lateral and basal drag transmitted upglacier via longitudinal stresses (Van der Veen, 1997), and may result from a range of features, including islands, coastline, submarine shoals or ice shelves. In overdeepenings, the adverse slope can act as a pinning-point that generates backstress; glacier retreat or flotation, such that the base is no longer in contact with the adverse slope, may therefore reduce backstress and enhance flow. Motyka et al. (2002) and Boyce et al. (2007) detailed rapid disintegration of Mendenhall Glacier as it retreated into an overdeepening, partly as a consequence of thinning, causing flotation and thus loss of basal drag, and partly as a consequence of retreat from an adverse slope, causing loss of backstress. Indeed, Mendenhall Glacier shows cyclic behaviour where periods of retreat are punctuated by periods of stability, interpreted by Boyce et al. (2007) as periodic pinning against bedrock-rises. Calving glaciers can also stabilise on a 'shoal' of sediment that has been eroded from or transported along the glacier bed that can advance with the glacier margin (Figure 3c) (e.g. Powell 1991; Hunter et al. 1996; Fischer and Powell 1998; Koppes et al., 2010). With sufficient sediment supply, a shoal can arrest retreat, or even promote advance (e.g. Warren 1992; Nick et al., 2007).

These influences on the calving-rate means that retreat across an overdeepening can be very rapid but advance into an overdeepening is typically slow (Meier and Post, 1987). For example, an advance of Columbia Glacier into an overdeepening (Nick et al., 2007) was an order of magnitude slower than the former rate of retreat (O'Neel et al., 2005), and similar behaviour has been observed for outlet glaciers of the Greenland ice sheet (e.g. Howat et al., 2008; Nick et al., 2009) (see section 5.1.3). Nevertheless, the influence of overdeepenings and, more broadly, bed topography on ice flow and calving processes is uncertain. For example, Nick et al. (2010) found that calving rate is not simply controlled by bed topography but is instead sensitive to the depth of crevasse penetration and the depth of water in those crevasses. The modelling study by Nick et al. (2010) further demonstrated that stable glacier positions could be reached even on reversed bed slopes; hence, irreversible glacier retreat into overdeepenings (as seen, for example, at Columbia Glacier, Mendenhall Glacier and outlet glaciers of Greenland) may not be a feature of all water-terminating glaciers.

5.1.3. Stability and climatic response of glacier and ice sheets

The significance of overdeepenings for the processes and spatial patterns of subglacial drainage and ice flow indicates that their presence in the ablation zones and/or terminal regions of valley and outlet glaciers could be important to the fate of such ice masses in response to recent climate change. This is supported by recent research that has demonstrated the rapid and unpredictable retreat of outlet glaciers in Greenland and Antarctica as a result of glacier grounding-lines being located on adverse subglacial slopes (e.g. Howat et al., 2008; Nick et al., 2009). However, many issues remain poorly understood, including the longer-term response of valley and outlet glaciers with terminal or multiple overdeepenings to surface runoff forcing, and, in the case of marine ice sheets, grounding-line retreat.

Recent research efforts, particularly in Greenland, have focussed on the potential influence on ice sheet and outlet glacier dynamics of the rapid transmission of surface runoff to the ice-sheet bed (e.g. Zwally et al. 2002; Solomon et al. 2007; Van de Wal et al., 2008). Given that overdeepenings are numerous beneath many outlet glaciers, and that they are likely to promote hydraulically inefficient subglacial drainage and therefore high basal water pressures, it is reasonable to speculate that surface melt that is intercepted by overdeepenings could have important consequences. For example, surface melt that traverses overdeepenings in an inefficient subglacial system could enhance ice-bed decoupling and the speed of glacier response to on-going climate change, particularly if it prevents the formation of seasonally present subglacial channels (cf. Sundal et al. 2011). On the other hand, if englacial pathways are preferred, increases in surface melt delivery may have little or no effect on ice dynamics. The influence of surface runoff could also be unpredictable owing to the distribution of overdeepened areas beneath the ablation zone (e.g. Figure 8e), and the implications of overdeepening could extend to the influence of silt-dominated basal ice formation and till deformation on the spatial pattern of ice flow (see above). Unfortunately, the impact of such processes on mass loss from ice sheets has yet to be investigated.

Work has instead focussed on the unpredictable advance and retreat cycles of lacustrine- and marine-terminating glaciers, which exhibit slow periods of advance (over hundreds to a thousand years) balanced by extremely rapid periods of retreat (over decades to a century) (e.g. Meier and Post, 1987). Oerlemans (1989) and Oerlemans et al. (2011) have used simple numerical models to demonstrate that this behaviour results from branching of the equilibrium state when overdeepenings are present in the glacier bed (cf. Figure 7d). This is especially important in calving situations, as Oerlemans et al. (2011) have illustrated using a numerical model of Hansbreen, Svalbard (Figure 13a), because a gradual lowering of the ELA causes the glacier to extend uniformly and then advance rapidly across the overdeepening (Figure 13a; point 3), whilst a gradual raising of the ELA causes initially uniform glacier recession followed by rapid collapse of the glacier tongue (Figure 13a, point 4). The same branching of equilibrium states is central to the marine ice sheet instability problem (Weertman, 1974; Thomas and Bentley, 1978), in which an ice sheet that is grounded and extends beyond an overdeepening can be forced into irreversible retreat when sea levels rise, accumulation rates drop, or mean ice temperature or bed slipperiness increase (Schoof, 2007). Each of these changes causes rapid recession because they cause ice discharge to exceed the balance flux (Figure 13b).

Numerical modelling has shown that such concepts help to explain the nonlinear behaviour of tidewater glaciers that rest on undulating submarine beds, such as Hansbreen, Svalbard (Vieli et al. 2001; Oerlemans et al. 2011), and those that terminate on submarine shoals. Oerlemans and Nick (2005, 2006) used a simple model incorporating calving and sediment transport to show that, for a typical tidewater glacier, slow periodic forcing of the equilibrium-line altitude results in a cycle of steady advance into the estuary, stability and the creation of a submarine terminal moraine (with an 'overdeepening' behind), and then very rapid retreat; Oerlemans et al. (2011) added that such behaviour will be further enhanced if the overdeepening is in-filled with sediment during ice advance and re-excavated subglacially during stability. Alley et al. (2007) argued that this sensitivity of tidewater glaciers to climate change might account for apparently synchronous behaviour of ice sheets on millennial time scales. Further, sedimentation beneath ice shelves should help to stabilize ice sheets against grounding-line retreat in response to sea-level rise, except under extremes of climate change. Similar self-stabilisation behaviour, resulting from overdeepening and moraine building, is likely to occur at terrestrial glacier margins (Figure 3a) (Alley et al., 2003b).

This non-linear yet synchronous behaviour has been observed in Greenland outlet glaciers by Howat *et al*. (2008) and Nick et al. (2009) (Figure 13c–f). These authors observed that, prior to rapid retreat, glaciers in southeast Greenland exhibited typically flat or reversed surface slopes within several kilometres of the calving front (e.g. Figure 13c), indicating grounding of ice on bathymetric highs between overdeepenings; minor retreat induced by warmer air and/or sea temperatures then produced rapid recession, accompanied by temporarily high ice velocities (e.g. Figure 13d), followed by stabilisation on new bathymetric highs, from which some glaciers thickened and re-advanced. Nick et al. (2009) successfully modelled the behaviour of Helheim Glacier (Figure 13e and f), demonstrating that rapid speed-up and retreat appear to result from a loss of backstress provided by the adverse bed slope, and concluded that such overdeepening-enhanced sensitivity to climate may have caused many outlet glaciers in Greenland to adjust rapidly and almost synchronously.

Figure 13. Dynamics of calving ice masses terminating in overdeepenings. (a) and (b) Modelled equilibrium states of (a) Hansbreen (redrawn from Oerlemans et al., 2011) and (b) a typical marine ice mass (Schoof, 2007; redrawn from Joughin and Alley, 2011). For the Hansbreen model, unstable equilibria associated with the overdeepening are indicated by the dotted line between glacier positions 3 and 4; changes in ELA therefore produce strongly asymmetric cycles of glacier advance and retreat (black arrows). For the marine ice mass model, unstable equilibria occur where ice discharge (q), which is dominated by calving and is therefore dependent largely on ice thickness at the grounding line, exceeds the ice flux (ax); retreat of the grounding line past position 2 therefore results in rapid re-stabilise of the grounding line at position 3. (c) to (f) Observed and modelled terminus positions and flow velocities for along-flow profiles of Helheim glacier, East Greenland, during retreat across an overdeepening in the glacier bed (thick dashed line) (redrawn from Nick et al., 2009): (c) and (e) observed and modelled retreat; (d) and (f) observed and modelled velocities.

The largest retreats of Greenland outlet glaciers appear to have caused the speed-up of ice from the ice draining from the ice sheet interior (Joughin et al. 2010), implying that loss of backstress enhances mass loss significantly. Nick et al. (2009) have therefore stressed the need for high-

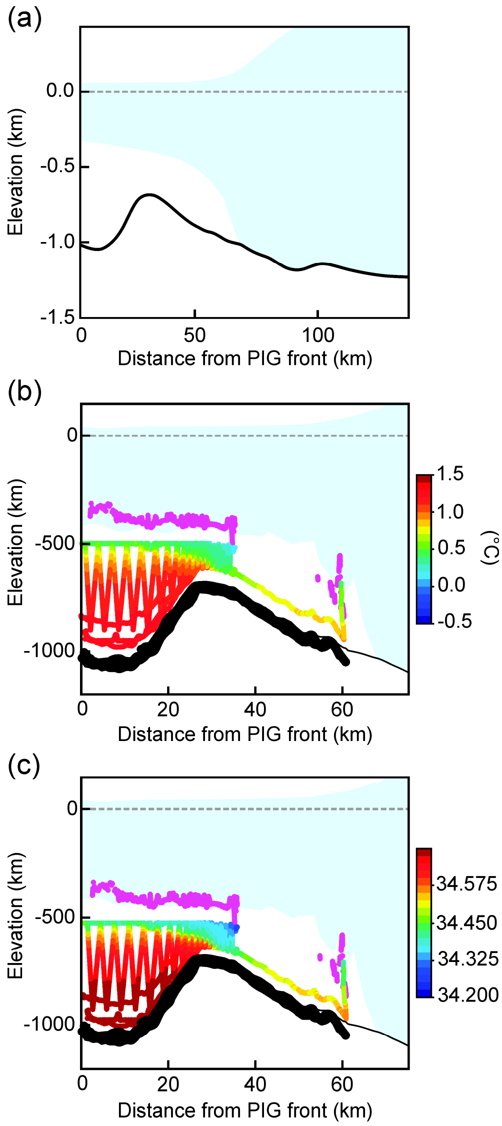

Figure 14. Characteristics of the sub-ice shelf cavity at Pine Island Glacier, West Antarctica. (a) Profile of the ice shelf and sub-ice cavity, which has enlarged as the grounding line has retreated from the prominent shoal or riegel. (b) and (c): Ingress of CDW into the cavity, shown by autonomous measurements of (b) potential temperature and (c) salinity. Reproduced with permission from Jenkins et al. (2010).

resolution bed topography and the development of a free-evolving calving terminus in ice-sheet models. However, Nick et al. (2009) and Howat and Eddy (2011) have speculated that, at least for some parts of Greenland, synchronous re-stabilisation at the inland end of overdeepenings should cause mass losses to diminish, hence present losses should be extrapolated cautiously. The future of the West Antarctic Ice Sheet (WAIS), in contrast, is less secure, owing to the centre of the ice sheet being grounded largely below sea level and at depths far deeper than that of present grounding line positions (Figure 4a). Importantly, many outlet glaciers possess deep troughs that extend almost to the centre of the ice sheet, such as Pine Island Glacier (Figure 4e), where numerical modelling indicates that unstable grounding-line recession may already be occurring (Katz and Worster, 2010).

Changes at Pine Island Glacier highlight another feedback that occurs when Antarctic tidewater glaciers retreat from grounded positions located on overdeepenings: ingress of relatively warm Circumpolar Deep Water (CDW) into the overdeepening via a widening gap between the overdeepening lip and the ice shelf above (Jenkins et al., 2010; Figure 14). The effect of retreat into the overdeepening is to expose a far greater surface area of ice to melting by CDW, and to initiate a

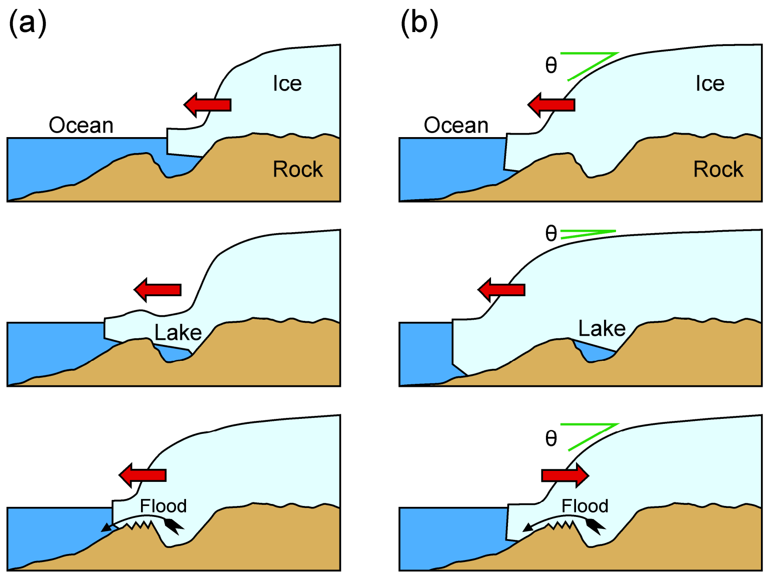

Figure 15. Alternative models for palaeo subglacial lake formation (redrawn from Jordan et al., 2010b). (a) The 'Captured Ice Shelf' or 'Trapped marine' model (Erlingsson, 1994; Alley et al., 2006). Water is trapped during ice-shelf advance across the basin; freeze-on of the shelf-base to the basin-lip (Alley et al., 2006) prevents flooding until the ice has advanced and thickened sufficiently for the shelf-base to have reached melting point. (b) The 'New lake' model (Jordan et al., 2010b). The surface slope (θ) of an advancing ice sheet permits lake formation in a basin when the gradient of the adverse slope exceeds 11θ; flooding is prevented until the ice sheet retreats and θ increases.

positive feedback loop of increased melting and thinning of the ice shelf, increased widening of the gap between the bed and ice shelf base, and increased ingress of CDW into the overdeepening. Such changes can produce strong thinning, extending inland for hundreds of kilometres, in just a few years (Joughin and Alley, 2011), and, combined with the associated effects of reduced ice-shelf buttressing and the natural instability of ice grounded on adverse topography, this process is likely to control the fate of the WAIS over the next 1000 years (Joughin and Alley, 2011).

5.1.4. Other phenomena

5.1.4.1 Surging, jökulhlaups, and Heinrich events

A number of studies indicate that basins can encourage surging phenomena because of their influence on basal water pressure and ice thickness. Björnsson et al. (2003) noted that overdeepenings are present beneath many Icelandic glaciers that exhibit surge-type behaviour, and observed that the importance of hydraulically inefficient subglacial drainage for surging (e.g. Kamb et al., 1985; Raymond, 1987) might enable overdeepening to facilitate surging. Flowers et al. (2011) suggested that overdeepening facilitates surging of a small glacier in the Donjek Range, Yukon, because the adverse slope resists ice-flow and enhances build-up of the surge 'reservoir' during the quiescent phase of the surge cycle. Flowers et al. (2011) further suggest that this arrangement of bed topography may permit surging when it would otherwise be impossible, and that, in addition to thermal and hydrological controls, topography could be an important control on surging more generally.

Overdeepenings are also implicated in the generation of jökulhlaups from beneath large ice sheets that have scoured vast regions and may potentially have triggered ice stream surges and associated Heinrich events. That an advancing ice sheet could create a large subglacial lake by trapping water within an existing overdeepening was first proposed by Erlingsson (1994) in his captured ice shelf (CIS) hypothesis, in which the ice front was envisaged to advance across a basin as a floating ice shelf that would form a hydrostatic seal on reaching the basin perimeter (or sill); the seal and weight of ice would then overpressurise the water-body, such that inflow to the lake sufficient to overcome this seal would produce a catastrophic jökulhlaup. Erlingsson (2006) later demonstrated that Lake Vostok was consistent with this hypothesis, speculating that it may currently be close to producing a catastrophic jökulhlaup, and asserted that this hypothesis could account for Laurentian jökulhlaups that have been implicated in dramatic climatic changes (e.g. Hulbe et al., 2004).

Alley et al. (2006) arrived independently at a similar model, but added that freezing-on of ice to the sill would cause thickening of ice above the lake, and that leakage of supercooled water, and thus basal ice accretion, would cause thickening of ice over the sill. Alley et al. speculated that these thickening processes would lead to pressure melting of ice overlying the sill, leading to outburst flooding, rapid ice flow, and drawdown of the ice sheet when a critical ice thickness was met (Figure 15a); thus, a cyclic process of advance, thickening and rapid drawdown could account for numerous jökulhlaups from the same overdeepening. An alternative model proposed by Jordan et al. (2010a) contends that flooding is triggered passively during ice retreat because the steepening ice surface slope increases the ponding threshold and evacuates the lake water (Figure 15b). Jordan et al. (2010a) used this model to explain evidence for a palaeo-jökulhlaup from Wilkes Subglacial Basin, East Antarctica, where the ice surface is currently too steep for the basin to host a subglacial lake; changes in surface slope have also been invoked to explain lake drainage at Crane Glacier, Antarctic Peninsula, during the rapid retreat of the calving front (Scambos et al., 2011).

Erlingsson (2006), Alley et al. (2006) and Jordan et al. (2010a) all agree that such processes may have been involved in Heinrich events and the generation of dramatic climate changes. Further, drainage from subglacial lakes is potentially an important mechanism for both triggering and sustaining ice stream phenomena (see section 5.2.3).

5.1.4.2 Glacier hazards

One of the most evident impacts of the near-synchronous retreat of glaciers globally over the past 100 years has been the formation of new proglacial lakes in former subglacial basins that increase substantially the risk of catastrophic floods. Once a lake forms, many other processes associated with glacier retreat affect the hazard situation (Frey et al., 2010), including: (i) the development of new ice-avalanche initiation zones; (ii) the destabilisation of adjacent hanging-glaciers; (iii) the debuttressing of adjacent rock walls; (iv) the exposure of morainic or other unconsolidated materials that are prone to mass-movement; (v) the exposure of further lakes, which increases the potential of a flood from one lake initiating flooding from several; and (vi) the destabilisation of frozen rock-walls and moraine dams as a result of the degradation of permafrost or dead-ice.

Frey et al. (2010) described a multi-level strategy for the identification of overdeepened parts of the glacier beds and therefore sites of future lake formation. Key to their approach is the automated estimation of ice thickness and hence glacier-bed topography (e.g. Farinotti et al., 2009; Linsbauer et al., 2009), which enables estimation of potential maximum lake area and volume and the topography of the lake-bottom and surrounding terrain. Frey et al. (2010) have shown that the method of Linsbauer et al. (2009) produces ice thicknesses that are in good agreement with measurements in the field, requiring only a DEM, glacier outlines, and a set of branch lines to calculate average basal shear stress per glacier and hence ice depth along the branch lines; thicknesses are interpolated spatially using an algorithm that mimics the typical parabolic shape of glacier beds. Frey et al. (2010) used this method to identify numerous overdeepenings beneath a number of Swiss glaciers (Figure 16), and the technique has since been extended to identify overdeepenings beneath the Bernese Alps (Haeberli et al., 2011) and Jostedalsbreen ice cap (Meister et al., 2011).

Particularly hazardous situations occur when glaciers retreat from overdeepenings that are partly or entirely sediment confined, forming highly unstable moraine-impounded lakes in close proximity to the retreating terminus (Lliboutry et al., 1977; McKillop and Clague, 2007; Bolch et al., 2008). Further, Frey et al. (2010) observed that glacier retreat from overdeepenings tended to culminate in glacier stabilisation on steeper parts of the bed directly above the newly formed lake, thus increasing the hazard-potential from ice avalanches. This work indicates that future lake development and

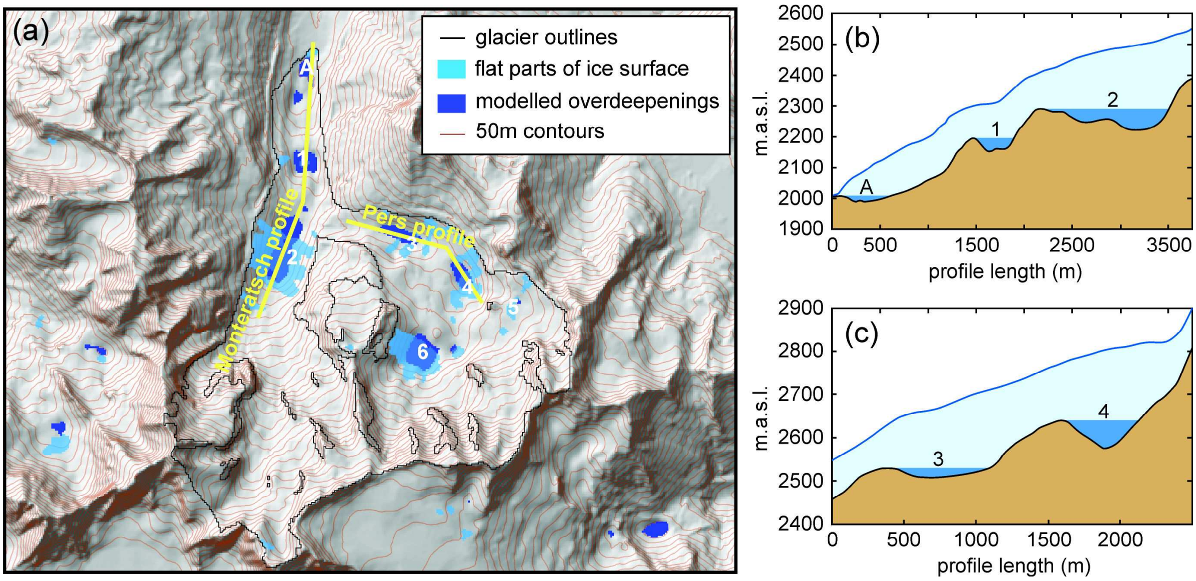

Figure 16. Overdeepenings beneath the Morteratsch and Pers Glaciers, Bernina region, Switzerland. (a) Overdeepenings identified using a threshold ice-surface slope criterion of < 5 ° (light blue areas) and modelling of ice thickness distribution (dark blue areas) (see text). (b) and (c): The ice surface and inferred subglacial topography along the (b) Morteratsch and (c) Pers profiles marked in (a). Overdeepenings (numbered) are shown as filled lakes in order to demonstrate the possible extent and depth of new-lake formation, should the ice-cover be removed. Reproduced with permission from Frey et al. (2010).

associated hazards may represent a significant problem in the future, although with prediction systems of this nature there is potential to be prepared for these developments in advance of their formation.

5.1.4.3 Sea level

Gomez et al. (2010a) highlighted the role of subglacial basins (or 'negative topography') in providing accommodation space for melt produced by the retreat of marine-based ice. For example, the volume of water stored in the Antarctic ice sheets equates to a global eustatic sea level rise of ~70 m, but negative topography beneath marine-based ice sheet sectors means the actual geographically-uniform increase in sea level, termed the effective eustatic value (EEV), would be closer to 55–60 m. If the WAIS alone is considered, the same values are ~8 and ~5 m, respectively. The latter value has been cited widely (e.g. IPCC, 2007, chapter 10), but Gomez et al. (2010a) note that the true value is unknown and varies according to the level of scientific sophistication used to generate the estimates; indeed, these figures have more recently been revised downward again owing to improved knowledge of the subglacial topography (e.g. Bamber et al., 2009). Further, in the WAIS case, Gomez et al. (2010a) argued that some locations could experience sea level rise ~40% higher than the EEV because ice sheet decline also causes: (i) loss of gravitational attraction (cf. Clark and Lingle, 1977); (ii) isostatic rebound; and (iii) changes in Earth rotation. Clearly, accurate knowledge of subglacial topography is also important to such calculations.

The issue of ice sheet stability in the presence of 'negative topography' (cf. Weertman, 1974) has been tackled by Gomez et al. (2010b), who show that deformational and gravitational effects can produce a sea-level fall at the margin of a rapidly-shrinking ice sheet that can stabilise ice sheet retreat even in the presence of an adverse bed slope. Assuming that the ice-sheet maintains a steady-state surface profile and mass loss affects in excess of 50% of the ice sheet, computation of gravitationally self-consistent sea-level (GSCSL) resulting from grounding line retreat shows that an ice sheet with radius greater than 1,400 km can remain stable on a bed with a slope of up to -0.25 m km^{-1}. Gomez et al. (2012) further demonstrated that by allowing sea level to vary in response to ice sheet retreat, shallow bed slopes, with gradients between 0 and -0.3 m km^{-1}, allowed grounding line position to stabilise within 200 to 300 km of its initial position. For steeper beds, with gradients -0.7 to -1.0 m km^{-1}, the grounding line would retreat catastrophically, but associated changes in sea level would delay collapse by 500 to 1000 years. Whilst valuable, these scenarios do not include the complex overdeepened topography of real ice sheet beds, where changes in bed slope are frequent and gradients exceed -100 m km^{-1}. The effects of CDW ingress and reduced ice-shelf buttressing, are also excluded, both of which focus mass loss at the ice margins (e.g. Joughin and Alley, 2011).

5.1.5. Synopsis

Subglacial basins have fundamental influence on the nature of water and ice flow within glacial systems and the nature of their response to climate changes. Basins appear to promote focused delivery of surface melt to the glacier bed and persistently high basal water pressures; glaciohydraulic supercooling is uniquely associated with subglacial basins and may elevate basal water pressures further by reducing the transmissivity of subglacial channels. These hydrological characteristics appear to enhance ice flow, most notably by encouraging basal sliding, and the tendency for basins to accumulate sediment, and for glaciohydraulic supercooling to produce silt-rich basal ice, may further enhance glacier motion through basal sediment and ice deformation. The details of the hydrology of such basins are, however, understood very poorly: whether terrestrial glaciers and ice sheets with substantial subglacial basins exhibit significantly different hydrological and ice dynamical behaviour from those without basins is largely unknown; hence, the implications of subglacial basins for the response of terrestrial glaciers and ice sheets to climate warming are difficult to predict. Nevertheless, basins have fundamental implications for the stability and dynamics of lacustrine and marine outlet glaciers that produces grossly enhanced sensitivity to climate. Whilst this enhanced sensitivity is unlikely to produce substantial losses of ice from the Greenland ice sheet, it is likely to dominate the future of the WAIS. Such basins provide accommodation space for water when ice masses retreat, form new lakes in mountainous areas that present a substantial hazard risk, and in ice sheet contexts enable water to be stored subglacially and thus to be catastrophically released.

5.2. Glacial geomorphology and long-term landscape and ice sheet evolution

5.2.1. Sediment entrainment and transfer in the glacial system

There is strong evidence that overdeepenings exert significant influence over the mechanisms and patterns of glacial sediment entrainment, transport and deposition. In particular, overdeepenings have come to be associated with the process of glaciohydraulic supercooling because of its potential to form metres to tens of metres of debris-rich basal ice, even in temperate locations where basal ice accretion would otherwise be limited (Alley et al., 1998; Lawson et al., 1998) (see section 4.1.1 for a description of the process). Basin-related hydrological processes and feedbacks also appear to influence greatly the quantity and nature of sediment available for entrainment, the mechanisms of entrainment, and the proportion of transport in glacial versus fluvial transport pathways.

Sediment produced at the ice-bed interface by the erosion is one source of sediment that is available for entrainment by glacial or fluvioglacial processes. The portioning of sediment between transport pathways depends on the relative efficiencies of the individual entrainment processes and issues of sediment availability: fluvial transport tends to dominate because water is efficient at entraining fine sediment, subglacial streams have large unsatisfied sediment transport capacity, and fluvial entrainment reduces the availability of sediment for glacial transport (Alley et al., 1997; Swift et al., 2005). Overdeepenings are likely to influence rates and processes of transport strongly because they suppress the efficiency of subglacial drainage, which encourages the deposition of sediment out of fluvial transport and discourages fluvial entrainment of sediment produced locally by erosion, thereby increasing the sediment available for glacial entrainment and transport (e.g. Kirkbride and Spedding, 1996). Glacial sediment transport will be favoured especially if englacial pathways inhibit access of surface melt to the bed (e.g. Hooke and Pohjola, 1994), thereby reducing the likelihood that hydraulically-efficient channels are present, and when water within subglacial channels is forced to distribute across the bed (e.g. Röthlisberger and Lang, 1987). The absence of hydraulically efficient channels will also reduce the likelihood that sediment in glacial transport, especially in basal ice or till layers, will be captured or re-captured by fluvial transport processes (Spedding and Evans, 2002).

The most widely reported glacial sediment entrainment mechanism associated with overdeepenings is the formation of debris-rich basal ice as a result of glaciohydraulic supercooling within subglacial channels (aka 'supercool facies' basal ice) (Figure 12c). On the basis of work undertaken at Matanuska Glacier, Alaska, Lawson et al. (1998) proposed that freezing of supercooled water produces porous aggregates of anchor- and frazil-ice that adhere to the glacier base and filter sediment from percolating subglacial water; subsequent ice-sediment segregation thus produces thick layers of debris-rich, stratified basal ice. Bands of similar debris-rich ice occur englacially at Matanuska Glacier (Ensminger et al. 1999, 2001), apparently because of the injection, and subsequent freezing, of water into basal crevasses, indicating that glaciohydraulic supercooling is capable also of entraining sediment into englacial positions. Similar structures have been observed in crevasses and around subglacial vents at Svínafellsjökull (Figure 12h–j) and other Icelandic glaciers (Larson et al., 2010), including Skeiðarárjökull, where they have been observed to have been emplaced in englacial fractures created during jökulhlaups (Roberts et al., 2002). Fluvial debris has been observed within abandoned englacial conduits at overdeepened glaciers (Kirkbride and Spedding, 1996; Spedding, 2000), and this may be indicative of glaciohydraulic supercooling encouraging subglacial flows to adopt alternative englacial pathways (e.g. Lliboutry, 1983; Hooke & Pohjola, 1994).

Despite the implied importance of supercooling for basal ice formation and sediment transfer at a few well-studied glaciers (e.g. Lawson et al., 1998; Roberts et al. 2002), there is uncertainty as to the importance of supercooling more generally (Spedding and Evans, 2002; Cook et al., 2006). Roberts et al. (2002), for example, suggested that supercooling was an important mechanism of basal ice formation and sediment transfer at glaciers in southeast Iceland; however, having undertaken detailed work on sediment transport processes at Kvíárjökull in southeast Iceland, Spedding and Evans (2002) noted that there appeared to be only limited evidence for glaciohydraulic supercooling having contributed to basal ice formation, despite ice surface and bed slope conditions being favourable. Further, isotopic and sedimentological analysis of basal ice and debris bands at Kvíárjökull by Swift et al. (2006) have provided little support for sediment entrainment by supercooling. In an attempt to resolve this controversy, Cook et al. (2007, 2010) have quantified the proportion of basal ice formed by various processes at Svínafellsjökull, southern Iceland, and demonstrated that only ~42% of stratified-facies basal ice was formed by supercooling.

The hydrological conditions within overdeepenings have other consequences for sediment transfer, particularly the size and form of sediment that is transported in specific pathways. Although there is evidence that coarse debris (gravel, cobbles and boulders) can be transported in englacial waterways across overdeepenings (e.g. Kirkbride and Spedding, 1996), observations by Pearce et al. (2003) at Matanuska Glacier demonstrate that subglacial flowpaths within overdeepenings struggle to evacuate coarse debris. Notably, measurements of sediment calibre at subglacial discharge vents by Pearce et al. (2003) revealed negligible bedload component, even after significant increases (by a factor of 2.5) in discharge, whereas basal ice formed by the freeze-on of supercooled water contained at least some clasts, all of which were bigger than those measured in the vents. These results support the hypothesis that supercooling causes clogging of subglacial flowpaths (e.g. Lawson et al., 1998) and that the reduced competence and capacity of subglacial flows will reduce the evacuation of coarse sediments (Alley et al., 2003a), leading to the development of coarse lag deposits within overdeepenings (Spedding and Evans, 2002; Alley et al., 2003a). These processes indicate the existence of distinctive sediment sizes in contrasting sediment transport pathways, but also highlight the potential for mixing of sediment between pathways.

Despite the focus on supercooling, there is growing evidence for direct glacial entrainment of sediment within overdeepenings as a result of the enhanced longitudinal flow-compression provided by the adverse slope (e.g. Spedding and Evans, 2002; Swift et al., 2006; Evans, 2009; Moore et al., 2011). Spedding and Evans (2002) suggested that exposures of basal ice at Kvíárjökull were consistent with formation by processes such as regelation or flow of sediment-laden water through the intergranular vein network (Lliboutry, 1993; Knight and Knight 1994), but that the thickness of basal ice layers (up to 10 m) and their intercalation with englacial ice indicated stacking of basal ice layers as a result of folding and thrusting under compressive ice-flow; this process preserves basal ice layers that may otherwise be destroyed by basal melting or subglacial meltwater flow (Spedding and Evans, 2002). Swift et al. (2006) concluded that the same strongly compressive flow-regime had produced transverse thrusting of basal ice and sediment to the surface of Kvíárjökull along ogive-parallel foliation (Figure 12c). Notably, the bands contained both basal debris and very large water-worked clasts, the latter displaying evidence of transport at the ice-bed interface in the form of striations, indicating that the debris had been scavenged from a highly heterogeneous lag-deposit within the overdeepening (Figure 12e). Evidence of the same process of debris elevation has also been found at Svínafellsjökull (Swift et al., in prep).

It must be noted that the processes described above may exhibit high temporal and spatial variability, leading to switching between transport mechanisms and pathways; changes in the dominant transport pathway; and changes in the source or character of transported sediment. For example, the tendency for advancing glaciers to override sediment (Alley et al., 2003b) means that sediment transport during advance might be dominated by fluvial recycling of non-glacial material, after which stabilisation of overdeepening geometry at the supercooling threshold (Alley et al., 2003a) will produce a switch to dominantly glacial transport of sediment of fluvioglacial and glacial origin. Further, subtle changes in the ratio between ice surface and bed slopes, for example as a result of enhanced ablation or accumulation, or changes in the position of the glacier terminus relative to a subglacial basin, should result in alternation between glacially and fluvially dominated sediment transport. Finally, spatial variation might be expected as a result of glacier and basin geometry arising from the position of major subglacial channels or the supply of sediment in fluvial and glacial pathways.

5.2.2. Ice-marginal sedimentation and the Quaternary record

The issues reviewed above highlight the need to consider glacial and fluvial sediment production and transport systems as coupled and not separate and spatially exclusive process realms, where, for example, material eroded subglacially is simply retained at the glacier bed (cf. Spedding and Evans, 2002). Even more importantly, the presumed dominance of fluvial sediment transport (e.g. Hallet et al., 1996) may not apply where overdeepenings are present (Swift et al., 2002), which has far-reaching implications for understanding of rates of glacier erosion and rates and styles of ice-marginal sedimentation (Figure 17). Swift et al. (2002) and Spedding and Evans (2002) have suggested that ice-marginal landsystem models (e.g. Benn et al., 2003), which structure palaeo-glaciological environments according to climatologically based process-realms that are reflected in the sediment-landform evidence, are weakened by their inability to appreciate internal glaciological feedbacks that have vastly different implications for the sediment-landform record.

Modern landsystems approaches are increasingly subtle in their representation of process-landform relationships (e.g. Hambrey and Glasser, 2012). An example of the need for such subtlety is highlighted by Swift et al. (2002), who question the notion that temperate glaciers should produce the largest moraines (cf. Andrews, 1972), a feature that is implicit in some temperate glacier landsystem models (e.g. Eyles, 1979, 1983; see Spedding and Evans, 2002). Swift et al. (2002) drew upon evidence for highly efficient fluvioglacial sediment evacuation (e.g. Swift et al., 2005) to argue that, though temperate glaciers should possess the highest balance gradients and erosional potential, direct glacial sedimentation (and hence moraine formation) will dominate only if supraglacial debris transport is very significant and/or the efficiency of the subglacial drainage system is restricted. Inspired by the work of Spedding and Evans (2002) at Kvíárjökull, Swift et al. (2002) proposed that temperate glaciers with terminal overdeepenings would produce the largest moraines because of their high erosional potential but and the strong influence of the overdeepening on the transport pathways of sediment. This influence is exerted through suppression of efficient subglacial drainage near the glacier terminus and the enhancement of glacier terminus flow-compression, enabling supercooling

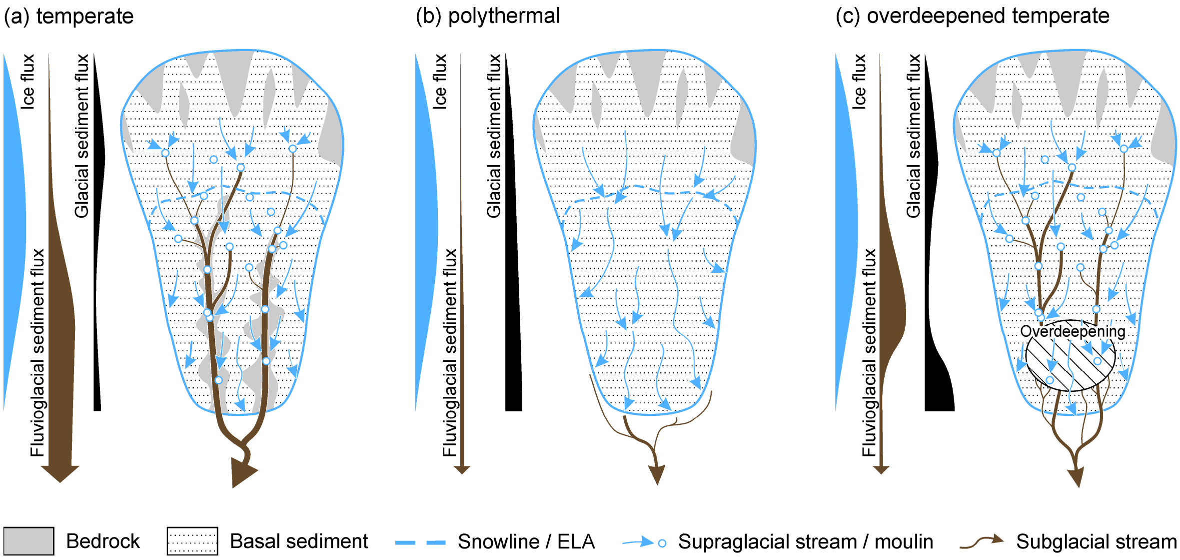

Figure 17. Schematic diagrams showing hypothesised proportions of sediment in glacial versus fluvioglacial pathways for various glacier types (after Swift, 2002). Glacial transport is assumed to occur in basal ice and sediment layers only and sediment production is assumed to scale positively with ice flux but negatively with basal sediment thickness because thick basal sediment layers protect bedrock from glacial erosion. (a) and (b): The total sediment transported by the glacier system is greater at (a) temperate glaciers than at (b) polythermal glaciers because seasonally present hydraulically efficient channels capture and transport large quantities of glacial sediment, which reduces the thickness and extent of basal sediment layers and thus promotes high rates of subglacial erosion. (c) The greatest total fluxes of sediment and the greatest fluxes in glacial transport pathways are likely to occur at temperate glaciers with terminal overdeepenings because the adverse slope reduces the sediment transporting capacity of the subglacial drainage system.

and thrusting to elevate large quantities of formerly fluvially transported debris into englacial and supraglacial transport (e.g. Cook et al., 2010; Swift et al., 2006).

The presence of overdeepenings beneath both glaciers and ice sheets is likely to influence the type and distribution of glacial sediments and landforms across a wide range of scales. Notably, Larson et al. (2006) argue that landform-sediment assemblages in areas formerly covered by the Laurentide and Scandinavian ice sheets are similar to distinctive assemblages at Matanuska Glacier that are produced by the unusually large glacial sediment fluxes that are associated with supercooled-facies basal ice. At the valley-glacier scale, the sediment-landform assemblage included: (1) silt-rich melt-out till derived from *in situ* melting of supercool-facies basal ice; (2) stacked debris flow deposits derived from the re-mobilisation of melt-out deposits; and (3) hummocky terrain, including ice-contact ridges, hummocks, kettle depressions and debris-flow fans. At the ice sheet scale, Larson et al. (2006) observed a similar signature of debris-flow deposits, melt-out tills and hummocky terrain, located both along and just beyond the distal rim of overdeepenings and the crests of steep adverse slopes near former ice margins. Larson et al. (2006) conceded that their observations do not provide unequivocal proof of supercooling; nevertheless, it seems unlikely that such landscapes were not at least partly conditioned by supercooling where bed- and ice-surface surface slopes were favourable.

A major limitation in identifying ‘overdeepened’ landsystems is that the precise geomorphological and sedimentological implications of overdeepening remain uncertain. For example, rates of sediment transfer by supercool-facies basal ice, which is arguably *the* defining characteristics of such landsystems, are highly uncertain. Estimates of debris discharge associated with this process vary from 4.8 to 9.6 $m^3\ m^{-1}\ a^{-1}$, measured by Cook et al. (2010) at Svínafellsjökull (the range of values reflecting uncertainty associated with the estimation of ice velocity), to 56 $m^3\ m^{-1}\ a^{-1}$ measured by Larson et al. (2006) at Matanuska Glacier. Nevertheless, for Svínafellsjökull, this value was sufficient to account for 83% of sediment discharge by stratified-facies basal ice (Cook et al., 2010). Some work has been undertaken to address this uncertainty from a sedimentological perspective by identifying supercool-facies sediment-signatures in the landscape. For example, Larson et al. (2006) considered the dominance of silt within supercooled basal ice to produce silt-dominated melt-out sediments that are unique to sites with overdeepenings. Cook et al. (2011) have provided further evidence for silt-dominance in sediments that are associated with the melt-out of supercool-facies basal ice and have demonstrated that they are distinguishable from regelation-derived deposits. However, Cook et al. (2011) have also drawn attention to the potential for silt to be over-represented in ice-marginal deposits given that overdeepenings act as traps for fine sediment during glacier retreat and that glaciers tend to override such sediments during glacier advance.

Arguably, a second defining characteristic of 'overdeepened' landsystems is the elevation of debris into englacial locations, principally by supercooling, folding and/or thrusting, to produce 'controlled moraine': supraglacially derived hummocky moraine with a strong linearity that is imparted by the pattern of debris concentration within the ice and hence the process(es) of entrainment (Evans, 2009). Evans (2009) argued that supercooling is 'clearly very effective in the production of thick sequences of debris-rich basal ice and therefore could conceivably be manifest in the landform record as controlled moraine'. However, Evans doubted whether many such processes can be recognised from geomorphic evidence, highlighting the supercool-conditioned landform signature described by Larson et al. (2006), which Evans argues is insufficiently distinct from landform-sediment assemblages at glaciers without overdeepenings. Further, Evans argued that the widely-published englacial-thrust explanation of moraine formation, which has been applied largely in non-overdeepened contexts where thrusting is assumed to be produced by a thermal transition (e.g. Hambrey et al., 1996, 1997), is unlikely given the very low preservation potential of englacial thrust structures during deposition. Regardless of whether this is true, there does appear to be potential to recognise such processes from sedimentological evidence, such as particle size (e.g. Cook et al., 2011) and clast-form properties (e.g. Swift et al., 2006). Further, whilst the geomorphic expression of individual processes may be impossible to define, there is good evidence that unusually high rates of glacial sediment transport that are associated with overdeepenings significantly enhance moraine-building potential (Spedding and Evans, 2002; Swift et al., 2002).

Further evidence links conditions within overdeepenings to the sediment-landform record, providing opportunities for process and palaeoglaciological reconstruction that in some cases are yet to be fully realised. The potential role of subglacial basins as the physical cause of Heinrich events and the source of sediment in North Atlantic ice-rafted-debris (IRD) layers is one such example (e.g. Lawson et al. 1998; Roberts et al. 2002; Alley et al., 2006). As described above, Alley et al. (2006) have hypothesised that leakage of supercooled water across the lip of a 'captured' basin will accrete debris-rich basal ice onto the ice shelf base; several studies agree that supercooling did enhance the entrainment of sediment by Laurentide ice streams (e.g. Alley et al. 1999; Andrews & MacLean 2003; Hemming 2004; Hulbe et al. 2004), including entrainment resulting from the injection of supercooled water into basal crevasses during jökulhlaups (Roberts et al., 2002). Other work by Shreve (1985) has demonstrated that changes in esker shape are consistent with predicted changes in R-channel shape in the presence of supercooling, and supercooling has been suggested to have produced up-valley dipping clinoform structures within tunnel valleys (Kristensen et al., 2008). Tunnel valleys themselves, which appear to be a highly specific form of overdeepening related to intense and highly localised erosion of bedrock or sediment by water (e.g. Dürst Stucki et al., 2010), have also been used for palaeoglaciological reconstruction because the direction of subglacial water flow is dominantly controlled by the direction of ice-surface slope (e.g. Stewart and Lonergan, 2011).

Overdeepenings also play an important passive role as glacial and post-glacial sediment sinks, which frequently yield environmental records extending back to the last glacial-maximum (LGM) and occasionally through many glacial cycles (e.g. Dehnert et al., 2012). Such sequences, which may be in excess of 500m in thickness, can also be important sources of groundwater and drinking water (e.g. Seiler, 1990). Many studies have interpreted environmental changes from purely mineral deposits. Pugin (1989), who studied the fill of two overdeepenings in the Sarine valley, Switzerland, observed that palynological evidence associated with glacial deposition was often poor, and that

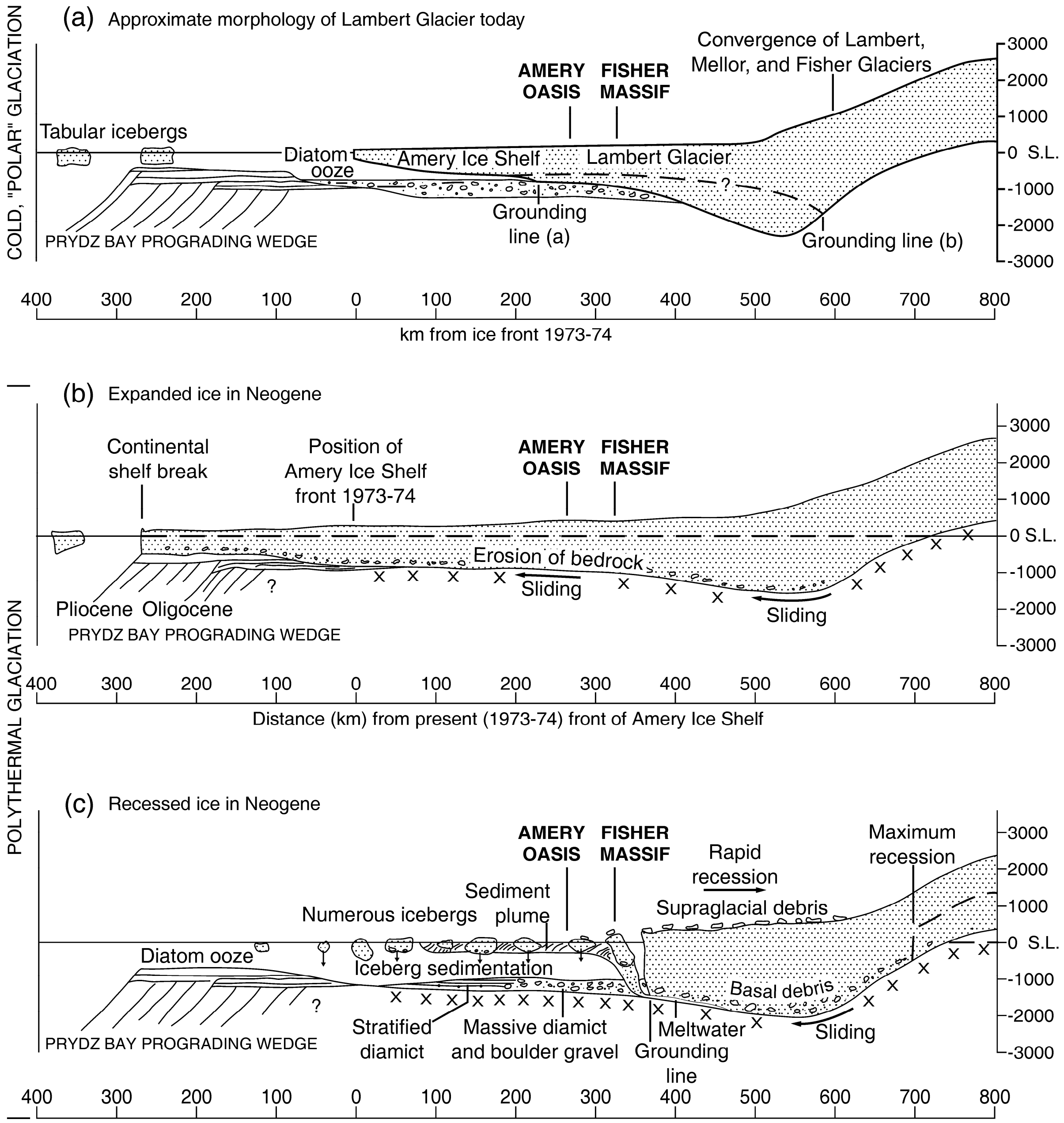

Figure 18. Conceptual model for the growth and decay of the tidewater Lambert Glacier–Amery ice shelf system, East Antarctica, including patterns of erosion and sedimentation within the overdeepened fjord basin and at the shelf edge. (A) Present configuration, indicating recent positions of the grounding-line within the overdeepened basin that are controlled by glacial erosion and sedimentation. (B) Maximum glacier and ice shelf extent. Re-excavation and renewed erosion of the overdeepening feeds a prograding sequence or 'shoal' of sediment, deposited at or near the shelf edge, which stabilises the position of the glacier terminus. (C) Intermediate configuration, illustrated during rapid recession from the maximum glacial extent and prior to re-stabilisation of the grounding line at the maximum recession position (inferred from the inland extent of the overdeepened basin). Reproduced with permission from Hambrey and McKelvey (2000).

sediment-facies associations can yield a richer record of environmental change. Hansen et al. (2009) have used 3D interpretation of rich datasets obtained using ground-penetrating radar, seismic refraction profiling and auger drilling to demonstrate that the sedimentary fill of terrestrial bedrock-confined overdeepenings passively reflects climatically controlled variation in sedimentation rates and sources. Nevertheless, Brookfield and Martini (1999) have demonstrated that such sequences can be enormously complex when deposition is predominantly glaciolacustrine and characterised by frequent changes in both glacier terminus position and water outflow level, as would have been common along the terrestrial margins of former ice sheets. Notably, Eyles et al. (1990, 1991) have used such evidence at Okanagan Lake, British Columbia, to infer ice-dynamical changes associated with rapid disintegration of the Cordilleran Ice Sheet as it retreated into overdeepened basins.

A few studies of overdeepening sediment fills that have discovered fossil or organic material have indicated a minimum age for overdeepening. In the Lambert Glacier trough and Prydz Bay, Eastern Antarctica, glacial deepening (Figure 18) has uplifted glacigenic sediments to positions above present sea level, where they record major oscillations of the ice sheet since the early Oligocene (Hambrey and McKelvey, 2000); this agrees with the minimum age of overdeepening inferred from seismic data and the first appearance (at 14 Ma) of recycled Palaeogene fossils in glacially-deposited sediments on the lower slope and rise (Cooper and O'Brien, 2004). In the Alps, interglacial pollen assemblages at the base of two overdeepenings in the Aare Valley indicate minimum ages of ~ 430–330 ka and ~ 280–250 ka, respectively, (see discussion in Anselmetti et al. 2010), but minimum ages elsewhere are typically much younger (Preusser et al., 2010; Anselmetti et al. 2010, Reitner et al., 2010), which may reflect different temporal patterns of erosion as a result of a particular location's geomorphological and glaciological setting (Reitner et al., 2010). These latter studies highlight the preservation of pre-LGM sediments at the base of many overdeepenings, implying not all glaciations possess similar erosional intensity. Perhaps unsurprisingly, a number of studies indicate the presence of Holocene and LGM sediments in overdeepenings beneath contemporary glaciers, such as the Tschierva and Pasterze glaciers in the European Alps (Joerin et al., 2008; Brückl et al., 2010).

5.2.3. Long-term landscape and ice sheet evolution

As a fundamental expression of glaciation, overdeepenings contribute significantly to the development of characteristic relief, valley cross-sectional area and valley long-profile in glaciated landscapes, and it is therefore important to consider their role in the glacial and post-glacial evolution of such landscapes. Notably, the influence of overdeepenings on sediment transport pathways and the retention of sediment at the glacier bed should influence long-term rates and patterns of glacial erosion and sediment production; the deepening and overdeepening of fluvial valleys should have important implications for extraglacial and postglacial hill-slope processes and landscape response; and the widening and deepening of valleys should have implications for ice sheet dynamics and extent over successive glacial cycles.

It is widely accepted that erosion rates in glaciated basins surpass those of comparable fluvial systems (e.g. Hallet et al., 1996), but the tendency for glaciers to produce overdeepened bed-profiles that restrict the efficiency of subglacial drainage implies that such high rates of glacial erosion are not always sustainable. For example, in non-active orogens, glacial modification of formerly fluvial valleys should, over time, cause sediment production to decline as valleys become 'tuned' to glacial fluxes (e.g. Charreau et al, 2011; Herman et al., 2011); overdeepening should reinforce this phenomenon because it suppresses the flushing of debris from the ice-bed interface and therefore inhibits processes of glacial erosion (cf. Alley et al., 1997; Alley et al., 2003a; see also section 3). In active orogens, uplift should preclude or reduce overdeepening by maintaining steeper catchment gradients (e.g. Brocklehurst and Whipple 2007; Whipple, 2009), which should permit more efficient subglacial drainage and therefore more efficient flushing, greater ice-bed interaction, and higher rates of subglacial erosion and debris comminution (cf. Swift et al., 2002); this sediment in subglacial transport will be augmented by sediment in supraglacial transport supplied by above-glacier erosion of interfluves and peaks. As a result, contemporary sediment fluxes from glaciated catchments (e.g. Hallet et al., 1996; Koppes and Montgomery, 2009), which are generally assumed to be dominated by ice flux and hence erosional potential (Hallet et al, 1996), are likely to depend also on both the tectonic context and extent of glacial landscape modification.

The role of overdeepenings in the evolution of glacial landscapes is similarly poorly understood. Recent research has emphasised the role of cirque erosion in controlling the elevation of glaciated mountain ranges because the cirque-floor sets the base-level for hill-slope processes acting on adjacent mountain peaks (e.g. Oskin and Burbank, 2005; Anders et al., 2010; Egholm et al., 2009);

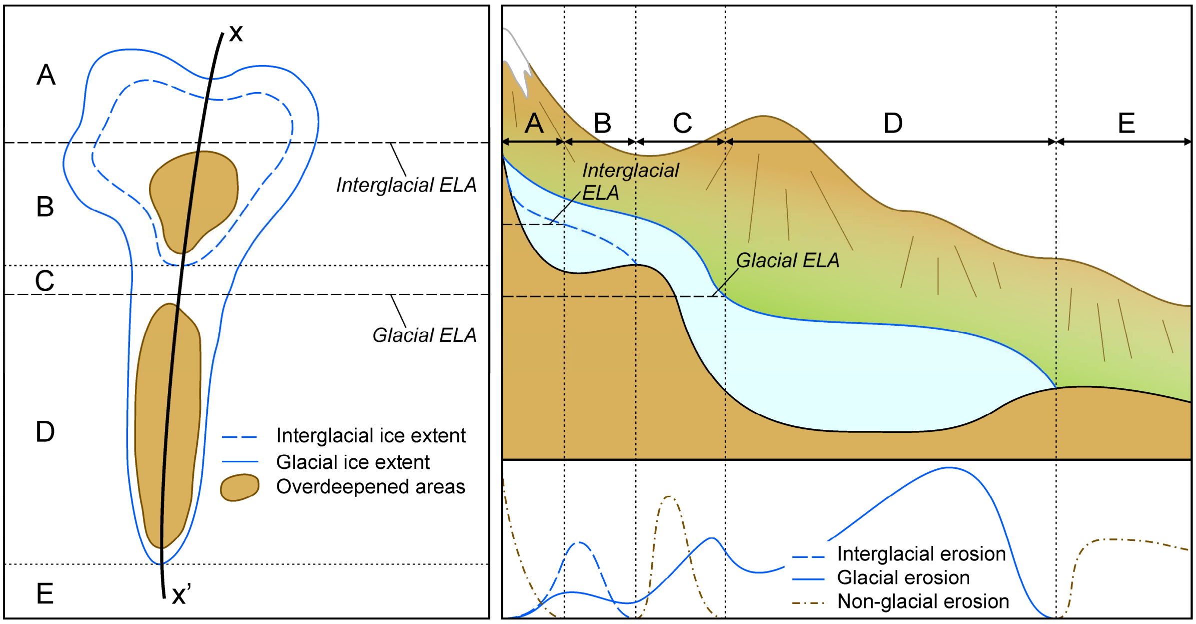

Figure 19. Schematic diagrams illustrating the hypothetical pattern of valley-deepening and overdeepening resulting from glacial erosion under a time-varying climate. (a) Simple plan-view of interglacial and glacial ice extent and anticipated areas of overdeepening. Horizontal lines indicate ELA positions and divisions of the along-flow transect x–x'. (b) Profile-view of the ice surface and bed topography along transect x–x' and the adjacent topography (vertical scale is exaggerated). The anticipated contribution of different erosional processes to landscape development during interglacial and glacial periods is shown qualitatively below. Zones A to E of the along-flow transect indicated in both (a) and (b) are defined according to their glacial context and the anticipated relative contributions to their evolution of the various interglacial and glacial erosional processes (see text for explanation).

this hypothesis provides a mechanistic explanation for the "glacial buzzsaw" effect, whereby the topography of mountain ranges is observed to be curtailed at or around the contemporary ELA (e.g. Brozovich et al., 1997). It is widely accepted that cirques form at or near the long-term average ELA, but it is not certain why cirque-glaciation alone should control the elevation of mountain ranges (Anders et al., 2010). Notably, glacial deepening and widening of trunk valleys during periods of more extensive glaciation also contributes substantially to relief production and should also therefore stimulate erosion of adjacent peaks (e.g. Amerson et al., 2008; Champagnac et al. 2007; van der Beek and Bourbon, 2008; Norton et al., 2010; Herman et al., 2011). For example, in the context of the Swiss Alps, Champagnac et al. (2007) has shown that greatest relief is focussed around wide trunk valleys, not near cirques or valley-heads, whilst Norton et al. (2010) has demonstrated that glacial valley-deepening has been more significant than cirque erosion in producing 'over-steepened' valley profiles. Clearly, a glacially overdeepened profile is significant because it represents the maximum possible lowering of base level that can be achieved by any subaerial process. Further, although both cirques and overdeepenings reflect similar focussing of erosion beneath the ELA under different ELA positions, many factors support valley deepening and overdeepening being far more efficacious than cirque erosion (c.f. Herman and Braun, 2008), and modelling shows that the spatial focussing of erosion that is produced by overdeepening-related feedbacks is essential in producing characteristic glacial valley relief (Herman et al., 2011; Egholm et al., 2012). Trunk-valley erosion and overdeepening should therefore play an important role in landscape evolution despite cumulative glacial occupancy of trunk-valleys being lower than that of their cirques and valley-heads.

The factors that support valley deepening and overdeepening being more efficacious than cirque erosion are summarised in Figure 19, where a typical glacier and valley long-profile has been subdivided into zones according to the relative contribution of interglacial and glacial processes to erosion and landscape evolution. Zones B and D are zones of very efficient glacial erosion because abundant surface melt promotes efficient sliding and sediment flushing. Erosion is most efficient in zone D during glacial conditions because: (i) larger accumulation and ablation areas produce higher balance fluxes; (ii) the larger ablation area results in more abundant surface melt; and (iii) topographic focussing of ice flow increases the flux of ice per unit area of the bed. Erosion is least efficient in zone A because of the absence of surface runoff and absence of strong topographic focussing of ice flow during both glacial and interglacial conditions. Compared with zone D, zones B and C are also areas of less efficient glacial erosion, for similar reasons. Nevertheless, deep erosion in zone D, and to a lesser extent in zone B, will steepen the bed of zones A and C, such that some erosion of these latter zones may occur through ice-erosion feedback (section 4). Headwall erosion in zones B and D will be limited by height-mass balance feedback (section 4).

The above discussion indicates that, over many glacial-interglacial cycles, there will be intense trunk-valley deepening and only moderate cirque-type deepening, with areas of overdeepening separated by a prominent valley-step. This is despite glacial occupation of zones A-C being much greater than that of zones D and E, which further reinforces the argument that cirque erosion, with respect to glacial valley erosion, is far less efficacious. The depth of deepening in zones B and D, which eventually produces closed-basin overdeepenings, is limited by the steepness of their respective adverse slopes, which limit the efficiency of water and sediment transport from the glacier system. Hence, the depth of erosion in zones B and D is limited by that in zones C and E, respectively, which are zones of little or no glacial erosion, and may be dominated by non-glacial (i.e. fluvial) erosion.

The rate of landscape response to the lowering of base-levels and steepening of valley-sides that is associated with cirque and overdeepening formation will be greatest as deglaciation exposes previously subglacial valleys and basins (e.g. Meigs and Sauber, 2000; Hinderer, 2001; Dadson and Church, 2005), destabilising over-steepened slopes (cf. Burbank et al., 1996). In addition, in the postglacial landscape, overdeepenings become important sinks for sediments produced by erosion of adjacent valley-sides and mountain peaks (Hinderer, 2001; Straumann and Korup, 2009). For example, Straumann and Korup (2010) have estimated that 90% of sediment storage in mountain belts occurs in the lowest 25% of the elevation, reflecting valley-fill storage as a result of the downstream decrease in fluvial transport capacity, but also filling of closed-basins formed by overdeepening; notably, the proportion of basin area covered by valley-fill increased with basin size, reflecting the greater degree of glacial modification that had occurred in larger basins. Champagnac et al. (2009) have illustrated the importance of such storage, demonstrating that loading as a result of sedimentation within the large overdeepenings that occupy the Swiss Alpine Foreland has balanced exhumation stimulated by post-orogenic decay of the Alpine mountain belt. Nevertheless, Straumann and Korup (2010) point out that the processes and distribution of postglacial sediment storage are still poorly known and are not explicitly considered by numerical models of landscape evolution.

Another significant yet little explored role of glacial valley-deepening and overdeepening is in the co-evolution of ice sheets and their topography, in which erosion-induced feedbacks over many glacial cycles might be expected to cause evolution in ice dynamics and extent. Such evolution is illustrated by the branching of equilibrium states and subsequent hysteresis in ice mass behaviour that occurs when valley glaciers erode overdeepenings (e.g. Oerlemans, 1989). Eyles et al. (1991) argued that overdeepening preconditioned the Cordilleran ice sheet towards dynamic and rapid disintegration similar to that exhibited by tidewater glaciers (e.g. Howat et al., 2008). More recently, Kaplan et al. (2009) has speculated that cirque erosion and U-shape valley development during the Quaternary period, which has led to the lowering of mass-accumulation areas, is likely to have affected the mass balance profile and mass-flux during succeeding glaciations, resulting in a reduction in the maximum extent of glaciation over successive glacial cycles. Numerical modelling of ice sheet growth over a simulated landscape with and without mature outlet valleys supports this idea (Kessler et al., 2008), and further demonstrates that more efficient discharge of ice through widened and overdeepened troughs means that successive ice sheets should take longer to reach a steady-state geometry.

Similar theories have been proposed to support observations of the long-term behaviour of marine ice sheets and ice streams (e.g. Hambrey and McKelvey, 2000; Figure 18). ten Brink and Schneider

(1995) have argued that an overdeepened continental-shelf morphology is a general feature of long-term ice sheet glaciation because, during glacial periods, the preferred location of the grounding-line is at the shelf-edge, causing erosion to be focussed on the inner shelf, and deposition to be focussed on the outer shelf and shelf-edge (Figure 18b); overdeepening of the inner shelf reinforces this pattern because grounding-line positions behind the shelf-edge are unstable (Figure 18c; cf. Figure 13a). In the Lambert Glacier-Amery ice shelf system, East Antarctica, the onset of overdeepening appears to have been associated with expansion of the ice to the shelf edge (Cooper and O'Brien, 2004), but the consequence has been to increase the volume of ice needed to reach the shelf edge (Taylor et al., 2004; cf. Kessler et al., 2008) and to reduce the number of ice advances to the shelf edge during the Late Pleistocene, even though global ice volumes have increased (O'Brien et al., 2007). Nevertheless, Domack et al. (2006) have hypothesised that overdeepening beneath the Palmer Deep outlet system, West Antarctica, has promoted advance of ice to the continental shelf-edge by enabling the development and drainage of a subglacial lake and associated ice streaming. This highlights the potentially important role of overdeepenings in promoting fast flow (see section 5.1.2.1).

5.2.4. Synopsis

Subglacial basins exert significant influence over rates and patterns of glacier erosion, sediment transport and deposition because hydraulically inefficient subglacial drainage, glaciohydraulic supercooling and enhanced terminal flow compression promote thicker basal sediment and basal ice layers and the entrainment and transport of sediment by glacial processes. Glaciers terminating in overdeepenings may therefore possess elevated moraine building potential, and the important role of supercooling and thrusting means ice-marginal landsystems may be characterised by unique but as yet poorly understood process-landform associations. Basins are also important sedimentary sinks, and play an important role in storing sediment post-glacially.

On longer timescales, the erosion and filling of overdeepenings must play an important role in glacial and post-glacial landscape evolution and the evolution of ice mass geometry and dynamics. Importantly, landscape analyses and theoretical considerations indicate that overdeepening is more efficacious than cirque erosion and should play a dominant but as yet poorly understood role in glacial and postglacial landscape evolution. Further, overdeepening of outlet glacier and ice sheet beds over many glacial cycles changes the geometry and dynamics of subsequent ice masses, increasing the sensitivity of successive ice masses to climate change (cf. Section 5.1).

6. DISCUSSION: KEY QUESTIONS

Whilst some aspects of overdeepening and its importance have received significant attention, many aspects remain poorly researched and understood. In this section, we raise key questions concerning the origin and importance of overdeepening in glacial systems and landscapes, and seek to highlight those that are worthy of further investigation.

6.1. Overdeepening: why, how, where, when?

Almost all aspects of the origin of overdeepened basins remain poorly known, which severely restricts understanding and prediction of their importance in glacial systems. Key unresolved questions are: (i) why [does overdeepening occur]? (ii) how? (i.e. what are the main processes?) (iii) where? (e.g. how terminal overdeepenings form and what controls their distribution?) and (iv) when? (e.g. is overdeepening characteristic of a specific stage of glaciation or evolution of the landscape?). The solutions to these questions are of course linked. For example, robust understanding of the processes of overdeepening and the conditions under which they occur should help to elucidate why, when, and where. A great many avenues of research are therefore likely to provide valuable insight. It is further likely, however, that there exists more than one type of overdeepening.

Ideas concerning why overdeepening occurs are principally that: (1) an overdeepened bed geometry is an equilibrium form that all glacier beds should tend towards; and (2) an overdeepening is the product of an initial instability in ice flow that is amplified by ice-erosion (and hydrological) feedbacks. The equilibrium hypothesis has been articulated in its clearest form by Alley et al. (2003a), who have argued that ice mass geometry dictates that excess energy for erosion and sediment transport is only satisfied when the bed slope is negative, the key limiting process being the supercooling of water ascending adverse slopes in excess of ~ 1.2–1.7 times the ice-surface slope. Few glaciers or glaciated valleys, however, appear to have uniform overdeepened bed profiles that reflect the anticipated pattern of along-flow erosion and sediment transport potential. It is possible that more complex bed profiles reflect: (1) the diachronous nature of glaciation, which will cause contrasting patterns of erosion and sedimentation to be superimposed over time; (2) the focussing of erosion and sediment transport in multiple locations, such as at glacial confluences; or (3) the absence of sufficient time for glacial erosion processes to produce an equilibrium bed geometry. Further, some studies (e.g. Swift et al., 2008; Jordan, 2010; Preusser et al., 2010) indicate that overdeepening is promoted by changes in geological structure or lithology; these may initially produce small bed irregularities that are enhanced by ice-water-erosion feedbacks, and, given sufficient time, may coalesce into a single graded overdeepening.

The processes of overdeepening are poorly known because they are difficult to observe in nature and cannot be replicated in the laboratory. Numerical modelling indicates that patterns of ice flow and glacial occupation (e.g. Oerlemans, 1984; Macgregor et al., 2000; Anderson et al., 2006) should alone produce features analogous to overdeepenings, and certain patterns of overdeepening appear to reflect instabilities that are purely related to ice flow (e.g. Mazo, 1989; Comeau, 2009). However, the importance of subglacial water pressure for basal sliding and quarrying, and the requirement for the products of erosion to be evacuated from the glacier system, indicates that realistic models require ice, water and sediment transport processes to be coupled (Herman et al., 2011; Egholm et al., 2012). Which processes and feedbacks are most important? Alley et al. (1997, 2003a) have argued that overdeepening is produced by bedrock abrasion that is sustained by efficient flushing of sediment from the ice-bed interface; no positive reinforcement of this process is specified, although focussing of ice and water flow into the incipient overdeepening would provide an obvious feedback. Hooke (1991) argued that overdeepening is produced by quarrying sustained by focussed contributions of surface runoff: positive reinforcement here occurs through the influence of bed topography on the location of crevasses that intercept and direct surface runoff to the bed. As has been described, these models lead to contrasting expectations of overdeepening location and geometry, yet they have been addressed largely in a theoretical manner only, and their implications have yet to be tested against numerical predictions (e.g. Egholm et al., 2012) or statistical analysis of real phenomena (as has been undertaken for drumlins; e.g. Clark et al. 2009). Further, these models are based primarily on observations at only two glaciers (Storglaciären and Matanuska), and more recent studies have directly challenged such models (e.g. Creyts and Clarke, 2010), or have advocated additional feedback processes of as yet unknown influence (e.g. Swift et al., 2006).

Some insight into overdeepening location and therefore possible origin can be drawn from the large number of examples that have been documented, either intentionally or unintentionally, in published studies (Table 1). These indicate that overdeepening is favoured in certain contexts and locations, notably: (1) in cirques; (2) at confluences and valley-constrictions; (3) at lithological changes or changes in bedrock strength; and (4) beneath glacier termini. Cirque- and confluence- and/or constriction-type overdeepenings appear to arise because of the strong dependence of erosion on ice velocity (particularly basal sliding velocity), such that erosion is at a maximum where ice flux is greatest; this generally occurs beneath the ELA, explaining the focussed erosion of cirque basins, but fluxes will also peak locally when ice flow is focussed by topography. Lithological changes or bedrock weaknesses are likely to be exploited by erosion, producing changes in bed topography that are amplified by ice-erosion feedbacks. Overdeepenings in terminal regions are, however, more problematic, because ice flux and hence erosional potential should decrease towards the terminus, especially in the context of diffluent flow; the influence of hydrology (e.g. basal sediment evacuation by meltwater) and/or ice-erosion feedbacks in terminal regions must therefore be very strong. Some terminal overdeepenings are located entirely in unconsolidated sediments; for example, piedmont-type glaciers are often overdeepened where the terminus rests on glacial outwash (e.g. Svínafellsjökull and Kvíárjökull, Iceland). Nevertheless, terminal (or formerly terminal) bedrock overdeepenings are not uncommon, such as the well-documented examples that occupy the Swiss Alpine Foreland (Figure 4g). Numerical modelling (Egholm et al., 2012) indicates that full understanding of overdeepened basin hydrology, sediment transport and ice-water-sediment feedbacks is essential to understand the origin and distribution of these landforms.

The apparent significance of hydrology indicates that overdeepening should be favoured during periods of stability and active retreat, when contributions of surface melt to the ice-bed interface are greatest and should sustain the highest rates of basal sliding, basal sediment evacuation, and erosion. Conditions during ice advance, in contrast, are likely to be associated with reduced water availability but enhanced availability of sediment, resulting in glacial overriding and reworking of sediment-filled overdeepenings comprising sediment of non-glacial and glaciofluvial origin (e.g. Alley et al., 2003a). Nevertheless, the preservation of pre-LGM sediments within many overdeepenings (e.g. Dehnert et al., 2012) indicates that not all glaciations afford sufficient time, achieve the necessary geometry, or are characterised by sufficiently efficient subglacial drainage systems that have sufficient sediment transporting capacity, to produce renewed overdeepening. Preservation of sediments is perhaps most likely in terminal overdeepenings, in which many of the oldest sediments appear to have been found, because excavation and renewed overdeepening will depend on climate enabling a glacier geometry to be reached and maintained and that is characteristic of, and at least as extensive as, the previous most extensive glaciation. Preservation is further likely because flow of ice that has re-advanced over an ‘old’ sediment-filled overdeepening or basin is likely to be characterised by significant rates of basal sediment deformation and hence lower ice surface slopes.

This raises the related question of how much overdeepening is possible, the answer to which requires greater understanding of the processes that enable ice, water and sediment transport through overdeepened basins and their fundamental limits. Ice dynamics alone indicate no such limit because, providing sufficient ice flux is maintained, any resistance to flow that is produced by the formation of an adverse slope should produce thickening of ice above the overdeepening until driving-stresses are sufficient for basal resistance to flow to be overcome; the only potential limit is for flow separation to produce immobile ice at the glacier bed (e.g. Gudmundsson, 1997), a phenomenon that has yet to be supported by field observation. Instead, subglacial hydrology is likely to be most important in governing patterns of sediment transport and erosion within an overdeepened bed. Importantly, the tendency for basal water pressures within overdeepenings to be at or near overburden, and for ice occupying overdeepenings to tend toward lower surface slopes, indicates that hydrological conditions within overdeepenings enhance basal sliding and hence erosional potential. However, the inverse relationship between drainage efficiency and steepness of the adverse slope, which is responsible for the high basal water pressures with overdeepenings, will reduce drainage transmissivity and the competence and capacity of subglacial flowpaths, and thus encourage retention of sediment at the glacier bed. It is therefore likely that the supercooling threshold, or some other threshold that promotes englacial rather than subglacial water flow, provides the ultimate constraint.

In the absence of a supercooling threshold, a further and very strong constraint on the limit of overdeepening is that, continued overdeepening will ultimately result in ice within the overdeepening becoming buoyant. For a glacier that terminates within the overdeepening, such a situation would cause acceleration and thinning and thus rapid calving and retreat; however, if the ice mass extended beyond the overdeepening, acceleration and thinning should instead lead to ponding of water and the formation of a subglacial lake (cf. Livingstone et al., in press). This latter outcome is perhaps unlikely at ice masses where surface runoff access the glacier bed and subglacial erosion rates are therefore relatively high, because glacial and fluvioglacial sediment transported from non-overdeepened areas would likely accumulate in the incipient lake and therefore maintain contact between the ice and its bed. Height-mass balance feedbacks will also be important in limiting overdeepening, particularly its upglacier extent. However, for large ice masses, where overdeepening will not necessarily impinge on the accumulation area, the depth of overdeepening is likely to be dependent on the extent to which ice-water-sediment feedbacks (section 4) are able to both focus erosion and maintain the evacuation of water and sediment.

Because erosion requires ice-bed contact, efficiencies of water and sediment evacuation are likely to be key limiting factors. Thus, even if overdeepening is dominated by localised quarrying-related ice-water-erosion feedbacks, glacier beds should tend toward overdeepened geometries that reflect the broad controls on water and sediment transport. This highlights the need for numerical models that incorporate coupled ice-water-sediment-erosion processes (cf. Herman et al., 2011; Egholm et al., 2012). At present, many of these processes and feedbacks, including thrusting, supercooling and the transport capacity and erosive power of hydraulically efficient channels, are absent from or are represented poorly in numerical glacier models, largely because their mechanisms and the controls on their efficiency are still poorly known. Improved models are therefore needed, as well as improved data on overdeepening depth and location, to enable testing of modelled process implications in regions where former ice dynamics and extent are well known (e.g. the Norwegian fjordland and the Northern Alpine foreland of Switzerland).

6.2. How does overdeepening affect the hydrology of glaciers and ice sheets?

The nature of subglacial drainage and magnitude of surface water contributions to the glacier bed are critical in determining basal water pressures and hence glacier dynamics, but also the mechanisms and spatial patterns of erosion and sediment entrainment, deposition and transport. There is consensus that basal water pressures within overdeepenings are generally higher than other areas of the bed and often approach ice-overburden pressure. However, many crucial questions remain, including: What proportion of flow remains subglacial when water traverses an overdeepening? How and under what conditions can subglacial flowpaths, especially efficient channels, traverse overdeepenings? Does the hydrology of the overdeepened bed evolve seasonally? How do overdeepenings affect the nature and seasonal evolution of adjacent (e.g. upglacier) regions that are not overdeepened?

Although research into the nature of drainage in overdeepened areas has been undertaken at remarkably few glaciers, most evidence available indicates that flow at the bed is limited and occurs mainly via low transmissivity, hydraulically inefficient and predominantly distributed pathways (e.g. Hodge, 1976, 1979; Hantz and Lliboutry, 1983; Röthlisberger and Lang, 1987; Hooke, 1991; Fountain, 1994; Alley et al., 1998; Lawson et al., 1998). This observation is supported by theoretical studies, which further indicate that flow should largely bypass overdeepenings, either englacially or laterally (Lliboutry, 1983; Hooke and Pohjola, 1994). This has implications for the nature and temporal pattern of basal water pressures within the overdeepening: specifically, pressures should remain high, but should not be responsive to diurnal or seasonal trends, and may remain below the temporarily high pressures experienced near to subglacial channels that are fed directly by diurnally peaked surface melt (cf. Hubbard et al., 1995). Field data, mainly from Storglaciären, indicates that this picture is at least partially true: some water does reach the bed, but channels that convey large volumes of melt are not present, and do not form in response to seasonal increases in melt. Instead, englacial flowpaths seem to be preferred, and these capture surface runoff that is produced in the region of the overdeepening and convey it to areas downglacier of the overdeepened bed.

What is known less well are how such englacial networks originate and what the implications are for the wider hydrological evolution of the subglacial drainage system, including whether subglacial channels are able to form upglacier of overdeepened areas, and what happens to channels on reaching an overdeepening. This is of fundamental importance to glacier dynamics because the seasonal formation of hydraulically efficient channels limits the potential for melt-induced acceleration of ice flow during spring and summer (e.g. Sundal et al., 2011), except perhaps where diurnal melt is sufficiently peaked to cause water pressures to exceed overburden (e.g. Mair et al., 2003; Swift et al., 2005; Bartholomous et al., 2007). At non-overdeepened glaciers, the channel network extends upglacier as the snowline retreats and runoff from low-albedo glacier ice produces strong diurnally peaked water pressure fluctuations at the glacier bed. These pressure fluctuations locally destabilise the distributed drainage system (e.g. Kamb, 1987; Nienow et al., 1998), resulting in the formation of a channel that presumably 'grows' downglacier to join the incipient, upglacier-extending channelised drainage network. Overdeepenings present an intriguing problem for this model because thicker ice in the overdeepening should mean the distributed system is more resistant to destabilisation by the same meltwater inputs (Kamb, 1987). Further, even if sufficient inputs reached the bed of areas upglacier of the overdeepening, any channel would have to ‘grow’ downglacier through the overdeepening. The presence of overdeepening-bridging englacial conduits might enable upglacier drainage evolution unfettered by overdeepening, but, like channels, such conduits should close by inward deformation as melt subsides, and are therefore unlikely to persist over winter.

Assuming that channels form at the bed of overdeepenings, further uncertainties exist regarding the conditions within such channels. For example, it is unclear whether channel size and the water pressure within such channels would respond to seasonal and diurnal variations in melt. Further, little is known about the configuration of the subglacial drainage system (i.e. distributed or channelized) and the associated impact on ice motion. Creyts and Clarke (2010) demonstrated that where the adverse slope is sufficiently steep to allow the supercooling of subglacial water, ice will form preferentially within channels, especially overnight when channels remain full but hydraulic gradients, and hence flow velocity and viscous dissipation, are low. Creyts and Clarke (2010) therefore concluded that daytime flows should reach overburden pressure quickly and distribute laterally as a sheet. Below this threshold slope, there is no reason for water to distribute as a sheet. Nevertheless, because channels will always be flooded, it would appear likely that channel shape should be semi-circular or almost circular in cross-section, as opposed to being broad and low, which is characteristic of channels in non-overdeepened areas where flow alternates diurnally between open and closed conditions (cf. Hooke and Pohjola, 1994). As a result of this channel shape, viscous dissipation will be less effective (cf. Creyts and Clarke, 2010), and hence the conductivity of the channel system, especially its ability to transmit peak daily melt, should be more limited than that of channels in non-overdeepened areas. Hence, daytime flows will reach overburden pressures more rapidly than in non-overdeepened areas, and flow will distribute more readily across the bed.

The limited evidence that is available therefore indicates that drainage within overdeepenings is predominantly inefficient and that, in the absence of efficient englacial and/or latero-subglacial flowpaths, areas upglacier of overdeepenings will be similarly inefficient. Even if channels are present within and upglacier of overdeepenings, it seems likely that they will develop more slowly and remain less efficient than at glaciers where overdeepenings are not present. It follows that, even without the possible influence of supercooling, the nature and seasonal pattern of evolution of basal water pressures at glaciers with overdeepenings is likely to be very different to those characteristic of non-overdeepened systems. Notably, delayed or restricted evolution of an efficient channel system should cause widespread high basal water pressures that are present at the onset of the melt season to persist for longer, and, later in the melt season, for diurnally-high water pressures within channels (if present) to exceed overburden pressures more frequently.

The possible implications of overdeepening for basal water pressure and sliding rates are summarised in Figure 20. For a typical non-overdeepened bed, water pressure/sliding is initially low, reflecting the negligible volume of surface runoff, and increases steadily during periods 1 and 2 as surface runoff reaches a hydraulically-inefficient distributed drainage system; water pressure/sliding then begins to decline during period 3, despite the sustained increase in surface runoff, as a consequence of the formation of hydraulically-efficient subglacial channels. For the overdeepened bed, however, water pressure/sliding is persistently high as a result of the low transmissivity of subglacial drainage within overdeepenings, and rises still further during periods 1 and 2; water pressure/sliding then begins to decline in period 3 if subglacial water can exploit alternative englacial flowpaths (curve i), and may fall below spring values, if only temporarily, if conditions allow the formation of efficient subglacial channels. During periods 3 and 4, water pressure/sliding remains above that of the non-overdeepened bed as a result of the low transmissivity of englacial flowpaths with respect to that of subglacial channels, and the reduced efficiency of subglacial channel-flow in the presence of an adverse slope. In the absence of englacial flowpaths or subglacial channels, water pressure/sliding may continue to rise and subsequently decline in lock step with surface runoff (curve ii).

There also may be implications for areas of the bed that are upglacier of overdeepenings. Most obviously, if overdeepenings delay or inhibit the seasonal evolution of hydraulically efficient

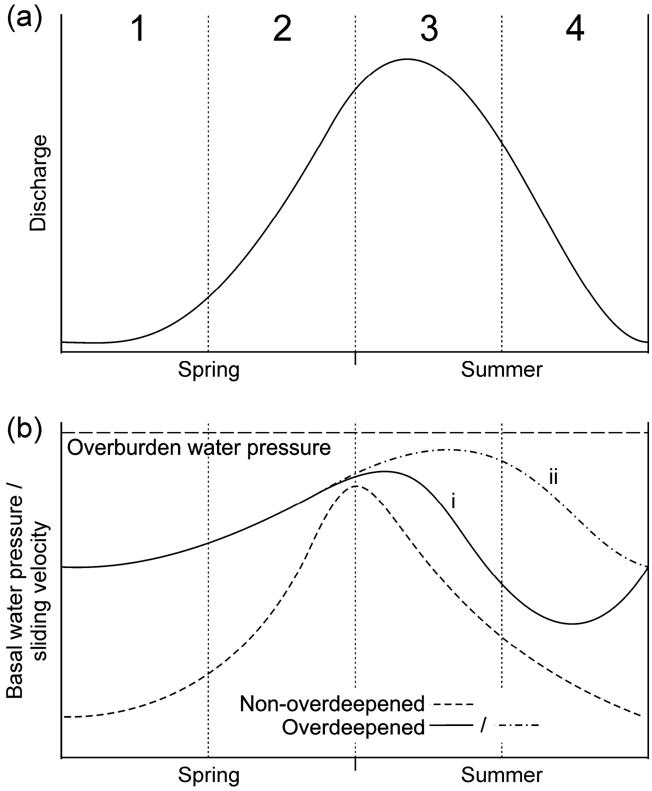

Figure 20. Schematic diagram showing basal water pressure and sliding velocity over a melt season for a valley or outlet glacier with an overdeepened or non-overdeepened bed. Water supply from surface runoff and basal melting is shown in (a) and hypothesised water pressure and sliding rate in (b). For the non-overdeepened bed, water pressure and sliding initially increases with water supply and then declines rapidly as subglacial flow switches from hydraulically inefficient flowpaths to hydraulically efficient subglacial channels. For the overdeepened bed, water pressure and sliding are persistently higher as a result of the lower transmissivity of subglacial flowpaths that must ascend the adverse slope. Higher pressures and sliding rates may also be sustained for longer because drainage system evolution, including switching of flow to alternative englacial flowpaths, is likely to be slow (curve i) or severely limited (curve ii).

subglacial drainage systems then upglacier areas should also retain high basal water pressures, provided alternative englacial or lateral flowpaths that traverse or avoid the overdeepening are also absent. Importantly, these areas of the bed may be highly sensitive to seasonal and diurnal melt variability, which might be communicated downglacier within a distributed subglacial system as a pressure wave, as has been observed at Storglaciären (Hooke et al. 1992; Jansson, 1995, 1996).

Clearly, it is undesirable that there are so few detailed analyses of the hydrology of overdeepenings. Studies at Storglaciären in particular offer an insight into the possible configuration and time-evolving behaviour of water flow in overdeepenings, and hence the opportunity to develop testable hypotheses and predictions of how water negotiates overdeepenings. The core theme for future hydrological research should be to evaluate the extent to which the hydrology of overdeepened areas differs from non-overdeepened areas and the implications for ice motion, dynamics and sediment transfer (sections 6.4, 6.5 and 6.6). Field data on the hydrological configuration within overdeepenings and its evolution over seasonal and diurnal timescales are lacking, and ideally, such studies should be undertaken at a range of other glaciers and augmented by modelling studies. There are also questions about whether the hydrological conditions experienced within overdeepenings beneath valley glaciers (such as Storglaciären) are also experienced beneath outlet glaciers and ice streams. Overdeepenings are common and appear to have very different hydrological characteristics compared to non-overdeepened beds, and it is apparent that water flow beneath ice sheets fundamentally alters their behaviour. This knowledge suggests that overdeepenings could have a profound influence on the behaviour of valley and outlet glaciers, yet we have barely begun to quantify the influence of overdeepenings on hydrology in such settings.

6.3. How pervasive is glaciohydraulic supercooling and what is its significance?

Glaciohydraulic supercooling has received much attention as a process of sediment transfer, most notably because of its potential to produce debris-rich 'stratified facies' basal ice (e.g. Lawson et al., 1998; Cook et al., 2010). However, many other aspects that are relevant to ice dynamics, such as the potential contribution of melt-out from fine-grained supercool-facies basal ice to till continuity and ice stream dynamics (see section 5.1.2.1), have been relatively overlooked. Further, little attention has been paid to spatial and temporal patterns of supercooling, especially in light of likely seasonal and diurnal evolution of subglacial water flow and drainage pathways, the possibility that englacial pathways may also convey water across overdeepened areas, and the changes in bed and ice-mass geometry and melt regimes that are associated with cycles of glacial advance, stability and retreat.

The idea that supercooling is a pervasive process within overdeepenings seems at present to be widely accepted and has influenced recent research on overdeepening formation and geometry significantly. Supercooling is the central stabilising-feedback process in the model of Alley et al. (1999, 2003a,b) because it presents efficient erosion and sediment evacuation when overdeepening causes the adverse slope to meet a certain threshold slope. However, there is little *prima facie* evidence that sediment entrainment and transport by the formation of debris-rich supercool-facies basal-ice is any less efficient than evacuation by subglacial drainage, except in theoretical terms (e.g. Alley et al., 1997). Even if it were not as efficient, the formation of supercool-facies basal-ice could still result in net erosion of the adverse slope, and therefore eliminate any stabilising-effect, if sediment supply was sufficiently limited; for example, if, for sediment availability reasons, subglacial channels entering the overdeepening carried little sediment. It is, therefore, important to know where supercool-facies formation takes place, the rate at which formation takes place and entrains sediment, and how the rates of transport by supercool-facies basal ice compare with the rates of fluvial transport.

Field evidence regarding the spatial and temporal pervasiveness of supercooling at individual glaciers is sparse. Based on work at Matanuska Glacier, Evenson et al. (1999) provided a detailed account of the features that can be used to identify the existence of supercooling; the presence of many of these features, including fractures filled with platy ice, upwelling supercooled water containing frazil ice, and silt-laden basal ice, have been used to support the existence of supercooling elsewhere (e.g. Tweed et al., 2005; Cook et al., 2007, 2010). However, such studies have generally demonstrated that field evidence for supercooling is restricted only to small portions of the glacier margins (Cook et al., 2007, 2010), or that discharge of subglacial supercooled water does not persist continuously over time (Tweed et al., 2005). Overall, field evidence for glaciohydraulic supercooling appears to be sparse at most glaciers other than Matanuska.

Recent modelling work by Creyts and Clarke (2010) supports the idea that supercooling should be spatially and temporally limited because supercooling should occur in large channels only, which cover only a small proportion of the glacier bed, are only metres wide, and will be restricted to periods when the hydraulic gradient is low (i.e. during night-time and morning). Further, evidence of the process will be even more limited because the majority of ice formed by supercooling in large channels will be destroyed when the hydraulic gradient is high and flow velocities in channels increase (i.e. during midday and afternoon). The implication is that supercooling may be very common but glacier beds will continue to evolve toward greater supercooling rather than stability and that physical evidence of supercooling will be very sparse. The model also overturns the idea that supercooled-facies freeze-on occurs mainly within the inefficient distributed subglacial system (e.g. Röthlisberger and Lang, 1987; Hooke, 1991; Alley et al., 1998) because viscous dissipation per unit area is higher in a broad, flat flowpath, such as a sheet, than in a semi-circular or circular channel. Nonetheless, field evidence of supercool-facies basal ice does seem to indicate broad zones of accretion that are consistent with distributed flow (e.g. Lawson et al., 1998).

Uncertainties regarding the nature of flow across overdeepenings and the nature of seasonal evolution of such flowpaths are also significant limitations. For example, Creyts and Clarke (2010) did not allow the subglacial drainage in their model the freedom to open or exploit alternative englacial pathways. If such pathways are possible, then closure of incipient channels by supercool-facies ice formation could drive water into higher englacial pathways (e.g. Hooke and Pohjola, 1994) that would be less susceptible to supercooling and would undoubtedly grow more efficiently. As a result, supercooling might take place only in a very limited period when spring melt reaches the bed and before flow switches to a predominantly englacial system. Clues as to the operation of such flow switching come from ice motion studies such as Hooke et al. (1992) where the greatest forcing of ice motion occurred in spring when melt reached a 'winter' drainage system on the adverse slope of the overdeepening. Such flow switching might lead to even more supercooling in channels at the bed if the effect of englacial drainage is to reduce the hydraulic head.

The pervasiveness of supercooling is important because it has been linked to the rapid and widespread accretion of thick debris-rich basal ice that may be responsible for a wide range of unusual geomorphic and sedimentological phenomena, including the possibility of creating anomalously thick melt-out tills (e.g. Larsen et al. 2004) and Heinrich Layers (e.g. Roberts *et al.* 2002; Andrews & Maclean 2003; Hulbe *et al*. 2004). However, it may also have ice dynamic implications, especially at ice sheet scales. Notably, the sediment content of ice has been suggested to influence ice rheology (e.g. Hubbard and Sharp, 1989; Alley et al., 1999), and melt-out of sediment from debris-rich supercool-facies, basal ice could contribute importantly to till continuity (Alley et al., 2003b). Nevertheless, few studies have attempted to quantify rates of freeze-on or their wider ice-dynamic implications. Estimates of extent and sediment entrainment efficiency of supercool-facies basal ice formation vary greatly (e.g. Larson et al. 2006; Cook et al., 2010) and it is clear that the process controls are poorly understood. Modelling work by Creyts and Clarke (2010) has made an important step towards understanding spatial and temporal patterns of supercooling and its significance, producing supercool-facies basal ice accretion rates of 3–8 cm a^{-1} (i.e. 3–8 m per century) that are reasonably consistent with rates derived from field studies (e.g. Lawson et al., 1998), but crucially their model did not include processes of sediment entrainment and transport.

A recent study has further highlighted the impact of supercooling at ice sheet scales. Bell et al. (2011) estimated that ~24% of the area beneath Dome A, East Antarctic Ice Sheet, experienced freeze-on processes, of which ~16% could be attributed to glaciohydraulic supercooling. These packages of freeze-on ice ranged from 200 to 500m in thickness, could be traced for up to 30km along flow direction, and represented at least 10000 to 20000 years of ice accretion. Such estimates of the pervasiveness of freeze-on processes (including supercooling) are important because, in the case of Dome A, the additional ice thickens the ice column alters basal ice rheology and fabric, upwarps the ice sheet, and changes flow behaviour. However, it is likely that such processes will only have a significant mass balance effect beneath larger ice masses such as ice sheets (rather than beneath valley glaciers) because only here is the ice sufficiently stable over long periods of time to allow gradual accretion to glacier-scale thicknesses. For many valley glaciers, much of the accreted ice may be melted by subglacial water erosion and frictional heating from sliding.

The operation and importance of supercooling may be determined by cycles of glacier advance and retreat. We speculate that supercooling is likely to be more prevalent when a glacier is advancing across an overdeepening and the ice presses against the adverse slope. In the initial stages of advance the glacier flows into the overdeepening, and with no adverse slope there cannot be any supercooling. The same is true for retreat out of an overdeepening. Where an overdeepening exists some distance behind a retreating terminus it might be expected that more abundant melt would enhance the potential for supercooling to operate (notwithstanding Creyts and Clarke, 2010, and the possibility of steeper hydraulic gradients and more viscous dissipation leading to more melt than freeze-on).

Overall, there is a growing recognition of the importance of supercooling, not only for sediment transfer and basal ice formation, but also for glacier hydrology and processes of ice motion. Recognising this influence has proved challenging because the field evidence for supercooling is sparse, relatively few field studies have been undertaken to understand the spatial and temporal patterns of supercooling, and modelling studies, which offer important insights into the largely inaccessible subglacial environment, are few and require further testing against field evidence. These shortcomings in our existing knowledge offer opportunities for future research. As a starting point we need greater understanding of when and where supercooling takes place. We need to understand what controls supercooling, and explain why some overdeepened beds promote the operation of this process, whilst others do not seem to do so. These issues lead us then into research to understand when and where supercooling is important, and what the nature of its influence might be.

6.4. How does overdeepening affect glacier and ice sheet motion?

It has been demonstrated that overdeepenings exert at least some influence over the mechanisms and spatial and temporal patterns of glacier motion (section 5.1.2). Notably, although the majority of surface runoff may in fact be diverted away from the bed of overdeepenings by englacial drainage, the suppression of efficient subglacial drainage within overdeepenings appears to enhance the sensitivity of ice motion to seasonal, diurnal and even transient variations in melt. One could assume therefore that basal water pressures are persistently close to ice overburden pressure and that the system is poised for even minor changes in melt to produce rapid and widespread increases in ice-bed separation and hence basal sliding. However, it is unclear whether this is a consistent feature of overdeepening and whether it is significant in terms of overall rates and processes of ice flow. Most of what is known about this phenomenon is from combined field observation of water pressure and ice motion at Storglaciären, Sweden; few modelling studies exist and few comparable studies have been undertaken at other glaciers. A further limitation is the uncertainty over the nature of the glacial drainage system where an overdeepening is present.

There is general agreement that the drainage system at the bed of overdeepenings remains inefficient and therefore in summer the transmissivity of the system is quite different from that of non-overdeepened areas. However, the origin and implications of this difference in transmissivity are uncertain: Creyts and Clarke (2010) argue that at Matanuska glacier it is a result of supercool-facies ice formation in larger channels, whereas at Storglaciären they indicate that it is the result of a distributed system, in which Hooke and Pohjola (1994) have suggested that flow occurs ‘sporadically’ via a network of interlinked ‘storage pockets’ between the ice and thick till. If channels exist, then seasonal increases in surface runoff might lead to some increase in drainage efficiency and decrease in water pressure, but supercooling is likely to limit channel growth (Creyts and Clarke, 2010) (see section 6.2), such that basal water pressures and ice motion should respond strongly to seasonal and diurnal variability in melt. If channels are not present then flow must occur through a distributed system, in which case basal water pressures and ice motion should respond strongly to changes in melt, or through englacial or lateral channels that bypass the overdeepening, in which case there may be few noticeable effects. Nevertheless, even where there is good evidence for englacial drainage across the overdeepening, for example at Storglaciären, there is still evidence that seasonal and diurnal changes in melt have a significant effect on ice motion (e.g. Hooke et al., 1989). There is even evidence to suspect some kind of seasonal evolution of the distributed system at Storglaciären, such that ice motion effects are greatest in spring when melt reaches a ‘winter’ system (Hooke et al., 1992), although the changes in subglacial transmissivity that these observations imply may simply reflect seasonal evolution of greater transmissivity in the englacial system. The sliding mechanism is also uncertain: data from Storglaciären indicate that the high basal water pressures within overdeepenings, whilst giving rise to weak water-saturated tills (Hooke et al., 1989), tend to reduce effective pressure on the bed and hence reduce the potential for glacier motion through deformation of subglacial sediments (Iverson et al., 1995) and basal ice.

Other processes of ice motion may also be influenced by overdeepening. Notably, the enhancement of longitudinal flow-compression may lead to folding and/or thrust faulting of glacier ice. Thrusts have been observed at several overdeepened glaciers where they appear to have transported basal debris to the surface of the ice (e.g. Swift et al., 2006), and recent work has imaged debris-rich thrusts emanating from the base of glaciers (Murray and Booth, 2010), but the processes that encourage and allow thrusting on glacier-thickness scales remain elusive and the magnitude of displacement is difficult to constrain (e.g. it may be many glacier thicknesses). Whether tectonic processes of this nature operate within overdeepenings beneath ice sheets is more uncertain, but is certainly likely where fast-flowing outlet glaciers flow against adverse bed slopes. Flow compression and its accommodation through thrusting and/or folding could be important dynamical effects, perhaps with temporary accommodation of ice flow until failure along thrusts, or progressive deformation for folding. Gudmundsson (1997) has indicated that ice in the deeper parts of overdeepenings may be stagnant, but the characteristics of sediment within thrusts indicates that material is entrained from the bed, often from the lag deposit that presumably occupies the deepest part (Spedding & Evans, 2002; Swift et al., 2006). Nonetheless, it seems plausible that the vertical velocity distribution within an overdeepening could be very different compared to ice-flow across non-overdeepened beds.

Another uncertain process is the role of overdeepening in trapping sediment and therefore promoting thick subglacial sediment layers that enable enhanced rates of till-deformation and glacier flow, and may even initiate fast ice flow on ice-sheet scales. The key mechanisms here are the suppression of efficient subglacial drainage, which should encourage high basal water pressures, the deposition of sediment from fluvial transport, and the retention and transport of sediment in basal sediment layers (Alley et al., 1999, 2003a,b); these factors could encourage a switch to faster ice flow immediately downglacier of an overdeepening (e.g. Peters et al., 2006), especially because faster ice flow should discourage the re-formation of hydraulically efficient channels that would lower basal water pressure and erode till thickness. Further, although entrainment and transport of sediment by supercooling may dominate over till layer accumulation within the overdeepening, the subsequent melt-out of such facies is likely to support and enhance till continuity beneath fast-flowing ice (Alley et al., 2003b). Thick sequences of debris-rich supercool-facies basal ice may also enhance the deformability of basal ice layers (Alley et al., 1999).

Despite some processes remaining speculative, the evidence available indicates that overdeepenings have significant influence over patterns and processes of ice motion. Whether or not large volumes of surface melt reach the bed of overdeepened areas, the higher basal water pressures within overdeepenings imply that basal sliding and/or till deformation should be more efficient than for other areas of the bed. This is supported by field observations at Storglaciären (e.g. Hooke et al., 1989), and implies that the ice surface profile above overdeepenings should be shallower than that above non-overdeepened areas. In the extreme example of a subglacial lake being present, complete ice-bed separation allows the surface slope to reduce to near zero (e.g. Figure 4b), providing the lake is of sufficient size (e.g. Pattyn et al., 2004). The presence of 'true' subglacial lakes in valley or outlet glacier contexts has been observed rarely (e.g. Scambos et al., 2011), but 'ponding' of a kind must occur diurnally within overdeepenings when surface melt ceases and the hydraulic gradient diminishes to zero, such that ice does not float but remains 'flooded' to an elevation determined by the overdeepening 'lip'. Further, the thickness and extent of basal sediment layers will increase as the limits of overdeepening are reached, and it is probable that overdeepening ultimately produces a transition from hard bedrock to a predominantly soft-sediment bed.

The principal outcome of overdeepening formation would therefore appear to be the substantial reduction of basal drag, resulting in greater efficiency of ice flow, and, for mature overdeepenings, the lowering of ice surface slope, which should encourage further sedimentation and even more efficient ice flow. Nevertheless, many aspects of the influence of overdeepenings on ice motion require further study. Little is known about the vertical velocity distribution within overdeepenings, including the potential for flow separation, flow stagnation with depth, or tectonic processes such as englacial thrusting. The hydrology of overdeepenings is also poorly understood, and as yet we know little about the influence of overdeepenings on sliding rates and how they might vary spatially (e.g. between overdeepened and non-overdeepened beds) and temporally (e.g. seasonally and diurnally). Above all, the net influence of overdeepenings on ice motion processes of all kinds is poorly unknown, and high-resolution studies of horizontal and vertical velocity and strain across overdeepenings, integrated with hydrological studies and assessment of bed geometry and composition, are required.

6.5. Do overdeepenings have a stabilising or destabilising effect on ice dynamics?

Increasingly, the (non-overdeepened) alpine valley glacier model of coupled seasonal evolution of glacier hydrology and motion is being invoked to explain the dynamics of ice sheets and their outlets (e.g. Zwally et al., 2002; Van de Wal et al., 2008; Sundal et al., 2011). However, this model is incomplete in terms of its understanding of hydrology and motion in areas of the glacier bed that are overdeepened: a situation that appears to be extremely common for present and former ice masses at all scales. Although understanding is indeed limited, it is clear that overdeepenings have both stabilising and destabilising effects on ice motion, and it is pertinent to consider whether their presence might amplify or even mitigate ice dynamic responses to climatic change. In doing so, it is important to consider the temporal and spatial influence and significance of such effects and, because overdeepening may be isolated, whether these effects are dampened or amplified by processes that operate in non-overdeepened areas of the bed.

An obvious mitigating factor is the tendency for high glacial sediment fluxes to build moraines and sediment shoals that stabilise the terminal regions of ice sheets and glaciers, which can greatly reduce ice-mass sensitivity to climatic and/or sea level change (e.g. Alley et al., 2007; Anandakrishnan et al., 2007; Koppes et al., 2010). Back-stresses imposed by ice flowing along or terminating on adverse bed slopes should also have a stabilising influence on ice motion. These effects are particularly important where glaciers terminate in water because the moraine or shoal influences significantly the rate and pattern of calving by controlling back-stress and water depth across the glacier terminus. Further, the self-stabilising potential of moraine building is important because observations (e.g. Howat et al., 2008; Meier and Post, 1987) and modelling (e.g. Schoof, 2007; Oerlemans and Nick, 2005, 2006; Nick et al., 2009; Oerlemans et al., 2010) demonstrate the fundamental instability of ice grounded on an adverse slope. A great deal of work has observed and described (e.g. Hunter et al., 1996; Anandakrishnan et al., 2007; Koppes et al., 2010) and modelled (e.g. Alley, 1991; Oerlemans and Nick, 2006; Alley et al., 2007) the effects of a sediment shoal on glacier dynamics. There has been little consideration, however, of how moraine-building potential might be enhanced by the overdeepening itself, for example by the suppression of efficient drainage and formation of debris-rich supercool-facies basal ice. It may therefore be difficult to predict which glaciers transfer sediment sufficient to build large stabilising moraines, and to what extent the effect of building a moraine shoal might negate or inhibit unstable grounding-line retreat.

For ice that is grounded, the most important implications of overdeepening for ice motion are likely to be caused by hydrological factors. Foremost, the tendency for overdeepenings to trap sediment and suppress hydraulically efficient drainage means that they should function in effect as ‘slippery spots’ where ice motion is enhanced by sliding and till-layer deformation. However, it is unclear to what extent this effect is mitigated by other processes, such as longitudinal stress coupling with non-overdeepened areas (Hooke et al., 1992; Jansson, 1995, 1996) and the additional back-stress that is imposed by flow against the adverse slope. Further, the possibility that runoff bypasses overdeepenings through englacial or lateral pathways indicates that overdeepened areas may be ‘immune’ to strong perturbations in basal water pressure that are required for efficient sliding and till deformation, such that during spring and summer they remain areas of relatively slow ice flow. Nevertheless, the persistence of basal water pressures in overdeepenings that are near overburden implies that even a small fraction of melt reaching the bed might be sufficient to produce strong ice motion effects. Even more importantly, the likelihood that the presence of an overdeepening restricts the efficiency of the subglacial drainage system more widely, especially by limiting the growth of efficient channels upglacier of the overdeepened area, implies that ice motion effects will be wider and more persistent than otherwise anticipated.

Actual observations indicate that large areas of the Greenland ice sheet do indeed show a seasonal increase in ice velocity associated with the onset of melt (e.g. Zwally et al., 2002; Van de Wal et al., 2008) that is followed by a dramatic decrease in velocity that implies the widespread evolution of efficient channelized drainage system (e.g. Howat et al., 2010). This suggests that the challenges presented by the lack of understanding of the nature of ice and water flow through overdeepenings do not as yet present a serious challenge to the dominant ‘alpine valley glacier’ model. Nevertheless, even slight delays or constraints on the wider evolution of efficient subglacial drainage, perhaps as a result of supercooling in an overdeepening near to the terminus, may be responsible for higher and more sustained periods of fast ice motion than would otherwise be the case. Further, the mechanisms underlying changes in ice motion may produce observations that agree with model expectations but, if they are not in fact accurately represented in the model, our ability to understand and predict future changes in ice mass behaviour or response may be subtly undermined. Notably, the influence of ice surface slope and ELA position on the processes that operate within overdeepenings (e.g., Figure 8e), and the existence of multiple overdeepenings beneath many outlet glaciers, indicate that glacier retreat could result in sharply non-linear changes in glacier behaviour and response.

It is clearly unsatisfactory that our understanding of hydrology and ice motion where glacier beds are overdeepened is both incomplete and limited largely to just a few valley glacier situations, and whether processes observed at valley glaciers are applicable at outlet glacier and ice sheet scales is an important issue for future research. Nonetheless, the information available indicates that overdeepenings influence glacier dynamics and stability greatly, and hence we should seek to integrate detailed bed topography and accurate representations of glacier hydrology and sediment transport into glacier and ice sheet models. Bed topography data in particular are important for predicting the response of tidewater glaciers to future sea level and climate change as a result of overdeepening-enhanced climate sensitivity, and sediment transport processes are important to understand the potential stabilising influence of lacustrine and marine sediment shoals.

6.6. Is there a landform-sediment signature that is unique to overdeepened glacial systems?

Tests of ice sheet models that employ reconstructions of palaeo-ice mass extent and dynamics are a key tool used to understand past and future climatic and environmental change. Recognition of overdeepening-related processes in the landscape, many of which have ice-dynamic implications, is therefore an essential component of such reconstructions. Unfortunately, the effectiveness of overdeepenings as depositional sinks, especially during deglaciation (e.g. Bennett et al., 2010), means that instances of overdeepening may be overlooked. This is especially likely where overdeepening has occurred in glacial or pre-glacial sediment and where this sediment is of a similar nature to the post-overdeepening or post-glacial sediment-fill. Key, therefore, is whether evidence of overdeepening-related processes, ideally ones that reflect the depth and extent of overdeepening, can be found in sediments and landforms beyond the immediate area of overdeepening.

A recent study at Kvíárjökull (Bennett et al., 2010) highlights the problem of landform preservation (or lack thereof) for glacial process reconstruction, especially where an enclosed foreland provided by an overdeepening encourages aggradation of fluvial outwash, fluvial reworking of glacially deposited sediments and landforms, and deposition of sediment on top of decaying glacier ice. Nevertheless, many instances of overdeepening, including at Kvíárjökull, appear to be associated with unusually large moraine complexes that seem to reflect the enhanced moraine-building potential that is produced by the suppression of efficient subglacial drainage and the presence of supercooling (e.g. Spedding and Evans, 2002), but also the stabilising influence of the adverse slope on the location of the glacier terminus. The size of such moraine systems means that their form and structure is difficult to investigate; however, observations at Kvíárjökull (Spedding and Evans, 2002) and Svínafellsjökull (Swift et al., in prep) indicate that material is accreted largely on the adverse slope in a cyclical process comprising the stacking of thick basal till layers during periods of ice advance followed by melt-out of thick basal ice sequences and englacial debris septa during subsequent retreat.

Other than the likelihood that overdeepening encourages large moraine complexes, it appears that no sedimentary landforms have yet been identified that are specific to overdeepening-related processes. Evans (2009) has highlighted the potential role of supercooling and thrusting in producing controlled-moraine, but the hummocky topography produced by this processes is not distinct and has little potential for preservation (see also Bennett et al., 2010). Similarly, the melt-out-till-dominated landform-sediment assemblage described by Larson et al. (2006) does not appear to be unique, and

Bennett et al. (2010) have remarked that tills are uncommon as a result of much sediment being elevated into englacial or supraglacial positions and being reworked subsequently by fluvial processes. Nevertheless, recognition of supercooling as a mechanism for producing metres-thick sequences of debris-rich basal ice has renewed interest in a melt-out origin for Quaternary till sequences (Evans et al. 2006), the thickness of which is difficult to explain using traditional mechanisms of basal ice formation (e.g. Paul and Eyles, 1990; Hart, 1998) without appealing to thickening as a result of subglacial sediment deformation (e.g. Piotrowski et al., 2001). It could be argued also that overdeepening might help to explain the presence of thick deformation tills owing to the tendency of overdeepening to encourage glacial, rather than fluvial, sediment transport.

Poor preservation potential of landforms and lack of distinct landform features indicates that features diagnostic of overdeepening may be restricted to the characteristics of the sediment itself, notably size and form. Importantly, sediment in till, moraine, basal ice and englacial debris septa at Kvíárjökull and Svínafellsjökull, whilst being structurally distinct according to the process of entrainment and/or deposition, is commonly comprised of sediment characteristic of both fluvial and glacial transport pathways sourced from lag deposits within the overdeepening. Further, as a result of undergoing glacial transport prior to deposition, fluvial material is never sorted, consists of a very wide range of sizes, and is commonly striated; only in melt-out from supercool basal ice facies, which is presumably formed in larger channels close to the terminus of the overdeepening, does fluvial debris display a degree of sorting and no evidence of striae. Dominance of silt in melt-out has been argued to be diagnostic of supercooling (Larsen et al., 2006), but Cook et al. (2011, 2012) have argued that silt-dominance may be a wider signature of overdeepening rather than of supercooling as a result of the likelihood that overdeepenings act as sinks for fine sediment during periods of glacial retreat. Indeed, laboratory experiments undertaken by Cook et al. (in press) demonstrated that sandy textures were more likely to be entrained as supercooled water freezes, rather than silt-sized sediments.

Further research is required to evaluate so-called diagnostic signatures of overdeepenings and the processes that operate within them, and to assess whether such signatures have been preserved in the geological record of past glaciations. Diagnostic evidence of supercooling may be able to reveal overdeepening depth and extent beneath former northern hemisphere ice sheets, and may therefore provide valuable information on former overdeepening distribution and ice sheet dynamics.

6.7. What is the context of subglacial lake and sedimentary environments?

Isolated from external change, subglacial lakes may provide important refugia for pre-glacial life (e.g. Priscu et al., 1999; Anesio and Laybourn-Parry, 2012) and their sediments may host unrivalled palaeoenvironmental archives. At least some subglacial lakes, such as Lake Ellsworth (Figure 4d), appear to owe their existence to the overdeepening of pre-glacial valleys, which implies that they may not be viable locations for pre-Quaternary sediments or refugia (cf. Livingstone et al., submitted). In addition, the origin of such lakes is ambiguous given that overdeepening is not possible without contact between ice and bedrock: ice that is insufficiently thick to maintain contact with the bed, or an ice-surface slope that is insufficiently steep to drive water up the adverse slope, will allow water and sediment to pond, and bedrock erosion will cease. This indicates that such lakes occupy overdeepenings eroded during ice sheet advance, when the surface slope of ice immediately above the overdeepening would have been steeper, and were formed as ice advanced beyond the overdeepening and hence the ice surface slope above shallowed. In this way, an overdeepening might be progressively deepened during successive glacial advances, and a lake might re-form at each glacial maximum. However, the lake will drain catastrophically when ice retreats, and, because lake stability is dependent on the surface slope of overlying ice (assuming that the lake is not filling; Evatt et al., 2006), it may drain even if retreat is relatively minor (see Jordan et al., 2010a).

The instability of subglacial lakes of glacial origin contrasts with the potential stability of lakes of a tectonic origin. The captured ice shelf model (Erlingsson, 1994, 2006), in which advancing ice floats across a lake occupying a pre-glacial basin, provides a mechanism for capturing the pre-glacial contents of the lake beneath an ice mass. Grounding and freeze-on of ice at the far side of the basin will cause ice above the basin to thicken (Alley et al., 2006), leading to pressurisation of the water in the basin and catastrophic drainage. However, not all the lake water will drain provided that the ice mass is sufficiently thin to remain buoyant or the ratio of the ice surface slope to the basin adverse slope exceeds the 'ponding' threshold (Paterson, 1994; Clarke, 2005). Further, the influence of thermal regime on glacier erosional potential indicates a glaciological control on the potential for subglacial lake formation. Specifically, at warm-based ice masses where abundant surface melt reaches the bed, and therefore subglacial erosion rates are relatively high (e.g. Cowton et al., 2012), glacial erosion will modify pre-glacial valleys and tectonic basins such that adverse slopes will tend towards the supercooling threshold and few basins will host lakes. In contrast, where the bed is cold or receives little surface melt, erosion will be less efficient and suitably deep pre-glacial basins are likely to host lakes for many millions of years.

6.8. What is the role of overdeepening in landscape and ice sheet evolution?

The paucity of work on the role of overdeepening in landscape evolution means that any discussion of this topic is necessarily highly speculative. Overdeepening contributes importantly to the production of both relief and 'missing mass' and may be strongly localised in space and time; consequently, it is likely to influence mountain-scale patterns of isostatic uplift, topographic evolution, and crustal deformation (Herman et al., 2011). However, the location and volume of overdeepening at different stages in the development of such landscapes are important details that are still poorly known. There is undoubtedly a tectonic context to overdeepening: high rates of tectonic uplift mean overdeepening is less likely because strong uplift will maintain steeper valley gradients and higher rates of above-glacier erosion, meaning that subglacial channels will not have the unsatisfied sediment transport capacity needed to evacuate basal debris and sustain focused subglacial erosion. It follows that 'mature' overdeepenings are most likely to be present in tectonically passive regions and that overdeepening in active settings is unlikely, except where hydrology and erosion are very efficient. Further, the mechanisms that link overdeepening to hill-slope processes acting on adjacent interfluves and peaks are unclear, and are complicated by the unknown relative contributions of glacial versus non-glacial landscape response. For example, inter- and post-glacial sedimentary in-filling of overdeepenings will counter overdeepening-induced isostatic uplift (e.g. Champagnac et al., 2009).

Another key unknown in mountain landscape evolution is the relative efficacy of cirque erosion versus overdeepening, and their respective limits. Many factors support glacial valley deepening and overdeepening being the most efficacious. Cirque glaciers are typically thin, have low balance gradients, and do not develop efficient subglacial drainage, whereas valley glaciers focus ice flow, ice is thicker and faster flowing, and, crucially, a larger ablation area enables the development of efficient subglacial drainage (Figure 19). Hence, it is not entirely clear why cirque erosion alone should control the elevation of mountain ranges (cf. Anders et al., 2010). Overdeepening may ultimately be less important because of the presence of stabilising feedbacks: notably, headward erosion should be limited because of height-mass balance feedbacks (e.g. Herman et al., 2011); depth is also limited by ice flux, because erosion of a sufficiently deep overdeepening will cause ice to float; and finally, ice-surface steepness limits on the steepness of the adverse slope. Thus, the geometry of an ice mass and its bed should evolve toward an equilibrium configuration (Figure 20a) that will infill with sediment under glacier thinning or retreat (Figure 20b); continued overdeepening is impossible unless lowering of the ELA causes ice to thicken and steepen (Figure 20c) or fluvial erosion lowers the bedrock threshold at the glacier terminus (Alley et al., 2003b), assuming further erosion does not introduce height-mass balance feedback (Figure 20d). From this perspective, overdeepening can be viewed as a passive phenomenon, the depth of which is dependent on non-glacial factors; however, cirque erosion should be subject to similar controls, and the nature of stabilising factors that should limit overdeepening are still debated (e.g. Creyts and Clarke, 2010).

Fundamental limits on the depth of overdeepening further support the idea that, at the onset of large-scale glaciation, erosion and sediment production rates should demonstrate an initial peak that, as

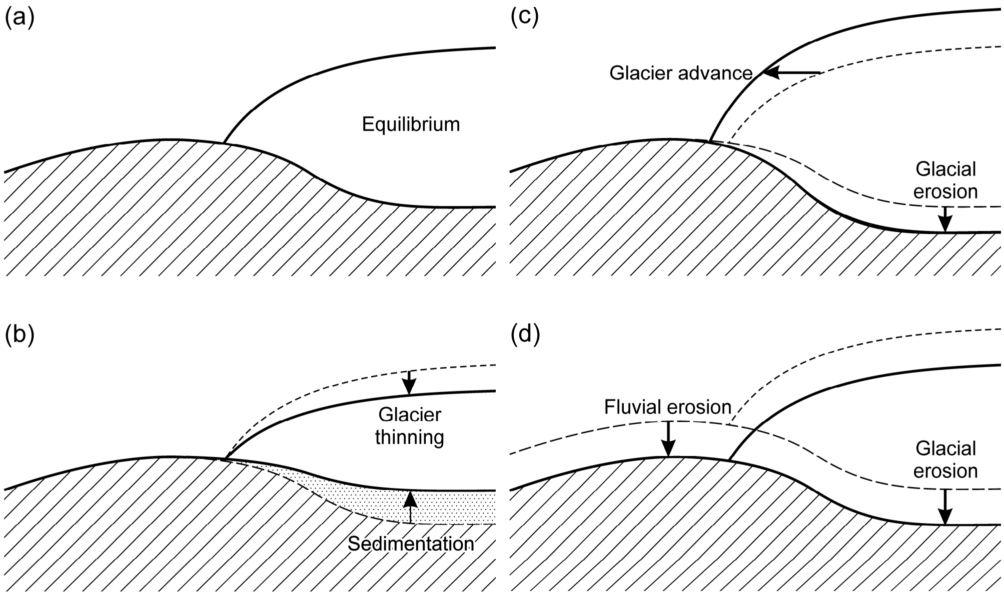

Figure 21. (a) Ice mass and bed geometry in equilibrium: erosion and sedimentation are in balance (i.e. there is no net erosion of the bed) because ice within the overdeepening is nearing flotation and the gradient of the adverse slope is such that the ice-surface slope to bed slope ratio is at the threshold for glaciohydraulic supercooling. (b) Thinning of the glacier shown in (a) would cause ice to float or, more likely, the deposition of sediment within the overdeepening. (c) Lowering of the ELA would cause the glacier shown in (a) to thicken and hence the ice-surface slope to bed slope ratio to reduce, producing renewed overdeepening. (d) Renewed overdeepening can also be produced by fluvial erosion of the proglacial moraine or bedrock threshold, which again causes the ice-surface slope to bed slope ratio to diminish.

topography switches from a 'fluvial' to a 'glacial' landscape, declines to some 'background' level (e.g. Charreau et al., 2011). This pattern will be reinforced strongly by the tendency of overdeepening to suppress the efficiency of the subglacial drainage system, which strongly limits glacial sediment transport and therefore erosion, but also encourages sediment to be entrained by glacial transport, causing sediment to be deposited in proximal locations (i.e. moraines) as opposed to being transported distally by proglacial streams (Swift et al., 2002). Further, overdeepening is likely to play a significant role in ice-sheet evolution over successive glacial cycles, both through feedbacks between valley form and ice dynamics (e.g. Kessler et al., 2008) and erosion of elevation (Kaplan et al., 2009), both of which are likely to influence the extent and behaviour of ice masses. These factors are even more important for ice sheets that rest below sea level or where outlet glaciers terminate in marine or lacustrine environments, since the development of a negative slope behind the ice margin has a destabilising affect that is highly significant for ice mass behaviour (e.g. Schoof, 2007).

Ultimately, the overdeepening of outlet glacier and ice sheet beds increases the efficiency of ice discharge and significantly enhances the sensitivity of ice masses to climate; marine ice masses in particular, such as the present WAIS, become vulnerable to catastrophic collapse. Nevertheless, models of landscape evolution typically do not include the coupled ice-hydrology-sediment processes that are necessary to simulate patterns of erosion and sediment transport as a result of overdeepening (e.g. Herman et al., 2011), and models of long-term ice sheet behaviour typically have not considered the effects of coupled evolution of ice dynamics, hydrology, and bed topography (e.g. Pollard and DeConto 2009). Elucidation of the links between landscape and ice sheet evolution require more accurate studies of overdeepening distribution and depth and its incorporation in landscape and ice sheet models. Currently, such efforts are limited by lack of large-scale quantitative studies of overdeepening distribution and geometry as well as lack of understanding of the erosional processes and feedbacks that produce and limit such focussed glacial erosion.

7. CONCLUSIONS – KEY QUESTIONS

Subglacial basins and especially overdeepenings are fundamental component of glacier systems and landscapes. Glacier beds tend towards overdeepened geometries because ice and water are able to transport sediment along adverse subglacial slopes, enabling the erosion of subglacial bedrock or sediment to depths significantly below fluviatile base level or sea level. Terminal overdeepening may be enhanced the construction of moraines or sediment shoals. Whilst deglaciated landscapes and geophysical investigation demonstrate the ubiquity of subglacial basins and overdeepenings, the origin and evolution of such basins, and their influence on the processes of glacier hydrology, motion, and landscape evolution, remain poorly understood.

From our synthesis of research concerning the origin and significance of subglacial basins, we advance the following conclusions:

1. Overdeepening of the glacier bed is extremely common in all glaciated environments where warm-bed conditions exist or have existed. Overdeepenings are typically found beneath present/former ablation zones where subglacial water is abundant, but overdeepening of the bed can also be highly localised: isolated overdeepenings are often found in cirques, at valley confluences, and in terminal zones (e.g. beneath former piedmont-style lobes). Overdeepenings range from hundreds to thousands of metres in length and from a few tens of metres to many hundreds of metres in depth, and their key dimensions appear to scale with ice flux and the length of glacial occupancy.

2. An overdeepened bed is an equilibrium form that all warm-based ice masses will evolve towards as a result of the capacity for ice and water to transport sediment along adverse subglacial slopes. Overdeepening geometry is a product of the local balance along the glacier bed profile between sediment supply, which prevents subglacial erosion, and the capacity of ice and subglacial water to move sediment downglacier. Because subglacial channels have large unsatisfied transport capacity that increases towards the glacier terminus, sediment transport by channels dominates, but transport capacity decreases as channels flow upslope and more energy has to be used for channel melting. Thus, the bed deepens until flow capacity is matched by the rate of sediment supply and net erosion is zero.

3. Although feedbacks between hydrology and sediment transport indicate that glacier beds should grade toward a uniform overdeepened profile, overdeepening may be focussed strongly by ice-water-sediment feedbacks in which water and sediment transport play key roles. The location of overdeepening is also influenced by the focussing of ice flow and glacial, interglacial and intraglacial changes in ELA, which control glacier and ablation area extent. Further, overdeepening may be initiated by lithological changes or pre-glacial bed irregularities that stimulate ice-erosion feedbacks. Glacial occupancy and tectonic uplift will influence the development of 'mature' overdeepened bed profiles.

4. Overdeepening is self-limiting as a result of height-mass balance feedbacks that reduce ice flux and cause thinning and retreat of ice occupying the overdeepened basin. Thereafter, overdeepening is limited by the rate of lowering of the ELA and/or the basin lip. The geometry of the overdeepening is controlled by height-mass balance feedback and the efficiency of water and sediment evacuation from the glacier system: adverse slopes greater than ~1.2 times the ice surface slope promote the freezing of water in subglacial channels and the accumulation of sediment in a basal till layer; adverse slopes in excess of ~11 times the ice surface slope cause 'ponding' of water in a subglacial lake.

5. Overdeepening restricts the transmissivity of subglacial drainage, which raises basal water pressures and may enhances the likelihood that subglacial water will follow englacial pathways across the overdeepening. Reduced transmissivity reflects a reduction in the energy available for melting and, for adverse slopes in excess of ~1.2 times the ice surface slope, the

freezing of water in subglacial channels (i.e. glaciohydraulic supercooling). These processes may cause a switch from channelised to distributed drainage within the overdeepening.

6. High basal water pressures within subglacial basins and overdeepenings enhance local ice motion and its sensitivity to diurnal and seasonal variations in melt. Overdeepening is also likely to influence ice motion by encouraging the retention of sediment at the ice-bed interface in the form of till and basal ice layers. Further, adverse slopes provide backstress that resists ice flow and can encourage ice deformation on glacier-thickness scales.

7. The likelihood that overdeepenings focus subglacial water flow, restrict the efficiency of subglacial drainage, and produce layers of deformable till and soft basal ice, supports observations of ice stream onset zones above bed overdeepenings.

8. Overdeepening increases ice mass sensitivity to climate because it causes a branching of equilibrium states that allows even moderate warming to cause catastrophic ice loss (cf Oerlemans et al. 2011; section 5.1.3). This is particularly important in lacustrine and marine environments, where retreat and floatation results in loss of backstress and rapid calving. In the context of the WAIS, retreat also exposes an increasingly large area of ice to warm ocean water, promoting strong positive feedbacks. The building of a moraine or sediment shoal may enhance the stability of ice masses that terminate on adverse slopes, but ultimately this increases the potential for catastrophic ice loss if warming is sustained.

9. New lakes that form in emergent overdeepenings in glaciated regions represent a significant hazard. There is evidence that glacier readjustments to anthropogenic warming and the end of the Little Ice Age are resulting in the formation of new lakes at an increasing rate. The hazard potential, especially for bedrock-confined lakes, is enhanced significantly by the tendency for the glacier terminus position to re-stabilise at the overdeepening head.

10. Overdeepening will impact rates, patterns and processes of erosion and sediment transfer in glacial systems and the proportion of sediment in competing transport pathways. Notably, overdeepening will suppress glacial erosion and sediment transport by subglacial drainage and will enhance the proportion of sediment in subglacial and englacial transport. These processes should encourage the formation of large moraines and produce distinctive landform-sediment associations and characteristics.

11. Subglacial basins and overdeepenings are important sediment sinks and archives of past environmental change that exist in both formerly glaciated landscapes and beneath present ice sheets. The preservation of pre-glacial sediment archives is most likely where overdeepenings have a non-glacial, tectonic origin, and are therefore sufficiently deep for ice sheet advance to capture pre-glacial lake waters and sediments.

12. Overdeepenings contribute significantly to relief production and missing-mass in glaciated landscapes. They are therefore likely to play an important role in linking tectonics and climate and in mountain-scale patterns of isostatic uplift, topographic evolution, and crustal deformation. The inter- and post-glacial filling of overdeepenings with sediment is also likely to play an important role in such linkages.

13. Subglacial basins increase the efficiency of ice flow by substantially reducing basal drag. On long time scales, the development of an equilibrium-geometry overdeepened bed is therefore likely to exert substantial influence on the patterns and dynamics of ice flow. Further, for marine outlet glaciers and ice sheets, overdeepening of the bed substantially reduces ice extent over successive glacial cycles and markedly increases ice mass sensitivity to climate.

Consequently, glaciers with overdeepened beds should exhibit substantially different dynamic behaviour to those with non-overdeepened beds, and the focussed excavation of bedrock to many hundreds of metres below fluviatile base level is likely to play a central role in the evolution of glaciated landscapes and the ice masses that occupy them. As such, overdeepening and related processes need to be integrated accurately into numerical models into numerical glacial and landscape evolution models. However, research that has addressed explicitly the processes and implications of overdeepening is extremely limited, and many questions remain.

We therefore highlight some key questions that future research should address:

1. Why do glacier-beds tend towards overdeepened geometries, and how, where, and when does this occur? Specifically, we need greater understanding of the processes that control and limit focussed glacier-bed erosion, the timescales for overdeepening formation, and the occurrence of conditions favourable to bed overdeepening. It is also unclear whether all overdeepenings reflect the same erosional process(es). Further insights into these issues may be gained through glacier erosion models that incorporate ice-water-sediment feedbacks.

2. How does overdeepening affect the hydrology of glaciers and ice sheets? Specifically, we need greater understanding of the style and transmissivity of subglacial drainage within overdeepenings; whether or not efficient subglacial channels are present and the conditions under which they might form; whether or not water may also follow lateral or englacial flowpaths and how these pathways might develop; and, the proportions of water in englacial and subglacial drainage pathways and whether or not the volume of water in these different pathways evolves seasonally in response to changes in surface runoff.

3. How pervasive is glaciohydraulic supercooling and what is its significance? Specifically, we need to understand the spatial and temporal controls on supercooling; whether or not model predictions and field evidence agree; the extent to which the presence of alternative englacial flowpaths affects the spatial and temporal distribution of supercooling; and the rates of basal ice formation by supercooling and its implications for basal ice rheology and till continuity.

4. How does overdeepening affect glacier and ice sheet motion? Specifically, we require more information on basal water pressure fluctuations and their implications for basal sliding; the significance of a till layer at the base of an overdeepening; the effect of different subglacial drainage system morphologies and their seasonal evolution; the implications of glacier flow within overdeepenings for neighbouring non-overdeepened areas; and the nature of, and conditions required for, deformation on ice-thickness scales.

5. Do overdeepenings have a stabilising/destabilising effect on ice dynamics? Specifically, it is unclear whether the presence of an overdeepening enhances net ice flux, or whether non-overdeepened areas 'restrain' flow; whether seasonal deceleration of ice, induced by subglacial channel or englacial conduit formation, will occur when overdeepenings are present; whether overdeepenings are implicated in fast ice flow and the nature of the process(es) responsible; and whether sediment-transport and moraine building can stabilise outlet glaciers against present warming trends.

6. Is there a landform-sediment signature that is unique to overdeepened glacial systems? Specifically, we know very little about whether or not there are unique landform-sediment associations associated with overdeepenings, or whether there are uniquely characteristic proportions of sediment in competing transport pathways. It is unclear to what extent such sedimentary characteristics and stratigraphy are diagnostic of overdeepenings or the processes that operate within them.

7. What is the context of subglacial lake and sedimentary environments? Specifically, it is unclear whether 'true' glacially overdeepened basins are secure sites for pre-glacial refugia and sedimentary archives. Further, the preservation potential of waters and sediments within overridden basins of different origin is also poorly understood.

8. What is the role of overdeepening in landscape and ice sheet evolution? Specifically, we do not yet understand the controls on mountain-scale patterns of overdeepening and inter- and post-glacial sedimentation; the significance of overdeepenings for missing-mass and associated crustal, isostatic and landscape response; and the extent to which overdeepening affects ice sheet inception and long-term evolution, dynamics and stability.

In the context of current warming trends, concern must focus on the hydrology of overdeepenings and their effect on valley glacier, ice sheet and outlet glacier dynamics, as well as grounding-line stability and the retreat processes of lacustrine and marine outlet glaciers and ice sheets. For the Greenland ice sheet and the WAIS especially, overdeepening-enhanced climate sensitivity is likely to dominate future ice sheet response and the glacial contribution to global sea level change.

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Table 1: Selected examples of subglacial basins beneath modern day ice masses from a range of locations and environments.

Site	Study	Ice mass type	Glacier Length (km)	Glacier area (km^2)	Basin type and location	Basin length (km)	Basin area (km^2)	Basin depth (m)	Maximum ice depth in basin (m)
Aletschgletscher, Switzerland	Hock et al. (1999)	Temperate valley glacier	24.7	86.7	Overdeepening at junction of two tributaries	2.0	~2	590	890
Gornergletscher, Switzerland	Iken et al. (1996)	Temperate valley glacier	13.5	59.7	Overdeepening at junction of two tributaries	3.0	3.6	~100	350
Glacier d'Argentiere, France	Hantz and Lliboutry (1983)	Temperate valley glacier	9	14	Overdeepening in mid-valley bounded by riegel	0.9	0.3	62	250
Kvíárjökull, Iceland	Spedding (2000)	Temperate valley glacier	13.5	25.0	Terminal overdeepening	>1.6	-	120	175
Matanuska, Alaska	Lawson et al. (1998)	Temperate valley glacier	45.0	~200	Terminal overdeepening	0.4	0.1	25	25
South Cascade, Washington State	Fountain (1994)	Temperate valley glacier	3.0	3.0	Overdeepening approximately 0.7 km from terminus	0.7	~0.4	55	195
Storglaciären, Sweden	Hooke and Pohjola (1994)	Polythermal valley glacier	3.2	3.0	Overdeepening approximately half way along glacier length	1.2	0.5	40	245
Aldousbreen, Austfonna, Svalbard	Dowdeswell (1986)	Ice cap outlet glacier (polythermal?)	20	150	Terminal overdeepening	3	-	30	160
Taylor Glacier, Dry Valleys, Antarctica	Hubbard et al. (2004)	Polar valley glacier (polythermal)	100	510	Overdeepening 3 to 6 km upglacier from terminus	8.5	12	100	470
Helheim Glacier, Greenland	Nick et al. (2009)	Ice sheet outlet glacier	>55	515	Glacially deepened rift beneath tributary glacier	7	-	200	850
Gamburtsev Mountains, Antarctica	Bo et al. (2009)	Ice sheet interior	-	-	Pre-existing river valleys beneath ice sheet interior that have become glacially overdeepened	16.2	64	400	-
Vostok	Siegert (2005)	Ice sheet interior	1515 (Totten)	60000	Pre-existing tectonic structure along a geological boundary. Possibly glacially overdeepened.	24	130	510	4000
Astrolabe	Siegert (2005)	Ice sheet interior	200	4000	Glacially overdeepened rift valley beneath ice sheet interior	350	1500	2000	4776
Lambert	Siegert (2005)	Ice sheet interior	400	1.6×10^6	Pre-existing graben that has become glacially overdeepened	500	2000	2500	3500
Byrd	Siegert (2005)	Ice sheet interior	136	1.1×10^6	Geologically controlled fjord that has become glacially overdeepened	350	2500	1000	3500

Table 2: Implications of the diurnal cycle of discharge and surface-slope to bed-slope ratios for the pattern and magnitude of supercooling and the formation of 'supercool facies' basal ice (after Creyts and Clarke, 2010).

Hydraulic forcing	**Surface:bed slope ratio above the 'supercooling' threshold**	**Surface:bed slope ratio at the 'supercooling' threshold**	**Surface:bed slope ratio below the 'supercooling' threshold**	**Implications**
Constant (no diurnal cycle of discharge variation); it is assumed that a subglacial channel system is present	Some supercooling of water flowing in large channels occurs because melting enlarges channels and thereby reduces the hydraulic gradient below that of the ice surface slope. Basal water pressures remain well-below flotation.	Supercooling occurs in large channels and limits the growth and efficiency of a subglacial channel network. Ice accretion in channels causes water to distribute locally across the bed and therefore local basal water pressures ~~to~~ reach flotation. Ice does not accrete in the distributed system because these flowpaths conduct less water per unit wetted perimeter and hence viscous dissipation is higher than in the channel system.	Strong supercooling in large channels causes water to distribute widely across the bed and basal water pressures to reach flotation. Ice does not accrete in the distributed system because these flowpaths conduct less water per unit wetted perimeter and hence viscous dissipation is higher than in the channel system.	Supercooling occurs under all simulations and is confined to larger flowpaths (i.e. channels), but visible ice accretion is only likely for threshold and below-threshold simulations. For these simulations, net ice accretion is greater than under strong diurnal forcing because hydraulic gradients are lower and invariant, meaning there are no major changes in flow velocity that result in the melting of accreted ice.
As above, but imposing a strong diurnal cycle of discharge	Supercooling occurs in large channels during the evening and night-time when hydraulic gradients are low. Very strong melting during the morning and daytime when hydraulic gradients are high removes all of the accreted ice. Melting is nevertheless insufficient to transmit peak discharge, causing water to distribute locally across the bed and local basal water pressures to reach flotation.	Supercooling within large channels during the evening and night-time is sufficient for ice accretion to cause significant rates of channel closure and for water to distribute locally across the bed and thereby raise local basal water pressures to flotation. The reduced transmissivity of channels produces very high hydraulic gradients during the morning and daytime and therefore strong melting that removes the majority, if not all, of the accreted ice; in addition, water continues to distribute locally and therefore local basal water pressures continue to reach flotation. Water retreats to existing and incipient channels at night-time when discharge is very low.	Strong ice accretion occurs in channels during the evening and night-time but very strong melting during the morning and daytime, as a result of very high hydraulic gradients, melts a significant proportion of the accreted ice and maintains a rudimentary channel system of very low transmissivity. Low transmissivity of channels cause water to distribute across the bed very widely during the morning and daytime, and basal water pressures across the bed are at or near flotation almost constantly, except at night-time when discharge is very low.	Supercooling occurs under all simulations, but visible ice accretion is only likely for the below-threshold simulation. For threshold and below-threshold simulations, net ice accretion is lower than under constant forcing because hydraulic gradients are higher and are raised even further by channel closure, such that diurnally-high flow velocities cause melting of a significant proportion of the accreted ice. Because there is net ice-loss along the glacier bed for both above-threshold and threshold simulations, the ice-surface slope should flatten-out, leading to stronger supercooling, as opposed to stabilisation of the glacier-bed at a given supercooling threshold.