A review of topographic controls on moraine distribution

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Abstract

Ice-marginal moraines are often used to reconstruct the dimensions of former ice masses, which are then used as proxies for palaeoclimate. This approach relies on the assumption that the distribution of moraines in the modern landscape is an accurate reflection of former ice margin positions during climatically controlled periods of ice margin stability. However, the validity of this assumption is open to question, as a number of additional, nonclimatic factors are known to influence moraine distribution. This review considers the role played by topography in this process, with specific focus on moraine formation, preservation, and ease of identification (topoclimatic controls are not considered). Published literature indicates that the importance of topography in regulating moraine distribution varies spatially, temporally, and as a function of the ice mass type responsible for moraine deposition. In particular, in the case of ice sheets and ice caps (> 1000 km^2), one potentially important topographic control on where in a landscape moraines are deposited is erosional feedback, whereby subglacial erosion causes ice masses to become less extensive over successive glacial cycles. For the marine-terminating outlets of such ice masses, fjord geometry also exerts a strong control on where moraines are deposited, promoting their deposition in proximity to valley narrowings, bends, bifurcations, where basins are shallow, and/or in the vicinity of topographic bumps. Moraines formed at the margins of ice sheets and ice caps are likely to be large and readily identifiable in the modern landscape. In the case of icefields and valley glaciers (10–1000 km^2), erosional feedback may well play some role in regulating where moraines are deposited, but other factors, including variations in accumulation area topography and the propensity for moraines to form at topographic pinning points, are also likely to be important. This is particularly relevant where land-terminating glaciers extend into piedmont zones (unconfined plains, adjacent to mountain ranges) where large and readily identifiable moraines can be deposited. In the case of cirque glaciers (< 10 km^2), erosional feedback is less important, but factors such as topographic controls on the accumulation of redistributed snow and ice and the availability of surface debris, regulate glacier dimensions and thereby determine where moraines are deposited. In such cases, moraines are likely to be small and particularly susceptible to post-depositional modification, sometimes making them difficult to identify in the modern landscape. Based on this review, we suggest that, despite often being difficult to identify, quantify, and mitigate, topographic controls on moraine distribution should be explicitly considered when reconstructing the dimensions of palaeoglaciers and that moraines should be judiciously chosen before being used as indirect proxies for palaeoclimate (i.e., palaeoclimatic inferences should only be drawn from moraines when topographic controls on moraine distribution are considered insignificant).

1. Introduction

Glacial landforms are a fundamental source of information about the extent and dynamics of former glaciers and ice sheets (Dyke and Prest, 1987; Kleman and Borgström, 1996; Clark, 1997; Kleman et al., 1997, 2006). Many studies are undertaken with the express purpose of obtaining palaeoclimatic data from these landforms through the reconstruction of former ice-mass dimensions (e.g., Sutherland, 1984; Benn and Ballantyne, 2005). By far the most useful and widely used landforms for this purpose are ice-marginal moraines, which provide direct evidence of former ice margin positions (see Dyke and Prest, 1987; Svendsen et al., 2004). In providing this information, the modern distribution of ice-marginal moraines has effectively been used as an indirect proxy for palaeoclimate (e.g., Benn and Ballantyne, 2005; Ballantyne et al., 2007). However, the validity of this approach is open to question as a number of additional, nonclimatic factors are known to influence moraine formation (see Mercer, 1961; Funder, 1972, 1989; Punkari, 1980; Warren and Hulton, 1990; Warren, 1991; Kessler et al., 2006; Kaplan et al., 2009; Pratt-Sitaula et al., 2011; Anderson et al., 2012; Barr and Clark, 2012b), preservation (see Putkonen and O'Neal, 2006; Anderson et al., 2012; Kirkbride and Winkler, 2012), and ease of identification (see Barr and Clark, 2012b). Perhaps the most important of these factors is topography, which can lead the modern distribution of moraines to 'appear to carry palaeoclimatic significance which it does not have' (Warren, 1991, p. 14). Despite this importance, the role of topography in regulating the distribution of moraines is rarely explicitly addressed in palaeoglaciological reconstructions (i.e., moraines are often directly dated and used to infer former ice margin positions without consideration of landform origin or topographic context). The aim of this review is to address this shortcoming and to encourage critical discussion by focusing specifically on the role played by topography in regulating moraine formation, preservation, and ease of identification. This paper is primarily a review of published literature and is divided into four sections: (i) a background to ice-marginal moraines and their use in palaeoclimate reconstructions; (ii) a consideration of topographic controls on moraine formation; (iii) a consideration of topographic controls on moraine preservation and ease of identification; and (iv) an assessment of the implications that topographic controls on moraine distribution have for palaeoglacier and palaeoclimate reconstructions. This paper focuses on ice-marginal moraines, though much of the material also applies to other associated landforms such as controlled or hummocky moraines (see Lukas, 2005; Evans, 2009).

2. Background

2.1. Moraine properties

Ice-marginal moraines are ridge-like formations, which are typically classified according to their location as either end (deposited around glacier termini), lateral (deposited along glacier lateral margins), or latero-frontal (Fig. 1). They are formed through a number of processes, as supra-, en-, and subglacial debris is dumped at glacier margins (Eyles, 1983; Benn, 1992); proglacial debris is bulldozed during advance (see Boulton, 1986; Bennett, 2001); subglacial sediment is squeezed from beneath glacier margins (see Price, 1970); and bedrock and unconsolidated sediments are thrust into imbricate ridges during ice advance (see Evans and England, 1991; Hambrey and Huddart, 1995). Individual moraine ridges are often produced through a combination of these processes and can form subaerially and subaqueously (see Krzyszkowski and Zeliński, 2002; Ottesen and Dowdeswell, 2006). The internal structure and composition of ice-marginal moraines is largely determined by their mode of formation and is therefore highly variable. Some are almost entirely composed of glacial diamicton, whilst others contain a variety of glacigenic and glaciofluvial materials, often preserving sedimentary and/or deformation structures (see Boulton et al., 1999; Evans, 2009). A key factor in moraine formation is that debris must accumulate at glacier margins. This debris is typically transported englacially or supraglacially (though proglacial sediment can also be bulldozed), and the volume of material available for moraine formation therefore depends on ice velocity, the volume of debris within/upon a glacier, and the duration of ice margin stability (i.e., still-stand duration) (Andrews, 1972; Kirkbride and Winkler, 2012). For this reason, conditions that favour the formation of large moraines are often found at the margins of dynamic and erosive (temperate) glaciers which are able to entrain/accumulate large quantities of debris, but which also occupy stable positions for prolonged periods (Spedding and Evans, 2002; Swift et al., 2002; Cook and Swift, 2012). The largest ice-marginal moraines are 100s of kilometres long and 10s of kilometres wide (Fig. 2). Some have an altitudinal-range (relief) of hundreds of metres, whilst others show very little topographic expression (Barr and Clark, 2012a,b). Though margin stability is likely to encourage the development of large moraines, Anderson et al. (2014) demonstrated that 20-m-high examples can form in < 20 years; and in front of modern surge-type glaciers, the formation of similarly sized end moraines can occur in a matter of days (Benediktsson et al., 2008).

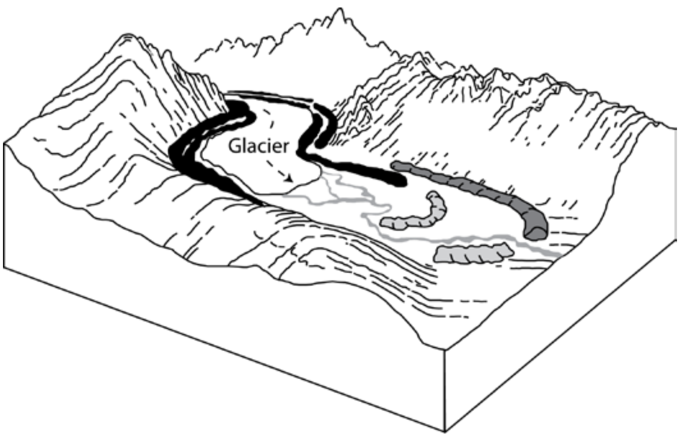

Fig. 1. Schematic illustration of end moraines (light grey), lateral moraines (black), and latero-frontal moraines (dark grey) formed by a valley glacier. The dashed arrow shows ice-flow direction (image modified from Huber, 1987).

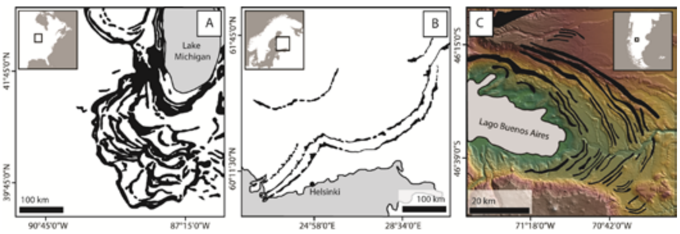

Fig. 2. Examples of extensive, arcuate end moraines. (A) Moraines marking the southern extent of the Lake Michigan lobe of the Laurentide Ice Sheet, between 22 and 13.5 ka (redrawn from Frye and Willman, 1973). (B) The Salpausselkä (I, II, III), North Karelia, Koitere, and Pielisjärvi end moraines, demarcating former ice sheet extent in southern Finland and western Russia (redrawn from Rainio, 1998). (C) Moraines demarcating the extent of the Lago Buenos Aires lobe of the former Patagonian Ice Sheet (redrawn from Glasser and Jansson, 2008).

2.2. *Moraines as indicators of palaeoclimate*

Ice-marginal moraines are ubiquitous in glaciated landscapes (e.g., Heyman et al., 2008; Lovell et al., 2011; Fredin et al., 2012), with some of the largest known examples deposited by the vast ice sheets that occupied North

America and northern Europe during the Last Glacial Maximum (LGM) (Figs. 2A-B). Some of the oldest moraine sequences on Earth date back over 1 Ma (Singer et al., 2004; Kaplan et al., 2009) (Fig. 2C); and a global population of offshore examples is significant, and expanding (e.g., Bradwell et al., 2008; Spagnolo and Clark, 2009; Winsborrow et al., 2010). Sedimentological and morphometric analyses of moraines provide information about former glacier flow dynamics, thermal regime, and debris content (amongst other factors) (see Boulton, 1986; Lukas, 2005; Evans, 2009), but moraines are most commonly used simply as indicators of former ice margin positions (e.g., Svendsen et al., 2004; Fredin et al., 2012). On the assumption that fluctuations in ice margins are driven by variations in climate (Oerlemans et al., 1998; Dyurgerov and Meier, 2000; Putnam et al., 2012), this approach potentially allows moraines to be utilised as proxies for palaeoclimate (e.g., Benn and Ballantyne, 2005; Ballantyne et al., 2007). This link between moraines and palaeoclimate is usually made in one of three ways: (i) moraine positions are used to infer variations in the areal extent of glaciers, and to provide a qualitative understanding of palaeoclimatic variation between periods (e.g., Lasalle and Elson, 1975). (ii) Moraines (and other landforms) are used to generate three-dimensional (3D) palaeoglacier reconstructions, from which steady-state palaeo equilibrium-line altitude (ELA) estimates can be derived (e.g., Benn and Ballantyne, 2005; Benn and Lukas, 2006; Rea and Evans, 2007; Finlayson et al., 2011). This approach is based on the assumption that the ELA can be linked to climate because it marks the point on a glacier where net annual accumulation and ablation are exactly equal (i.e., it is governed by variations in temperature and precipitation) (see Braithwaite et al., 2006; Hughes and Braithwaite, 2008; Golledge et al., 2010). (iii) Moraines are used to approximate palaeo ELAs (e.g., Leonard, 1989), based on simple measures such as the maximum elevation of lateral moraines (MELM), or a ratio between moraine altitudes and a specified point within a drainage basin (e.g., the toe-to-headwall altitude ratio method; THAR) (see Benn and Lehmkuhl, 2000; Benn and Evans, 2010). Each of the above approaches relies on the assumption that the distribution of moraines in the modern landscape is an accurate reflection of former ice margin positions during climatically-controlled periods of ice margin stability. However, the validity of this assumption varies both spatially and temporally, and moraine distribution is known to reflect a number of additional, nonclimatic controls, particularly topography.

3. Topographic controls on moraine formation

Topography has the capacity to regulate where and when moraines are deposited. This effectively represents a control on glacier dimensions, dynamics, and margin stability (i.e., where and when glacial still-stands occur). The role of topography as a regulator of modern ice-mass stability is a topic of ongoing research interest (e.g., Mercer, 1961; Hughes, 1987; Payne and Sugden, 1990; Kerr, 1993; Tomkin, 2003; Singer et al., 2004; Taylor et al., 2004; Kessler et al., 2006), particularly when considering marine-terminating outlets of the Greenland and Antarctic ice sheets (e.g., Bennett, 2003; Schoof, 2007; Jamieson et al., 2012, 2014; Carr et al., 2013), but associated implications for the palaeorecord are rarely discussed (c.f. Warren and Hulton, 1990; Warren, 1991; Kaplan et al., 2009; Anderson et al., 2012; Pedersen and Egholm, 2013). To address this, topographic controls on glacier dimensions, dynamics, and margin stability ─ which have direct implications for the location and timing of moraine deposition ─ are considered here. Specific focus is placed on the role of accumulation area topography; evolving basin topography; and topographic pinning points. Topoclimatic factors, such as orographic precipitation gradients, controls on solar radiation, or the channelling of precipitation through valley systems are not discussed (though they clearly have an impact on glacier dimensions and, by extension, moraine positions, and are sometimes difficult to differentiate from purely topographic controls), as these are considered elsewhere (e.g., Evans, 1990; Dahl and Nesje, 1992; Hulton and Sugden, 1995, 1997; Lehmkuhl et al., 2011; Lehmkuhl, 2012; Loibl et al., 2014) and reflect local climatic, rather than purely topographic, controls (the focus of this paper).

3.1. Accumulation area topography as a control on moraine formation

In a given region, the topography above the regional climatic-ELA (i.e., in the accumulation zone) can vary significantly, from subdued plateaus to high relief peaks (Ward et al., 2012). This often regulates the size and shape of local glaciers (Manley, 1955; Ives et al., 1975; Sugden and John, 1976; Golledge, 2007) and thereby determines where in a landscape they terminate and deposit moraines (Kaplan et al., 2009). Here, we discuss this relationship, with a focus on four principal factors: (i) topographic controls on glacial hypsometry (i.e., the altitudinal-distribution of glacier surface area); (ii) topographic controls on indirect accumulation; (iii) topographic controls on debris supply; and (iv) evolving upland topography and implications for glacier extent.

3.1.1. Accumulation area topography and glacier hypsometry

In order for glaciers to develop in a particular region there must be sufficient land surface area above the ELA upon which snow and ice can accumulate and persist interannually (Kessler et al., 2006; Kaplan et al., 2009). Thus, the initiation of glacier growth requires topography that exceeds the regional climatic-ELA (or is in close proximity to it if accumulation is derived from snowblow or avalanching) but which also has favourable local topographic characteristics (such as aspect and gradient). The land surface area above the ELA on which glacier ice can accumulate regulates total ice-mass accumulation and thereby governs the distance below the regional climatic-ELA that glaciers are able to extend (Furbish and Andrews, 1984; Kerr, 1993; Kaplan et al., 2009; Chenet et al., 2010; Pratt-Sitaula et al., 2011). In situations where upland topography varies from one region (or drainage basin) to the next, this can allow glaciers to have very different dimensions and to deposit moraines at very different locations and altitudes, despite experiencing similar climatic conditions (Andrews et al., 1970; Barr and Clark, 2012b). This is apparent when glaciers occupy mountains of notably different altitudes (Fig. 3A) and is particularly important when plateau and nonplateau topography are considered (Figs. 3B and C) (Manley, 1955; Sugden and John, 1976; Rea et al., 1998, 1999; McDougall, 2001). For example, upland plateaus provide large areas for snow and ice accumulation, whilst steep mountain peaks provide little. Thus, plateau topography potentially allows extensive glaciers to develop, whilst nonplateau topography will restrict glaciation to smaller ice masses (Sugden and John, 1976) (Fig. 3). The Saltfjellet Mountains of western Norway (Fig. 3C) provide a modern example of this scenario where the Østre Svartisen icefield occupies an upland plateau and extends from ~ 1550 to ~ 300 m asl, with outlets > 10 km in length. By contrast, adjacent nonplateau topography has peaks > 1600 m asl but only supports small (< 3 km in length) glaciers, which fail to extend below 700 m asl. This example illustrates how, under a similar climate regime (Hijmans et al., 2005), ice masses occupying plateau topography can be more extensive and can deposit moraines at lower altitudes than corresponding glaciers occupying adjacent nonplateau surfaces. In fact, where plateau and nonplateau topography exists, high altitude peaks may remain ice-free whilst glaciers occupy adjacent but lower altitude plateaus and valleys (Manley, 1955; Ward et al., 2012). Thus, variations in upland topography clearly have an impact on where in a landscape moraines are deposited.

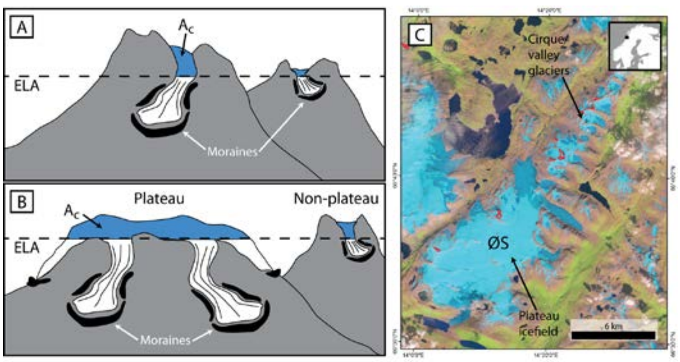

Fig. 3. Illustration of how, under uniform climatic conditions (reflected by a uniform ELA), (A) mountain height and (B) plateau and nonplateau topography can lead to variations in glacier dimensions and thereby control moraine location. Blue zones represent glacier accumulation areas (A_c). (C) Example from the Saltfjellet Mountains (western Norway) of how the plateau topography occupied by the Østre Svartisen icefield (ØS) allows outlet glaciers to extend

farther than adjacent noncoalesced cirque and valley glaciers, despite occupying equivalently high topography and experiencing a similar climate regime.

Hypsometry and glacier response to climate: In addition to exerting a control on the downvalley extent of glaciers under fixed climate regimes (as outlined above), accumulation area topography and, by extension, ice mass hypsometry can have an impact on how glaciers respond to fluctuations in climate (Oerlemans, 1989; Kerr, 1993; Chenet et al., 2010; Pratt-Sitaula et al., 2011; Pedersen and Egholm, 2013; Loibl et al., 2014). This is illustrated in the Annapurna region of central Nepal where Pratt-Sitaula et al. (2011) found a shift from cryo-arid late glacial conditions (with comparatively low ELA but also low glacier mass turnover) to warmer and wetter early Holocene conditions (with a higher ELA but greater mass turnover), to result in the advance of glaciers with high altitude source areas but the retreat of lower altitude examples (Fig. 4). This pattern is considered to have developed because glaciers with high altitude source areas experienced only a slight reduction in the size of their accumulation zones as ELA increased, and this was more than compensated by increased precipitation under warmer and wetter conditions (Fig. 4B). By contrast, glaciers with low altitude source areas experienced a significant reduction in the size of their accumulation zones in response to the elevated ELA, and increased precipitation under the wetter regime was unable to compensate for this loss. Such topographically induced variation in glacier response to climate, and associated asynchrony in when glaciers reached their maximum extent, is likely to result in the preservation of very different moraine sequences from one valley to the next. For example, where glaciers reached their maximum extent as ELA increased toward the early Holocene, moraines preserved in the landscape are likely to relate to this phase (Fig. 4B), with preexisting deposits destroyed by obliterative overlap (where glacial advances remove evidence of previous ice extent) (Kirkbride and Brazier, 1998; see section 4.1.1). By contrast, where glaciers experienced net retreat in response to increased ELA, a more detailed moraine record might be preserved, perhaps reflecting ice extent during the late glacial and during the early Holocene (Fig. 4B).

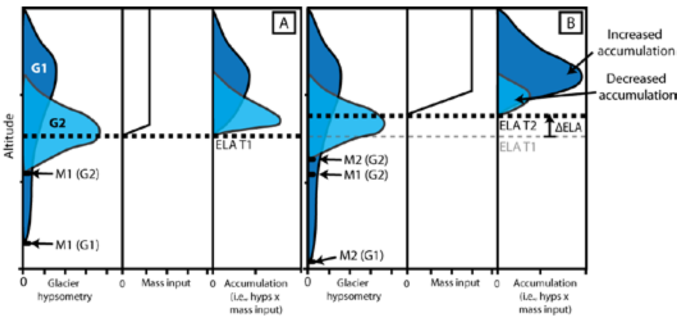

Fig. 4. Schematic illustration of how glacier hypsometry (as controlled by accumulation area topography) can regulate the nature of glacier response to fluctuations in climate. In this example, a shift from (A) cryo-arid conditions with comparatively low ELA but low mass turnover (i.e., nondynamic glaciers) to (B) warmer, wetter conditions with slightly higher ELA (rise of ~ 120 m) and greater mass turnover (more dynamic glaciers), can lead a glacier (G1) with a high altitude source area to advance, whilst a lower-altitude glacier (G2) recedes. G2 reaches its maximum extent during phase A, whilst G1 reaches its maximum extent during phase B. Thus, accumulation area topography results in asynchronous glacial maxima and very different moraine records preserved in each basin. For example, the moraine record preserved by G1 will reflect ice extent during the more recent, higher ELA, phase of advance, whilst G2 may preserve a more detailed moraine record, reflecting ice extent before and after the increase in ELA. Figure adapted from Pratt-Sitaula et al. (2011).

3.1.2. Accumulation area topography and controls on indirect accumulation

The accumulation of glacier ice occurs through direct snowfall, the superimposition of ice, and through the redistribution of snow and ice accumulated elsewhere in a glacier's catchment (i.e., not on the ice surface) (Koerner, 1970; Hagen et al., 2003). Glacier-to-glacier variations in total direct snowfall and superimposed ice are largely governed by climate (including topoclimatic factors) (Oerlemans et al., 1998; Dyurgerov and Meier, 2000; Wadham and Nuttall, 2002; Putnam et al., 2012), whilst the redistribution of snow and ice generally occurs through snowblow and avalanching and is controlled by accumulation area topography. Specifically, the volume of redistributed material accumulating on a glacier's surface is determined by the area, slope, and aspect of surrounding topography (Benn and Lehmkuhl, 2000). For example, extensive plateau topography provides the potential for significant buildup of snow and ice, which might then be blown onto a glacier (Kerr, 1993; Ballantyne 2007a,b); steep slopes (> 20-30°) overlooking glacier accumulation zones may supply snow and ice through avalanching (Benn and Lehmkuhl, 2000); whilst slopes facing dominant wind directions, or slopes with pre-existing hollows, might act as a barrier to windblown snow and ice, encouraging this material to lodge and accumulate (Sugden and John, 1976). Thus, where a series of glaciers occupy equally high mountains and experience the same climate regime, variations in accumulation area topography can regulate the volume of redistributed snow and ice supplied to their surfaces and thereby govern their total accumulation and downvalley extent (see Luckman, 1977; Ageta and Higuchi, 1984; Benn and Lehmkuhl, 2000). Little attention has been paid to the impact this might have on the distribution of moraines in the modern landscape, though some have attempted to account for the contribution of redistributed snow and ice when calculating palaeoglacier ELAs (e.g., Sissons and Sutherland, 1976; Ballantyne, 2007a,b).

3.1.3. Accumulation area topography and controls on debris supply

Where steep, high-relief slopes surround a glacier, there is not only potential for increased snow and ice accumulation from avalanching (see section 3.1.2) but also potential for increased overall debris supply to the ice surface (see Anderson, 2000). This accumulating debris serves as a source of material for moraine construction (Reznichenko et al., 2011, 2012) but can also insulate a glacier and dampen its response to climate (Kirkbride and Warren, 1999; Anderson, 2000; Scherler et al., 2011) and can even trigger phases of glacier advance (see Tovar et al., 2008; Hewitt, 2009). Where steep slopes overlook a glacier's surface, debris accumulation can be considerable, and glaciers can persist far below the regional climatic ELA (Scherler et al., 2011) — with a direct impact on where moraines are deposited. In addition, because of the availability of material, moraines associated with debris-rich glaciers may be large (Andrews, 1972), but the total number of moraines is likely to be limited as debris-covered glaciers are unlikely to respond to short-lived periods of climatic forcing (see Scherler et al., 2011). For this reason, debris-covered glaciers may only preserve moraine records reflecting larger variations in climate, whilst debris-poor glaciers preserve detailed records of smaller advances (Kirkbride and Winkler, 2012).

3.1.4. Ice extent and evolving upland topography

As outlined in the sections above, accumulation area topography exerts a control on glacier extent and, by extension, regulates where in the landscape moraines are deposited. However, the nature of this control varies during and between phases of glaciation (Singer et al., 2004; Kaplan et al., 2009). Specifically, under given climatic conditions, uplift and/or isostatic rebound will increase the overall land surface area available for ice accumulation (i.e., increasing the area above the ELA), resulting in more extensive glaciers; whilst denudation and/or subsidence will do the opposite (Singer et al., 2004; Anderson et al., 2012). As a result, ice-mass dimensions might differ from one phase of glaciation to the next because of intervening periods of uplift and/or subsidence (Fig. 5). When uplift outpaces denudation and glacier extent increases with time, only the most recently deposited moraines will be preserved in the modern landscape, as older examples are destroyed by obliterative overlap (Fig. 5A) (section 4.1.1). Conversely, where denudation or subsidence outpace uplift and the downvalley extent of glaciers diminishes with time, a detailed record of different-aged moraines might be preserved (Kaplan et al., 2009; Anderson et al., 2012;

Pedersen and Egholm, 2013) (Fig. 5B). This latter scenario is particularly relevant as glacial occupation of mountain ranges increases rates of erosion (denudation), generally reducing the elevation of accumulation areas (see Brozović et al., 1997; Whipple et al., 1999; Tomkin, 2003; Mitchell and Montgomery, 2006; Egholm et al., 2009). Thus, where the erosion of upland topography is efficient, and outpaces uplift, glaciers may become smaller over successive glacial cycles (Kaplan et al., 2009).This has been referred to as a self-defeating mechanism (MacGregor et al., 2000), or erosional feedback (Anderson et al., 2012), and is likely to operate over millennial time scales and be more effective in regions occupied by warm-based, dynamic, and erosive glaciers that are not frozen to their beds (see section 3.2).

In addition to regulating regional ice extent during and between phases of glaciation (as outlined above), the evolution of upland topography can regulate the distribution of ice between individual drainage basins or from one side of a mountain divide to another (i.e., through ice piracy or drainage capture). For example, uplift, erosion, or subsidence of topography (particularly interfluves) can lead to variations in flow directions, and focus (or steer) ice from one catchment to another (e.g., Joughin and Tulaczyk, 2002; Taylor et al., 2004). This has the effect of increasing ice supply to one basin and reducing it in another, which, under the same climatic regime, theoretically allows the expansion of one glacier whilst an adjacent or contiguous glacier is more restricted in extent. In this way, the focusing of ice flow into a particular valley not only determines where moraines are deposited but can lead to intervalley variations in ice velocity, thickness, and thermal regime, which govern the extent and intensity of subglacial erosion and can influence the efficacy of erosional feedbacks (see section 3.2) (Glasser, 1995; Taylor et al., 2004; Kessler et al., 2006; Jamieson et al., 2008; Kaplan et al., 2009).

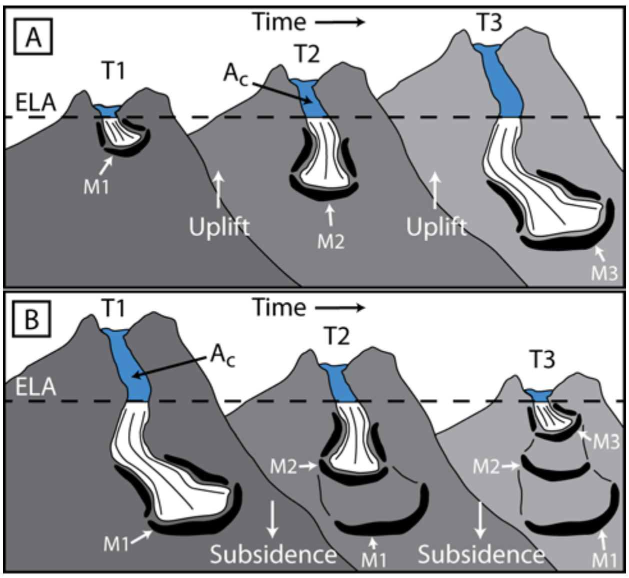

Fig. 5. Illustration of the impact mountain uplift and subsidence (or denudation) can have on glacier extent, and therefore moraine location, under uniform climatic conditions (fixed ELA). (A) Where uplift occurs between phases of glaciation (marked by time-steps T1, T2, and T3), glacier length progressively increases. At each time-step, maximum ice extent is marked by a deposited moraine (M1-3). In this scenario, only the outermost moraine (i.e., M3) is preserved, as older moraines are destroyed by obliterative overlap. (B) Where upland subsidence (or denudation) occurs between phases of glaciation, glacier length progressively decreases, and a detailed record of different-aged moraines is preserved in the landscape (reflecting various periods of ice advance). Blue zones represent glacier accumulation areas (A_c).

3.2. Evolving basin topography as a control on moraine formation

Although accumulation area topography can play a significant role in regulating glacier extent (as outlined in section 3.1), the shape of valley basins (i.e., ablation area topography) is also known to be important (Oerlemans, 1989; Taylor et al., 2004; Kessler et al., 2006; Kaplan et al., 2009; Anderson et al., 2012; Pedersen and Egholm, 2013). Specifically, a contrast is often made between glacially modified basins, which are deep — wide and punctuated by topographic steps and overdeepenings (see Harbor, 1992; MacGregor et al., 2000; Montgomery, 2002; Cook and Swift, 2012) — and nonglacial basins — which have comparatively smooth longitudinal profiles that gradually decrease in slope downvalley (MacGregor et al., 2000). These topographic characteristics have important implications for the extent of the ice masses that come to occupy such basins (see Oerlemans, 1984; MacGregor et al., 2000). For example, as nonglacial basins are not deeply eroded, they effectively act as regions of plateau topography; allowing sizable ice masses to develop (Fig. 6A) (see section 3.1.1). By contrast, ice masses occupying deep, glacially eroded basins will rapidly extend to low altitudes. Such glaciers will be comparatively small, and terminate, depositing moraines closer to glacier source areas (see Small and Anderson, 1998; MacGregor et al., 2000; Kaplan et al., 2009; Anderson et al., 2012) (Fig. 6A). Over a series of glacial cycles, subglacial erosion acts to progressively erode basin topography, pushing the mean elevation of valley profiles toward increasingly low altitudes (Harbor, 1992; Barr and Spagnolo, 2014). Thus, even when climatic conditions (and global ice volumes) remain essentially the same from one glacial maximum to the next, ice masses can experience a reduction in aerial extent over successive glacial cycles (Fig. 6A) because of this erosional feedback as basins are skewed toward lower altitudes (see Oerlemans, 1984; MacGregor et al., 2000; Taylor et al., 2004; Kessler et al., 2006; Kaplan et al., 2009; Anderson et al., 2012). Some have argued that this leads to a sequence of moraines that become progressively younger toward ice source areas, but where moraine position is largely a reflection of cumulative basin erosion rather than climate (see Kaplan et al., 2009; Anderson et al., 2012) (Fig. 6A). Anderson et al. (2012) referred to the outermost moraines in such a sequence as far-flung moraines (often deposited well beyond regional LGM ice limits) (Fig. 7) and demonstrated that they are an inevitable consequence of repeated glaciations, assuming that the rate of glacial erosion outpaces uplift and that climatic conditions remain similar from one glacial maximum to the next — as suggested by global ice volume estimates from the benthic $\delta^{18}O$ record (Fig. 8).

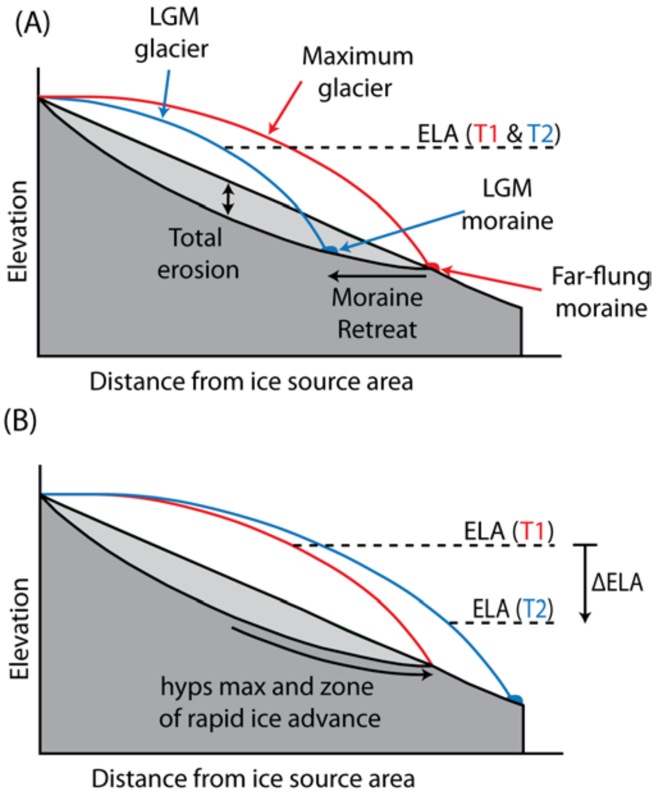

Fig. 6. (A) Illustration of the impact subglacial erosion (erosional feedback) can have on glacier length during successive phases of ice advance, despite climatic conditions remaining essentially the same from one glacial maximum to the next (i.e., the ELA remains constant). In this scenario, the maximum glacier (in red) advanced into a nonglacial drainage basin and extended a considerable distance downvalley before terminating and depositing a far-flung moraine (in red). The LGM glacier (in blue) advanced into a landscape deepened by the maximum glacier and rapidly extended below the regional climatic ELA. This glacier therefore terminated earlier and deposited a moraine (in blue) closer to the ice source area. Figure modified from Anderson et al. (2012). (B) As in (A), but where ELA is lowered following the initial stage of ice advance. In this instance, the glacially eroded landscape allows rapid ice advance once the ELA reaches the hypsometric maximum (hyps max) — resulting in the removal of the T1 moraine. Figure based on concepts outlined in Pedersen and Egholm (2013).

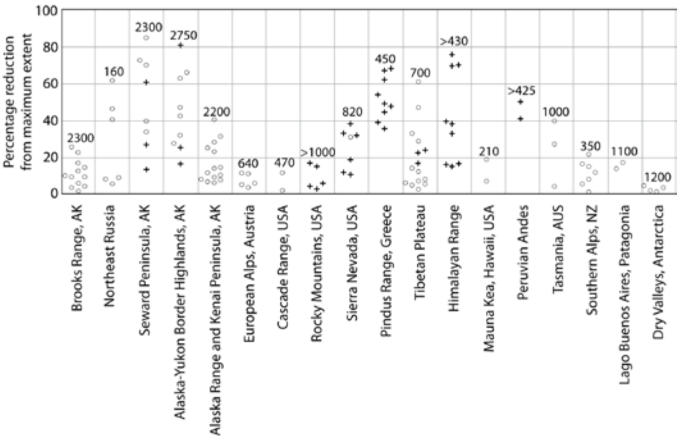

Fig. 7. Global compilation of far-flung moraines. Plus markers represent moraines produced by valley glaciers. Circles represent moraines produced by ice cap outlet glaciers. The figure shows the percentage reduction in glacier length between the greatest glacier extent and documented glacier extents prior to, or including, the LGM in each region or range. Dates (in ka) above each column are the oldest moraine age for each region or range. Figure redrawn from Anderson et al. (2012).

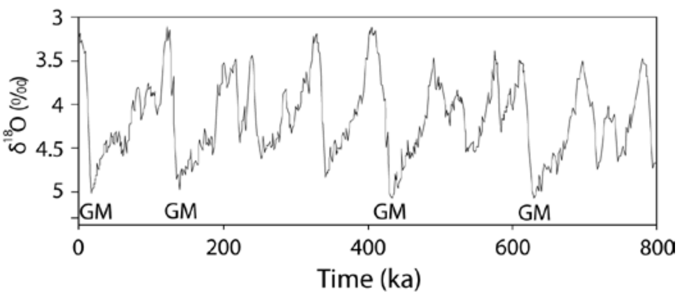

Fig. 8. Benthic $\delta^{18}O$ record (considered a proxy for global ice volumes) covering the past 800,000 years (from Lisiecki and Raymo, 2005). These data are used by Anderson et al. (2012) to support the argument that, even when climatic conditions (as reflected by global ice volumes) remain essentially the same from one glacial maximum (GM) to the next, ice masses can experience a reduction in areal extent over successive glacial cycles because glacial erosion skews basin topography toward lower altitudes over time (i.e., erosional feedback).

3.2.1. Evidence of erosional feedback

The global prevalence of far-flung moraines, apparently recording a reduction in ice extent over successive glacial cycles, was highlighted by Anderson et al. (2012) (Fig. 7). However, whether this trend reflects the influence of erosional feedback, climatic forcing, or some other control (such as progressive basin subsidence) is difficult to establish. As noted in section 3.1.4, subglacial erosion is likely to be promoted beneath warm-based, dynamic glaciers that are not frozen to their beds (Delmas et al., 2009; Kaplan et al., 2009), and erosional feedback might therefore be the leading explanation for far-flung moraines in mid-latitude, temperate regions. This certainly appears to be true in the southern Andes of South America where Singer et al. (2004) hypothesised that uplift prior to ~ 1.1 Ma led to significant glaciation and that erosional feedback has resulted in progressively less extensive ice coverage during subsequent glaciations. This notion was supported by Kaplan et al. (2009) who attributed the nested sequences of moraines associated with outlet glaciers of the former Patagonia Ice Sheet (see Fig. 2C) to a successive reduction in ice extent over the past ~ 1 Ma, caused by feedback resulting from the erosion of accumulation areas and drainage basin topography. However, far-flung moraines are also found in regions where erosional feedback might be expected to be less efficient. One example is the Dry Valleys, Antarctica (see Fig. 7) where evidence suggests a decrease in ice extent during successive glaciations despite cold-based glaciation (Brook et al., 1993; Schaefer et al., 1999; Margerison et al., 2005). In such areas, the specific processes responsible for the decrease in ice extent (e.g., erosion, tectonics, or climate) might be difficult to establish (Kaplan et al., 2009). However, notably, the reduction in the extent experienced by these glaciers is typically less pronounced than in other regions (likely occupied by more erosive glaciers) (see Fig. 7), as would be expected if erosional feedback was the major control on glacier dimensions.

3.2.2. Confounding factors

Despite evidence to suggest that erosional feedback acts to limit glacier extent, and thereby control moraine distribution (see section 3.2.1), a number of factors potentially counter this mechanism, and introduce complexity to the record. These include the following: (i) erosional feedback is unlikely to be effective over short time scales or in the case of small glaciers. (ii) Bedrock lithology and geological structure can exert considerable control on erosion rates (Delmas et al., 2009). (iii) As noted in section 3.2.1, cold-based glaciers are often considered to be minimally erosive (Dyke, 1993; Davis et al., 2006). (iv) Basin uplift between phases of glaciation can act to counter glacial erosional feedback. (v) As a landscape becomes tuned to glacial occupation, the extent to which glaciers deepen/erode their basins is likely to diminish (e.g., Charreau et al., 2011; Herman et al., 2011). This is likely to be reinforced in regions where overdeepened basins prevent the efficient removal of subglacial debris (Cook and Swift, 2012). In this instance, the intensity of erosion (and sediment production) is likely to peak during the initial transition from a largely fluvial to glacial landscape (Charreau et al., 2011) and diminish with time thereafter. (vi) Where the magnitude of climatic deterioration increases from one glacial maximum to the next (i.e., lowering of ELA), erosional feedback is likely to be of limited importance. Pedersen and Egholm (2013) considered this final scenario and found that deterioration in climate from one phase of glaciation to the next in fact stimulates very rapid ice advance in basins that have been glacially modified. This occurs because during an initial phase of glaciation, basin topography is eroded and deepened. This erosion is typically most intense at the ELA where total ice-flux is maximised (Boulton, 1996; Hallet et al., 1996; MacGregor et al., 2000; Anderson et al., 2006), and land surface area therefore becomes concentrated at just this altitude (i.e., a hypsometric maximum develops) (Egholm et al., 2009; Pedersen et al., 2010; Pedersen and Egholm, 2013) (Fig. 6B). During a subsequent phase of glaciation (following a period of retreat), ice slowly advances and extends down the glacially modified landscape. Once the glacier reaches the flat topography associated with the glacially induced hypsometric maximum (i.e., just below the previous ELA), rapid advance occurs (Fig. 6B). Thus, the interaction of the glacier and the hypsometric maximum generated during an earlier phase of glaciation allows a threshold of stability. This can also happen when glaciers are able to advance, and coalesce, on intermassif upland basins (see Payne and Sugden, 1990; Kerr, 1993; Sugden et al., 2002). For example, where glaciers develop on isolated, comparatively high-relief massifs, not surrounded by upland basins, ice quickly advances to low altitudes where margin stability ensues (Fig. 9A). Because of the absence of intermassif upland basins (effectively high altitude hypsometric maxima), such ice masses are unable to readily coalesce, and their dimensions remain comparatively restricted (Fig. 9A). By contrast, where ice mass development occurs on low-relief massifs, glaciers are able to extend into surrounding high altitude basins and coalesce (Fig. 9B). At this point, ice thickening leads to a positive feedback, as much of the ice surface is elevated above the ELA. This leads to a rapid growth in ice-mass dimensions and considerable ice advance, as a threshold of stability is crossed. In this way, total relief and the distribution of upland basins (hypsometry) are important regulators of ice mass development (see Payne and Sugden, 1990; Hulton and Sugden, 1997; Sugden et al., 2002).

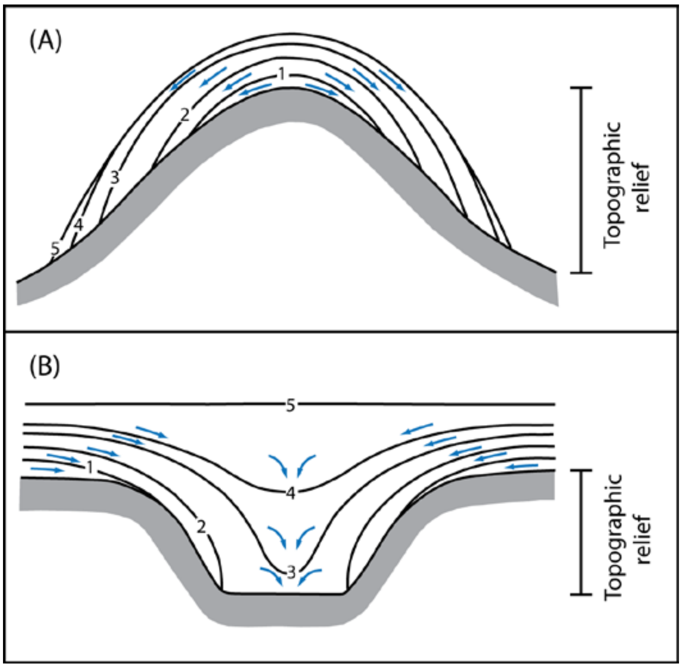

Fig. 9. Schematic cross section showing different phases of ice mass growth in response to a constant temperature depression (ELA lowering). Numbered surfaces reflect stages of growth — from small (1) to large (5). (A) Where ice mass development occurs on an isolated, comparatively high relief massif, ice flows to low altitudes where margins stabilise. The isolation and high relief prevents coalescence, and glaciers remain comparatively small. (B) Where ice mass development occurs on comparatively low-relief uplands, and glaciers are able to extend into surrounding basins, a threshold of stability is crossed between growth stages 2 and 3, where ablation is unable to balance accumulation and ice masses coalesce. Ice thickening at this point leads to a positive feedback, as much of the ice surface is elevated above the ELA. This leads to a rapid growth in ice-mass dimensions. Arrows indicate ice flow directions. Figure modified from Payne and Sugden (1990).

3.3. Topographic pinning points as controls on moraine formation

Pinning points are regions where topography either physically restricts ice advance or causes margin stability through an impact on glacier mass balance (see Hillaire-Marcel et al., 1981; Burbank and Fort, 1985; Warren and Hulton, 1990; Warren, 1991). Such locations often result in the deposition of moraines (Warren, 1991), and here we focus on four different factors: (i) physical restrictions to glacier advance — where variations in lithology, preexisting landforms, or topographic steps physically restrict ice movement; (ii) bed slopes — where longitudinal variations in bed topography regulate glacier mass balance and ice margin stability; (iii) variations in valley width — where variations in cross-sectional topography regulate the mass balance and stability of land-terminating glaciers; and (iv) fjord topography — where topographic factors regulate the mass balance and stability of lake- and marine-terminating glaciers.

3.3.1. Physical restrictions to glacier advance

In certain circumstances, geology, geomorphology, and/or topography can act to physically restrict the downvalley extent of glaciers and can control where in a landscape moraines are deposited (Burbank and Fort, 1985; Kirkbride, 2000; Spedding and Evans, 2002; Evans et al., 2010). One example of a geological restriction comes from the Zanskar Range in the northwestern Himalaya, where Burbank and Fort (1985) found evidence that the downvalley extent of late Pleistocene glaciers was restricted by vertical strata of sandstones and conglomerates (molasse), rising 400-1000 m above valley floors (Fig. 10). This upstanding bedrock is intersected by a number of narrow steep-walled canyons, which restricted ice advance and led to margin stabilisation and moraine deposition at canyon entrances (Figs. 10 and 11A). This is a pattern characteristic of a number of sites in the northwestern Himalaya (Burbank and Fort, 1985). Similar constraints on downvalley ice extent can be imposed by geomorphology, particularly where glaciers advance into valleys already occupied by large moraines or other landforms (e.g., Kirkbride, 2000; Spedding and Evans, 2002; Evans et al., 2010) (Fig. 11B). In this instance, preexisting deposits can cause glaciers to terminate at similar locations during successive phases of advance, resulting in the formation of large, complex, tightly nested moraine sequences (Kirkbride and Winkler, 2012). Modern examples are the multicrested push moraine complexes produced by multiple surges of valley glaciers on Svalbard where moraines formed by successively younger advances are piggy-backed on to the ice-proximal side of older moraines (Hart and Watts, 1997) (Fig. 12A). This control on glacier extent is likely to be most influential where geomorphological barriers are large, where glaciers are small and able to extend laterally on meeting obstructions, or when climatic differences between successive glacial advances are not particularly large. Where glaciers are large (relative to the size of the geomorphological barrier), are unable to expand laterally, or when climatic differences between phases of advance are considerable, moraines are likely be overridden, resulting in obliterative overlap (see section 4.1.1). The final physical restrictions to downvalley ice extent considered here are topographic steps in longitudinal valley profiles. For example, reverse bed slopes, such as the downvalley sides of overdeepenings (Fig. 11C), require glaciers to advance uphill (see Cook and Swift, 2012). These topographic steps therefore act as obstacles to flow and are likely to encourage ice margin stability, leading to moraine deposition (Kerr, 1993). An example of this comes from the Skyring and Otway lobes of the former Patagonian Ice Sheet in southernmost South America, which are delineated by closely nested sequences of moraines located on higher ground (Lovell et al., 2011; Darvill et al., 2014), assumed to relate to several glaciations (Kaplan et al., 2009). Here, the ice lobes have been forced to flow through depressions and up reverse slopes, depositing latero-frontal moraines on the edge of the higher ground on several occasions (Fig. 12B). The depressions were subsequently occupied by ice-marginal lakes as ice retreated from the higher ground (Lovell et al., 2012) and are likely to have been infilled during and following deglaciation (see section 5.2.), which may explain why they are not particularly noticeable in the modern landscape.

Fig. 10. Entrance to Stok Canyon (Zanskar Range, northwestern Himalaya) where vertical strata of sandstones and conglomerates (labelled A) rise 400-1000 m above valley floors and acted as physical barriers to late Pleistocene ice advance (according to Burbank and Fort, 1985). During this period, glaciers terminated and deposited moraines at the canyon entrance. Remnants of these moraines (labelled B), with exposed till, can be seen in the left foreground. Description based on Burbank and Fort (1985). Photo courtesy of P. Chladek.

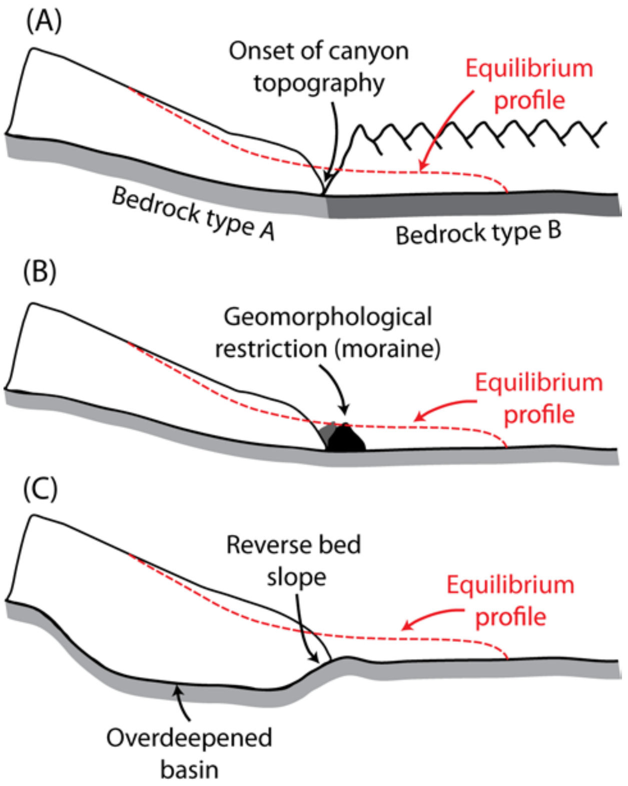

Fig. 11. Illustration of geological, geomorphological, and topographic barriers to ice advance; each of which potentially results in moraine formation. (A) Where bedrock configuration (i.e., vertical strata dissected by canyons) restricts ice advance (based on ideas in Burbank and Fort, 1985). (B) Where preexisting deposits (in this case moraines) act as geomorphological restrictions to ice advance, resulting in closelynested moraine sequences as younger moraines (in dark grey) are piggy-backed on the ice-proximal side of older moraines (in black). (C) Where a reverse bed slope (on the downvalley side of an overdeepening) acts as a physical barrier to ice advance. In each example, the restriction to downvalley advance generally leads to over-steepened and over-thickened ice margins as glaciers become disconnected from their equilibrium profiles (illustrated schematically by dashed lines).

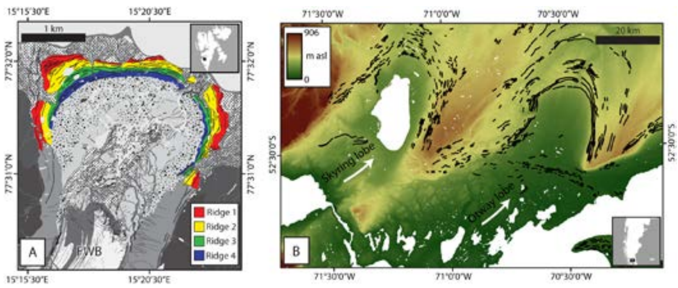

Fig. 12. (A) Multicrested push moraine complex at the margin of Finsterwalderbreen (FWB), a surge-type glacier on Svalbard. The individually coloured ridges were formed by separate advances, with each successive ridge piggy-backed on to the ice-proximal side of a ridge formed during a previous advance. (B) Nested latero-frontal moraine sequences of the former Skyring and Otway lobes in southernmost Patagonia displayed on an SRTM DEM. Note how

the moraines are largely located on the edge of the higher ground (> 200 m asl) rather than in the depressions. White arrows show inferred palaeo ice flow directions. Mapping from Lovell et al. (2011) and Darvill et al. (2014).

3.3.2. Bed slopes, glacier mass balance, and ice margin stability

As noted in section 3.3.1, topographic steps in longitudinal bed profiles can act as physical barriers to ice advance and thereby regulate where moraines are deposited. More subtle variations in bed topography can also exert a control on moraine formation, through an impact on glacier mass balance and margin stability (here, the focus is on land-terminating glaciers, whilst lake- and marine-terminating glaciers are considered in section 3.3.4) (Oerlemans, 1989; Warren and Hulton, 1990; Warren, 1991, 1992; Kerr, 1993). For example, where bed topography is comparatively shallow, glaciers typically have correspondingly low surface gradients and are therefore sensitive to climate forcing (Oerlemans, 1989). This sensitivity occurs because vertical displacements in ELA result in significant variations in the size of glacier accumulation areas (Kerr, 1993; Pedersen and Egholm, 2013) (Fig. 13A). By contrast, glaciers occupying steep bed slopes are likely to be comparatively insensitive to variations in climate, particularly when the regional ELA coincides with topographic steps (creating ice falls), and so vertical displacements in ELA will have limited impact on the size of glacier accumulation areas (see Dugmore, 1989; Oerlemans, 1989) (Fig. 13B). Glaciers on relatively shallow bed slopes are therefore likely to deposit a large number of comparatively small moraines, as ice margins fluctuate in response to minor variations in climate but rarely experience prolonged periods of stability. By contrast, glaciers on steep bed slopes are likely to deposit a small number of large moraines, as margins are comparatively stable (i.e., they do not respond to minor variations in climate). Thus, variations in bed slope can have a direct influence on the number and characteristics of moraines preserved in a given area, often making the task of correlating moraines across a region, or even between adjacent valleys, very difficult (see section 4.1.1).

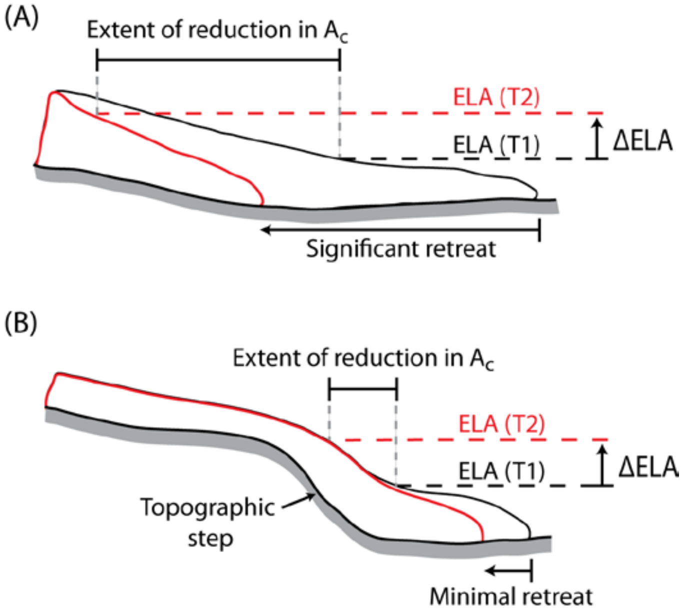

Fig. 13. Schematic illustration of the impact bed slopes can have on glacier dimensions and stability. Glacier (A) occupies comparatively shallow (low gradient) topography, whilst glacier (B) occupies a topographic step. In both (A) and (B), glacier dimensions during two time periods are shown, (T1) and (T2), separated by a period of climatic amelioration (resulting in an increase in ELA). The ELA displacement in (A) and (B) is equal, but resulting variations in terminus position vary significantly. In (A), the glacier experiences significant retreat between T1 and T2; whilst in (B), retreat is minimal because the topographic step means that the displacement in ELA has limited impact on the size of the glacier's accumulation area (A_c). Glacier (A) may deposit a number of small moraines, whilst the (comparative) margin stability experienced by glacier (B) is likely to promote the formation of few, but comparatively large, moraines.

3.3.3. Variations in valley width

Valley width can play an important role in regulating glacier mass balance and margin stability (again, the focus here is on land-terminating glaciers, whilst lake- and marine-terminating examples are considered in section 3.3.4). The nature of this relationship varies spatially and temporally and partly depends on whether a glacier is undergoing a period of net advance or retreat (Kerr, 1993; Lukas, 2007; Barr and Clark, 2012b). For example, in response to a period of positive mass balance, glaciers confined to narrow basins will rapidly advance, as (assuming no change in the mass balance gradient) the increased total accumulation will only be counterbalanced by an increase in the length of the ablation area (i.e., ice is unable to extend laterally) (see stages 1-3 in Fig. 14A). Upon reaching unconfined plains, these glaciers will spread laterally, forming piedmont lobes (Fig. 14C), and terminus advance will begin to slow or even stop (see stages 4-7 in Fig. 14A) because of a rapid increase in the size of glacier ablation areas.

This effectively corresponds to a topographically induced period of margin stability and is likely to result in the formation of large, lobate piedmont moraines (Dugmore, 1989; Barr and Clark, 2012b) (Fig. 14D). During periods of retreat, mass loss from these piedmont glaciers will often be rapid (because of their large ablation areas), until their termini have returned to the confines of sheltered valleys (Fig. 14B). At this point, retreat will slow, partly because of a reduction in the size of ablation areas but also because of increased topographic shading (i.e., limiting ablation through direct insolation). The comparative margin stability experienced by glaciers that have retreated to the confines of sheltered valleys is likely to result in the deposition of moraines. For example, in Krundalen, south-central Norway, Lukas (2007) found a moraine complex located just upvalley from the junction of Krundalen with the main valley of Jostedalen. This complex is considered to reflect a period of margin stability during net retreat from the Younger Dryas position, and Lukas (2007) attributed its location to a topographically induced reduction in ablation, as the glacier retreated into the confined and sheltered Krundalen valley. Thus, as a result of variations in valley width, during periods of glacier advance moraine formation might be favoured in piedmont regions; whilst during retreat, moraine formation might be favoured just inside laterally confined valleys (see Funder, 1972; Warren and Hulton, 1990; Lukas, 2007; Barr and Clark, 2012b). This suggestion is supported by a data set of 8414 east Siberian moraines mapped by Barr and Clark (2012a,b) where ~ 67% (n = 5677) are found in piedmont regions (see example in Fig. 14D and details in Table 1), and ~ 9% (n = 757) are located within 1 km upvalley from basin junctions.

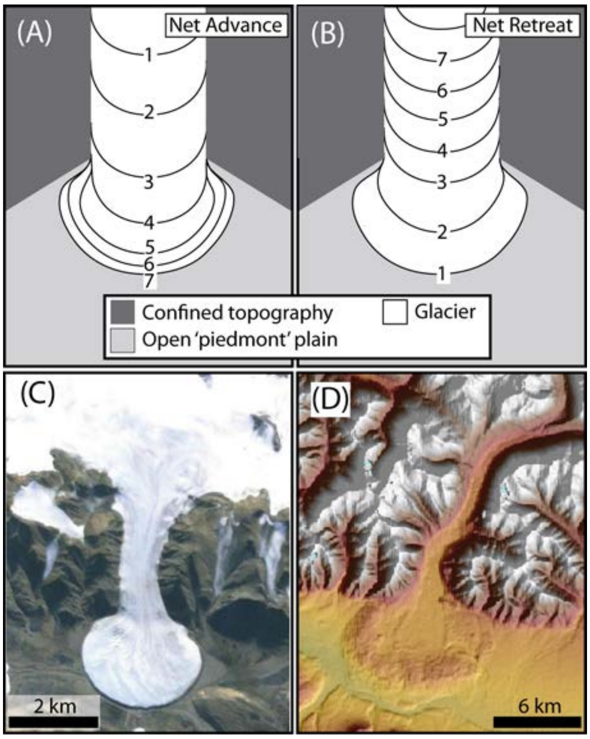

Fig. 14. Illustration of the impact variations in valley width can have on glacier margin stability and moraine location. (A) Schematic planview of different phases of glacier growth during net advance (ELA lowering). Numbers reflect stages of growth and illustrate how glaciers confined to narrow basins advance rapidly (stages 1-3), but upon emerging from valley confines, expand laterally, leading to comparative stability (stages 3-7). (B) Schematic planview of different phases of glacier retreat. Numbers reflect stages of retreat and illustrate how piedmont glacier retreat is initially rapid (stages 1-3) but begins to stabilise (stages 3-7) once the ice terminus retreats to a comparatively confined valley. The periods of margin stability in (A) and (B) are considered to promote moraine formation. (C) Example of a piedmont glacier on Axel Heiberg Island, Arctic Canada (glacier location: 80.34° N., 93.52° W.) (Landsat image ETM+ image, displayed in Google Earth™). (D) Piedmont moraine in southern Kamchatka (moraine location: 53.46° N., 157.53° E.) (SRTM DEM). Moraines are often located in this piedmont zone, potentially reflecting topographic control on ice margin stability.

Table 1. Attributes of 8414 east Siberian moraines mapped by Barr and Clark (2012a,b); moraines are separated into piedmont and nonpiedmont examples

	Piedmont (n = 5677)			Nonpiedmont (n = 2737)		
	Min	Max	Mean	Min	Max	Mean
Area (km^2)	0.01	428.60	5.20	0.01	477.51	3.10
L (km)	0.11	67.00	3.70	0.05	108.9	2.80
W (km)	0.04	7.97	0.81	0.02	5.10	0.63
Relief (m)	0.52	325.01	28.69	1.00	268.16	28.28

3.3.4. Fjord topography

In the case of lake- and marine-terminating glaciers, ablation is dominated by calving, and ice mass stability (or lack thereof) is dictated by the position of the grounding-line where moraines are generated (Warren, 1992; Benn et al., 2007; Schoof, 2007). Calving glaciers are prone to fluctuations, but stability is typically favoured where the calving front is comparatively small (i.e., ablation is limited); where the terminus is grounded; and/or where basal-drag, lateral drag, and backstress are comparatively high (O'Neel et al., 2005; Benn et al., 2007). Topography (fjord/trough geometry) plays a key role in governing these factors and is therefore fundamental to the stability of lake- and marine-terminating glaciers (more so than in the case of land-terminating examples) (Warren and Hulton, 1990; Warren, 1991, 1992). In particular, topographic pinning points represent potentially stable locations where moraine formation can occur (Warren, 1991; O'Neel et al., 2005; Jamieson et al., 2012, 2014; Carr et al., 2013). These pinning points often consist of laterally confined shallow basins, subaerial topographic obstructions (i.e., islands) or bathymetric highs (such as the adverse slopes of overdeepenings) (Warren and Hulton, 1990; Warren, 1991, 1992; Cook and Swift, 2012). During retreat from stable locations into wider/deeper sections of troughs and/or away from islands or bathymetric highs, ablation will increase significantly (perhaps enhanced by the incursion of warm water into previously enclosed basins — see Jenkins et al., 2010), and glaciers will undergo rapid retreat until the next topographic pinning point is reached (Warren and Hulton, 1990; Boyce et al., 2007; Nick et al., 2009; Jamieson et al., 2012; Carr et al., 2013). This propensity for periodic stability during retreat is highlighted by Jamieson et al. (2012, 2014) who used a numerical model of ice-stream flow, constrained by mapped grounding-zone wedges, to analyse the post-LGM retreat of the Marguerite Bay Ice Stream (western Antarctic Peninsula). These data (Fig. 15) indicate that the ice stream experienced several retreat rate slowdowns (i.e., grounding-line stabilisations) at notable topographic highs and on reverse bed slopes (where lateral drag was enhanced at trough narrowings) (Jamieson et al., 2012). Since grounding-line stability is key to moraine formation, these topographic controls are generally considered to result in the preferential deposition of moraines (and grounding-zone wedges) in proximity to valley narrowings (i.e., associated with the transition to wider locations where calving is increased: Mercer, 1961; Funder, 1972); bends; bifurcations; where basins are shallow; and/or in the vicinity of topographic bumps (Warren and Hulton, 1990; Warren, 1991, 1992).

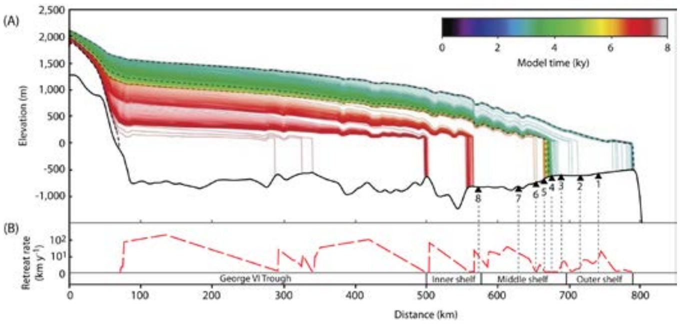

Fig. 15. Modelled ice stream retreat characteristics along the Marguerite Bay ice stream (western Antarctic Peninsula), illustrating the comparative stability of the grounding-line on topographic highs and on reverse bed slopes (where the trough narrows). (A) Along-flow profiles of surface elevation and terminus positions. Coloured lines reflect ice surface profiles at 5-year intervals, over an 8000-year period. Dotted black lines show the ice surface every 2000 years. The along-flow bed geometry is shown by the solid black line. (B) Modelled grounding-line retreat rate

(dashed red line). Numbered triangles and dashed grey lines show the locations of mapped grounding-zone wedges. Figure redrawn from Jamieson et al. (2012).

When lake- or marine-terminating glaciers retreat on to land (up a bedrock slope or prominent step), they often experience margin stabilisation, or even slight advance, as calving ceases (Hillaire-Marcel and Occhietti, 1977, 1980; Hillaire-Marcel et al., 1981; Funder, 1989; Warren and Hulton, 1990; Warren, 1991). This transition from a calving to terrestrial margin often results in moraine formation and can occur gradually during retreat or extremely rapidly in response to lake drainage (Hillaire-Marcel et al., 1981; Warren, 1991). In this latter scenario, a floating ice tongue, calving into a lake, can become instantly grounded during partial or complete lake drainage (Hillaire-Marcel et al., 1981) (Fig. 16). At this point, the glacier becomes disconnected from its equilibrium profile (i.e., the shape the glacier would assume in the absence of the lake) and will experience margin stability or slight advance until the equilibrium profile is restored (Fig. 16). During this period, a moraine (often referred to as a re-equilibration moraine) may be formed (see Andrews, 1973; Thomas, 1977; Hillaire-Marcel et al., 1981; Eyles and Eyles, 1984; Larcombe and Jago, 1994; Vorren and Plassen, 2002; Occhietti et al., 2011). Hillaire-Marcel et al. (1981) considered the 500-km-long Sakami moraine complex in eastern Canada to be an example of such a moraine, formed during the stepwise retreat of the Laurentide Ice Sheet, as Lake Ojibway drained and the lake level dropped by ~ 200 m. Moraines formed during the transition from lake- or marine-terminating to terrestrial margins are therefore topographically controlled (with little climatological significance) and often identifiable because they lie roughly parallel to coastlines or palaeolake shorelines (Sugden and John, 1976; Hillaire-Marcel et al., 1981; Warren and Hulton, 1990).

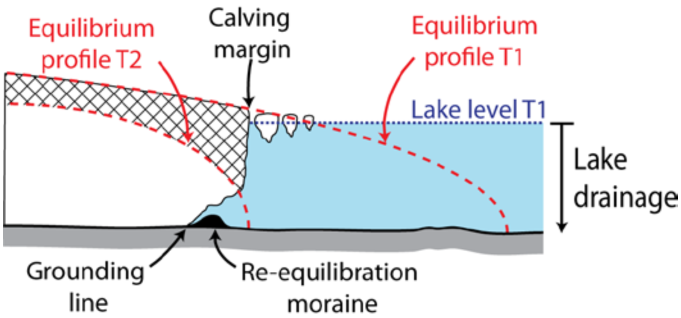

Fig. 16. Schematic illustration of how re-equilibration moraines can form following a lake drainage event. In this example, the lake level at time-step 1 (T1) results in calving. The glacier is in climatic equilibrium but, because of the efficiency of calving, does not assume the dimensions of its equilibrium profile, which reflects a theoretical ice surface in the absence of the lake (under a given climate). Between time-steps 1 and 2, lake drainage occurs. This results in a

very rapid reduction in ablation (because calving is stopped) and ice margin stabilisation. Active retreat will only occur once the volume of ice between equilibrium profiles T1 and T2 (shown here as the hatched area) has been lost through ablation (i.e., until the glacier assumes the dimensions of the equilibrium profile at T2). During this period (whilst ice in the hatched area is melting), a re-equilibration moraine may form. Figure based on Hillaire-Marcel et al. (1981).

4. Topographic controls on moraine preservation and identification

The record of moraines, as mapped in a given landscape (e.g., Heyman et al., 2008; Lovell et al., 2011; Barr and Clark, 2012a; Fredin et al., 2012; Darvill et al., 2014), is not only an indication of where features were deposited (as outlined in section 3), but also reflects their preservation and ease of identification. A number of moraines have been lost from the record or are unobservable for a number of reasons. Here, the focus is on topographic controls on these factors.

4.1. Topographic controls on moraine preservation

The moraines identifiable in the modern landscape are only a small sample of those that have formerly existed, as the majority have been eradicated by post-depositional erosion or burial (see Gibbons et al., 1984; Kirkbride and Brazier, 1998; Kirkbride and Winkler, 2012). This removal of landforms has been referred to as erosional-censoring and limits the inferences that can be drawn from the moraine record (Kirkbride and Winkler, 2012). The older the record, the greater the proportion of moraines likely to have been censored (Kirkbride and Brazier, 1998), and the surviving sample is therefore skewed toward modern examples (Gibbons et al., 1984; Kirkbride and Winkler, 2012). The principal factors acting to censor moraines, either through erosion or burial, are summarised in Table 2; and a distinction is often made between self-censoring mechanisms, such as the removal of preexisting moraines by subsequent glacier advance, and external-censoring mechanisms where moraines are removed by nonglacial processes (Gibbons et al., 1984; Kirkbride and Brazier, 1998; Kirkbride and Winkler, 2012). The intensity of these processes and the impact they have on the moraine record is partly topographically controlled, and this is the focus here.

Table 2. Processes that result in the censoring (removal) of moraines (table modified from Kirkbride and Winkler, 2012)

	Process	Effect	Timescale
Self-censoring	Obliterative overlap	Later glacial advances of larger extent destroy moraines deposited during earlier, less extensive, advances.	10^1-10^5 yr Occasional
	Dead-ice melting	Moraine surfaces gradually lowered as ice cores melt (only applies to ice-cored moraines).	10^2-10^3 yr Continuous
External censoring	Fluvial erosion	Lateral migration of river channels removes moraines by undercutting. Particularly effective on large proglacial fans with multiple avulsing channels and high peak discharges.	10^1-10^3 yr Continuous
	Alluvial deposition	Moraines on glacier forelands are buried by proglacial aggradation. Effective at ice margins where glacial-to-proglacial sediment transfer is efficient.	10^2-10^4 yr Continuous
	Jökulhlaups	Wholesale destruction or burial of foreland moraines by exceptional glaciovolcanic (subglacial) floods. Partial censoring by smaller floods may strip surface sediment from moraines and modify preexisting forms.	Hours to days. Episodic, rare at global scale, but regionally occasional.
	Lake-drainage floods	Outburst floods from failed ice- and moraine-dammed lakes or overwash events of displaced water by avalanching into lakes. Erosion of proximal deposits, possible burial in more distal locations.	Hours to days. Episodic, rare at global scale, but regionally occasional.

Rock avalanches	Incorporation of moraines into moving avalanche and/or burial in avalanche runout zones.	Minutes. Rare, but regionally occasional.
Paraglacial slope processes	Erosion of unconsolidated, steep lateral moraine material that may be stripped down to bedrock. Interaction with fluvial/glacifluvial erosion possible. Downslope redeposition buries valley-floor moraines.	Repetitive over 10^1-10^3 yr Episodic, common

4.1.1. Self-censoring

Glacial advances can override, and remove, evidence of previous, less extensive ice extent through a process known as obliterative overlap (Gibbons et al., 1984; Kirkbride and Brazier, 1998; Kirkbride and Winkler, 2012). As a result of this process, Gibbons et al. (1984) suggested that from 10 episodes of glacier advance, only 2–3 moraines might be expected to be preserved per catchment (i.e., only 20–30% of former phases of ice advance will be reflected in the moraine record). This estimate is based on the assumption that the extent of glacial advance is random over time. However, as demonstrated in section 3.2, some evidence indicates that glacier extent decreased episodically (and nonrandomly) throughout the Quaternary in response to erosional feedback (Kaplan et al., 2009; Anderson et al., 2012).This evidence (for erosional feedback) would suggest that the number of moraine sequences identifiable in a given landscape is partly determined by the number of glacial cycles but also by the timing (in this sequence of cycles) of maximum ice extent. For example, if maximum ice extent occurred at the LGM, the sequence will be limited to moraines deposited at this time and since. By contrast, if maximum ice extent occurred earlier in the Quaternary (as in Patagonia), then a detailed and continuous sequence of moraines might be preserved (e.g., Fig. 2C). However, as noted in section 3.2, the effectiveness of erosional feedback is controlled by subglacial erosion, suggesting that the preservation of a detailed moraine record is likely to be promoted in regions formerly occupied by warm-based, dynamic glaciers that were not frozen to their beds (Kaplan et al., 2009). Given that glacier dynamics can vary from basin to basin, and certainly across mountain divides (Clapperton, 1993), this might result in detailed moraine records being preserved in one drainage basin or on one side of a mountain chain (formerly occupied by dynamic, erosive glaciers) but not in/on another (formerly occupied by cold-based, minimally erosive glaciers) (Clapperton, 1993; Kaplan et al., 2009). Thus, the view that moraine preservation is controlled by erosional feedback has merit, but a number of other factors are also likely to be important. For example, during a period of advance, glaciers occupying laterally confined, comparatively flat basins are likely to readily overlap and remove previously deposited moraines; whilst glaciers in wide, unconfined basins (e.g., in piedmont zones) may by slow-moving and able to extend laterally upon encountering geomorphological obstacles, leading to the formation of large, complex, tightly-nested moraine sequences (see Dugmore, 1989; Kirkbride and Winkler, 2012). This mechanism potentially contributes to the dominance of piedmont moraines in the modern record (see section 3.3.3), as features are not only preferentially formed in such locations but are also preferentially preserved.

4.1.2. External censoring

Following deposition, moraines become vulnerable to erosion and censoring by a number of external (i.e., nonglacial) processes (see Table 2). During deglaciation, valley side debuttressing and meltwater discharge (amongst other factors) generally result in landscape instability and low moraine preservation potential (Ballantyne and Benn, 1994; Ballantyne, 2002; Curry et al., 2006; Kirkbride and Winkler, 2012). This is particularly true of moraines deposited on, or in proximity to, steep valley slopes where rock avalanches and other slope processes may be frequent (Ballantyne and Benn, 1994; Ballantyne, 2002; Curry et al., 2006). Similarly, moraines in valley basins may be buried by aggrading outwash fans and sandur plains or eroded by glaciofluvial drainage (including jökulhlaups and lake-drainage events).Thus, external censoring of moraines is typically more intense in narrow, laterally confined valleys where there are few locations with high moraine preservation potential. Again, this might partly explain the apparent prevalence of piedmont moraines in the modern record (see section 3.3.3), as such landforms likely experience less intense external censoring. Attempts have been made to isolate the impact of censoring from moraine sequences (e.g., using age-difference curves or by creating decay curves of moraine ages), but these approaches often fail to unambiguously identify periods of moraine removal (see Kirkbride and Brazier, 1998; Kirkbride and Winkler, 2012).

4.2. Topographic controls on moraine identification

The record of mapped moraines is partly a reflection of those which are identifiable in a given landscape and depends on a number of factors, many of which relate (either directly or indirectly) to topography. The focus here is on the role played by topography in regulating moraine size (area and relief), surface slope and surface cover — each of which impacts identifiability.

4.2.1. Moraine size

In attempting to unambiguously identify moraines, their size (area and relief) is an important factor, with larger examples typically easier to identify in the field and, critically, from remote sensing sources (see Barr and Clark, 2012a; Napieralski et al., 2013). Three principal factors determine moraine size: (i) the lateral extent of glacier termini; (ii) the total supply of debris to glacier margins; and (iii) the degree of post-depositional modification.

Lateral extent of glacier termini: As outlined in section 3.3.3, the lateral extent of glacier termini is partly determined by topography, with narrow basins preventing the development of extensive glacier lobes; whilst larger glacier tongues, and associated moraines, can develop in unconfined sectors. Thus, larger moraines are likely to be deposited in lowland areas away from mountain centres. This is apparent in the case of piedmont moraines, which, according to the data set of 8414 moraines mapped by Barr and Clark (2012a,b), are generally larger than nonpiedmont examples (see Table 1). The formation of large moraines in piedmont zones is likely enhanced by the tendency for ice margin stability in such areas (see section 3.3.3), leading to the formation of composite, nested moraines as successive glaciers stabilise at a similar location (see Dugmore, 1989; Alley et al., 2007; Anandakrishnan et al., 2007; Koppes et al., 2010; Putnam et al., 2013).

Supply of debris to glacier margins: The total supply of debris to glacier margins is controlled by a number of factors but is generally promoted by temperate conditions where glaciers have steep mass balance gradients, and are therefore dynamic and erosive (Andrews, 1972; Kaser and Osmaston, 2002). Debris availability and transport to the ice surface is also important and is promoted by steep, unstable slopes flanking glaciers (see section 3.1.3). The entrainment of debris at glacier beds, and its subsequent transport, also provides an important supply of material to the margins. Such entrainment, elevation, and transport is likely to be most effective under warm-based conditions, which facilitate the development of debris-rich basal ice sequences (Sharp et al., 1994; Knight, 1997), debris-rich shear planes (Glasser et al., 1998), and injections of debris into basal crevasses (Rea and Evans, 2011). It is suggested that the entrainment of subglacial debris is particularly efficient on the adverse slopes of overdeepenings, where supercooling processes result in the formation of debris-rich basal ice (Cook et al., 2010) and debris is entrained and elevated via shear planes (Swift et al., 2002). As a consequence, some of the largest moraines may well form in front of temperate glaciers with terminal overdeepenings (see Swift et al., 2002; Cook and Swift, 2012). This assertion is often difficult to test, as terminal overdeepenings can be infilled and are sometimes difficult to identify in the modern landscape (Cook and Swift, 2012). Where glaciers have significant debris cover at their margins, they can sometimes lead to the formation of ice-cored moraines (Lukas, 2011). The cores of such deposits often continue to melt for decades to millennia after deposition (Carrara, 1975; Driscoll, 1980; Everest and Bradwell, 2003; Schomacker and Kjaer, 2007; Lüthgens et al., 2011). During this period, moraine surfaces are gradually lowered to a point where features may be difficult (or impossible) to identify (though the identification of such moraines might be promoted as thermokarst lakes develop on their surfaces — see section 4.2.3). This is effectively a form of self-censoring (Kirkbride and Winkler, 2012) (Table 2) and means that originally ice-cored moraines are likely to be underrepresented in the modern record

Post-depositional modification: The post-depositional modification of moraines can consist of censoring, surface degradation, or fluvial dissection. Censoring (as outlined in section 4.1) results in the complete removal of moraines and therefore has no specific impact on their size. By contrast, surface degradation will have some impact on moraine size (particularly relief) and is discussed in section 4.2.2. Perhaps the most important post-depositional factor in regulating moraine size is dissection, whereby fluvial drainage cuts moraines into segments, producing fragmentary moraine records (e.g., Lovell et al., 2011; Barr and Clark, 2012a,b; Darvill et al., 2014). Generally speaking, dissection is promoted in recently deglaciated landscapes where drainage pathways are poorly established, but is also likely to be more common in laterally unconfined basins where rivers have the opportunity to migrate and bifurcate. This dissection may make moraines difficult to identify, leaving them with indistinct plan-morphology when mapped individually (particularly when viewed from remote sensing sources), and only appearing arcuate (or to reflect the shape of former glacier termini) when viewed as a continuous sequence (e.g., Fig. 17).

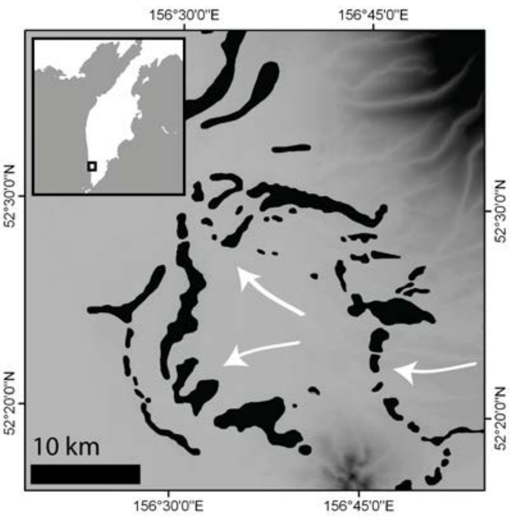

Fig. 17. Moraine sequences in southwestern Kamchatka. Individual moraine segments have indistinct plan morphology but appear arcuate when viewed as a continuous sequence. White arrows show inferred palaeo ice-flow directions. Mapping from Barr and Clark (2012a,b).

4.2.2. Surface slope angles

Slope angles are important when attempting to identify moraines, particularly from remote sensing sources (Kaufman and Calkin, 1988; Barr and Clark, 2012a). Recently deposited, fresh-looking moraines often have steep slopes, making them readily identifiable, whilst older examples are more subdued (Heiser and Roush, 2001; Kaplan et al., 2005; Hall et al., 2008; Putkonen et al., 2008). Part of this ability to identify moraines on the basis of their slope angles relies on the fact that they stand-out from the surrounding landscape. As a result, the general surface slope of a region is also important. For example, moraines occupying steep slopes may be difficult to identify, whilst those in flat areas might be prominent even when they have relatively shallow surface slopes. Moraines occupying steep topography (or surrounded by steep topography) might also be partially buried by rockfall and other slope processes and may, therefore, appear subdued (Warren and Hulton, 1990) or be difficult to identify from remote sensing sources because of local topographic shading. Each of these factors might favour the identification of moraines deposited on flat, open lowlands.

4.2.3. Surface cover

Moraine surface cover (particularly vegetation) typically varies with local climate and landform age (Birks, 1980; Matthews and Whittaker, 1987) and can be important when attempting to identify and differentiate moraines (Barr and Clark, 2012a). Specifically, moraines are often easier to identify when their surface vegetation is notably different from the surrounding landscape (Fig. 18A) or where the transition from recently deglaciated areas to areas that have been glacier-free for a prolonged period is marked by a shift from dense to sparse, varied to less-varied, and developed to less-developed vegetation (see Goward et al., 1985) (Fig. 18B). These botanical contrasts are often readily identifiable from satellite images (Figs. 18A-B) and may be particularly apparent in flat lowland sites where a contrast exists between poorly drained lowlands and well-drained vegetated moraines and where climatic conditions favour the development of a range of vegetation types. Thus, moraines in lowland environments may exhibit greater botanical contrasts and be easier to identify than their upland counterparts. In permafrost environments, the identification of lowland moraines might also be favoured by a contrast between thermokarst lake-dominated moraines and comparatively lake-free intervening topography (Fig. 18C). Such lakes occupy depressions formed as buried ice melts (Washburn, 1980) and rarely develop in steeper mountain topography where drainage is more efficient (Kääb and Haeberli, 2001). Again, these factors may make moraines deposited on flat, open lowlands easier to identify than upland examples.

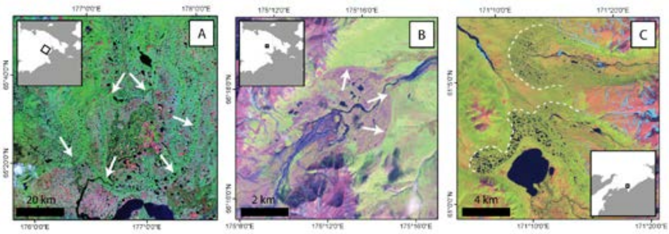

Fig. 18. Variations in moraine surface cover in eastern Siberia (displayed in Landsat satellite images). (A) End moraines (marked with white arrows) upon the Anadyr lowland, identifiable (at least partly) because they are colonised by vegetation that differs from the surrounding landscape. (B) Vegetation contrasts in the Pekulney Mountains, highlighting the transition from a recently deglaciated region (to the lower left of the image) to an area that has remained glacier-free for a prolonged period. This transition coincides with a moraine (marked by arrows). (C) Moraines and thermokarst lakes in the Koryak Mountains. The dashed white lines mark the transition from thermokarst lake-dominated to thermokarst lake-free environments, coinciding with the outer boundaries of moraines (former glaciers were generally flowing from right to left across the image).

5. Implications for palaeoglacier and palaeoclimate reconstructions

As noted in section 2.2, effort is often made to link past glacier dimensions to climate (e.g., Dyke and Savelle, 2000; Kerschner and Ivy-Ochs, 2008; Barr and Clark, 2012b). However, the moraine record upon which many reconstructions are based is not a reflection of palaeoclimate alone (Spedding and Evans, 2002; Swift et al., 2002; Kirkbride and Winkler, 2012; Lovell et al., 2012). In particular, topography exerts a significant control on moraine formation, preservation, and ease of identification (Warren and Hulton, 1990; Warren, 1991; Kaplan et al., 2009; Anderson et al., 2012). With this in mind, here we consider two fundamental questions: (i) exactly how important are topographic controls in governing the distribution of moraines in the modern landscape? and (ii) can uncertainty introduced by topographic controls (to palaeoglacier and associated palaeoclimate reconstructions) be mitigated?

5.1. How important are topographic controls?

Topography is likely to play some role (however small) in regulating the distribution of all mapped moraines, either through an impact on their formation, preservation, or ease of identification. However, the nature and extent of this control is likely to vary according to a number of factors and particularly in response to the type of glacier under consideration. As a result, here we discuss the importance of topographic controls by grouping glaciers into three different types (though in reality they are a continuum of forms): (i) ice sheets and ice caps; (ii) icefields and valley glaciers; and (iii) cirque glaciers. This discussion is summarised in Table 3.

Table 3

Importance of topography in governing moraine formation, preservation, and identification; listed according to ice-mass type[a]

		Controlling factor	Ice sheet	Ice cap	Icefield	Valley glacier	Cirque glacier
Topographic controls on moraine formation	Accumulation area topography	Glacier hypsometry	VL	M	VH	H	M
		Indirect accumulation	VL	L	M	H	VH
		Debris supply	VL	L	M	H	VH
		Evolving upland topography	VL	L	H	H	H
	Evolving basin topography	Erosional feedback	H	H	M	M	VL
	Topographic pinning points	Physical restrictions to advance	L	M	M	M	VH
		Bed slopes	L	L	M	M	L
		Valley width	M	M	H	H	VL
		Fjord topography	VH	M	M	M	VL
Topographic controls on moraine preservation and identification	Preservation	Self-censoring	M	M	H	H	VH
		External-censoring	H	M	H	H	VH
	Identification	Moraine size	H	M	M	M	H
		Surface slope angles	H	M	M	M	H
		Surface cover	H	M	M	M	M

[a]Classifications of importance: VL = very low; L = low; M = moderate; H = high; VH = very high.

5.1.1. *Ice sheets and ice caps*

Ice sheets and ice caps consist of central domes (accumulation areas) with outlets draining radially from ice divides and dispersal centres (Benn and Evans, 2010). A key characteristic of these ice masses is that, because of their considerable size (> 10,000 km^2), they often submerge the underlying landscape (at least in central sectors) and, as a consequence, flow directions are largely determined by the shape of the ice rather than the underlying topography (Sugden and John, 1976; Glasser, 1997; Golledge, 2007). Despite this, because outlets often occupy troughs, topography exerts some control on ice-mass dimensions and flow dynamics.

Controls on moraine formation: At the scale of ice sheets and ice caps, evolving basin topography (in cases where outlets are land-terminating) and fjord geometry (in cases where outlets are marine-terminating) are likely to be the dominant topographic controls on where in a landscape moraines are deposited; and other factors may be of limited importance. For example, because the central sectors of ice sheets and larger ice caps largely submerge the underlying landscape (sometimes with nunataks protruding through the ice surface), accumulation area topography (section 3.1) plays a limited role in regulating ice mass hypsometry (section 3.1.1), indirect accumulation (section 3.1.2), or debris supply (section 3.1.3). In scenarios where upland topography evolves, either during or between phases of glaciation (section 3.1.4.), this is again unlikely to have significant impact on the dimensions of ice sheets and ice caps because their accumulation zone altitudes are largely independent of the underlying topography (Fretwell et al., 2013). By contrast, because ice sheets and ice caps are often drained by dynamic and erosive outlets (Alley et al., 1986; Smith et al., 2007), evolving basin topography (section 3.2) ― specifically erosional feedback whereby ice masses become successively less extensive over time ― is likely to play a role in regulating where moraines are deposited over a series of glacial cycles. This assertion is supported by the data set of far-flung moraines compiled by Anderson et al. (2012) (Fig. 7), the majority of which (~ 67%) were deposited by ice cap outlet glaciers. Despite this, the role played by erosional feedback in governing moraine distribution is often difficult to assess, as the precise mechanisms responsible for any observed decrease in ice extent over time are often unknown and potentially involve a combination of erosional, climatic, and tectonic factors (Kaplan et al., 2009). Topographic pinning points (section 3.3) likely play some role in regulating where and when land-terminating outlets of ice sheets and ice caps deposit moraines but are key to regulating moraine deposition by marine-terminating examples. In the case of land-terminating outlets, physical restrictions to ice advance (section 3.3.1) or variations in bed slopes (section 3.3.2) are unlikely to be large enough to have any significant impact. Variations in valley width (section 3.3.3) may regulate where ice margins stabilise, promoting moraine deposition in piedmont regions, but the topographic independence of such ice masses is likely to limit the importance of this factor. By contrast, fjord topography (section 3.3.4) plays a key role in regulating the stability of marine-terminating outlets of ice sheets and ice caps — promoting the deposition of moraines in proximity to valley narrowings, bends, bifurcations, where basins are shallow, and/or in the vicinity of topographic bumps (Warren and Hulton, 1990; Warren, 1991, 1992). As such outlets retreat onto land, re-equilibration moraines can potentially be generated (see section 3.3.4), though establishing whether a moraine has been formed in this way is often difficult to conclusively judge (see Larcombe and Jago, 1994; Vorren and Plassen, 2002; Occhietti et al., 2011). In retreating from their former maximum extents, many large ice masses, such as the Laurentide Ice Sheet, terminated in a series of ice-marginal lakes (Teller et al., 2002), potentially leading to the formation of numerous re-equilibration moraines during deglaciation (Hillaire-Marcel et al., 1981).

Controls on moraine preservation and identification: As with other ice mass types, moraines formed by ice sheets and ice caps are subject to post-depositional processes that limit their preservation (section 4.1). However, the intensity of these processes is partly governed by topography. In particular, erosional feedback, whereby ice masses become less extensive over successive glacial cycles, will promote moraine preservation by limiting self-censoring through obliterative overlap (i.e., section 4.1.1). Because erosional feedback is likely to be at its most efficient beneath the dynamic outlets of ice sheets and ice caps (where ice velocities and total mass turnover are high), such ice masses might favour the preservation of detailed sequences of different-aged moraines (Anderson et al., 2012). In terms of topographic controls on moraine identification (section 4.2), ice sheets and ice caps often have extensive topographically unconstrained margins, allowing them to produce some of the largest moraines (section 4.2.1). In addition, moraines deposited by ice sheets and large ice caps often occupy lowland areas away from mountain centres (Figs. 2A-B), meaning that the intensity of external-censoring mechanisms (section 4.1.2), such as rock avalanching or paraglacial slope processes, is limited. Such moraines might also have surface slope angles (section 4.2.2) or surface cover (section 4.2.3), that differ from the surrounding shallow topography. For these reasons, moraines deposited by ice sheets and ice caps may be particularly easy to identify in the modern landscape. However, as a counter to this, many former ice sheets and ice caps terminated in regions that were submerged during post-LGM sea level rise, and their identification therefore relies on obtaining bathymetric data. This is particularly true of moraines deposited by marine-terminating glaciers or re-equilibration moraines deposited along former coastlines.

5.1.2. Icefields and valley glaciers

Icefields and valley glaciers are typically distinguished from ice sheets and ice caps by their size, but also by the way they interact with topography (Golledge, 2007). Icefields have upland accumulation zones from which outlet glaciers drain into surrounding basins. Valley glaciers can drain icefields or can be independent (with noncoalesced accumulation zones), discharging from upland cirques (Sugden and John, 1976; Benn and Evans, 2010). As a result of their comparatively restricted dimensions (i.e., < 1000 km^2), the morphologies and flow directions of icefields and valley glaciers are strongly controlled by topography (Sugden and John, 1976; Golledge, 2007; Benn and Evans, 2010), and this exerts some control on the distribution and characteristics of the moraines they deposit.

Controls on moraine formation: Numerous topographic factors influence the dimensions and flow dynamics of icefields and valley glaciers and thereby regulate moraine formation. For example (and in contrast to ice sheets and ice caps), accumulation area topography (section 3.1) exerts a strong control on ice-mass hypsometry (section 3.1.1) and effectively determines whether a region is occupied by icefields or a series of valley glaciers (Rea et al., 1998, 1999; McDougall, 2001) (Fig. 3C). In addition, a key characteristic of valley glaciers and icefield outlets is that they are overlooked by ice-free slopes, which provide important sources of indirect accumulation (section 3.1.2) and debris (section 3.1.3) to glacier surfaces. The supply of this material has an impact on glacier extent and dynamics and thereby impacts moraine formation (see section 3.1). Icefields and valley glaciers are also likely to respond to evolving upland topography (section 3.1.4), as uplift leads to more extensive glaciers (because of increased total accumulation) and vice versa. By contrast, the control exerted by evolving basin topography (i.e., erosional feedback) (section 3.2) is likely be less important than in the case of ice sheets and ice caps where total ice discharge and subglacial erosion are typically greater. This assertion is supported by the data set of far-flung moraines compiled by Anderson et al., (2012), with only ~ 33% deposited by valley glaciers (Fig 7). By contrast, and again because of ice-mass size, topographic pinning points (section 3.3) are likely to be more important in regulating the dimension and flow dynamics of icefields and valley glaciers than in the case of ice sheets and ice caps. Physical restrictions to ice advance (section 3.3.1) and variations in bed slopes play some role, and evidence certainly indicates that variations in valley width (section 3.3.3) are important (Lukas, 2007). This assertion is supported by the data set of Siberian moraines mapped by Barr and Clark (2012a,b), which indicates that piedmont moraines (deposited in wide, laterally unconfined piedmont zones) are heavily represented in the modern record. The importance of fjord topography (section 3.3.4) will vary regionally (e.g., a large proportion of Svalbard, Canadian Arctic, and Alaskan valley glaciers are currently marine-terminating), and examples of lake- and marine-terminating valley glaciers and icefields were likely more widespread during past glaciations.

Controls on moraine preservation and identification: As in the case of ice sheets and ice caps, moraines formed by icefields and valley glaciers will be subject to post-depositional processes that limit their preservation (section 4.1). Again, however, self-censoring will be partly mitigated by erosional feedback (though this may be less true than in the case of ice sheets and ice caps). External censoring (section 4.1.2), particularly rock avalanching and paraglacial slope failures, are likely to be important, as moraines are often deposited in upland areas, surrounded by steep and unstable slopes (Kirkbride and Winkler, 2012). In terms of topographic controls on moraine identification (section 4.2), a distinction can be made between moraines deposited in upland valleys and those deposited in open lowlands. The former are deposited by glaciers constrained by topography and are therefore likely to be small (because the lateral extent of glacier termini is restricted) (section 4.2.1) and surrounded by steep valley sides, meaning that moraine surface slope angles (section 4.2.2) and moraine surface cover (section 4.2.3) may be difficult to distinguish from the surrounding slopes. By contrast, moraines deposited in open lowlands (e.g., in piedmont regions) may be comparatively large (as a result of limited restriction to the lateral extent of glacier termini), experience little topographic shading, and have slope angles and surface cover that stand out from the surrounding lowland topography.

5.1.3. Cirque glaciers

Cirque glaciers are defined by their size, location, and morphology. They are typically < 10 km^2 (Golledge, 2007) and are located in semicircular hollows, bounded upslope by steep headwalls. Some glaciers can be entirely confined by cirque topography (i.e., they are fully enclosed), but they can also be open in a downslope direction (Benn and Evans, 2010). Cirque glacier morphology is strongly controlled by the underlying landscape (Sugden and John, 1976; Golledge, 2007), and topography therefore plays an important role in regulating the location and properties of moraines they deposit.

Controls on moraine formation: The distribution and dimensions of cirque glaciers are strongly controlled by topography (Mills et al., 2009; Bendle and Glasser, 2012). For the purpose of this paper, a distinction is made between topoclimatic factors (such as aspect, which regulates the receipt of solar radiation or allows snow to accumulate in the lee of prevailing winds), which directly impact local climatic conditions, and topographic factors (such as ice and snow avalanching from surrounding peaks), which regulate glacier size without a direct impact on local climate. Topoclimatic factors are important for glacier development and thereby exert a control on moraine position but are not the focus of this paper (see section 3); however, topography can impact cirque glacier dimensions/distribution and moraine location in a number of other ways. For example, accumulation area topography regulates cirque glacier hypsometry (section 3.1.1) through an impact of where ice can accumulate and by regulating the supply of windblown and avalanched snow and ice (section 3.1.2) — the dominant sources of accumulation for many cirque glaciers (see Sissons and Sutherland, 1976; Dahl and Nesje, 1992). A significant supply of redistributed snow and ice can allow cirque glaciers to persist and to deposit moraines below the regional climatic ELA (Dahl et al., 1997), as can the supply of debris to glacier surfaces. Because cirque glaciers occupy marginal zones (i.e., where peaks extend just above the climatic ELA), evolving upland topography (section 3.1.4) can also have a considerable impact on their extent, with uplift allowing cirque glaciers to develop into valley glaciers, and subsidence leading to glacier recession and disappearance — with a clear impact on where moraines are deposited. By contrast, at the scale of cirque glaciers, evolving basin topography (section 3.2) is unlikely to exert a control on moraine distribution, as glaciers are often too small to initiate significant erosional feedback. The importance of topographic pinning points in regulating cirque glacier dimensions and moraine position is likely to vary according to the type of pinning point in question. For example, physical restrictions to ice advance (section 3.3.1) are likely to be very important — particularly in the case of topographic steps (cirque thresholds) and geomorphological barriers (preexisting moraines), which can regulate cirque glacier extent (Dahl and Nesje, 1992). By contrast, bed slopes (section 3.3.2) and variations in valley width (section 3.3.3) are likely to be of very limited significance because glaciers are so small. Similarly, very few cirque glaciers are lake-terminating, and none are marine-terminating, so fjord topography (section 3.3.4) is of very little importance.

Controls on the preservation and identification of moraines: Moraines formed by cirque glaciers are likely to be particularly susceptible to processes that limit their post-depositional preservation (section 4.1). For example, because erosional feedback is unlikely to be efficient for such small glaciers, moraines might be particularly susceptible to obliterative overlap (self-censoring) and are only likely to be preserved if deposited during the final stages of deglaciation. In addition, moraines deposited by cirque glaciers are likely to occupy and/or be surrounded by steep unstable slopes, making them particularly susceptible to external censoring (section 4.1.2). As a result, moraines deposited by cirque glaciers can be difficult to unambiguously identify in the modern landscape and are not always easy to differentiate from surrounding slopes or other similar landforms (e.g., landslide deposits) (Shakesby and Matthews, 1996). This is a particular problem when identification is based on remote sensing sources alone, which often have coarse resolution relative to moraine size (Napieralski et al., 2013).

5.2. Can uncertainty introduced by topographic controls be mitigated?

As noted in this paper, the influence of topography brings the assumption that moraines can be readily used as indirect proxies for palaeoclimate into question. However, in theory, much of the uncertainty introduced by topographic factors can be mitigated by only obtaining palaeoclimatic data from three-dimensional reconstructions of palaeoglacier form. This approach has validity in that glacial geomorphology and land surface topography are taken into direct consideration to yield estimates of palaeo-ELA (e.g., Benn and Ballantyne, 2005; Rea and Evans, 2007; Finlayson et al., 2011). However, in reality, even when such reconstructions are generated, topographic factors continue to introduce uncertainty and limit the validity of resulting palaeoclimatic data. One key limitation stems from the fact that palaeoglacier reconstructions often rely on directly dating a small number of moraines and extrapolating these age-estimates to a wider population of landforms (e.g., Finlayson et al., 2011; Bendle and Glasser, 2012; Nawaz Ali et al., 2013). This approach relies on chronologically grouping moraines that are assumed to have been deposited synchronously. The processes involved in this grouping are rarely formalised but often involve consideration of moraine morphostratigraphy, location within the landscape, or relative position in a regional sequence (Kerschner and Ivy-Ochs, 2008; Barr and Clark, 2011; Heyman et al., 2011). However, this approach relies on assumptions about moraine deposition and preservation that, as demonstrated in this paper, are not always valid. For example, the moraine record can be censored, resulting in gaps in a regional sequence (Kirkbride and Winkler, 2012), or topographic controls can result in asynchronous moraine deposition in neighbouring valleys or regions (Fig. 4). As a consequence, fully considering topographic controls in this chronological grouping process is vital, and moraines used for reconstruction should be judiciously chosen (i.e., palaeoclimatic inferences should only be drawn from moraines when topographic controls on moraine distribution are considered insignificant) (Warren, 1991). An additional limitation of yielding palaeoclimatic data from three-dimensional glacier reconstructions is that in instances where topographic factors act to physically restrict ice advance (e.g., Fig. 11), where glaciers derive significant accumulated mass from redistributed snow and ice or where surface debris cover is significant, even robust ELA estimates derived from accurately reconstructed ice masses will be a poor reflection of palaeoclimate (i.e., glacier dimensions are not always a good reflection of palaeoclimate). This is a recurring problem in palaeoglacier reconstructions and demonstrates that, when using ELA estimates as proxies for palaeoclimate, care should be taken to select glaciers that appear to have received limited accumulation from redistributed snow and ice, with limited supraglacial debris-cover, and that appear to have been comparatively unrestricted in their advance (Burbank and Fort, 1985). One approach in attempting to isolate this topographic control on ELA is to obtain estimates from a number of glaciers (i.e., a synoptic view of moraine distribution should be taken) in different topographic settings within a region in the hope that the climatic ELA will be exposed (e.g., Bacon et al., 2001; Balascio et al., 2005) or to attempt to estimate the contribution of redistributed snow and ice to accumulation (e.g., Benn and Ballantyne, 2005; Ballantyne, 2007a,b). A final confounding factor in using palaeoglacier reconstructions as indicators of climate is that modern topography is not necessarily a reliable reflection of topography during, or before, glacial occupation, meaning that the role played by topography in regulating the extent of former glaciers (and, by extension, regulating moraine deposition) may not be apparent. For example, glacial overdeepenings are known to be efficient sediment sinks (Bennett et al., 2010) and may therefore fill during and after deglaciation, meaning that they are sometimes unidentifiable in the modern landscape (Cook and Swift, 2012). In such cases, the control exerted by overdeepenings on moraine distribution may not be evident. This demonstrates some of the fundamental difficulties associated with identifying, quantifying, and mitigating topographic controls on moraine distribution, and ideal solutions to many of these difficulties are currently lacking. The approach advocated here is to consider landform context (e.g., topographic setting) and to judiciously select moraines before making direct links to palaeoclimate (highlighting any associated uncertainties in reconstructions).

6. Conclusions

In this paper, topographic controls on moraine formation, preservation, and ease of identification are discussed, and the assumption that moraines can be readily used as indirect proxies for palaeoclimate is challenged. Some of the key points to be drawn from this review are the following:

- Topography has the capacity to regulate where and when moraines are deposited. This effectively represents a control on glacier dimensions (areal extent), dynamics, and margin stability (i.e., where and when glacial still-stands occur). Topographic factors that potentially have direct implications for the location and timing of moraine deposition include: accumulation area topography, evolving basin topography, and topographic pinning points.
- Topography not only exerts a control on where moraines are formed, but also regulates their preservation and ease of identification. The factors that influence moraine preservation are self and external censoring, and the intensity of these processes is partly determined by topographic controls on ice extent (in the case of obliterative overlap) and on the location of moraines within the landscape. Topography plays a role in regulating moraine identifiability through a control on moraine size (area and relief), surface slope, and surface cover.
- The importance of topographic controls in regulating moraine formation, preservation, and ease of identification varies spatially, temporally, and as a function of ice mass type. In particular, in the case of ice sheets and ice caps, one potentially important topographic control on where in a landscape moraines are deposited is erosional feedback, which can lead to the gradual reduction in ice extent over successive glacial cycles. For the marine-terminating outlets of such ice masses, fjord geometry also exerts a considerable (and perhaps dominant) control on ice margin stability and regulates where moraines are deposited. Moraines formed by such ice masses are likely to be large and readily identifiable in the modern landscape. In the case of icefields and valley glaciers, erosional feedback may well play some role in regulating where moraines are deposited, but other factors, including variations in accumulation area topography and the propensity for moraines to form at topographic pinning points, are also likely to be important. This is particularly relevant where land-terminating glaciers occupy piedmont zones (unconfined plains, adjacent to mountain ranges) where large and readily identifiable moraines can be deposited. In the case of cirque glaciers, erosional feedback is less important, but factors such as topographic controls on the accumulation of redistributed snow and ice, the availability of surface debris, and physical barriers to ice flow are likely to regulate glacier dimensions and thereby determine where moraines are deposited. In such cases, moraines are likely to be small and particularly susceptible to post-depositional erosion — sometimes making them difficult to identify in the modern landscape.

- Where links between moraines and palaeoclimate are made, some of the complexities introduced by topography can by mitigated by generating robust three-dimensional reconstructions of palaeoglaciers, from which ELA estimates may be derived (and linked to climate). However, these reconstructions still rely on assumptions about where and when moraines are deposited. For example, moraines are often chronologically grouped on the basis of their spatial distribution. This chronological grouping (or correlating) should be undertaken with caution (as suggested by Kirkbride and Winkler, 2012), and spatial correspondence (e.g., where moraines are found to lie at similar altitudes or in similar positions within a sequence) should not be assumed to reflect depositional synchrony. In some instances, even accurate reconstitutions of former glaciers will be unrepresentative of regional palaeoclimate, and the moraines used for palaeoenvironmental reconstruction should therefore be judiciously selected (i.e., palaeoclimatic inferences should only be drawn from moraines where topographic controls on moraine distribution are insignificant).

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