ISSN 2047-0371

3.4.9 Glacier Reconstruction

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Glacier reconstruction typically aims to establish the former extent of ice masses at any given period. Such reconstructions are important because they provide crucial information about past (palaeo) glacier changes over much longer timescales than the observational record permits. Reconstructing the dimensions and dynamics of palaeo-ice masses enables equilibrium line altitudes, and temperature or precipitation to be calculated, making glaciers an important palaeo-climate proxy. Given this utility, geomorphologically-based glacier reconstructions have been generated for many regions globally, although the specific methods employed are rarely described formally. To address this shortcoming, this chapter describes some of the methods employed in generating geomorphologically-based reconstructions for ice sheets and mountain-scale glaciers (< ~1,000 km^2).

KEYWORDS: Equilibrium Line Altitude (ELA), Glacier reconstruction, Geomorphology, Ice sheet, Palaeo-climate

Introduction

Glacier reconstruction involves the establishment of past (palaeo) changes in glaciated environments. Objectives of this research are typically to reconstruct glacier or ice sheet dimensions and dynamics for a particular point in time, analyse rates of ice-mass fluctuations, and better understand past climates (Ballantyne, 2007; Clark et al., 2012). Such reconstructions are important because they allow us to study glacier behaviour at much longer timescales than covered by the observational record (i.e., thousands vs. tens of years). This provides a better understanding of glacier and ice sheet response to past environmental changes, which, in turn, enables improved predictions of their behaviour in the future. Furthermore, geomorphological information about palaeo ice-mass dimensions and dynamics serve as key constraints for numerical ice sheet models (Stokes and Tarasov, 2010).

Reconstructions of mountain-scale glaciers (i.e. ice-masses <~1,000 km^2. Golledge, 2007) are often generated for the specific purpose of estimating former equilibrium line altitudes (ELAs), which represent a theoretical altitudinal zone where accumulation is balanced by ablation (Benn and Evans, 2010). Empirical data demonstrate that the ELA is determined by the relationship between air temperature and precipitation, with the ELA rising in response to decreasing snowfall and/or increasing temperatures, and vice versa (Ohmura *et al.,* 1992). This relationship means that if palaeo-temperatures are known independently, ELA estimates can be used to calculate former precipitation totals (e.g.

Ballantyne, 2007; Carr and Coleman, 2007), making derivation of ELAs a powerful tool for quantifying palaeo-climates. If analysed on a regional scale, palaeo-ELAs can be used to infer palaeo-precipitation gradients, allowing moisture sources and atmospheric circulation patterns to be reconstructed (e.g. Sissons and Sutherland, 1976; Sissons, 1979; 1980).

Given this utility, glacier reconstructions are widely utilised, but there is no standardised methodology for reconstructing glaciers from geomorphological evidence. Consequently, this chapter provides a framework that can be used as *guidance,* and considers methods for reconstructing the dynamics of past ice sheets, and the dimensions and ELAs of past mountain-scale glaciers.

Glacial geomorphological mapping

The type of ice-mass under consideration (e.g. ice sheet vs. mountain glacier) typically governs the specific methods employed. However, the first step in any reconstruction is the production of a geomorphological map (see Chapter 2, Sec. 6; Otto and Smith, 2013). Observations of contemporary glaciers reveal that they produce a variety of landforms, and one of the major challenges in geomorphological mapping is, therefore, the accurate recognition and delineation of these features. In light of this, it is essential that a detailed and robust geomorphological map is produced, prior to any glacier reconstruction (Figure 1A and 3A). Multiple lines of evidence should be mapped (e.g. glacial, periglacial, fluvial), which can then be used to infer glaciation style and extent (See plateau icefield landsystems in Rea and Evans, 2003; as well as McDougall, 2001; Boston et al., 2015). Landforms that are particularly useful for this purpose include ice-marginal features (e.g. moraines, terraces, fans and deltas) and sub-glacial bedforms (e.g. lineations and ribs).

Ice sheet reconstruction

The past extent, configuration, dynamics and retreat pattern of ice sheets can be reconstructed using a glacial inversion model - a set of assumptions that allow palaeo-glaciological inferences to be made from the geomorphological record (Boulton et al., 1985; Clark, 1999; Kleman et al., 2006). However, since the genesis (origin) of glacial landforms is often debated, these assumptions must be explicitly expressed when using this approach (Kleman and Borgström, 1996). Under the assumption that they are formed parallel to ice flow direction, subglacial lineations (e.g. drumlins and mega-scale glacial lineations) can be grouped into spatially coherent patterns known as *flow-sets* (Boulton and Clark, 1990; Clark, 1994; Stokes et al., 2009). Grouping lineations into flow-sets is based on their proximity, parallel concordance and morphological similarity (Clark, 1999; Figure 1B). Similar criteria can be applied to subglacial ribs, which form obliquely to ice flow direction (Hughes et al., 2014). A flow-set's dynamics can be determined from its geometry, bedform morphology and association with other glacial landforms (Table 1). Where the bedform imprint appears 'smudged', it is interpreted to have formed time-transgressively, behind a moving margin and/or beneath thickening/thinning ice. Clear, spatially coherent flow-sets are interpreted to form isochronously (Table 1).

Palaeo-ice streams (corridors of rapid ice-flow) are indicated by highly convergent flow patterns and elongate subglacial bedforms (Spagnolo et al., 2014; Margold et al., 2015). Conversely, subglacial ribs likely form under slower ice (Ely et al., 2016), which are thought to mark the transition between cold- and warm-based conditions (Hättestrand and Kleman, 1999), as well as zones of ice stream shutdown (Stokes et al., 2008). In this way, bedforms may provide an indication of basal thermal regime (i.e., whether cold-based, warm-based, or polythermal) and relative ice velocity. To account for situations where phases of cold-based ice have created a complex pattern of landform preservation, further subdivision of the landscape into 'glacial terrain zones' may be required (Trommelen et al., 2012).

Cross-cutting glacial bedforms record shifts in ice flow-direction during different phases, creating overlapping flow-sets (Clark, 1993; Livingstone et al., 2012). In such cases, a relative chronology of flow patterns can be built (Greenwood and Clark, 2009a; Hughes et al., 2014; Figure 1C). In addition, dated sediments or boulders can be used to constrain the relative chronology (Greenwood and Clark, 2009b; Clark et al., 2012). Moraines are taken to mark palaeo-ice margins, and their orientation can be used to infer retreat

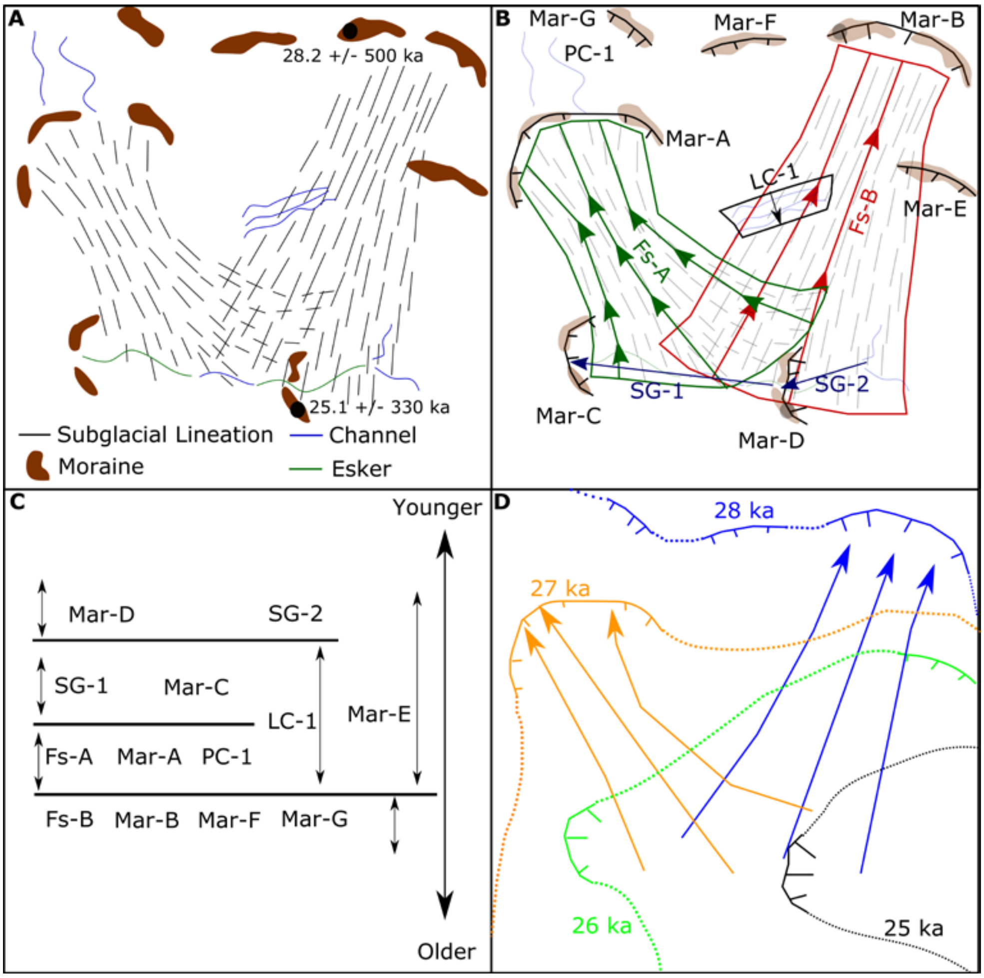

Figure 1: A) *Geomorphological map of a hypothetical study area. Points mark deglaciation dates of moraines. (B) Interpretation of landscape. Fs = flowset, Mar = margin, SG = subglacial assemblage, LC = lateral meltwater channels, PC = proglacial channels. (C) Construction of relative age sequence, based on cross-cutting relationships between landforms. For example Fs-A cross cuts Fs-B and therefore must be younger. Note, a lack of information means that some landform sets may fit into multiple relative stages (e.g. Mar-E). (D) Interpreted isochronous margins, constrained by the dates in (A). Note, that much of the margin is interpolated (dashed lines), and that other interpretations are possible due to a lack of absolute and relative dating (C).*

direction. Isochronous ice-margin positions can be constructed using relative and/or numerical dating methods. For example, dated ice margin positions are often interpolated to areas of adjacent moraines that lack corresponding dates using relative dating methods (Figure 1D). Landforms created by glacial meltwater are also used to reconstruct ice sheets. Eskers record patterns of channelised meltwater drainage, and their orientation is determined by the hydraulic potential gradient imposed by the ice sheet (Shreve, 1972; Storrar et al., 2014). They likely form time-transgressively, close to the ice margin, during deglaciation (Hooke and Fastook, 2007; Livingstone et al., 2015). Eskers can therefore be used to reconstruct ice margin retreat paths (Greenwood and Clark, 2009b). Meltwater channels require classification as either ‘subglacial’ or ‘lateral’ before use in ice sheet reconstruction (Greenwood et al. 2007; Table 2). Like eskers, subglacial meltwater channels form in relation to the hydraulic potential gradient imposed by the ice sheet (Sugden et al., 1991), but, unlike eskers, are not restricted to areas near the ice margin. Lateral meltwater channels form as water flows around the margin of ice-masses, and are common where cold-based ice prevents englacial or subglacial water flow. Lateral meltwater channels often form parallel sequences, indicating successive ice-margin positions during thinning (Greenwood et al., 2007). The occurrence of nested lateral meltwater channels is often used as a key indicator of the former presence of cold-based ice masses (Dyke, 1993; Kleman et al., 1997), although lateral meltwater channels can also form at temperate glacier margins (Syverson and Mickelson, 2009).

Table 1: Glaciodynamic classification of flowsets. Adapted from Hughes et al. (2014) and references therein.

Flowset Classification : Time-Transgressive

Bedform properties	Landform associations	Glaciodynamic context
Cross-cutting bedforms		Warm based ice
Low parallel conformity		Changing flow geometries a product of ice divide migration or outlet migration
Changes in bedform morphometry and distribution		
Bedform clustering and/or pattern changes around topography		Warm based ice Thinning/thickening ice
Bedform pattern corresponds to topographic obstacles	Eskers aligned with bedforms	Warm based ice
	End moraines	Thin ice Varying flow directions
Lobate pattern		Retreating wet margin
Few bedforms, predominately meltwater channels	Lateral meltwater channels conforming to topography	Topographically constrained Frozen Bed Thin ice Retreating dry margin

Flowset Classification : Isochronous

Bedform properties	Landform associations	Glaciodynamic context
No cross cutting	No aligned eskers	Warm based
Gradual downstream changes to bedform morphology	No end moraines	Stable flow configuration
Pattern ignores topography		Ice sheet interior flow
Parallel lineations		
Elongate bedforms	Shear margin moraines	Ice stream
Convergent flow patterns	Trough mouth fans	
Laterally confined bedform pattern		

Table 2: Criteria for classifying and interpreting meltwater channels, adapted from Greenwood et al. (2007) and references therein.

	Subglacial	Lateral		Proglacial	Supra/englacial
		Marginal	Submarginal		
Characteristics	Undulating thalweg	Parallel to contours		Meanders	Meanders on hill faces
	Steep chutes	Parallel sequences		Bifurcations	Gentle gradient
	Bifurcate	Low sinuosity		Flows downslope	Sinuous
	Sinuous	Found on valley sides		Wide and deep	Constant width
	Start and end abruptly	Terminate in chutes		Crater chains	
	Potholes	No networks	May be networked		
	Ungraded confluences	Gentle gradient	Steeper		
	Various sizes within the same system	Remain parallel for long distance	Change direction suddenly		
	Associated eskers	Abrupt terminations			
	No alluvial fans	May be isolated from other glacial geomorphology			
Interpretation	Follow subglacial hydraulic gradient and therefore approximate ice surface slope direction	Formed close to a cold based ice margin	Cold based ice Possible warm based transition where sudden changes of direction occur	Formed close to a warm based margin	Warm based ice

Mountain glacier reconstruction

Figure 2 outlines the reconstruction process and an example of its application is presented in Figure 3. Methods of reconstructing mountain-scale glaciers begin by using the mapped geomorphology to define the 2D extent (Figure 2, Step 1).

Two dimensional (2D) glacier extent

Table 3 provides a selection of landforms that can be used to define the lateral and vertical extent of the ice-mass. Typically, ice-marginal moraines (terminal and lateral) are the most useful and direct indicators of lateral extent. If these features are missing, or are heavily dissected due to paraglacial reworking, then ice-marginal fans, deltas, and kame terraces can be used, especially if they merge into moraine sequences. The lateral dimensions are then constrained by chronologically grouping the landforms, either through relative (e.g. landsystem) or direct (e.g. cosmogenic nuclide analysis) dating (Figure 3A).

Once lateral extents have been delimited, vertical dimensions can be reconstructed. In the lower reaches of glaciers (ablation zone), lateral moraines and drift limits are often used to constrain ice thickness (Lukas, 2006; Mills *et al*., 2012; Harrison *et al.,* 2010). The reconstruction process becomes more challenging in the upper reaches (accumulation zone), where vertical limits are based on tentative trimline evidence (See McCarroll, 2016), slopes covered in glacigenic deposits, or the density of glacially transported boulders (Lukas, 2006). Extrapolation and interpolation is often required in the absence of geomorphological evidence, in combination with surface profile modelling (Figure 3B).

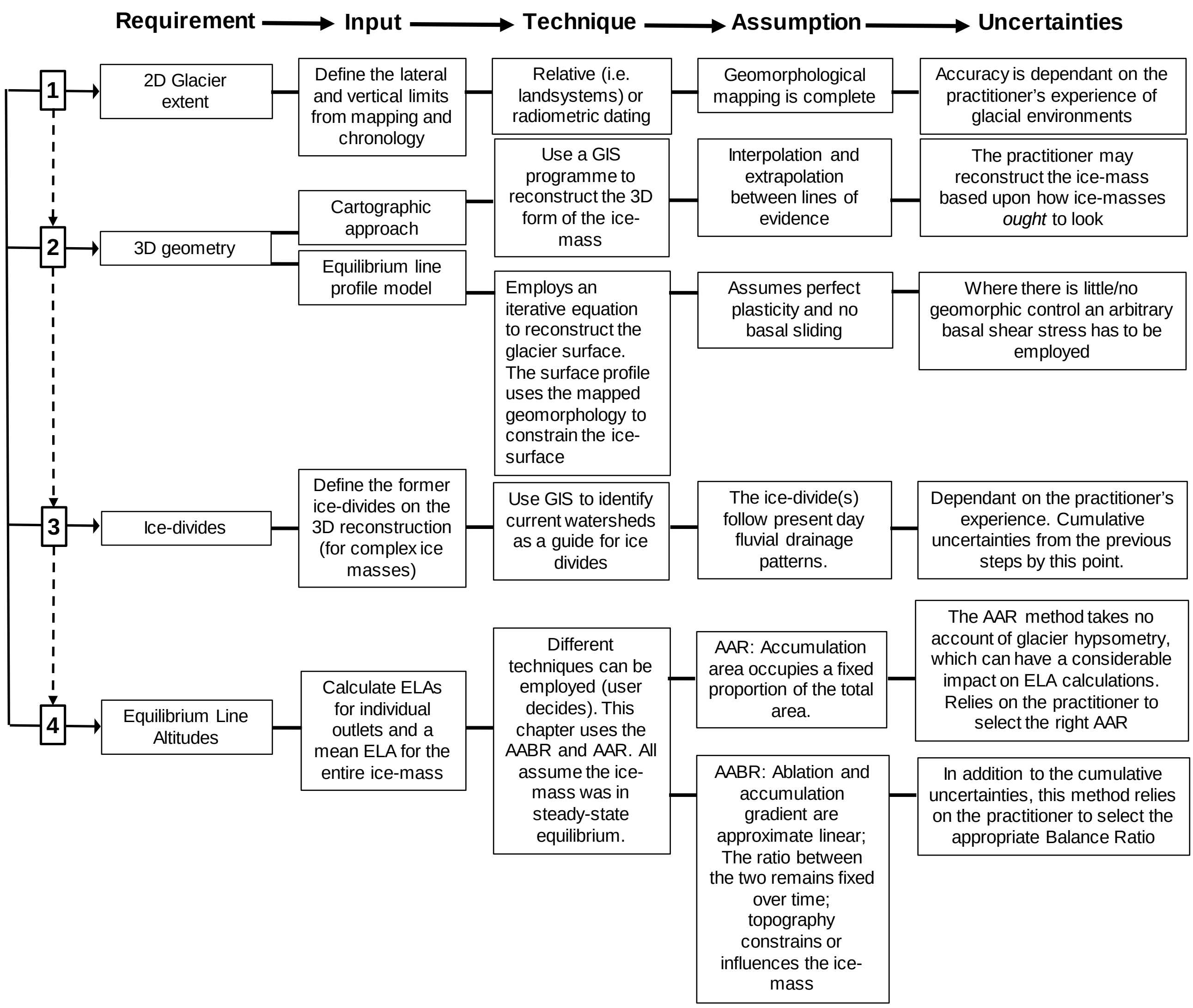

Figure 2: Suggested approach to reconstructing mountain-scale glaciation.

Table 3: Landforms that can be used to define the lateral and vertical dimensions of the palaeo-glacier. Adapted from Lukas (2006) and references therein.

	Landform/feature	Description	Indicates
Landforms to identify palaeo-glacier limits	Terminal (End) moraine	Ridge of unsorted sediment, often lying transverse to the land surface slope. May be fragmented.	Deposited along the frontal margin of glaciers. Indicate the former position of glacier terminus.
	Lateral moraine	Valley-side ridges of unsorted sediment, often lying sub-parallel to the land surface slope.	Deposited along the lateral margins of glaciers. Indicate former ice thickness.
	Drift/till limits.	The extent of thick glacial sediment. Boulders may be present.	Indicates the horizontal extent of former glaciation.
	Boulder limits	The spread of glacially-transported boulders, which may be present on valley sides, compared to adjacent slopes lacking boulders.	Indicates the horizontal extent of former glaciation.
	Trimlines (periglacial)	Boundary between smooth/polished bedrock inside the limits of former glaciers, and weathered bedrock and regolith outside.	*Possibly* indicates the horizontal extent of former glaciation. Note: Caution is required since trimlines can demarcate former thermal boundaries within the ice. Surrounding sediment-landform associations should be considered to aid interpretation (e.g. if a trimline 'merges' laterally into a drift limit , this suggests a former ice margin.
	Lateral meltwater channels	Fluvial channels (active or inactive), often sub-parallel to valley sides. Channels often start and end abruptly with no present-day source.	Formed around the margins of glaciers. Indicate former ice margin positions. Also important in identifying palaeo-plateau/summit icefields.
	Proglacial fans and deltas	Fan-shaped deposits, often with abrupt, formerly ice-proximal, slopes.	Formed at the margins of glaciers. Indicate former ice margin positions.
	Kame terraces	Flat-topped deposits, composed of sorted sand and gravel.	Formed at the margins of glaciers. Indicate former ice margin positions.
	Cirques	Semi-circular hollows, open in downslope directions, and bounded upslope by steep headwalls.	Cirques are sometimes assumed to have been completely filled with ice, with the upper ice limit generally following the highest most continuous contour of the cirque headwall or drawn to coincide with an elevation ~ 30 m below the top of the headwall.
Areas that remained outside glacier margins	Relict periglacial features	Various features, including frost-shattered rock, frost-weathered debris, scree and talus slopes, and solifluction lobes.	Indicate that a region was not ice covered. A sharp termination can potentially reflect former ice margin positions. Note: periglacial features have been preserved beneath cold-based plateau ice.
	Proglacial fluvial terrace.	Flat-topped fluvial terrace, often with abrupt up-valley termination.	Formed in a proglacial environment. Their up valley extent can indicate the former ice margin positions.
	River Terrace	Flat-topped deposits of an abandoned river surface, composed of sorted gravel, cobble and sand.	Up-valley extent often coincides with the position of a former ice margin.

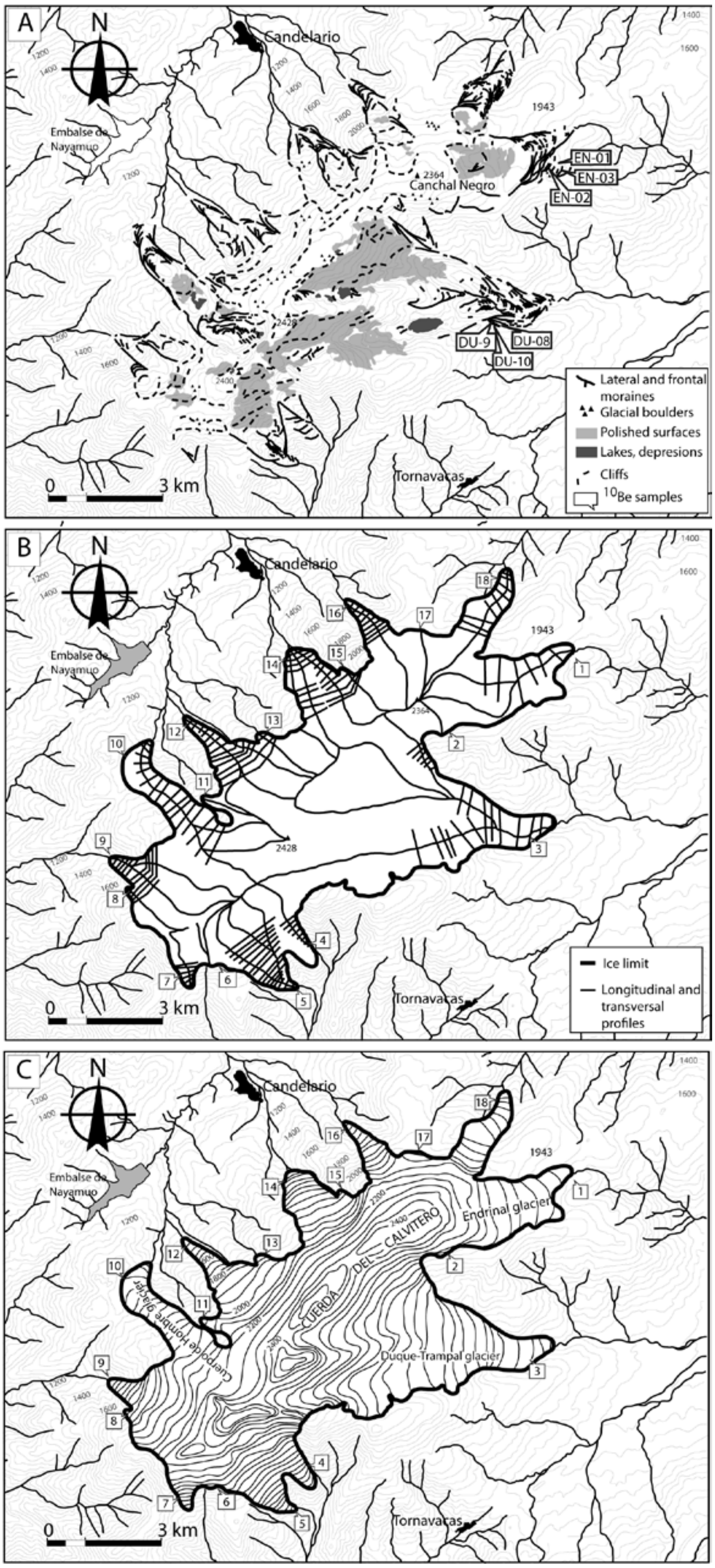

Figure 3: Example of the geomorphological approach to palaeo-glacier reconstruction illustrating the progression from (A) a geomorphological map, through (B) a reconstruction of the horizontal extent of former glaciation using a flowline model, to (C) a full three-dimensional palaeo-glacier reconstruction. Figure modified and redrawn from Carrasco et al. (2013).

Three dimensional (3D) geometry

Establishing the 3D geometry of the palaeo-glacier is important since subsequent reconstruction steps (including calculating ELAs) rely on an accurately constrained and contoured ice surface. Generating such a surface typically follows either a cartographic approach or by implementing a flowline model (Figure 2, Step 2).

The cartographic approach involves manually contouring the ice surface and ice-divides using drawing software (e.g. Adobe Illustrator) or a GIS (e.g. ESRI ArcMap). Contours are drawn through analogy with contemporary glaciers and, where ice flow direction is less clear, are drawn perpendicular to ice-flow direction indicators (e.g. ice-moulded bedrock, roches moutonnées and fluted moraines) (Sissons, 1980; Lukas and Bradwell, 2010). Generally, when viewed from above (i.e. in plan view), the lowest surface contour tends to be convex, bulging towards the glacier snout. Moving up-glacier, the contours become less convex, before becoming concave above the ELA, with increasing concavity towards the glacier headwall. Reconstructing contours and ice-divides using this manual approach has a long heritage (Sissons, 1974, 1980; Ballantyne and Andrews, 1989; Ballantyne, 2007; McDougall, 2001, 2013), but is largely dependent upon the quality of the landform record and the *individual's* understanding of present day glacier systems (used as analogues for palaeo-glaciers). The approach is better suited to valley glaciers, where margins are often clearly defined (e.g. Mills et al., 2012), since it becomes increasingly speculative for ice masses that are less topographically constrained (e.g. plateau icefields; Figure 3), or where geomorphological features are limited or lacking. In such circumstances, reconstructions often make use of simple flowline models, which help constrain ice surface elevations (Rea and Evans, 2007; Benn and Hulton, 2010; Pearce, 2014; Boston et al., 2015). The Excel spreadsheet program of Benn and Hulton (2010), based on the model of Schilling and Hollin (1981), is widely used for this purpose. This spreadsheet uses equation 1 to iteratively calculate ice-surface elevations along an inferred flowline up-valley from a glacier's terminus.

$$h_{i+1} = h_i + \frac{\tau_{au}}{s_i \rho g} \frac{\Delta x}{H} \qquad (1)$$

where *i* refers to the iteration number; τ_{au} is average basal shear stress; *s* is a shape factor (see below); Δx is step length (measured in metres); and *H* is ice thickness (Figure 4 shows this theoretical profile and coordinate system).

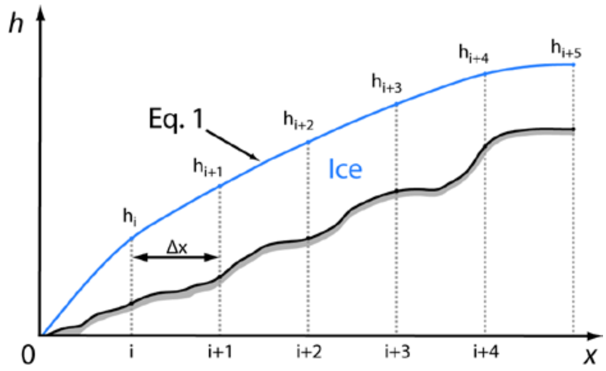

Figure 4: Example ice mass surface profile (and associated symbols) produced using an iterative flowline model, as given by Eq. 1. This model can be used to reconstruct ice masses of sub ice-cap scale (e.g. valley glaciers). i = denotes the first iteration with +n denoting the next cell in the calculation thereafter; Δx = step length (image adapted from Schilling and Hollin, 1981).

Solving equation 1 requires information about land-surface topography along a flowline (typically considered to coincide with a valley centreline) (Barr and Clark, 2011; Carrasco et al., 2013). This information can be readily extracted from a digital elevation model or topographic map. The user also needs to input a value for shape factor (*s*), which estimates the effect of lateral drag from the valley sides and is defined in equation 2.

$$s = \frac{A_r}{Hp} \qquad (2)$$

where A_r represents the estimated cross-sectional area of the reconstructed glacier, and *p* is the perimeter of the cross section at a given point on the glacier.

Hooke (2005) discusses the formula for valley-shape factor calculation in detail but early approximations followed Schilling and Hollin

(1981), where a value of 1 was used to represent an infinitely wide glacier (i.e. an ice-cap); 0.5, a semi-circular shaped channel, and 0.4; a cirque glacier. More recently, shape factors have been calculated from valley cross-profiles analysed at intervals considered to reflect changes in valley morphology (Rea and Evans, 2007; Pearce, 2014).

The flowline model can be implemented using a fixed value for average basal shear stress (typically 50–150 kPa; Monegato, 2012), but robust reconstructions rely on 'target elevations', derived from mapped geomorphological evidence (e.g. lateral moraines, drift limits, and/or trimlines) (Figure 5). The average basal shear stress can then be altered at relevant steps so that model output matches these 'target elevations' (Benn and Hulton, 2010). In this way, the model allows former surface elevations to be estimated along inferred flowlines. From this, the 3D ice-surface and ice-divides are typically reconstructed using a GIS, with ice-surface contours produced either physically or via geostatistical interpolation (spline or kriging, Figure 3C). In areas where geomorphological constraints on ice thickness are lacking, Boston et al. (2015) advocate the use of minimum and maximum thickness models to provide an estimate of uncertainty, which can be incorporated into ELA calculations through bracketing the range in ELA.

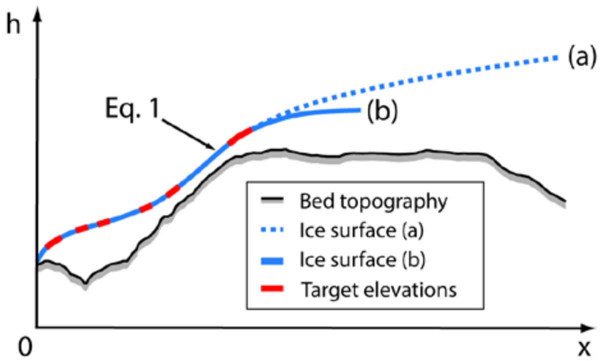

Figure 5: Profiles are constrained by geomorphological 'target elevations'. Profile (a) rises unrealistically on the plateau surface (at the ice divide), where 'targets' are lacking. In profile (b), the basal shear stress is attenuated to a value of zero the plateau (at the ice divide), producing more 'realistic' surface (figure based on Benn and Hulton, 2010).

Calculating palaeo-equilibrium line altitudes (ELAs)

A number of techniques are currently used to estimate ELAs of reconstructed glaciers (Meierding, 1982; Benn and Gemmell, 1997; Osmaston, 2005; Nie et al., 2014). In a given study, the decision about which procedure to use is often based on previous investigations (and the desire to make direct comparison between studies), simplicity of application, the type of glaciers under consideration, and technique precision and accuracy. It is common to adopt a variety of methods in a single study, often employing the most regionally-consistent to quantify final ELAs (i.e. the method resulting in the lowest inter-glacier variability in ELA) (Benn et al., 2005; Owen and Benn, 2005). An alternative is to use the mean, or weighted mean, of the various ELA estimates (e.g. Refsnider et al., 2008; Hughes, 2009). The two most widely used approaches are discussed here are the Accumulation Area Ratio (AAR) (e.g. Meier and Post, 1962; Benn and Lehmkuhl, 2000; Nesje and Dahl, 2000), and Area-Altitude Balance Ratio (AABR) (e.g. Furbish and Andrews, 1984; Benn and Lehmkuhl, 2000; Osmaston, 2005; Rea, 2009) both of which can be calculated using automated GIS tools (Pellitero et al., 2015). However, it is suggested that practitioners have an understanding of, and sometimes use, alternative methods to evaluate inter-method variability and allow comparison between studies.

The Accumulation Area Ratio method (AAR)

The AAR method assumes that the accumulation area of a steady-state glacier represents a fixed proportion of its total area. This is recorded as an accumulation area ratio (AAR; see Eq. 3), which typically lies in the 0.5–0.8 range, with 0.6 considered characteristic of cirque and valley glaciers (Meier and Post, 1962; Porter, 2001; Nesje, 1992; Chandler and Lukas, 2017).

$$AAR = \frac{A_c}{A_c + A_b} \qquad (3)$$

where A_c and A_b are the surface area of the accumulation and ablation zones, respectively.

This method often requires the user to select an 'appropriate' AAR for the type of glacier reconstructed. However, an alternative is to calculate ELAs for individual glaciers or ice catchments (where a coalesced ice mass is considered), using a series of AARs (between 0 and 1), then selecting the one which shows the lowest ELA variation between glaciers/ice-catchments (Hughes et al., 2007, 2011; Carrasco et al., 2013). A final option is to use a size specific AAR (ssAAR) for each glacier (Makos et al., 2012, László et al., 2013). This approach assumes a positive relationship between AAR and glacier surface area (outlined in Eq. 4), and is considered applicable for valley and cirque glaciers in the 0.1–40.0 km^2 size range (Kern and László, 2010).

$$ssAAR = 0.0648 \times \ln A + 0.483 \qquad (4)$$

where A is glacier area in km^2

The AAR method has been shown to produce consistent results (Porter, 2001), and the specific choice of AAR has limited impact on resulting ELA estimates (assuming a range of 0.5–0.8). For example, Balascio (2003) found that changing the AAR by ± 0.1, results in a ± 35 m variation in ELA. The major weakness of the method is that it takes no account of glacier hypsometry (the altitudinal distribution of glacier surface-area) (Osmaston, 2005), which can vary significantly from glacier-to-glacier and have a considerable impact on ELA estimates (Vieira, 2008; Rea and Evans, 2007). For this reason, it is often used in conjunction with the AABR method.

Area-Altitude Balance Ratio method (AABR)

The AABR method takes explicit account of both glacier hypsometry and altitudinal variations in mass-balance (Furbish and Andrews, 1984; Rea, 2009). Implementing the method for the purposes of palaeo-ELA estimation involves selecting an 'appropriate' BR, which reflects the ratio of the net mass balance gradient in a glacier's accumulation zone to the net mass balance gradient in its ablation zone (Eq. 5) (Osmaston, 2005).

$$\mathrm{AABR} = \frac{b_{\mathrm{nab}}}{b_{\mathrm{nac}}} = \frac{Z_{\mathrm{ac}} A_{\mathrm{ac}}}{Z_{\mathrm{ab}} A_{\mathrm{ab}}} \qquad (5)$$

Where b_n is the net mass balance gradient for the accumulation ($_{ac}$) and ablation ($_{ab}$) areas; Z is the area-weighted mean altitude (measured positively from the ELA) of the accumulation and ablation areas; A is the surface area of both the accumulation and ablation areas.

When using the method to estimate former ELAs, AABRs in the 1.67–2.2 range have typically been employed (Benn and Lehmkuhl, 2000; Benn and Ballantyne, 2005; Ballantyne, 2007). Until recently, 'representative' values were based on observations of only a small number of modern glaciers (Benn and Lehmkuhl, 2000), but Rea (2009) conducted a systematic analysis of modern AABRs from a global dataset of 66 glaciers, and found 'representative' values to be: 1.75 ± 0.71 for the total global dataset; 1.9 ± 0.81 for mid-latitude maritime glaciers; and 2.24 ± 0.85 for high-latitude glaciers. This dataset has been used in a number of recent studies (e.g. Santos-González et al., 2013; Boston et al., 2015) and currently serves as the most reliable source of information about appropriate AABRs. As with the AAR method, one technique for selecting an 'appropriate' AABR is to analyse a number of palaeo glaciers in a region, and use the values which show the lowest inter-glacier variability in ELAs. Again, however, the specific choice of AABR has been found to have a limited impact on calculated ELA estimates (Bendle and Glasser, 2012; Carrasco et al., 2013).

Linking the ELA to palaeo-climate

Though glacier response to climate is often complex and non-linear, established links between glacier ELAs and climate are often used to derive palaeo-climatic information from the 3-D glacier geometry (e.g. Sissons, 1979; Ballantyne, 1989, Lukas and Bradwell, 2010; Boston et al., 2015). In theory, glacio-meteorological models are the best tool for this purpose (Kuhn, 1981, 1989; Kaser, 2001; Kaser and Osmaston, 2002), but rely on assumptions about numerous variables (e.g. transfer coefficients for latent and sensible heat), which are typically unknown from the palaeo-record (Kerschner and Ivy-Ochs, 2008).

As a result, a number of simple, but effective, techniques are commonly used and, although a thorough review of these is beyond the scope of this chapter, these include: (1) the use of empirical precipitation/temperature (P/T) relationships; (2) the use of simple degree day models; and (3) the analysis of ELA depression relative to present (ΔELA). In some investigations, more than one of the above techniques may be used (e.g. Bendle and Glasser, 2012), allowing important methodological comparisons.

Conclusion

Geomorphologically reconstructed ice sheets and glaciers have a long heritage within the palaeo-community. Not only do they allow us to further our understanding of palaeo-ice dynamics but the reconstruction of mountain-scale glaciers also has the potential to yield ELAs and *quantitative* estimates of palaeo-climate, across remote areas where other sources of proxy data are often lacking.

Rapid advances have been made in this research area, largely due to the proliferation of geospatial data, which can be easily manipulated in GIS software and freely available spreadsheet-based models. Despite this progress, methods and sources of uncertainty are rarely explicitly stated. This chapter provides a *skeleton* framework, which can be used as guidance when reconstructing glaciers. Naturally, as reconstruction techniques continue to improve, some methods may become obsolete. However, the reconstructions *themselves* will remain important sources of information, often because field sites may never be revisited. Consequently, to ensure that these reconstructions can be utilised in the future, it is recommended that mountain-scale reconstructions involve detailed geomorphological mapping, the generation of 3-D reconstructions, and the calculation of ELAs (clearly detailing the methods used). Such data would provide important 'minimum benchmark' information, allowing reconstructions to be included in future studies.

Acknowledgments

We would like to thank Sven Lukas and Derek McDougall for their insightful and constructive reviews which improved this chapter.

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