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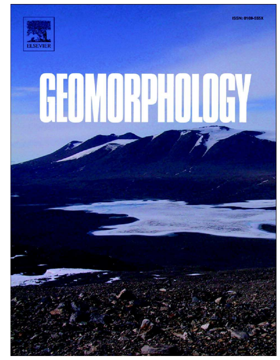

Short-term geomorphological evolution of proglacial systems

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Short-term geomorphological evolution of proglacial systems

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ABSTRACT

Proglacial systems are amongst the most rapidly changing landscapes on Earth, as glacier mass loss, permafrost degradation and more episodes of intense rainfall progress with climate change. This review addresses the urgent need to quantitatively define proglacial systems not only in terms of spatial extent but also in terms of functional processes. It firstly provides a critical appraisal of prevailing conceptual models of proglacial systems, and uses this to justify compiling data on rates of landform change in terms of planform, horizontal motion, elevation changes and sediment budgets. These data permit us to produce novel summary conceptual diagrams that consider proglacial landscape evolution in terms of a balance of longitudinal and lateral water and sediment fluxes. Throughout, we give examples of newly emerging datasets and data processing methods because these have the potential to assist with the issues of: (i) a lack of knowledge of proglacial systems within high-mountain, arctic and polar regions, (ii) considerable inter- and intra-catchment variability in the geomorphology and functioning of proglacial systems, (iii) problems with the magnitude of short-term geomorphological changes being at the threshold of detection, (iv) separating short-term variability from longer-term trends, and (v) of the representativeness of plot-scale field measurements for regionalisation and for upscaling. We consider that understanding of future climate change effects on proglacial systems requires holistic process-based modelling to explicitly consider feedbacks and linkages, especially between hillslope and valley-floor components. Such modelling must be informed by a new generation of repeated distributed topographic surveys to detect and quantify short-term geomorphological changes.

Keywords

paraglacial; periglacial; permafrost; glacier; geomorphology; landform; meltwater

Highlights

- Review of concepts and development of new holistic model
- Quantitative review of changes to landforms since Little Ice Age
- Presentation of emerging datasets and processing methods
- Knowledge gaps and pre-requisites for future process-based modelling

Contents

1. INTRODUCTION

Since the end of the Little Ice Age (LIA) c. 170 years ago, mountain glaciers and arctic glaciers have had retreating terminus positions (Grove, 2004; Matthews and Briffa, 2005). There are a few exceptions to this general retreat in some regions (e.g. Chinn et al., 2005; Hewitt, 2005). The general retreat of glacier termini by up to several kilometres in this relatively short time period, which has been accompanied by massive loss of ice volume, has been attributed to climate change. Both climate change and glacial volume loss have accelerated since the ~ 1980s (Zemp et al., 2009), especially in high mountain regions where a stronger warming trend is observed compared to the global mean (Zemp et al., 2015).

Glacier recession is associated with cryospheric changes (e.g. Haeberli et al., 2013), and the hydrological consequences of this ice mass loss (e.g. Moore et al., 2009) is most obviously dominated by a transient increase in runoff. However, soil development and ecological dynamics (i.e. the establishment and development of plant and animal communities) have also been extensively studied in recently deglaciated areas (Matthews, 1992; Hodkinson et al., 2003; Egli et al., 2006a, b; Moreau et al., 2008).

In contrast, geomorphological studies of deglaciating glacier forefields have been few and far between, although with a notable increase in recent years (e.g. Heckmann et al., 2016a and references therein). This recent increase in attention is due to a growing realisation that these landscapes exhibit a range of properties that make them a highly relevant subject of study: First, they are unique natural laboratories where the successional development of soil and vegetation can be investigated; recent studies emphasise the interaction of soil and vegetation development and geomorphological dynamics (Temme and Lange, 2014; Eichel et al., 2013, 2015; Temme et al., 2016). Second, due to the high rates of geomorphological activity, they form an ideal location for the application of mapping and measurement methods to document and quantify changes, and to test models, within a comparatively short time period (e.g. Carrivick et al., 2013; Staines et al., 2015; Micheletti et al., 2015; Lane et al., 2016). Finally, they are sources of potential hazardous events and processes, frequently with off-site effects that affect downstream valley sections (e.g. Vilimek et al., 2005; Haeberli et al., 2016). Indeed Knight and Harrison (2009, p. 230) state that 'paraglaciation will become the most significant process controlling sediment supply and landscape change in the mid- to high-latitudes over the next few hundred years' (see also Knight and Harrison, 2014).

Glacier forefields have been termed "proglacial", or "ice-marginal" areas or environments, defined by their proximity to a mountain glacier terminus, or to an ice cap margin or ice sheet margin (Slaymaker, 2011; using the definition of Penck and Brückner, 1909). Slaymaker (2011) considered proglacial environments as equilibrium systems adjusted to a typical climatic, hydrological and geomorphological regime with characteristic landforms. Daily and seasonal meltwater periodicity, and high-magnitude-low-frequency episodes of glacial meltwater dynamics and the associated changes in fluvial dynamics and sediment delivery (e.g., Marren, 2005; Milan et al., 2007; Baewert and Morche, 2014; Mao et al., 2014; Leggat et al., 2015), suggest that this 'equilibrium' is highly dynamic. Additionally, changes in (sub-)glacial drainage pattern (Willis et al., 1996; Nienow et al., 1998; Rippin et al., 2003) cause proglacial system changes upon crossing of internal thresholds, changes to system structure (Fig. 1), and/or exceptional events that are independent of climate change and glacier recession. Thus it is rather more instructive to consider that the vast majority of glacier forefields have become transient systems (Lane et al., 2016; Fig. 1) in which increasing areas of freshly exposed rock walls, hillslopes, moraine and glacifluvial landforms adjust to non-glacial conditions. In this context, the recognition of landform transience means that proglacial systems set a stage for paraglacial dynamics, i.e. the activity of a suite of non-glacial processes that are directly conditioned by deglaciation (Church and Ryder, 1972; Ballantyne, 2002).

The concept of paraglacial geomorphology includes a temporal component that is strongly related to spatial scale. On a large spatial scale, some landscapes are presently still adjusting to the termination of the last Pleistocene glaciation (e.g. Church and Slaymaker, 1989). At a hillslope, or landform scale, this adjustment is expected to be much more rapid (cf. Slaymaker, 2011; see also Curry et al., 2006); therefore, the state of imbalance, or 'disequilibrium' until glacially-conditioned sediment stores are either exhausted or stabilised (= the 'paraglacial period'; Ballantyne, 2002) is of a shorter duration and theoretically diminishes in intensity with distance from a receding glacier. Several studies have stressed that intense geomorphological activity can be concentrated on, if not restricted to, those areas that have been deglaciated since the end of the LIA (e.g. Bosson et al., 2015); morphodynamics induced by this comparatively recent deglaciation have been highlighted as 'neo-paraglacial' by Matthews and Shakesby (2004).

Following these findings, we use the often conspicuous LIA moraines as the boundary of a proglacial system (see also Schiefer and Gilbert, 2007; Heckmann et al., 2012) and adopt a 'short term' (on the temporal scale of decades) perspective on proglacial geomorphological dynamics in this paper. The hypothesis that 'proglacial systems are rapidly changing' needs to be robustly tested spatially, by comparing rates of change *within* proglacial systems to those *beyond* the proglacial area. Furthermore, the rate of change of these processes depends largely on major climatic, topographic settings. Thus, given this global, regional (inter-catchment) and even local (intra-catchment) diversity (that also refers to glacier dynamics, see Winkler et al., 2010), there is a pressing need for (i) a protocol for definition of proglacial system extent, (ii) an inventory of landform types, and (iii) quantification of (the rate of change of) earth surface processes both *beyond* and *within* proglacial systems.

The aims of this paper are to review existing studies on geomorphological changes and landscape evolution in proglacial areas, predominantly within LIA moraine limits, to synthesise quantitative estimates of these changes, and to identify research gaps and propose future research avenues.

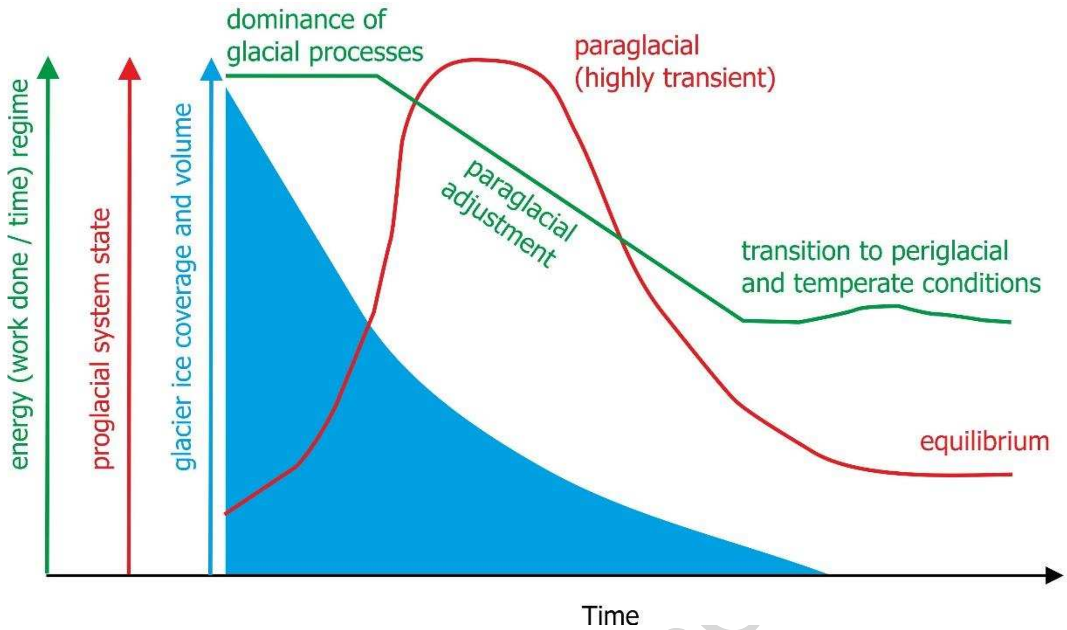

Figure 1: Conceptual model of proglacial system transition from domination by glacial processes, through a paraglacial period towards a periglacial and perhaps ultimately a temperate landscape. The non-linear decline of glacier ice coverage follows the results of GCM ensemble modelling by Radić et al., 2014. This diagram has been informed by Thorne and Loewenherz (1987), Slaymaker (2011), and Haeberli et al. (2016).

2. PREVAILING MODELS OF PROGLACIAL SYSTEM GEOMORPHOLOGICAL DYNAMICS

In this section, we review conceptual models that represent the effects of deglaciation on geomorphological processes. As a general framework, we use (i) an adaptation of Caine's (1974) concept of sediment systems (Fig. 2), and (ii) Benn et al.'s (2003) model of valley glacier landsystems (Fig. 3). While the former is focused more on sediment fluxes, the latter concentrates on geomorphology, i.e. the evolution of landforms and landform associations. Of course the two have common elements and complement each other.

We set LIA moraines and the present-day glacier extent as the boundaries of a proglacial system (cf. Schiefer and Gilbert, 2007; Heckmann et al. 2012; Carrivick et al., in review). We suppose that this area has experienced and continues to experience rapid geomorphological change. However, we concede that geomorphological evidence (moraine ridges) of previous (LIA) glacier extents may be ambiguous or obliterated (see Table 1 of Kirkbride and Winkler, 2012) and, as Barr and Lovell (2014) discuss, some caution must be exercised when relating moraines to former glacier dimensions.

Figure 2 depicts a proglacial system sediment cascade characterised by functionality (coarse and fine clastic sediments, and solutes; see also Fairchild et al., 1999; Etienne et al., 2003). The subsystems operate on different spatial and temporal scales. The corresponding sediment transfer pathways link sediment sources to storage landforms and to the river channel network. Detailed conceptual sediment budget models for the hillslope and river channel components were proposed by Caine (1974), Schrott et al. (2003) and Etzelmüller and Frauenfelder (2009), who all conceptualised the influence of (mountain) permafrost on the sediment transfer cascade system. They, in agreement with Benn et al. (2003), identified that the most important components of geomorphological systems in this context for determining landform assemblages are glacier thermal regime and permafrost.

Benn et al. (2003) conceptualised how topography, sediment supply to a glacier surface, and the efficiency of sediment transport from a glacier to its forefield control the development of its resultant landsystem associations (Fig. 3). A rather simpler alternative conceptual model has been suggested by Zemp et al. (2005). The landsystem associations can be placed in a continuum based on the proportions of ice and sediment supply, with (i) covered or uncovered glacifluvial outwash-head systems characterised by high ice supply, and (ii) moraine-dammed and rock glaciers representing the landform association end members. The configuration of landforms within such a landsystem has important implications for the degree of coupling between glacial and proglacial environments by the glacifluvial system (Benn et al., 2003), and thus with the efficiency of sediment cascades and the downslope and downstream transmission of geomorphological changes.

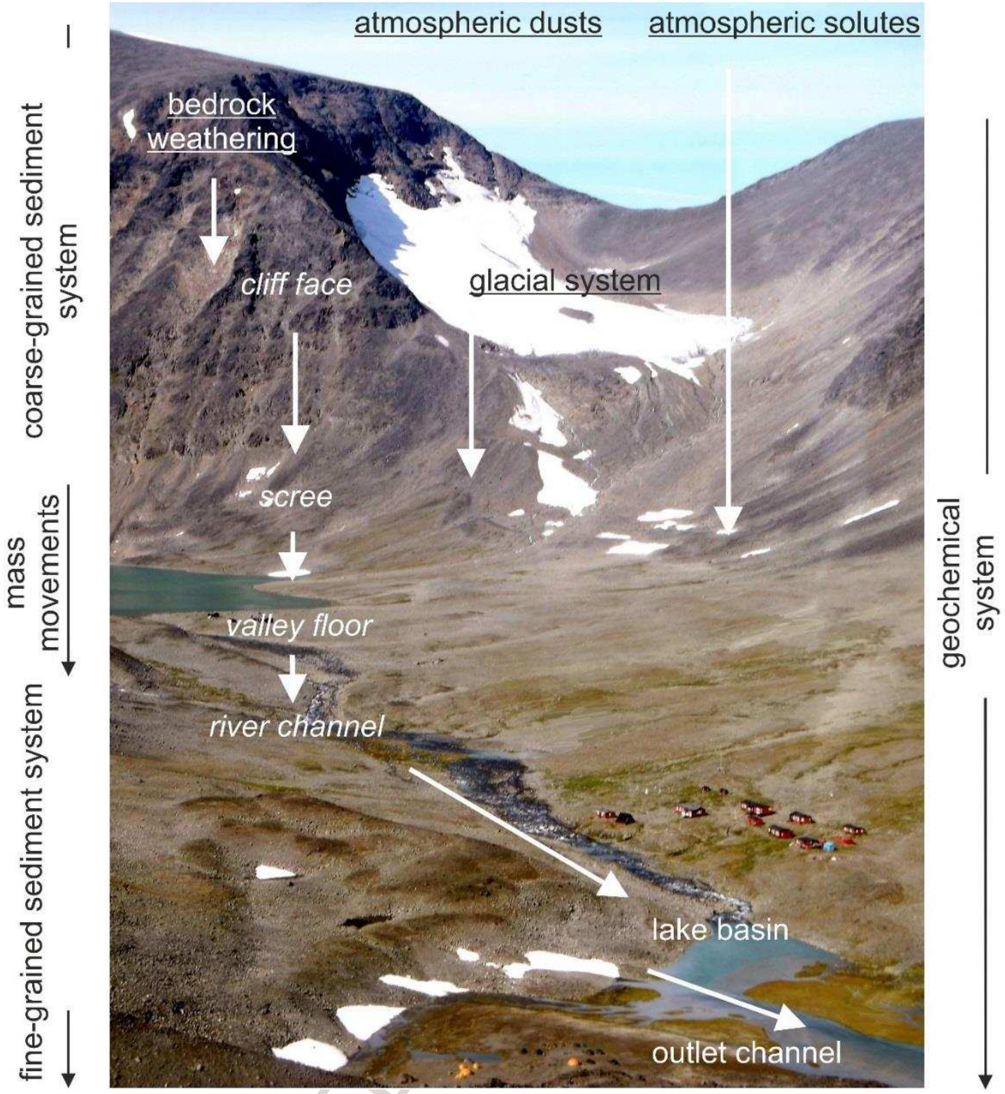

Figure 2: Field illustration of components of the sediment budget system in an alpine proglacial area, the upper Tarfala valley, arctic Sweden, following the model of Caine (1974) but with the addition of the 'glacial system' (cf. Barsch and Caine, 1984). Sediment system inputs are underlined and system transfers are in italics..

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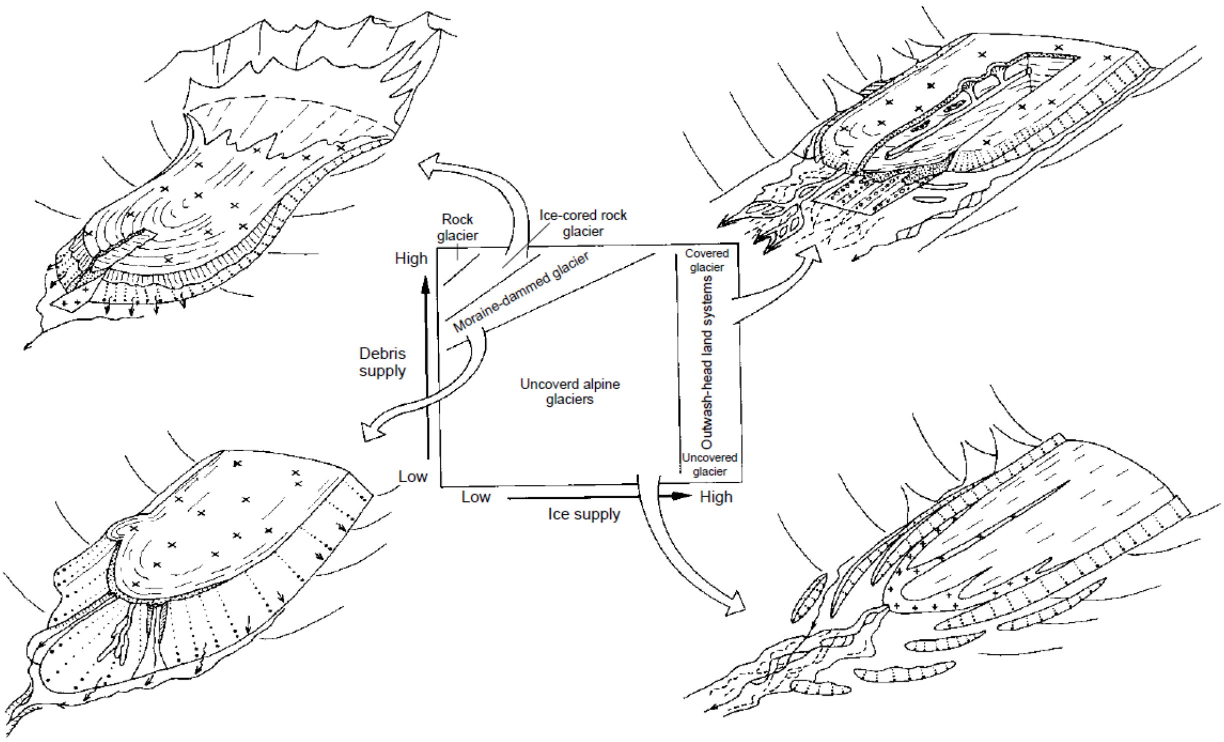

Figure 3: Conceptual relationships between constraints on landsystem development at valley glacier termini and developmental pathways into four landsystem associations. Source: Benn et al. (2003). Note that this model does not include hillslope or permafrost inputs to the proglacial system. This figure is a copy: it will need permission from publisher

2.1 Sediment sources

Three main sediment sources can be identified in a proglacial system: (i) sediment derived from glacial erosion and subglacial sediment storage, (ii) debris produced by the weathering and instability of deglaciated bedrock, and (iii) glacigenic debris covering hillslopes or accumulated in depositional landforms such as lateral and terminal moraines. Storage landforms derived from the re-working of other sediment sources, such as debris cones or fluvial terraces, can act as sediment sources themselves.

The first sediment source is linked to the erosional power of the glacier. Rates of (bedrock) erosion and sediment evacuation by glaciers have been shown to vary by orders of magnitude (0.01 mm yr^{-1} to 100 mm yr^{-1}), depending on the thermal state of the glacier and on lithology (Hallet et al., 1996, and references therein). Subglacial sediment storage has been thought to be small (Riihimaki et al., 2005) or quite substantial (e.g. Haeberli and Fisch, 1984; Zemp et al., 2005) but only a few measurements exist; Stocker-Waldhuber et al. (2015, in press) present a geophysics-based measurement from the Gepatsch glacier, documenting a subglacial sediment thickness of up to 20 +/- 5 m, and a rapid removal during a hydrometeorological extreme event (see also Baewert and Morche, 2014).

Second, deglaciated rockwalls steepened by glacial erosion are prone to mass movements triggered, for example, by glacial debuttressing, and include debris falls and large rock-slope failures (bergsturz). McColl (2012) gives an extensive review of such paraglacial rock-slope instability. Ballantyne and Stone (2013) dated 47 rock-slope failures in the Scottish Highlands and reported enhanced incidence immediately after deglaciation, decreasing to a more or less constant rate that has been maintained during the past c. 10 kyr (see also Ballantyne et al., 2014). Cossart et al. (2008) showed that c. 80 % of rock-slope failures mapped in a 1011 km² alpine study area were located below the Last Glacial Maximum (LGM) trimline. Holm et al. (2004) investigated slope instabilities following post-LIA glacier retreat in British Columbia; hillslopes in comparatively soft volcanic bedrock were markedly oversteepened and experienced intense rock-slope failures, whereas this was much less the case in the more stable granitic sections of the study area. Debris slides and debris avalanches were shown to dominate in deglaciated till-covered areas or colluvial deposits.

Such surficial materials on drift-mantled hillslopes and glacial depositional landforms (e.g. lateral moraines) form another very important sediment source. After deglaciation, they are subject to re-working by a wide range of geomorphological processes, including deformation (e.g., Hugenholtz et al., 2008), mass movements (e.g. Hürlimann et al., 2012; Oliva and Ruiz-Fernandez, 2015), debris flows on hillslopes (e.g. Curry et al., 2009; Haas et al., 2012) and in channels (Legg et al., 2014), and fluvial incision (e.g. Curry et al., 2006). Geomorphological dynamics and sediment transfer, and their development after deglaciation, have been described in conceptual models of paraglacial geomorphology. For example Ballantyne's (2002) exhaustion model and Church and Slaymaker's (1989) paraglacial cycle were discussed with reference to field measurements by Curry et al. (2006), Cossart and Fort (2008) and Friele and Clague (2009).

Depending on diverse factors, deglaciation can lead to localised patterns of aggrading and degrading permafrost (Kneisel and Kääb, 2007; Klug et al., 2016). Bosson et al. (2015) and Micheletti et al. (2015) stress the importance of the spatial distribution of (degrading) permafrost for landform dynamics in proglacial areas under conditions of climate change. Degradation of permafrost is especially associated with intense geomorphological activity, both in bedrock (Krautblatter et al., 2013) and in ice-cemented loose material (e.g. Bardou et al., 2011). Damm and Felderer (2013) concluded that almost half of mapped debris flows occurred in the area where permafrost or glacier ice disappeared since 1850, but only a minority of debris flows started within proglacial areas as defined in this review. Sattler et al. (2011) were sceptical about the direct linkage between unstable or decaying permafrost and debris flow activity. The linkage of permafrost and geomorphological processes in proglacial areas is obviously complex, encompassing multiple alternative or combining mechanisms (Bardou et al., 2011).

2.2 Proglacial rivers

Proglacial rivers have been in a focus of geomorphological research for a comparatively long time (Church and Gilbert, 1975; Maizels, 1983; Warburton, 1992). Their intense morphodynamics allow observation of diverse fluvial processes within short periods of time that lead to more general insights in gravel-bed river behaviour. The proximity to a glacier as a main source of meltwater implies a high temporal variability of discharge on multiple scales (Marren, 2005). Daily and seasonal periodicity is complemented by floods caused by meteorological conditions (e.g. temperature and precipitation), and exceptional glacier dynamics (e.g. Baewert and Morche, 2014). This variability is associated with a variability in sediment supply from different sources (glacier, sediment stores within the proglacial area) to the proglacial channel network (Hammer and Smith, 1983; Lane et al., 1996; Marren, 2005; Leggat et al., 2015). The potential of sediment derived from upslope and upstream sources to reach, for example, the outlet of the proglacial area, is termed sediment connectivity and is addressed in detail in subsection 2.4.

Changes in sediment supply and pathways, and effects of the sediment grain size distribution such as armouring (Ashworth and Ferguson, 1986; Warburton, 1992) lead to hysteretic behaviour of suspended sediment transport on different temporal scales (e.g. Hammer and Smith, 1983; Mao et al., 2014; Mao and Carillo, in press). Different studies have focused on suspended load (Gurnell, 1982; Willis et al., 1996; Orwin and Smart, 2004; Liermann et al., 2012; Leggat et al., 2015; Beylich et al., in press) or bedload (e.g., Warburton, 1992; Nicholas and Sambrook-Smith, 1998; Stott, 2002; Guillon et al., 2016). The measurement of sediment load in proglacial channels has been a primary source for estimations of glacial sediment budgets (cf. introduction of section 2). However, with the recent study by Beylich et al. (in press) being an exception, sediment delivery from hillslope sources is usually neglected because it is affected by non-glacial geomorphological processes, and by the undercutting and re-working of adjacent storage landforms. This is important for considering sediment budgets in proglacial areas because suspended sediment from non-glacial sources can dominate and that from glacial sources can show (much more) variability than that from non-glacial sources (Beylich et al. in press). That is not to say that glacial and non-glacial sources are entirely distinct though: Gurnell et al. (1999) highlighted that a reactive connectivity between a river and its surrounding valley sediments was one important characteristic of proglacial rivers.

Sediment transfer is strongly interrelated with river morphology. Typically, the conditions mentioned above lead to a braided river system. However, Maizels (1983) stated that areas experiencing long-term deglaciation had rivers in transition from predominantly braided, high gradient, low-sinuosity channels to single-thread, low-gradient deeper, more sinuous channels. Due to glacial meltwater discharge and a continuing sediment input by the glacier, proglacial systems (under equilibrium conditions?) are aggrading systems on the long term, experiencing cycles of aggradation and degradation. Incision and degradation are expected to dominate during and after deglaciation, respectively (Gurnell et al., 1999).

Marren and Toomath (2014) emphasised the role of topography (confinement by moraine breaches or fluvial terraces, reduction in slope in the upper part of the long profile) in the short-term response of proglacial channels to glacier retreat. This topographic forcing explains why proglacial channels may deviate from the 'expected' braided pattern. Milan et al. (2007), for example, analysed the amount and spatial distribution of scour and fill, and noted that whilst erosion and deposition varied considerably as discharge was driven by rainfall vs. periodic meltwater discharge, the ratio of erosion and deposition across the full length of the survey period was balanced. Other similar studies since (e.g. Carrivick et al., 2013) may be seen as some justification of the sediment budget approach to quantifying short-term changes in proglacial systems.

2.3 Proglacial lakes

Proglacial lakes are characteristic features of glacially formed landscapes. They can be classified according to their formation and their location relative to a glacier margin (Tweed and Russell, 1999; Clague and Evans, 2000; Carrivick and Tweed, 2013). Recent deglaciation has produced many new proglacial lakes in alpine settings (see list of references in Carrivick and Tweed, 2013) and on ice sheet margins (e.g. Carrivick and Quincey, 2014). Besides providing feedbacks affecting glacier mass loss (Carrivick and Tweed, 2013), they can affect geomorphological activity in the proglacial system and beyond, above all by buffering meltwater discharge dynamics and acting as effective sediment traps (e.g., Liermann et al., 2012; Geilhausen et al., 2013; Bogen et al., 2015). The sedimentary infill of proglacial lakes has been used as an archive from which to reconstruct past glacial, sediment, and ecological dynamics (e.g. Hicks et al., 1990; Hasholt et al., 2000; Loso et al., 2004; Schiefer and Gilbert, 2008; Xu et al., 2015). Proglacial lakes are a source of natural hazards (Werder et al., 2010; Frey et al., 2010; Schneider et al., 2014; Haeberli et al., 2016), which can deliver enormous amounts of sediment down valley.

2.4 Sediment connectivity

Sediment connectivity is a system property that describes the degree to which system components such as hillslopes, landforms and river channel reaches are coupled with respect to sediment transfer (Hooke, 2003; Brierley et al., 2006; Heckmann and Schwanghart, 2013). This encompasses both the lateral (i.e. hillslope to channel) and longitudinal (i.e. within channel, between channel reaches) transfer of sediment (Brierley et al., 2006), and is driven by water and gravity (Bracken et al., 2015).

Figure 4 depicts the components of a proglacial system together with sediment pathways and typical elements that effect (dis-)connectivity. In a well-connected catchment, sediment that is eroded or (re-)mobilised is effectively transferred along sediment cascades (Burt and Allison, 2010), i.e. from hillslopes to channels, and through the channel network down to the catchment outlet, instead of being deposited in long-term storage landforms. Besides being a part of sediment delivery (Fryirs, 2013), connectivity is related to a system's sensitivity to change (Brunsden and Thornes, 1979; Harvey, 2001; Fryirs, 2016), i.e. the effectiveness of the up- and downstream propagation of change. The significance of connectivity for proglacial sediment budgets and the sensitivity of proglacial areas in terms of transmission of geomorphological change (that is inherent to those areas) has been recently highlighted by Knight and Harrison (2014), Micheletti et al. (2015), and Lane et al. (2016). Overall, sediment connectivity governs to what degree a proglacial area functions as a sediment source or sink (Owens and Slaymaker, 2004).

Sediment connectivity emerges from the spatial configuration of landforms (structural connectivity) and the activity (spatial distribution, reach, frequency) of geomorphological processes and their spatiotemporal interactions (functional connectivity). Proglacial landforms can act as buffers to sediment flux (Cavalli et al, in review; Fig. 4). Deglaciated cirques and other overdeepened parts of glacial valleys form natural basins or host proglacial lakes that effectively decouple downstream valley sections from upstream sediment input (Geilhausen et al., 2013). Likewise, the degradation of ground ice forms depressions that represent sediment sinks. Storage landforms such as talus cones derived from proglacial rockwalls, or debris cones derived from the dissection of till-covered slopes and lateral moraines, can be sources of sediment. This is only effective where runout length is long enough to reach an active river channel, or where storage landforms are being undercut (Haas et al., 2012). Where lateral moraines are developed they provide space between their crests and an adjacent hillslope, intercepting rockfalls and debris flows (Heckmann et al., 2016b). The width and gradient of a (proglacial) valley, and the existence or lack of lateral confinement by moraines or stretches of bedrock has profound impacts on sediment transfer and geomorphological activity (Laute and Beylich, 2014; Temme and Lange, 2014; Beylich and Laute, 2015; Fryirs et al., 2016) and the accommodation space for intermediate storage landforms (e.g. Cavalli et al., in review). Perhaps most importantly, proglacial sediment connectivity can change as a consequence of glacier retreat (Goldin et al., 2016).

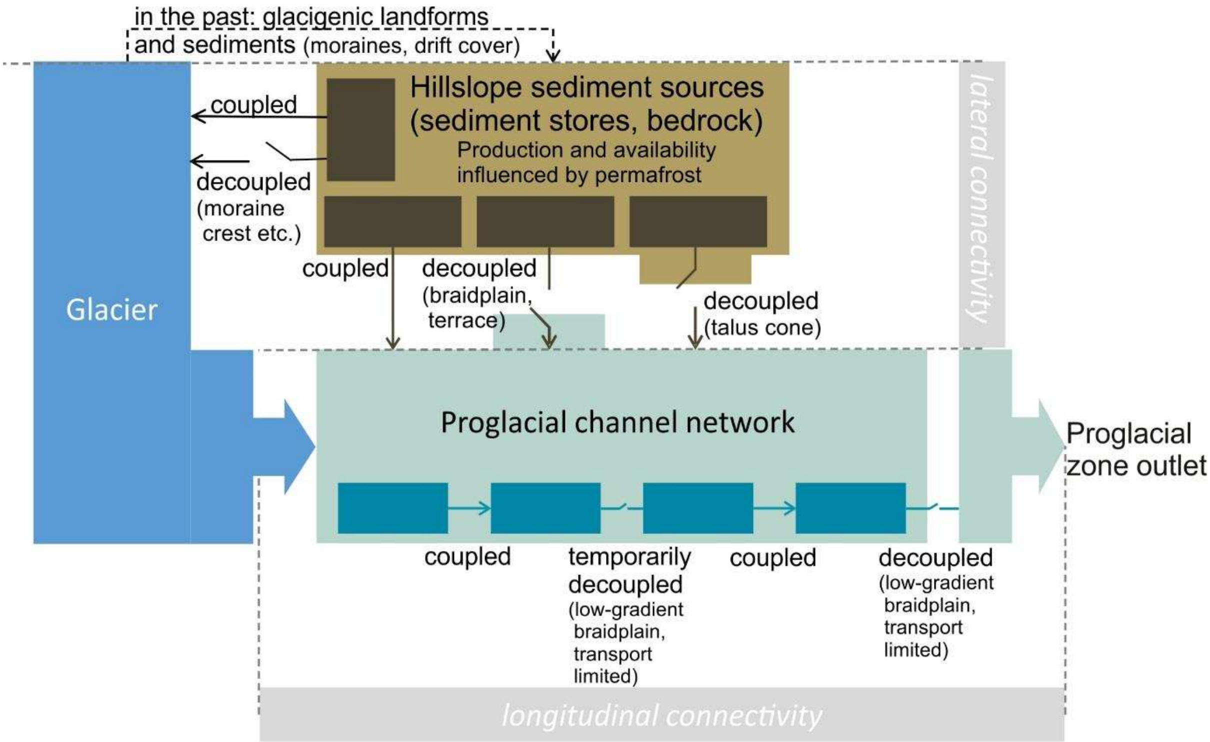

Figure 4: Conceptual model of sediment transfer through a proglacial system, highlighting the role of sediment connectivity between hillslopes, channels and the glacier (lateral connectivity), and along the channel network (longitudinal connectivity). Landforms in brackets are exemplary causes of disconnectivity (buffers and barriers sensu Fryirs et al., 2007).

2.5 Interactions of geomorphological dynamics, soil and vegetation development

The widespread geomorphological and historical evidence of a stepped retreat of glacier termini has led researchers to apply space-for-time substitution (ergodic reasoning) in studies of soil and vegetation development from an “initial” state after deglaciation. Vegetation properties such as diversity and areal coverage, and soil properties such as grain size distribution, pH and organic content, have been analysed on plots situated at different distances from contemporary glacier termini to form chronosequences (e.g. Egli et al., 2006a, b; Moreau et al., 2008, Vilmundardóttir et al., 2014). Recently, there has been a realisation that the assumption of a succession of vegetation or soil types is rather more complicated than this simple conceptual model and interactions of soil formation, vegetation development and geomorphological activity have become a new focus (Eichel et al., 2013, 2015; Temme and Lange, 2014; d'Amico et al., 2014; Temme et al., 2016).

3. QUANTIFYING SHORT-TERM GEOMORPHOLOGICAL CHANGES

Detection of geomorphological activity, especially that occurring over short time scales, is not trivial. Covering spatial *and* temporal scales is problematic and often small-scale to meso-scale activity has a low preservation potential. When geomorphological activity is detected, the question of cause and effect often remains open. Multiple processes interact or 'overwrite' the effect of a single process. This 'noise' is separate from a system trend and is the essence of quasi-stability in systems, and of course stability itself is a function of time scale of observation (cf. Schumm and Lichty, 1965; Figure 5). A detection of (elevation) change depends on the resolution and accuracy (level of detection) of a given method, whilst interpretation of that change in terms of geomorphological activity depends on its separation from a longer-term trend (Figure 5).

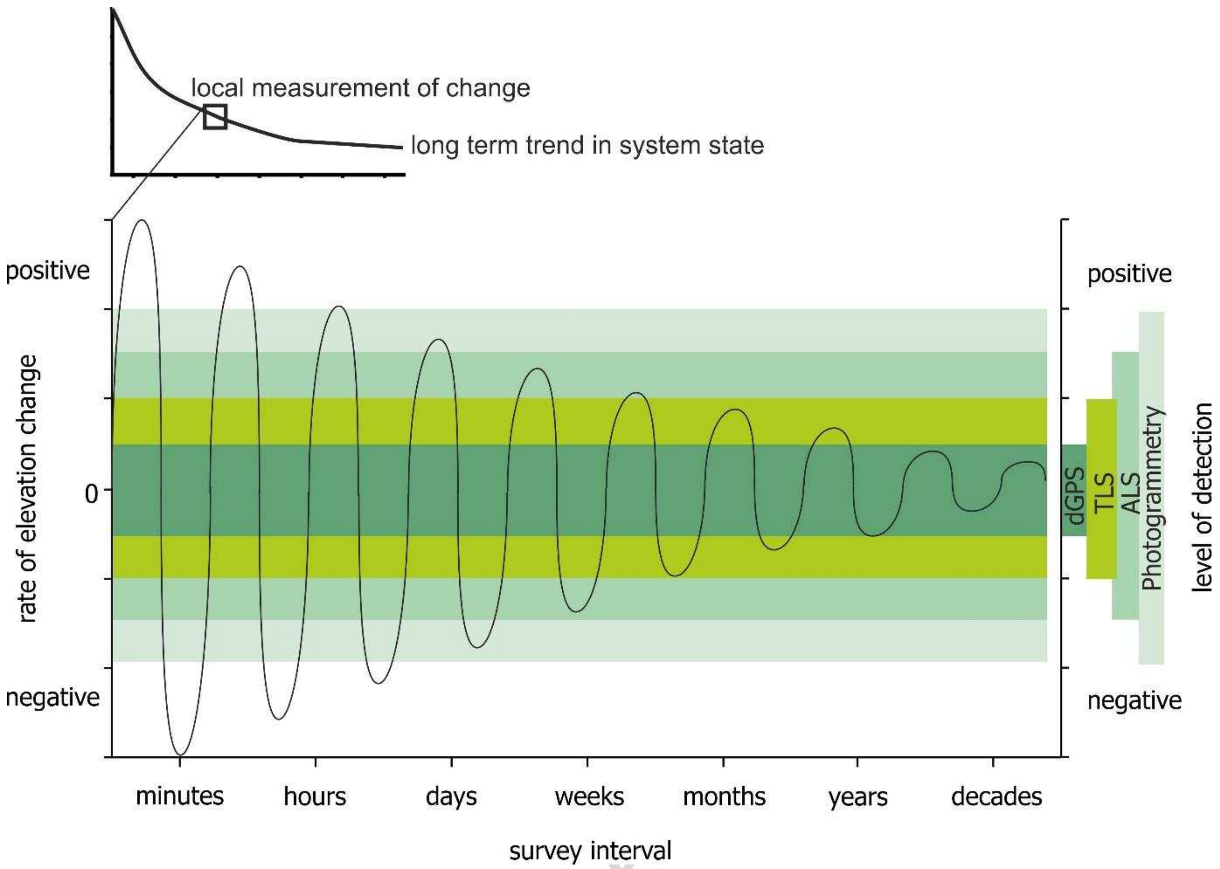

Figure 5: The survey methods of dGPS, terrestrial laser scanning (TLS), airborne laser scanning (ALS) and photogrammetry each have a resolution and accuracy (ve = vertical error) associated with acquisition and processing and differencing of successive surveys that determines a level of detection beyond which any (surface elevation) changes can be considered to be significant in the context of the system state

Additionally, a key question when considering quantification of rapid proglacial changes is what do we (should we) measure? There are two main categories of measurements that are made (Table 1). Spatially-integrated measurements consider the efflux of material *from* a (sub-)catchment in terms of (specific) sediment yield ($t.yr^{-1}$ or $t.km^{-2}yr^{-1}$), usually on the basis of in situ measurements such as of bedload within a river. Spatially distributed measurements *within* a catchment are usually based on volume changes to landforms. The prevailing methodological approaches for quantifying short-term geomorphological changes in proglacial systems can be subdivided regarding site-specific ‘targeted’ approaches and those that are spatially-distributed in both data acquisition and analysis.

Characterisation of Geomorphological activity	Typical time Scale	Typical spatial Resolution	Method	Selected references
Spatially-integrated: for a storage landform	Millennia	-	Volumetric analysis of dated storage landform (e.g. talus cone, rock glacier, etc)	Hinchcliffe and Ballantyne, 1999; Otto et al., 2009; Siewert et al., 2012
Spatially-integrated: from a catchment(s)	Millennia	-	Volumetric analysis of valley-fill and lake sediments	Hinderer, 2001; Schrott et al., 2003
			Cosmogenic nuclide (^{10}Be) analysis	Glotzbach et al., 2013
			Accelerator Mass Spectrometry (AMS) 14C dating of lake sediments	Xu et al. 2015
	Years to millenia	-	Suspended sediment record (lake seds.)	Hicks et al., 1990; Hinderer, 2012; Lane et al., 2016
	Decadal	-	Repeated lake bathymetry surveys	Geilhausen et al., 2012
Spatially-distributed: between catchment(s)	Years to millenia	Single points covering hundreds km^2	Manual mapping	Korup et al., 2010; Allen et al., 2011; Gorum et al., 2014
	Decadal		Digitisation of archived records	Fischer et al., 2012
Spatially-distributed: within a catchment	Decadal	metre	Aerial photographs and photogrammetry	Schiefer and Gilbert, 2007; Staines et al., 2015 Micheletti et al., 2015b
	Seasonal	Up to metre	DEMs from stereoscopic high resolution satellite images	Figures 8 and 12 in this study
		metre	ALS	Irvine-Fynn et al., 2011
		sub-metre	TLS	Carrivick et al., 2013; Baewert and Morche, 2014 Laute and Beylich, 2014; Kociuba, in press.
		sub-cm	SfM	Westoby et al., 2012

Table 1. Categorisation of how geomorphological activity in proglacial systems can be determined. The focus of this paper is on spatially-distributed analyses, which consider topographic changes and thus have capability to discriminate sediment sources, fluxes and sinks.

3.1 Planform analysis

Planform analysis requires definition of a perimeter. The planform extent of a proglacial system can be most usefully given as the difference in area between a LIA glacier extent and a contemporary glacier extent (Heckmann et al., 2012; Fischer et al., 2015; Carrivick et al., in review). This definition best suits alpine environments, because in arctic and polar regions many glacier termini are ‘moraine-dammed’ (c.f. Benn et al. 2003) and still close to their LIA extent, with most geomorphological activity occurring beyond this limit (Figure 6A, B). Delimiting contemporary glacier termini can be complicated by ice-marginal lakes, or by sediment-covered ice-cored moraine or dead ice, as supraglacial debris merges with till plain (debris-covered glacier example; Capt et al., 2016), as outwash head fan aggrade, and as glacier ice becomes submerged by an enlarging lake (Figure 6C).

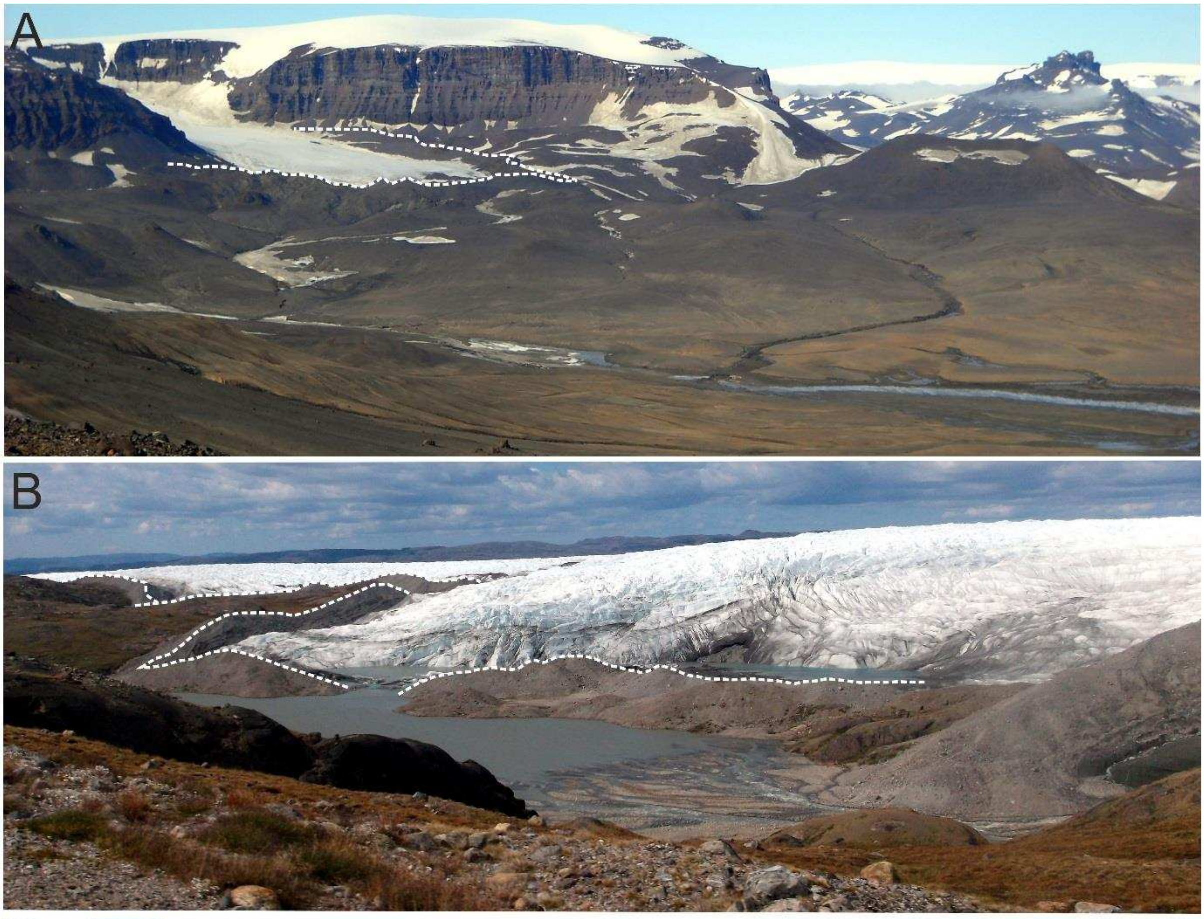

Figure 6: Glaciers on James Ross Island, Antarctic Peninsula, have termini positions that are very close to that of the late Holocene position (Carrivick et al., 2012) (A) and most geomorphological activity is beyond this. In west Greenland the ice margin is largely in contact with its LIA moraines and the

most prominent landscape legacy of geomorphological activity is beyond this (Carrivick et al., 2016a) (B). In the conceptual model of Benn et al. (2003) the glacier margins in A and B would be termed 'moraine-dammed'. In contrast, in many alpine environments glacier termini are often hundreds of meters, even a few kilometres behind LIA positions and within this space permafrost degradation and ice-marginal lakes are destabilising moraine ridge flanks as in the upper Godley valley, New Zealand (C). Some ice-contact lakes are not only rapidly expanding in area but also rising in water level, which when coinciding with a glacier retreating into an overdeepened basin means that the glacier terminus becomes submerged, such as at Sólheimajökull, Iceland (D). In all panels LIA moraine crests are marked by a dashed white line. Images in panel C are courtesy of Bruce Soper, and image in panel D is courtesy of Fiona Tweed.

3.1.1 Mapping the spatial components of proglacial systems

Consideration of multiple components of a proglacial system, especially for sediment budget analysis (see section 3.4) usually requires the derivation of a geomorphological map. Geomorphological mapping can be achieved with a combination of planform-based field surveys and expert judgement-driven on-screen digitisation of remotely-sensed data. The former is to identify landforms with sedimentological (composition) information and to 'ground-truth' landform properties such as position, size, shape and geomorphological associations as obtained by the latter.

Published geomorphological maps of proglacial systems are few, but some recent examples include Sletten et al. (2001), Evans et al. (2007) and Kjær et al. (2008) who all mapped and described the impact of glacier surge(s) on proglacial landscapes in Iceland, and Bennett et al. (2010) who produced a multi-temporal map of Kvíárjökull, Iceland as supported by multiple DEMs and historical aerial photographs. Elsewhere, Hasholt et al. (2008) examined landform and sediment associations at Mittivakkat Glacier in southeast Greenland, Geilhausen et al. (2012) mapped two proglacial systems in Austria to identify sediment sources, transfers and storages, and Colombo et al. (2016) and Lambiel et al. (2016) mapped valleys in the Italian and Swiss Alps, respectively, and both commented on the spatial distribution of glacial, periglacial, gravitational and fluvial processes. A few other geomorphological maps of proglacial systems are contained within reports on field monitoring process-based studies (e.g. Laute and Beylich, 2012; Eichel et al., 2013; Kociuba, in press).

Geomorphological mapping and associated planform analysis has been integral to gaining process-understanding on landform evolution in proglacial systems. For example, Marren and Toomath (2013, 2014) demonstrated the importance of topography in evolution of the planform of a proglacial river in southern Iceland over a decade. They went on to show that proglacial river pattern can change at timescales faster than hydrological or sediment budget changes usually occur, and in association with relatively minor changes in glacier mass balance. Micheletti et al. (2015) produced multi-temporal maps (and digital elevation models as discussed in section 3.3) spanning five decades for an alpine (not strictly proglacial as defined herein) Swiss mountain side and interpreted a distinct tendency for increased planform changes (as well as vertical displacements) during warmer periods. However, they also noted breaks in sediment pathways and thus valley system processes and landforms that no longer reflected climate perturbations. Dietrich and Krautblatter (in press) have used orthoimage analysis to detect and quantify debris flow activity, finding a higher activity in the last few decades than compared to the Holocene average.

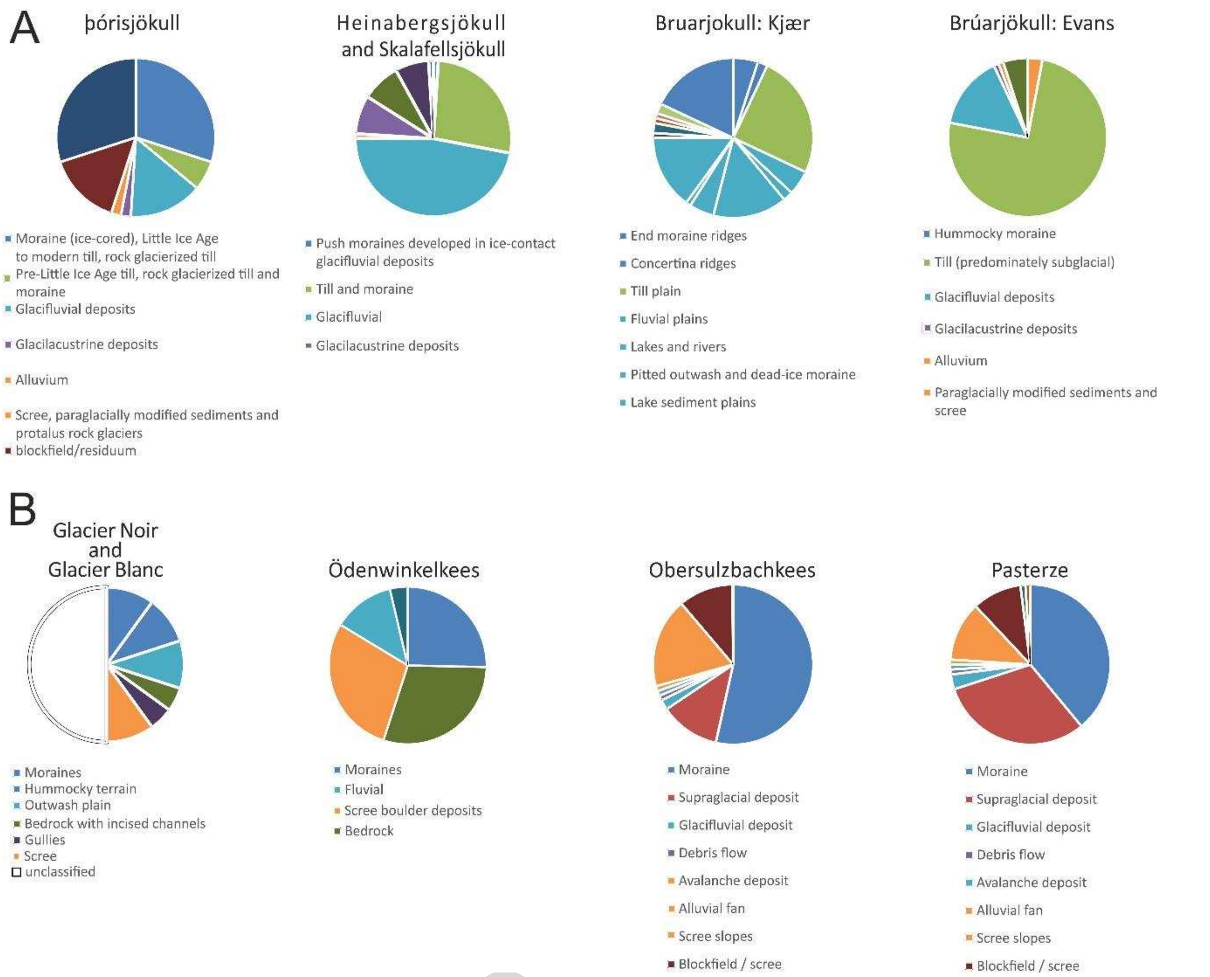

Figure 7: Percentage composition by area of some mapped proglacial systems illustrating variability not only between sites but also at a single site due to different authors' mapping conventions (see differing legend classifications and terminology) and planned usage. Note (i) some simplification has been made to permit comparisons, (ii) colours crudely denote major process types to permit comparisons: green represents till, light blue represented fluvial landforms, mid-blue represents moraine, red represents supraglacially-derived debris, orange and brown represent hillslope sediment, (iii) some landforms such as active rivers, gullies, eskers, drumlins, flutes and meltwater channels, are often mapped using vectors so do not contribute to areal coverage statistics, (iv) there are differences in these maps as to what constitutes the (mapped) proglacial system. Data derived from: Evans et al. (2006) for Þórisjökull, Iceland; from Evans and Orton (2015) for Heinabergsjökull and Skalafellsjökull, Iceland; from Kjær et al. (2008) and from Evans et al. (2007) for Brúarjökull, Iceland; from Lardeux et al. (2016) for Glacier Noir and Glacier Blanc, French Alps; from Carrivick et al. (2013) for Ödenwinkelkees, Austria; from Geilhausen et al. (2012) for Obersulzbachkees and Pasterze, Austria.

Overall these geomorphological maps and planform analysis of proglacial systems (i) provide an historical archive of landform development since the LIA, (ii) a monitoring tool for studying the temporal evolution of proglacial systems, and (iii) modern analogues for reconstructions of past glaciations. However, they are few and far between and usually only cover single sites, due to the investment of resources required. Where they are covering several sites (e.g. Geilhausen et al., 2012; Lardeux et al., 2015; Lambiel et al., 2016), they serve to emphasise inter-catchment variability in proglacial geomorphology (cf. Carrivick and Rushmer, 2009) (Figure 7). They are only at single time-frames. The paucity of consistent geomorphological mapping is a problem because knowledge of many catchments across domains of geology and climate is required if conceptual models of paraglaciation and proglacial landscape development are to be quantitatively tested.

Some of the differences between the geomorphological maps of proglacial systems are simply due to different map scales used. Additionally, these types of maps cannot adequately represent composite landforms (see map of Geilhausen et al., 2013) or landforms that become over-ridden / over-printed (see map of Evans and Orton, 2015). They do not usually consider linkages between landforms and they necessarily create artificial boundaries between landforms. Some maps, such as Lardeux et al. (2015), have very large amounts of unmapped 'space' between landforms. Assessing the compositional variability between these mapped proglacial systems (Figure 7) is therefore far from straight-forward.

In efforts to reduce the reliance on expert judgment, it is useful to note that landform 'maps' of proglacial systems have also been generated semi-automatically using object-based analysis (e.g. Anders et al., 2011). Automation helps produce comparable planform data from multiple sites quickly and efficiently, which is necessary of short-term changes are to be identified. Similar studies have been performed in mountainous, though not proglacial as defined in this study, landscapes (e.g. van Asselen and Seijmonsbergen, 2006). The number of landform classes and the classification accuracy vary between these semi-automatic methods but are typically ~ 10 and 70 %, respectively. Semi-automatic classifications of landforms have not yet been applied to detecting changes to proglacial systems.

3.1.2 Quantifying planform changes in proglacial systems

There has been very little quantification of contemporary planform changes to separate components of a proglacial system. However, there have been a few quantitative studies of the rate of landform changes in other alpine settings. Terraces and lobate parts of the upper parts of alpine (not proglacial) hillslopes have been measured to move at between 4 and 43 mm yr^{-1} (Caine, 1963; Benedict, 1970), which is an order of magnitude faster than the more stable parts of alpine interfluves (Caine, 1974).

Cliff retreat rates of up to 1 mm yr^{-1} have been estimated from volumetric analyses of scree accumulation in a number of alpine (though not necessarily proglacial) settings. This contemporary rate is often higher than longer-term estimates (Caine and Jennings, 1968; Rapp, 1960a, b; Haeberli et al., 1999; Hinchcliffe and Ballantyne, 1999; Humlum, 2000).

Scree slope planform movements encompass a continuum of processes and deposits from alluvial to avalanche, and are thus highly variable. An upper value of 20 cm yr^{-1} has been proposed for scree moving in an alpine (not proglacial) environment (Caine, 1974).

Recent quantitative analysis of the planform of proglacial systems has been dominated by studies focussing on braided rivers. The default measurements tend to be those of (i) channel network pattern, which can have pronounced expansion and contraction cycles (Ward et al., 2001), and (ii) sinuosity and degree of braiding, which tend to increase with an advance of alpine glacier terminus position(s), due to increased sediment supply (Maizels, 1983; Fenn and Gurnell, 1987). Both these sets of measurements have been performed on manually-digitised river planforms. Automatic classification of river channel networks is lacking and hindering spatio-temporal studies but might be achieved at least on larger river systems by using the Normalised Difference Water Index (NDWI) to classify (30 m pixel resolution) Landsat scenes, for example.

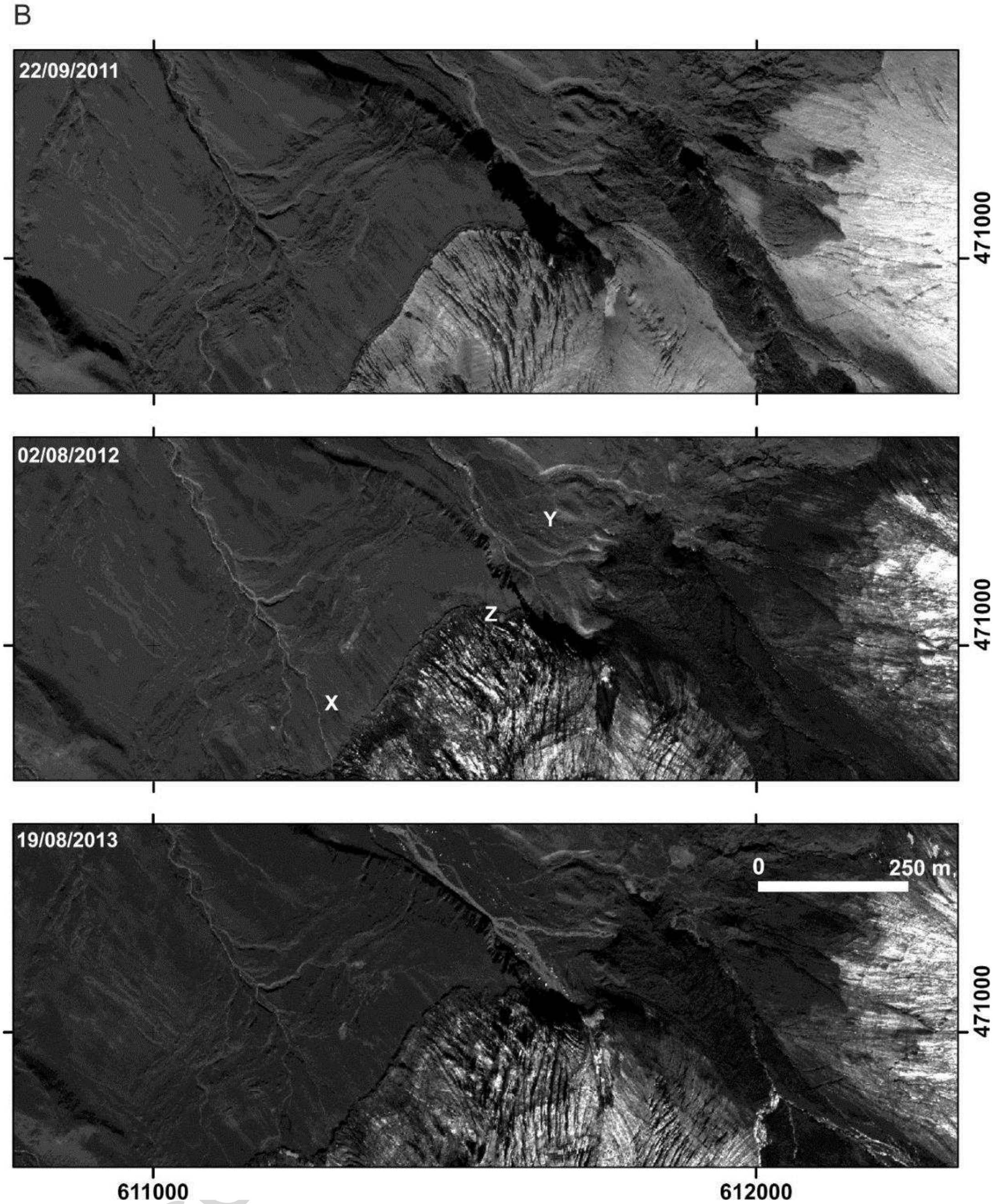

Figure 8: Satellite imagery (WorldView 1 and 2) at 0.5 m grid resolution of the terminus of Kverkjökull, north-central Iceland, in which X denotes route of a new proglacial stream, Y denotes area of proglacial stream avulsions (see light grey stream network expansion), Z denotes area of fragmentation of ice margin. Scale is the same in all panels.

On the time scale of a single alpine melt season, both increasing and decreasing braid intensity can be observed, again linked to sediment supply (e.g. Fahnestock, 1963; Warburton, 1994). In contrast, for two separate sites in Iceland both Nicholas and Sambrook-Smith (1998) and Staines et al. (2015) used braided river planform metrics to contend that low-frequency high-magnitude glacier floods, or 'jökulhlaups' are the dominant control on proglacial river planform. Debris flows from breaching of moraines and glacier lake outburst floods can similarly dominate proglacial river configurations and behaviour. Some field monitoring studies (see section 3.4.1) have determined that low frequency high-magnitude flood events can be the dominant control on proglacial sediment fluxes (see data and citations in Table 2 of Orwin et al., 2010; and see also Leggat et al., 2015).

Opportunities for quantifying short-term planform changes within proglacial systems are emerging in the form of spatially high-resolution satellite imagery (usually < 1 m: e.g. WorldView 1 and 2, IKONOS, KOMPSAT 3, Pleiades and QuickBird) that also has the advantage of being temporally high-resolution. These images can for example be used to identify subtle yet potentially important proglacial system changes such as newly emerging proglacial streams, avulsions of pre-existing proglacial river channels and ice-margin break-up, for example, on an inter-annual time-scale (e.g. Figure 8).

3.2 Horizontal movement

Horizontal movement is commonly analysed together with vertical movement. Orthoimage cross-correlation has been used on glacier surfaces (Berthier et al., 2005) and for landslides (Kääb, 2002; Lucieer et al., 2014). Within proglacial systems, horizontal movement of rock glacier surfaces, soil creep (e.g. Kääb and Kneisel, 2006; Dusik et al., 2015) and debris flows (e.g. Kenner et al., 2014) has also been studied using the same method. Whilst glacier movement is a relatively continuous process, albeit with seasonal fluctuations, landslides are episodic and so the time between images, the survey interval, is crucial if rates of change are to be determined. Larger spatial coverage can gain simultaneous measurement of displacements on multiple landforms, such as that on the surface of a rock glacier, across ice-cored moraine and on scree slopes and hillslopes in Iceland by Wangensteen et al. (2006). Image-correlation has also been achieved in studies of landslides (outside proglacial systems) using point clouds from a laser scanners (e.g. Travelletti et al., 2014).

Interferometry has also been used extensively outside of proglacial systems for glacier surface motion and landslide quantification, primarily using repeat Synthetic Aperture Radar (SAR) data. SAR data from a satellite platform is most common but ground-based platforms have also proven successful for studying mass movements. Interferometry has been used in proglacial systems to quantify river planform (horizontal) avulsions (Smith et al., 2000; Gomez et al., 2002), permafrost creep (e.g. Strozzi et al., 2004) and rockslides (Eriksen et al., 2015), for example. The resultant distributed velocity fields have permitted advancement in process-understanding; for example of sources, trajectories, pathways and sinks of sediment, volume changes and flow behaviour.

3.3 Elevation changes

Quantification of elevation changes between re-surveys can reveal the magnitude and the rate of local fluvial incision / deposition and/or erosion / aggradation due to hillslope processes. In the last thirty years there has been a progressive improvement in spatial density and geographical coverage. Thus there has been progress from studies of proglacial systems utilising transect-based targeted surveys and using optical levels and total stations (Table 1), to those employing differential Global Positioning Systems (dGPS), to those making fully-spatially-distributed or non-targeted surveys with oblique photogrammetry (e.g. Pyle et al., 2007), vertical photogrammetry (e.g. Staines et al., 2015), terrestrial laser scanners (TLS), airborne laser scanners (ALS) and structure from motion (SfM) methods (Table 1).

Irrespective of the method employed, a persistent problem with detecting elevation changes in proglacial systems is that they can be associated with sediment erosion/deposition, but they can also result from horizontal movements or from permafrost degradation/aggradation that is not associated with sediment transfer. The conflicting causes of these elevation changes can therefore lead to issues for sediment budgeting, for example in areas with negative surface change but no corresponding accumulation (because nothing was eroded or deposited). A potential way to distinguish these two factors could be the spatial pattern of changes, which in the case of ice meltout and (linear) erosion should be markedly different. A classification of a DEM of Difference (DoD), for example of the local/focal standard deviation or of the variogram of the DoD, should theoretically show a more homogeneous pattern in the case of ice melt, and less uniformity in the case of erosion. An additional problem in detecting and interpreting elevation changes as an index of geomorphological activity is that two different processes may spatially overlap, e.g. meltout plus debris flow erosion or deposition, so net rather than gross changes are evident.

Carefully targeted field studies employing repeat topographic surveys on cross-sections / transects have revealed that proglacial river systems in arctic settings may have a different set of morphodynamic controls than those in alpine environments (Nicholas and Sambrook-Smith, 1998). Furthermore, despite different environmental settings, dead-ice melting rates in high-arctic and humid sub-polar proglacial environments are very similar and thus driven by local topography and processes rather than by climate (Schomacker and Kjaer, 2008). This variability in inter- and intra-catchment controls on proglacial runoff and consequent geomorphological response makes comparing rates of change measured at one site to those at another difficult (Figure 9).

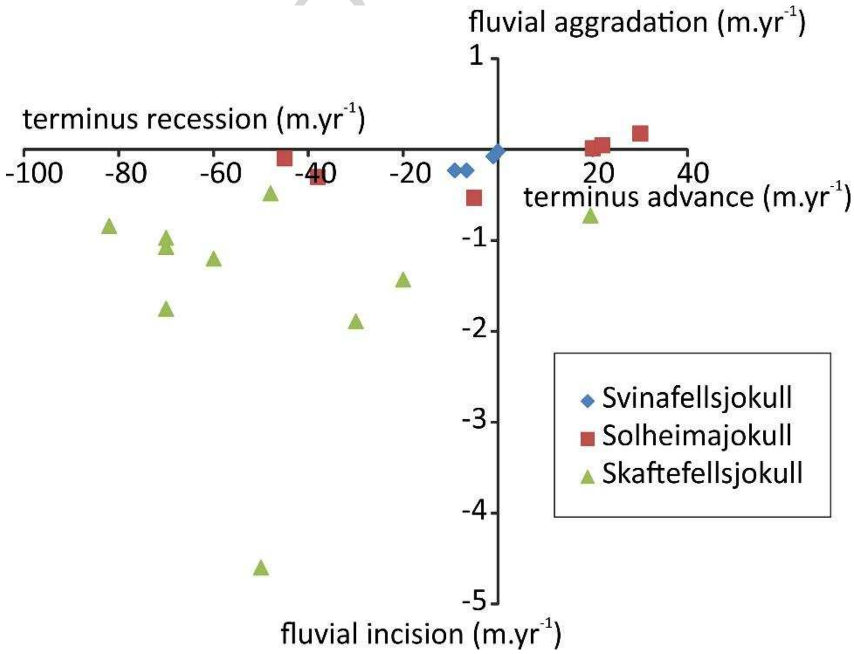

Figure 9: Proglacial fluvial erosion and deposition rates emphasising intra- and inter-catchment variability especially when related to rate of glacier terminus advance / retreat. Each point represents a different time period. Updated from Staines et al., 2015.

In contrast, most other repeat surveys within proglacial systems have focussed on using photogrammetric (e.g. Luchi et al., 2007), differential Global Positioning System (dGPS) (Brasington et al., 2000, 2003) and remote sensing data (e.g. Lane et al., 2003). River reaches have been the choice component of proglacial systems for study. Linkages between proglacial river system components such as proglacial fluvial channel avulsions and subsequent downstream bar growth, and system trends and discontinuities such as summer (high base flow) aggradation and winter (reduced base flow) incision have been identified (Carrivick et al., 2013), and provide a pervasive impression of geomorphological instability driven by climate change as manifest in continually adjusting water source contributions.

Where repeated proglacial surveys have been distributed over an area, then digital elevation models (DEMs) can be derived, differenced and hence a volume change can be calculated (Abermann et al., 2010). Increased spatial resolution and reduced point uncertainty has enabled relationships between grain-scale dynamics and landform evolution for gravel bars (e.g. Milan et al., 2007) and for hillslope-scale fluvial, hillslope and ice-marginal parts of a proglacial system (Haas et al., 2012; Kociuba, in press) to be developed. These TLS data acquisition and processing methods have been especially advanced in technicality with application to braided rivers, though not strictly proglacial as defined in this study, by Brasington et al. (2012) and Williams et al. (2014) for the Feshie river in Scotland and for the Rees river in New Zealand, respectively.

Notwithstanding the recent contributions of Haas et al. (2012) and Kociuba (in press), usage of TLS within proglacial systems for examining non-fluvial processes has yet to be fully exploited. However, outside proglacial systems TLS has quantified and explained changes on an alpine (not proglacial) talus cone over a four month period (Morche et al. 2008), identified elevation patterns and volume changes of a (not proglacial) debris-flow (Bremer and Sass, 2012; Blasone et al. 2014), and quantified short-term dynamics of ice-cored moraines on Svalbard (Ewertowski and Tomczyk, 2015). The comprehensive study by Davies et al. (2013) of ice-marginal landscape evolution on the largely ice-free Ulu Peninsula of James Ross Island, Antarctic Peninsula, and the recent study by Ruiz-Fernández and Oliva (2016) who reported qualitatively on the Holocene landscape evolution of a part of Livingston Island, Antarctica, are both notable for being amongst the very few studies on proglacial geomorphological evolution in polar environments.

Airborne Laser Scanning (ALS) data can assist macro-scale or catchment-scale considerations of geomorphology, but re-surveys by ALS that cover proglacial systems are rare and often not far enough apart in time for geomorphological changes to be discernible, at least beyond the level of detection as determined by the uncertainty within each individual survey. Notable exceptions are the studies by Bollmann et al. (2011) and Sailer et al. (2012) who both used annual ALS datasets acquired over nine years at the Hintereisferner, Austria, to quantify geomorphological activity including rock fall, fluvial incision and deposition, dead ice and permafrost degradation. Irvine-Fynn et al. (2011) quantified surface changes and associated processes between 2003 and 2005 in the proglacial system of Midtre Lovenbreen, Svalbard, noting the major changes being fluvial incision and the downwasting of a lateral moraine. Hilger et al. (in review) use multi-temporal ALS data for a comprehensive sediment budget of the Upper Kaunertal, Austria, with a focus on the proglacial area.

More commonly, ALS data have been used in conjunction with datasets obtained via other methods, usually photogrammetry, and in proglacial systems these types of studies have yielded a spatially-distributed landscape-scale impression of outwash plain, or ‘sandur’, response to a glacier outburst flood in southern Iceland (Smith et al., 2006), of a major rock fall in central Austria (Kellerer-Pirklbauer et al., 2012) and of debris flows (Legg et al., 2014). For example, Legg et al. (2014) were able to determine debris-flow dynamics in proglacial channels on Mt. Rainier using a single ALS elevation model, and pre-event aerial photos and by assuming parallel retreat of gully sidewalls, to infer volumetric change from width change. Carrivick et al. (2013) presented the only quantitative study to date to have reported multi-scale surveys across a whole proglacial system, and they demonstrated (i) coupling between upper and lower valley floor components of the fluvial system as a result of major glacier meltwater discharge events, (ii) spatial association between hillslope activity as responses to major snowmelt and precipitation events, and (iii) a likelihood of surveys at short intervals to detect more geomorphological activity per unit time than those surveys with longer time intervals between surveys. The latter finding means that longer time periods between surveys tend to reveal more balanced sediment budgets (Lindsay and Ashmore, 2002; Milan et al., 2007) rather than event-scale variability.

There is considerable variability in the rate of change in elevation of landforms between different proglacial systems (Figure 10). This variability is not simply a function of process and environment, landform position, age and connectivity, and survey time(s) and intervals. Indeed, different landform components and different environments have different paraglacial durations (Ballantyne, 2002). There are also cases of rather unusual activity being detected; Hubbard et al. (2005) showed (unusual) negative elevation changes due to the impact of a rock avalanche on a moraine-dammed proglacial lake at Laguna Safuna Alta, Cordillera Blanca, Peru.

If rates of change have been derived via differencing of DEMs there could be an artefact of point density, surface roughness and local slope (Bollmann et al., 2011; Sailer et al., 2014). Nonetheless, it is instructive to compile literature-reported values of the rates of landform changes in proglacial systems and to summarise the distribution of these values (Figure 10), because they can inform on regionalisation and up-scaling (e.g. Heckmann et al., 2016b). In brief, Figure 10 shows that the fastest changes occur with rockfall, up to 3 m yr^{-1}, that scree and rockfall talus are the only landforms with net gains of elevation, and that the fastest surface lowering occurs via debris fan incision and fluvial incision. We note that working with the high dispersion in these data can be eased by reducing to 2 classes; high versus low rates, e.g. Caine (1974), or to 3 classes; low, moderate, and high activity, e.g. Schrott et al. (2003). Then non-metric multidimensional scaling can be used to facilitate comparisons between groups, cf. Eichel et al. (2013, 2015).

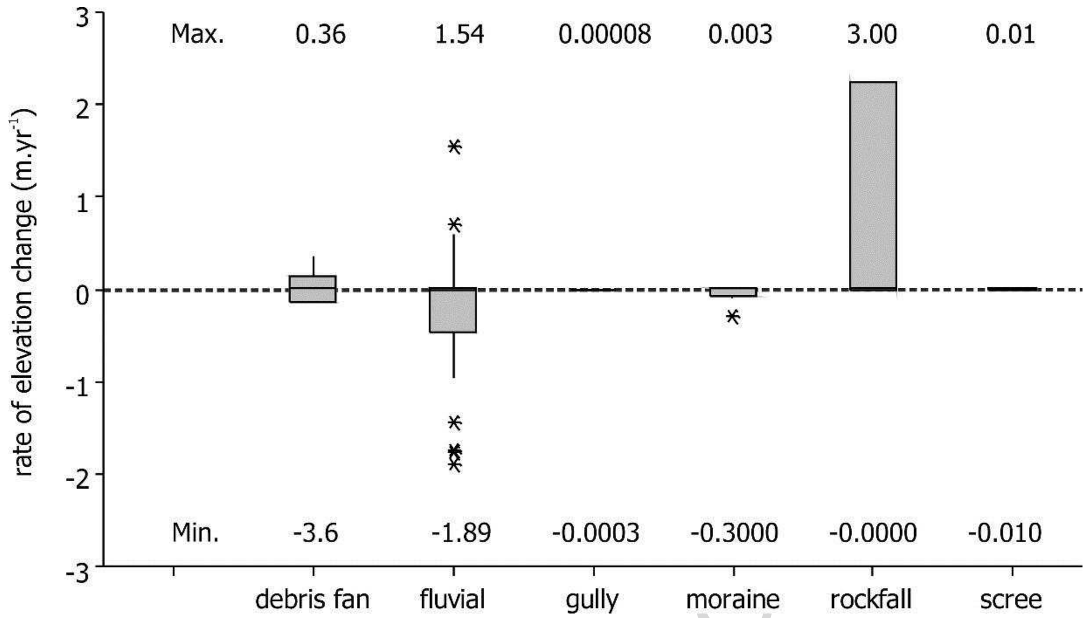

Figure 10: Rate of change in elevation reported for specific proglacial system landforms. Note that whilst all rates have been converted to the same units for comparison, (i) some processes causing these elevation changes are continuous and some are episodic, and (ii) the rates partly depend on the time between surveys (Sailer et al., 2012, 2014; Carrivick et al., 2013). Data compiled from Maizels, 1979; Thompson and Jones, 1986; Ballantyne and Benn, 1994; Ballantyne, 1995; Watanabe et al., 1998; Harrison and Winchester, 1997; Magilligan et al., 2002; Curry et al., 2006; Nicholas and Sambrook-Smith, 1998; Smith et al., 2000, 2006; Milan et al., 2007; Schomacker and Kjaer, 2008; Bollman et al., 2011; Irvine-Fynn et al., 2011; Bremer and Sass, 2012; Haas et al. 2012; Kellerer-Pirklbauer et al., 2012; Sailer et al., 2012; Carrivick et al., 2013; Marren and Toomath, 2013; Blasone et al., 2014; Ewertowski and Tomczyk, 2015; Heckmann et al., 2016b.

Emerging opportunities for DEM generation of proglacial systems include from-motion (SfM) and Surface Extraction with TIN-based Search-space Minimization (SETSM) methods. The former is a rapidly-developing workflow only requires multiple overlapping images of a surface of interest acquired multiple (camera) positions to generate a three-dimensional point cloud (Smith 2015; Carrivick et al., 2016b). To date, SfM has only been employed in systems for single surveys (Table 2), but a SfM-derived DEM can be compared DEM from another source, such as ALS (e.g.

Figure 11). The latter is automatic photogrammetric routine optimised for high-resolution (e.g. WorldView 1 and 2, IKONOS, KOMPSAT 3, Pleiades and QuickBird) satellite imagery obtained in high-relief environments and over highly-reflective and low-texture surfaces, such as snow, ice and lakes, as reported by Noh and Howatt (2015) (Fig. 12).

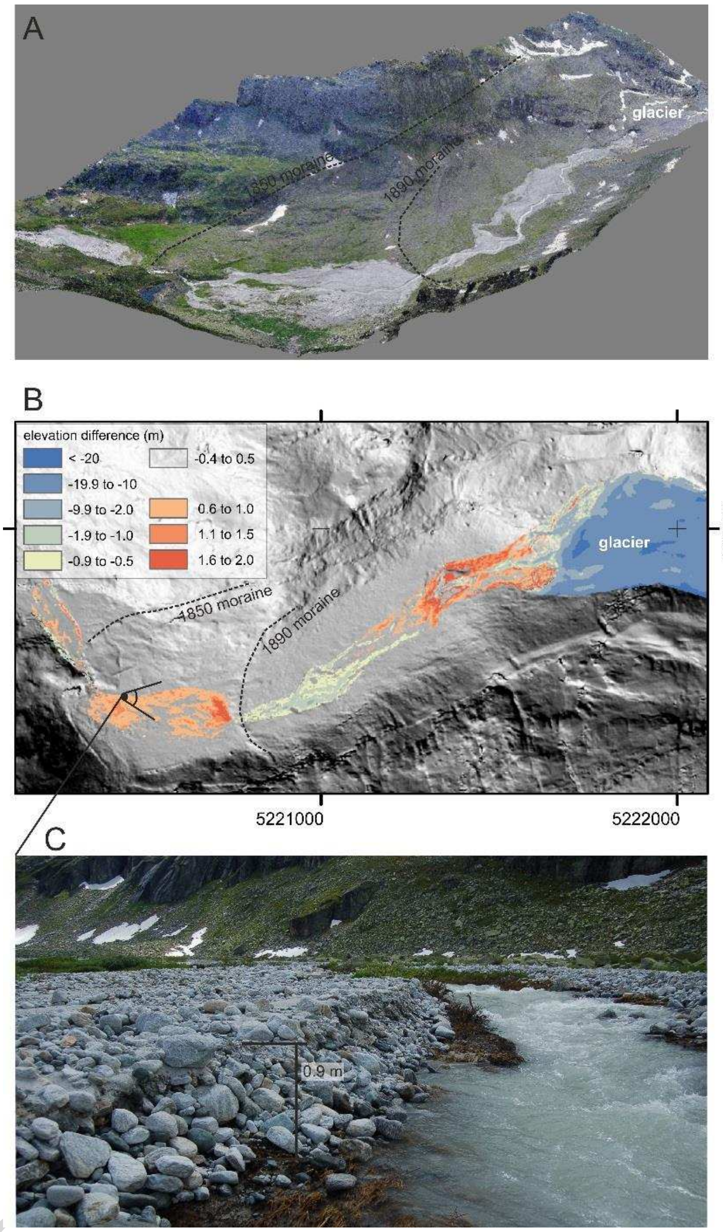

Figure 11: Example of structure-from-motion (SfM) used to re-survey a proglacial river channel at Ödenwinkelkees, central Austria in July 2015 (A), and the difference between the resulting elevation model and that from airborne laser-scanning seven years earlier in July 2008. The level of detection (of elevation change) is considered to be ± 0.5 m. Despite the time interval our annual (at least) visits to this site mean that we are quite certain that the pattern of elevation differences (B) is due to a single flood produced by intense rainfall in September 2014. Ground checks of the elevation differences show cobbles and boulders overlying a vegetated bar (C).

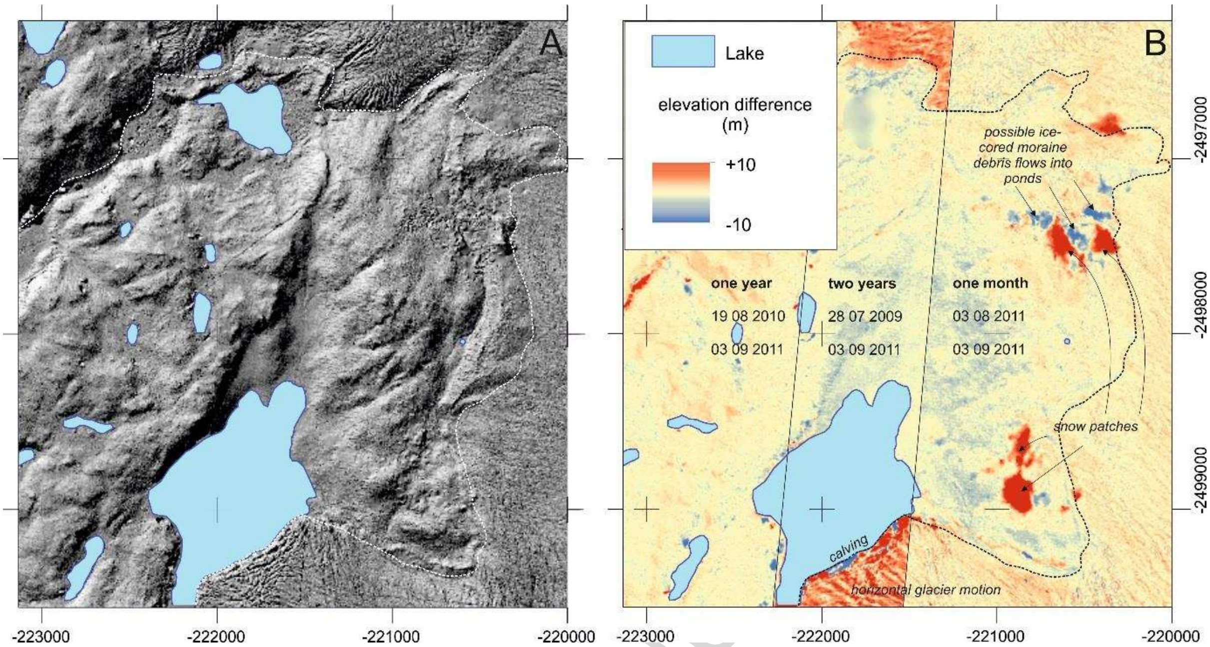

Figure 12: Example of 2 m grid resolution SETSM elevation data (hillshaded) from west Greenland (A) that has the potential to capture short-term elevation changes (B). However, caution must be exercised because these DEMs are composites derived from images on different dates and the intervals between surveys vary for different spatial entities (B).

Figure 13 shows an ALS DEM of difference (2006-2012) of a hillslope in the proglacial area of the Gepatsch glacier, Kaunertal, Austria. Between July and September, 2012, a rockslide of 18,000 m³ (cf. Heckmann et al., 2016b) occurred just below the 1850 trimline, depositing a debris cone on the slope foot and the adjacent glacier: the deep red area in the SW represents the downwasting of the glacier surface between 2006 and 2012, amounting to c. 20 m (Vehling et al., 2016, in press). SE of the conspicuous debris cone (blue), smaller rockslope failures pre-dating 2012 are visible; the positive surface changes SE of the rockslide pathway do not represent accumulation but a creeping movement of a c. 250 m wide rock-slope failure.

All of the above examples rely on expert judgment to interpret geomorphological process from spatial patterns of elevation changes. In contrast, objective methods of identifying landforms via spatial clustering of elevation metrics (range, standard deviation, skewness, kurtosis, coefficient of variation, standard deviation of curvature, rugosity, and volume:area ratio) across moving window sizes of varying size (10 to 1000 m in radius) and then via correlograms to identify thresholds separating different data (landform) groupings was championed in a study on a part of the Upper Mississippi River, by Scown et al. (2015). Such cluster analysis of elevation changes might help facilitate detection and understanding of surface changes without associated erosion-transfer/deposition, such as in the case of permafrost or debris-covered dead ice degradation, for example.

Proglacial system element	Predominant repeat survey method	Example references
Gravel river cross-sections	Levelling	Maizels, 1979; Ashworth and Ferguson, 1986; Thompson and Jones, 1986; Maizels, 1997; Nicholas and Sambrook-Smith, 1998
Gravel river cross-sections	Total Station	Warburton, 1992; Marren et al., 2002, 2005
Gravel bars and fan volume changes	dGPS+	Godin and Fortier, 2012; Carrivick et al., 2013
Sandur change is response to jökulhlaup Proglacial landform evolution	DEM from aerial photographs	Magilligan et al., 2002; Gomez et al., 2002 Bennett and Evans, 2012; Micheletti et al., 2015b
Gravel river reach Valley sedimentation due to a jökulhlaup Debris flow Ice-cored moraine dynamics	TLS	Milan et al., 2007; Brasington et al., 2012; Picco et al., 2013; Baewert and Morche, 2014 Dunning et al., 2013 Schürch et al., 2011; Bremer and Sass, 2012; Blasone et al., 2014 Ewertowski and Tomczyk, 2015
Rock fall, debris flow	ALS	Irvine-Fynn et al., 2011; Kellerer-Pirklbauer et al., 2012; Legg et al., 2014
Moraine Braided river reach Braided river reach Alluvial fan	SfM	Westoby et al., 2012* Javernick et al., 2014* Figure 11 this study Micheletti et al., 2015a*

Table 2. Topographic changes detected and analysed by repeated targeted campaign surveys, discriminated primarily by survey method and thus by spatial resolution. *Only single SfM surveys so not used for change-detection. +dGPS is often used to georeference distributed surveys with TLS, SfM and ALS. Note not all cited literature pertains to proglacial systems as defined in this study.

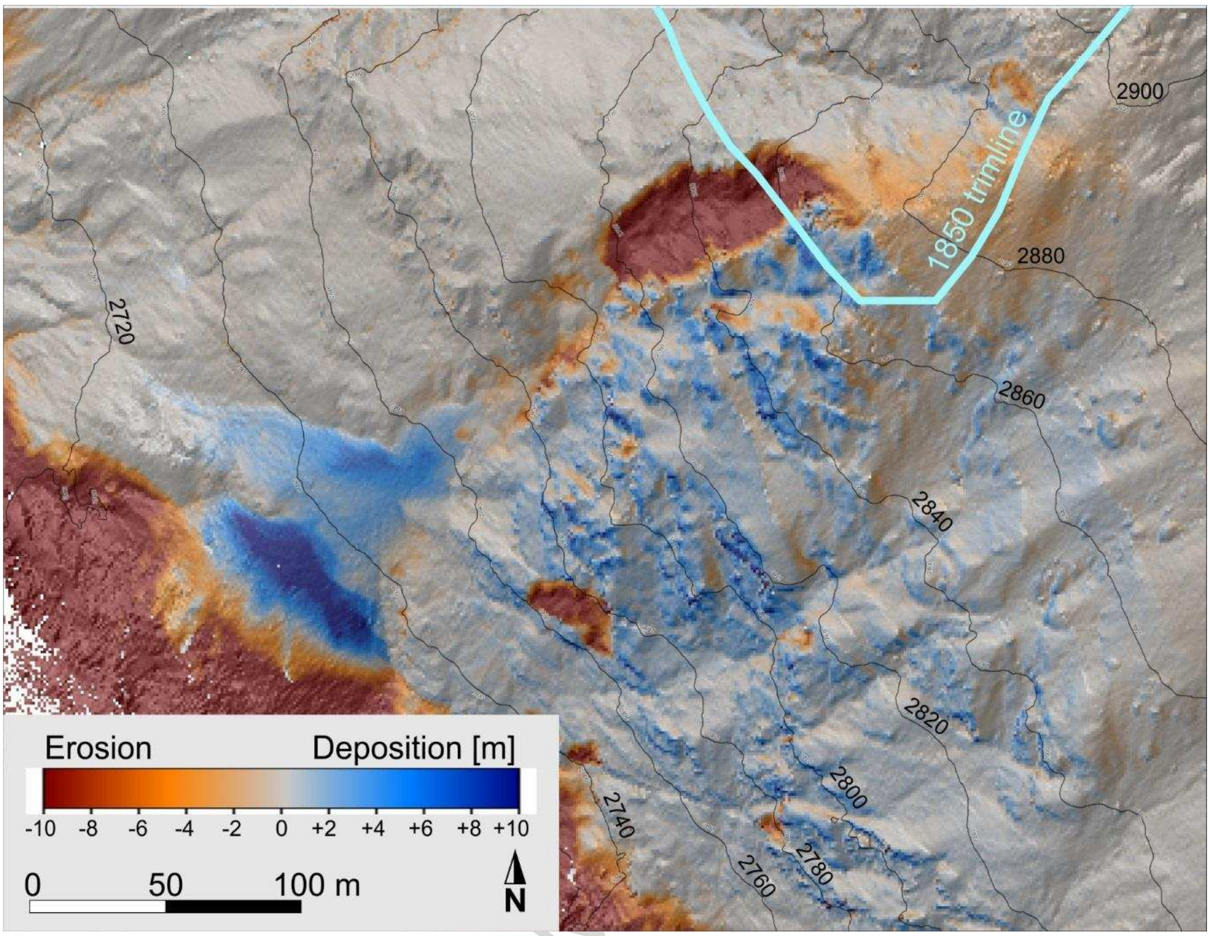

Figure 13: DEM of difference (ALS; 2006-2012) of a proglacial hillslope in Kaunertal showing a 18,000 m³ rockslide onto the glacier (modified after Heckmann et al., 2016b)

3.4 Sediment budgets

Determination of sediment sources, storages and fluxes within proglacial systems is challenging. Existing catchment-wide models (e.g. Caine, 1974; Dietrich and Dunne, 1978; Owens and Slaymaker, 2004) are qualitative yet are relied on heavily for designing field monitoring of water and sediment fluxes.

Proglacial systems with thick moraine accumulations are especially common in alpine environments because cirques or shallow-gradient valley floors that are inherited from previous glacier activity do not support an efficient link between the glacial and the hydrological transport systems (Benn et al., 2003). Proglacial systems with strong glacier–permafrost interactions are especially common in arctic and high-altitude environments where polythermal glaciers can have indistinct and even intercalated boundaries with surrounding hillslopes and glacier forefields (Cuffey and Paterson, 2010; Ribolini et al., 2010) and because permafrost in hillslopes introduces ice into the sediment system. These arctic and permafrost-influenced proglacial systems tend to typically contain massive ice, ice-cemented and ice-free sediments (e.g. Kneisel and Kääb, 2007; Ribolini et al., 2010) and landforms including rock glaciers, push moraines, and ice cored moraines (Bosson et al., 2015; Dusik et al., 2015). This global variety of sediments, landforms and geomorphological processes within proglacial systems makes standardised field monitoring of short-term changes essential.

Several studies have shown that there is considerable sediment delivery by non-glacial or periglacial processes from hillslopes to a glacier or directly to a proglacial system (Warburton, 1990; O'Farrell et al., 2009). Thus exclusive focus on sediment pathways from a glacier to (and out of) a glacifluvial system might create a knowledge gap. Hilger and Beylich (in review) concluded in their review that there was a negligence of system-internal sediment transfers and a lack of truly comprehensive (i.e. covering all processes, specifically those that are active outside the main fluvial system) sediment budgeting studies. A quantification of this problem was made by O'Farrell et al. (2009) who reported that up to 80 % of the proglacial sediment yield in their study area was derived from non-glacial sources. Despite being > 25 years old the proglacial systems study by Warburton (1990) is still exceptional in that it considered mass changes, rather than simply elevation changes and volume. So too is the consideration of geomorphological work and energy expenditure by Warburton (1993). The PROSA project (see Heckmann et al., 2012) has been quantifying mass fluxes affected by different geomorphological processes in the glaciated upper Kaunertal, Austria. Depending on the geomorphological process, different strategies were chosen to establish sediment budget contributions, e.g. by regionalising local-scale measurements using spatial modelling. The resulting sediment budget is described in detail in Hilger et al. (in review). An interesting part of that study is the fact that the proglacial stream enters a reservoir lake, the sediment infill of which could be quantified using the pre-reservoir topography and sequential TLS- and ALS DEMs acquired during controlled lake level lowering; hence an independent estimate was available to close the sediment budget.

Those studies that do have a more complete picture in terms of processes or subsystems either cover very short time periods (Warburton, 1990) or very small areas (Sanders et al., 2013), or adopt a long-term perspective that does not apply to the short-term reaction to rapid glacier retreat (Otto et al., 2009). Unfortunately, no single methodological approach can provide a holistic understanding of these sedimentary systems.

3.4.1 Field monitoring of fluvial water and sediment fluxes

In intensity of fieldwork and in consideration of the inter-dependent nature of fluvial and hillslope processes, field monitoring studies such as those in Canada by Orwin and Smart (2004), in Mittivakkat, Greenland, as summarised by Hasholt (2005), in Erdalen, Norway (Laute and Beylich, 2014; Beylich and Laute, 2015) and in the Val d'Herens, Switzerland (Micheletti and Lane, 2016), have further developed the proglacial sediment budget study of Warburton (1990, 1993). They combined their field monitoring with geophysical subsurface investigations, terrestrial laser scanning and spatial data analysis, for example. In general, these studies have (i) quantified the relative importance of sediment fluxes to the total budget, (ii) quantified rates of denudation, and (iii) have identified the control of local catchment morphometry on sediment delivery to the main channels and of main valley bottom profile for controlling sediment storage.

Equipment	Variable to be measured
Fluvial	
Continuous measurement of water stage (and construction of stage-discharge rating curve). Ultrasonic better than float for mitigating icing and braided stream channel shifts	Water discharge regime
Manual or automatic (e.g. ISCO) bottled and filtered water samples Continuous turbidity measurement	Suspended sediment flux
Particle tracers (e.g. painted gravel, magnetic particle tracers) Travel velocity of bedforms Helley–Smith type samples Underwater video filming Geophones Impact sensors	Bedload flux
Hillslope	
Collector nets	Rock fall
Hillslope profiles: manually-surveyed or extracted from a digital elevation model (DEM).	Balance of sediment movement

Slope debris thickness: either based on extensive field observations using natural exposures, or else using GPR or other geophysical method.	Volume of sediment

Table 3. Field monitoring equipment that can be targeted at measuring fluvial and hillslope water and sediment fluxes in proglacial systems.

There is a preference for measuring suspended load in proglacial streams, perhaps because it is the easiest sediment flux to measure (Table 3), especially using the proxy of turbidity, but also because it represents an efflux of material from the proglacial system. Suspended sediment discharge regimes have characteristic scales that range across six orders of magnitude with seasonal variations being the largest (Clifford et al., 1995; Leggat et al., 2015) and as conditioned by glacier hydrological system changes as well as proglacial channel (stability, network, wetness and temperature) conditions. Examples of the highest recorded suspended load in proglacial systems include 39,000 ppm (Klimek, 1972), > 1000 tonnes/day in east Greenland (Hasholt, 1992) and 12,000 mg/l at the Gangotri Glacier in the Himalaya (Haritashya et al., 2006). Kavan et al. (2016) have recorded suspended load up to ~ 3000 mg/l on the Ulu Peniunsula of James Ross Isalnd off the Antarcdtic Peninsula.

The different units of measurement highlight different measurement methods and corresponding problems of comparability. Suspended load can constitute up to 100 % of the total sediment transported within proglacial rivers (Fig. 14) and up to 60 % of suspended load has been reported to come from the glacier as opposed to from proglacial hillslopes (Leggat et al., 2015). Micheletti and Lane (2016) have recently shown an (expected) increase in meltwater yield and sediment transport capacity, but an (unexpected) disassociation of sediment yield with these parameters, which they attribute to missing connectivity between rockwalls, hillslopes and the proglacial channel network.

Bedload is very difficult to measure directly (e.g. Rachlewicz et al., in press) and a variety of different techniques addresses different facets of the problem by considering particle impacts, loads or volumes (Table 3). Bedload transport rates in proglacial rivers have been measured at up to 2.85 kg $m^{-3}s^{-1}$ (Goff and Ashmore,

1994) and 1.75 kg $m^{-3}s^{-1}$ (Warburton (1992) or 225 m^3hr^{-1} (Lenzi et al., 2004) in alpine systems, and 66 kg $m^{-1}day^{-1}$ in the arctic (Kociuba and Janicki, 2014), and can constitute up to 90 % of sediment transported in proglacial rivers (Fig. 14). However, (i) in arctic rivers instantaneous and hourly rates are several orders of magnitude lower (see Table 2 in Stott, 2002) and (ii) in supply-limited systems annual bedload yields are more likely < 15 t km^{-2} yr^{-1} (e.g. Beylich and Laute, 2015).

The characteristic high variability of bedload transport in proglacial rivers has been attributed to local channel sediment supply (Hammer and Smith, 1983; Warburton, 1990) as well as to glacier ablation and rainfall intensity (Østrem et al., 1967; Ashworth and Ferguson, 1986; Gurnell and Clark, 1987; Warburton, 1990). In both sub-arctic and alpine environments floods have been noted for being responsible for up to 90 % of all bedload that is discharged from proglacial systems (Nicholas and Sambrook-Smith, 1998; Orwin et al., 2010; Staines et al., 2015). Bedload transport rates can also be determined from repeated topographic surveys (Lane et al., 1995). Such surveys have the advantage of documenting longitudinal and lateral variability rather than just that bedload passing a single point or river cross-section.

Perhaps most importantly field monitoring studies including those of Liermann et al. (2012), Laute and Beylich (2014) and Beylich and Laute (2015) have reported order-of-magnitude differences of their calculated quantities of proglacial denudation and sedimentation rates when compared to those of other studies, and this is probably in part because of (their emphasised) intra- and inter-seasonal variability in proglacial runoff, sediment supply and hence in geomorphological activity (cf. Trimble, 1995). The short-term variability in geomorphological activity within proglacial systems is essentially dependent on sediment availability and triggering events (Schrott et al., 2003), and very important intermittent valley-floor and braidplain storage (e.g. Warburton, 1990; Orwin and Smart, 2004; Bertoldi, et al. 2009).

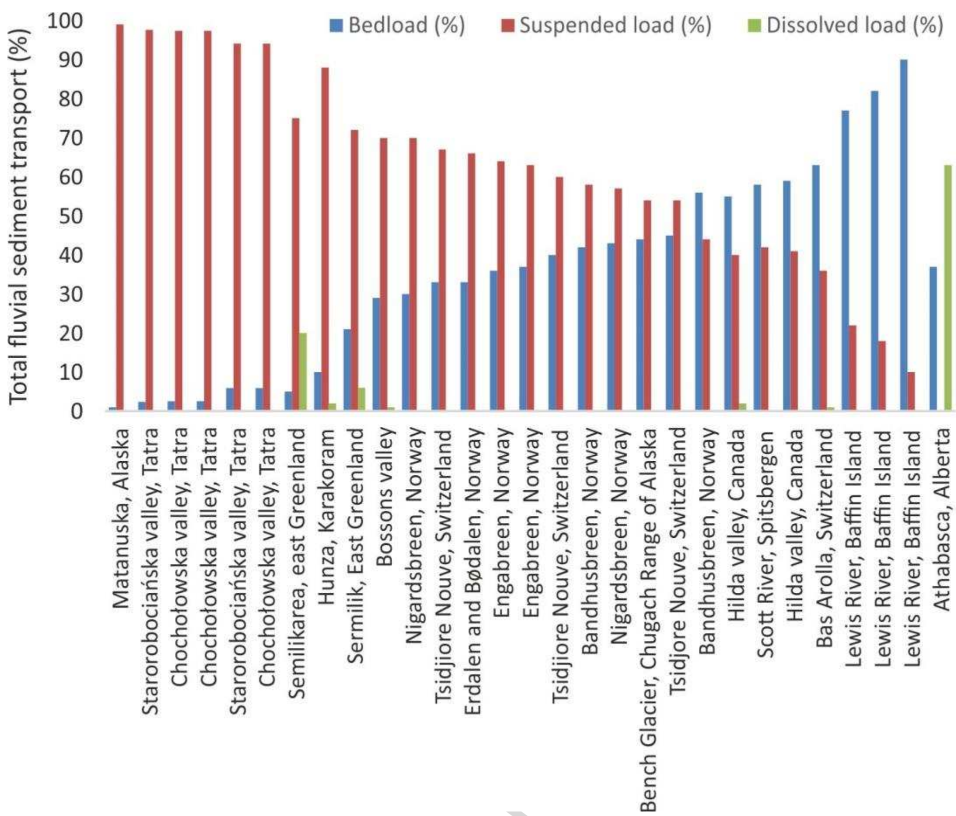

Figure 14: Comparison of the contribution of suspended load, bedload and dissolved load to total fluvial sediment transport in proglacial and alpine systems, as summarised in Table 1 of Orwin et al. (2010) and with additional information from Church (1972), Maizels (1979), Krzemień (1992), Hasholt (1992), Stott (2002), Lenzi et al. (2004), Riihimaki et al. (2005), Kociuba and Janicki (2014), Beylich and Laute (2015); Rachlewicz et al. (in press).

A few field monitoring-based studies have considered time-scales beyond decades by substituting space for time. They have quantified proglacial sediment volumes and sediment budget components both within and beyond LIA glacier limits and they have shown a higher intensity of geomorphological processes post-LIA (Burki et al., 2010; Laute and Beylich, 2012). Furthermore, as Seppi et al. (2015) have reported, some landforms have transitioned, e.g. from glacial to periglacial, in their geomorphological process types.

3.4.2 The geomorphological approach: re-surveying topography

Whilst field monitoring has focussed on selected points on proglacial rivers as pathways from upstream contributing area sediment sources, the best way to quantify geomorphological activity *within* proglacial systems, i.e. including hillslope activity, is to employ a geomorphological approach, i.e. to re-survey topography (Orwin et al., 2010) (Table 1). However, resurveying using traditional methods is exceptionally time-consuming and financially expensive for anything more than a few fixed cross-sections or valley profiles. Advancements in fully-distributed surveying technology such as TLS, ALS and SfM (section 3.2) usually target individual landforms, but have potential to be applied catchment-wide and thus to revolutionise spatio-temporal quantifications of sediment budgets in proglacial systems. They have been an integral methodological part of major multi-disciplinary catchment-wide projects such as PROSA (Heckmann et al., 2012; Hilger et al., in review).

Topographic surveying alone is usually insufficient to close a sediment budget because (temporary) sediment storage in rivers and lakes (within the proglacial system) are unaccounted for. Attempts at closure of the sediment budget of proglacial systems therefore tend to include sediment transport monitoring, either via river discharge data and sediment rating curves or via continuous records such as acoustic sensors for bedload (section 3.4.1; Table 2). Alternatively, inference of sediment transport (bedload) can be made from weirs or similar water intake structures that are used for discharge diversion (Bezinge et al., 1989; Lane et al., 2016), or sediment infill/delta progradation in natural (Liermann et al., 2012) or artificial lakes can be measured. Using records of sediment transport preserved within artificial lakes can be most suitable because the lake level is regularly lowered to increase available storage before the melt season.

Nonetheless, repeated topographic surveys can (i) identify spatio-temporally distributed sediment sources, pathways and sinks in proglacial systems, and (ii) quantify coupling between hillslope elements such as gully heads and debris fans, between valley-floor components such as gravel river banks and fan apexes, for example (Carrivick et al., 2013).

3.5 Geophysical investigations

Geophysical investigations of the composition and internal (sub-surface) structure of proglacial landforms have focussed on frozen ground and hillslopes (Scapozza et al., 2011; Otto et al., 2012; Schneider et al., 2013; Kneisel et al., 2014), ice-cored moraines (e.g. Langston et al., 2011; McClymont et al., 2011) and rock glaciers (Maurer and Hauck, 2007; Ribolini et al., 2010; Monnier et al., 2011, 2013, 2014; Onaca et al., 2013, Seppi et al., 2015). Both ice-cored moraines and rock glaciers can be considered as composite and transitional landforms as relicts from former glaciation (cf. Seppi et al., 2015). Many of these studies emphasise substantial subsurface heterogeneity despite relatively uniform surface conditions. This discrepancy between surface and subsurface conditions prompted Klug et al. (2016) to state that usage of ALS data alone, without geophysics data, is insufficient to detect and understand permafrost and hence proglacial dynamics. Perhaps these varying subsurface conditions, especially between frozen and thawed states, partly explain the apparently sporadic spatio-temporal nature of geomorphological activity within proglacial systems when surveyed solely from above-ground. Specifically, subsurface ice can both increase sediment flux by encouraging sliding and deformation of the ice-sediment mixture but it can also limit sediment flux via adhesion and cementation processes (Huggel et al., 2011; Haeberli et al., 2016).

The above-mentioned geophysical studies are single static surveys, targeted at a discrete landform and have considered landform evolution qualitatively. Two notable exceptions have however related proglacial system geomorphological activity to geophysical properties (of ground conditions) and they have done so by substituting space for time. Monnier et al. (2014) concluded on the basis of their geophysical data interpretations that a rock glacier in Chile is older (< 2000 years old) than the superimposed (LIA?) glacier. Bosson et al. (2015) revealed that accumulations of ridges and furrows composed of an ice-debris mixture have undergone continuous creep since the LIA to the present day, and that unconsolidated moraine and sediment had progressively accumulated on concave valley basin floors. It would be useful to understand the mean rates of change of these landforms, even in elevation terms if not volume and mass. That would require reconstruction of the LIA geometry either by considering the adoption of landform-driven geometrical approaches (e.g. Glasser et al., 2011; Carrivick et al., 2012; Hannesdóttir et al., 2015), or else using geophysically-obtained stratigraphy data to produce restored cross-sections via consideration of plain strain conditions (e.g. Williams et al., 2001).

4. DISCUSSION

Given the critical appraisal that we have made of prevailing conceptual models of proglacial systems (section 2) and the serious shortcomings of the quantitative data available on geomorphological properties of proglacial systems (section 3), we next discuss the challenges, opportunities and key research knowledge gaps with the intention of providing a signpost to future research efforts.

4.1 Challenges

4.1.1 Definition of proglacial systems in terms of geomorphological activity

The areal coverage of proglacial systems is progressively expanding in alpine areas as glacier termini retreat up-valley, but the rate of areal increase could be slowing as glaciers thin, slow down and perhaps even stagnate (e.g. Carrivick et al., 2015; Capt et al., 2016). Furthermore, the areal coverage of proglacial systems does not relate to a single static time frame, because for example in much of the Southern Hemisphere and especially on sub-Antarctic Islands and in Antarctica the occurrence and timing of the LIA is still much-debated. Indeed prominent moraines, such as those surrounding the termini of small glaciers on the Ulu Peninsula, James Ross Island, are most likely related to a mid-Holocene re-advance (Carrivick et al., 2012) (Fig. 6A). In west Greenland many glaciers are still in contact with LIA moraines (e.g. Forman et al., 2007) (Fig. 6B).

Definition of proglacial systems in terms of major landform types, earth surface process types and the rates of the associated surface elevation changes, volume changes and sediment fluxes remain yet to be clarified. Knowledge of whether these rates of change are significantly different from those in areas *beyond* LIA glacier extents is restricted to just a few catchments and to the fluvial domain (e.g. Burki et al., 2010; Laute and Beylich, 2012). There is therefore a challenge to define proglacial systems in terms of process types and rates of processes, relative to those beyond the proglacial area, whilst being mindful that 'proglacial' is defined by location as opposed to periglacial, which is a set of processes that are specifically related to frost action and/or the occurrence of permafrost, or paraglacial, which refers to morphodynamics, time and space (Slaymaker, 2011). Figure 15 illustrates the competing influences of longitudinal sediment fluxes supplied from a glacier and its meltwater, versus lateral fluxes of water and sediment to a valley floor from hillslopes as a conceptual start to this discussion.

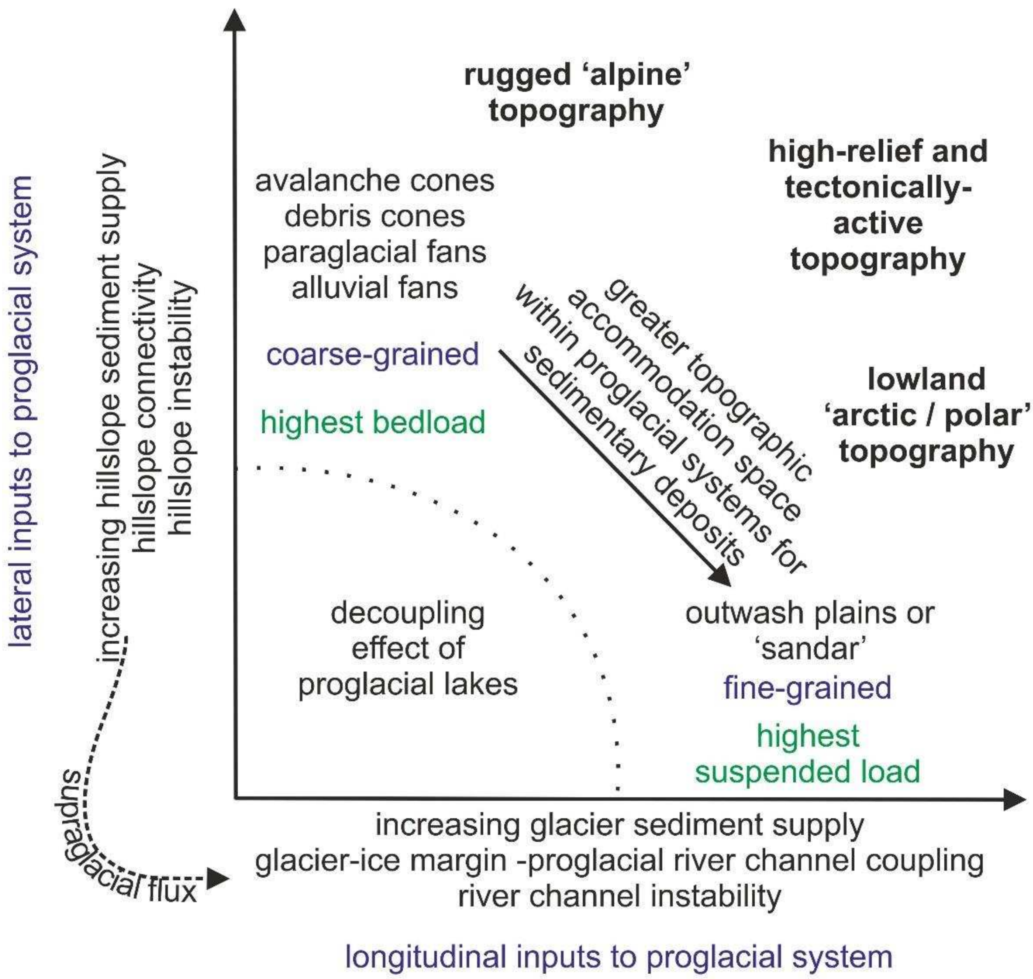

Figure 15 Conceptual definition of proglacial systems and major landform types by relative importance of longitudinal and lateral sediment fluxes from glaciers and from hillslopes, respectively.

There is an open question as to whether paraglacial adjustment of proglacial systems is accelerated/intensified or decelerated/attenuated by climate change. Paraglacial dynamics do not necessitate climate change except for the change that triggers deglaciation. To illustrate this argument, the paraglacial response could be represented within two scenarios, each considering the relative importance of longitudinal versus lateral sediment fluxes (Fig. 15): (A) Glacier mass loss increases water discharge in the proglacial river and this is a transient response so will decrease in the medium to long term. The increased meltwater discharge is likely to (i) intensify the undercutting of lateral sediment storage landforms, either moraines or primary paraglacial storage such as talus or debris cones, thereby keeping slopes steep and incision rates high. (ii) The short-term higher discharge (as long as the glacier exists) increases longitudinal connectivity (summer precipitation on top of higher glacial melt maxima). Alternatively, (B) if there is a trend of increasingly heavy precipitation, the paraglacial response could be exacerbated or intensified at least on hillslopes. Since the increasingly heavy precipitation would not likely compensate for a declining glacial component of the hydrograph, transport capacity of the proglacial river would progressively decrease, meaning a net increase in the relative importance of hillslope contributions to the proglacial system budget.

An opportunity for tackling this composition analysis is presented by slope-based classifications, such as that of Loye et al. (2009). Using such an approach for the Mattertal catchment (Fig. 16A) reveals that the 99 proglacial systems are likely composed of scree-dominated hillslopes (57 % of total area), moraines and debris fans (28 %), fluvially-dominated land surface (10 %) and bedrock (5 %). However, there is large variability in composition between these proglacial systems (Fig. 16A). Another benefit of such an inventory of proglacial systems would be to enable selection of representative ‘index sites’ for intensive field measurements, for example as per the sediment budget in cold environments (SEDIBUD) IAG/AIG Working Group protocols (Beylich, 2016), and as done with glacier mass balance (e.g. Zemp et al., 2009) and snow-line monitoring (e.g. Chinn, 1995), for example.

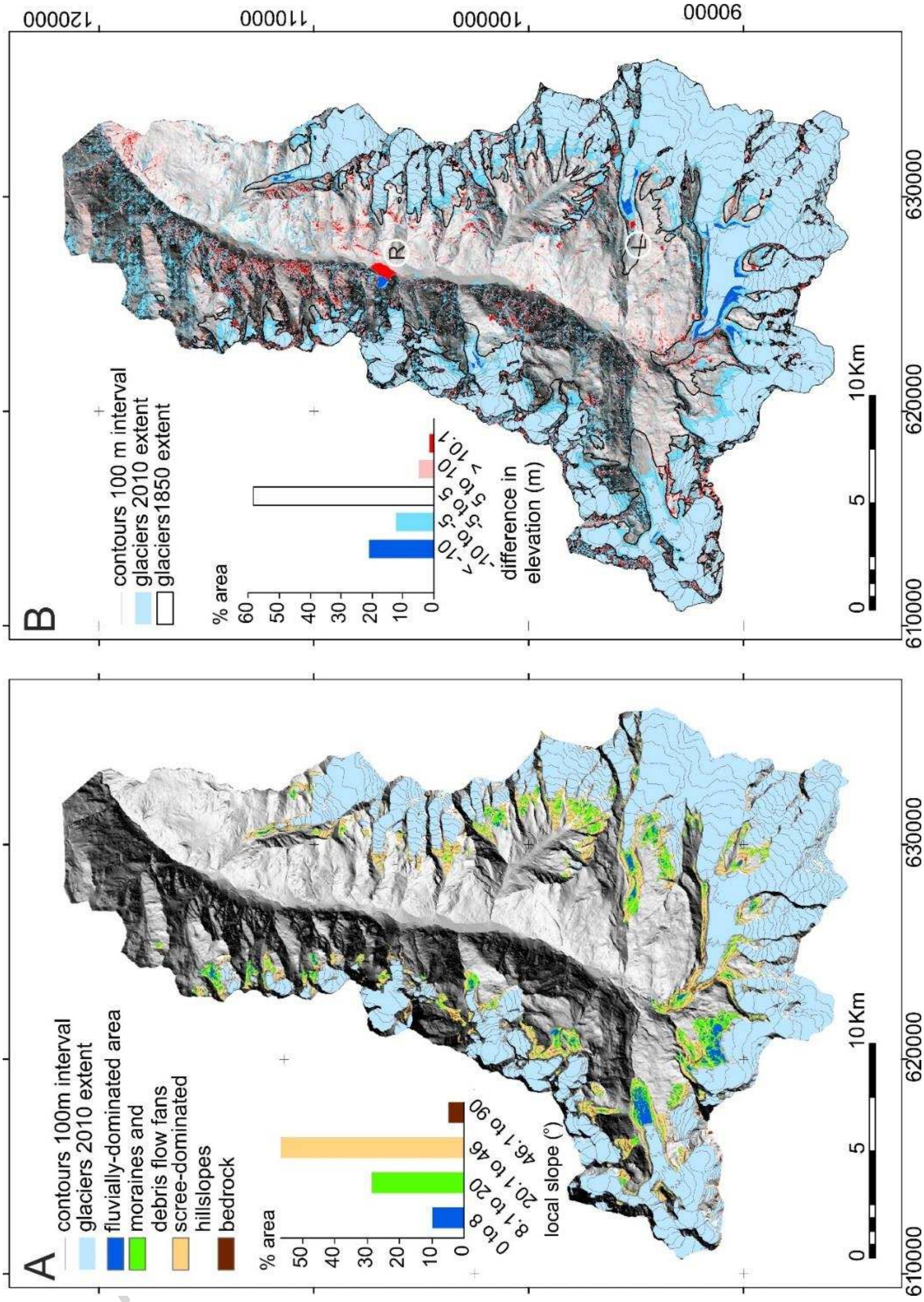

Figure 16: Planform extent and landform composition of proglacial systems in the Matteral catchment, Switzerland, as suggested by a slope-based classification (A). Differencing of two DEMs (1980s and 2010) courtesy of Mauro Fischer (Fischer et al., 2014a, b) reveals major geomorphological activity, for example the Randa rockslide (R), plus many other elevation changes that are above the ± 5m level of detection applied (B). Note that elevation changes *beyond* proglacial systems probably represent vegetation growth or errors in DEMs due to vegetation and that the inset graph is only for changes *within* proglacial systems.

4.1.2 Scaling-up of plot-scale field measurements

Plot-scale field measurements of surface changes (section 3.4.1) have highlighted several problems:

- The inference of sediment fluxes from surface changes is hampered by the fact that there are surface changes due to the melt-out of ground ice that are not associated with sediment transfer. Hence, as long as no morphometric approach exists to automatically infer ground-ice borne surface changes, surface changes have to be interpreted by an expert in order to detect (or reject) an associated geomorphological process and to compute the corresponding sediment flux accordingly.
- Where surface changes are due to sediment transport, it is still necessary to infer the spatial pattern of erosion and deposition, i.e. a DoD does not directly identify transport pathways and transport distance. Approaches towards identifying transport pathways could involve the downstream accumulation of DoDs using flow routing approaches (Pelletier and Orem, 2014; Neugirg et al., 2016), but have not been fully developed to our knowledge.

Nonlinearity and hysteresis in channel sediment transport affect the transferability of rating curves for suspended load and bedload to other times and places. Bedload hysteresis patterns have been interpreted in a glacier-fed catchment in Italy to be due to the availability of sediment within the channel network during snowmelt, or due to a time lag in major contributions of glacier-borne sediments being advected to the catchment outlet where sediment flux is measured (Mao et al., 2014; Mao and Carillo, in press).

A lack of spatially-distributed measurements (section 3.4.2) of geomorphological changes precludes establishment of catchment sediment budgets (i.e. including within-catchment sediment fluxes) beyond integration of the fluvial sediment flux measured at the catchment outlet. Quantification of geomorphological processes such as rockfall and debris flows has been limited spatially to the plot scale (e.g. Matsuoka, 2008; Sass, 2005). This means that predictive models of the relation of each transport process to its controls are necessary to generalize from a few measurement sites to a landscape (Dietrich et al., 1982). Accordingly, Heckmann et al. (2016b) transferred or 'regionalised' their plot scale rockfall measurements in the Kaunertal, Austria, by using spatial modelling to enable an evaluation of the contribution of rockfall to different landforms. Vehling et al. (2016) extended this approach to consider spatially-distributed rockfall rates in the same study area. Messenzehl et al. (in press) have concluded from two types of statistical model that regional permafrost distribution, time since deglaciation, and rock mechanics are the most important controls on paraglacial rockfall activity.

Whilst such efforts at up-scaling or statistical modelling are to be applauded for producing quantitative estimates of geomorphological composition, activity and controls, it must be noted that the plot-scale field measurements on which these models are built have highlighted problems of:

- considerable inter- and intra-catchment variability (e.g. Carrivick and Rushmer, 2009; Carrivick et al., 2013; Temme and Lange, 2014),
- targeted surveys pre-empting where geomorphological activity is going to occur and perhaps ignoring that elsewhere,
- the effect of survey interval on what type and intensity of geomorphological activity is detected (e.g. Sailer et al., 2012, 2014; Carrivick et al., 2013). Furthermore, we can expect seasonal differences in process activity in proglacial areas similar to what has been described by Bechet et al. (2016) and Neugirg et al. (in review).
- complexity in substituting space for time due to significant lateral (hillslope) fluxes as well as down-valley (fluvial) fluxes.

Ground-based surveys cannot be readily compared to elevation changes derived from airborne or satellite-based surveys because these are not sufficiently accurate to capture typical (inter-seasonal and inter-annual) geomorphological activity; the best (lowest) level of detection being between ± 1 m and ± 5 m from photogrammetry-derived products (Schiefer and Gilbert, 2007; Fischer et al., 2014a, b; Micheletti et al., 2015). Nonetheless, Figure 16B shows that major coherent elevation changes representing geomorphological activity such as the Randa rockslide can be detected. Detectable changes in Figure 16B that are within proglacial systems are less spatially coherent but generally represent net loss of material adjacent to contemporary ice margins and gains of material in some proglacial systems farther down-valley. With consideration of the likely geomorphology as in Figure 16A, these elevation changes can be interpreted to be ice-cored moraine degradation and fluvial sedimentation, respectively.

4.2 Knowledge gaps and research perspectives

4.2.1 Spatial coverage and representativeness

There are few baseline data from which short-term changes within proglacial systems can be assessed and a shortage of the number and variety of sites that have capability to support quantitative assessments of short-term changes in a decadal and centennial context. We argue that the emergence of photogrammetric and surveying techniques should provide researchers with an opportunity to turn to more holistic approaches. Firstly, geomorphological change can be monitored and quantified with high resolution and high accuracy, and at large spatial scales (extent) rather than limited to a few test plots. Secondly, high-resolution topography forms an excellent basis for the generation, interpretation and analysis of geomorphological maps. Thirdly, the geomorphological history of whole proglacial systems can be documented in more detail and more precise information on past topography using historical aerial photos dating back several decades; this allows for including path-dependence in geomorphological studies of proglacial landscape evolution. Specifically;

- Surveys must be made both within and beyond proglacial systems to test a hypothesis that proglacial areas are especially dynamic, or transient landscapes.
- Surveys must be made with distance from a glacier terminus to test the hypothesis that geomorphological activity in proglacial systems decreases with time due to landscape stabilisation.
- Surveys must be conducted on different landform types within proglacial systems so that indicative rates can be gained for upscaling or ‘regionalisation’ spatially, and thus for suggesting the nature of future responses of proglacial systems to climate change.

Polar regions, the glaciated mountains of North and South America and the Himalaya are all very under-represented in the proglacial geomorphological literature.

4.2.2 Temporal evolution

The hypothesis that ‘proglacial systems are rapidly changing’ needs to be more robustly tested temporally to elucidate the relative importance of glaciological versus weather controls, e.g. intense rainfall, snowmelt predominantly driven by air temperature, permafrost decay. Once rates of change in proglacial systems have been determined across regions for multiple catchments, they could be compared to the rates of change of glacier systems, either ice volume loss, or more simply terminus retreat rates or ice surface lowering rates.

Within catchments, rates of geomorphological processes associated with the temporal evolution of proglacial systems will require identification of chronosequences, most likely from historical maps and historical aerial photographs, and DEM analyses to derive topographical variables. Chronosequences are commonly used in proglacial areas but for soil development studies (e.g. Egli et al., 2006) that specifically target geomorphologically-stable parts of the landscape. Interestingly, Temme and Lange (2014) concluded in their cross-disciplinary bio-geomorphological studies on stable and unstable parts of proglacial chronosequences that time since glaciation and local topography could only explain half of the variance in soil conditions. The remaining half was probably due to ‘geomorphological history and regime’. Similarly, sediment delivery/transfer in proglacial systems has been suggested to be strongly affected by glacial inheritance (Dell’Agnese et al., 2015). These suggestions imply that topography, which can be inherited, favours or hampers sediment transfer from newly exposed proglacial sediment stores to the proglacial system outlet.

Additionally, we might consider that rates of glacier retreat (at least at the present-day) are greater than the rate of paraglacial adjustment, so the proportion of proglacial area that is subject to heavy reworking as opposed to the proportion that is approaching stability should be increasing. For example, if deglaciated lateral moraines are incised to a maximum within a period of c. 50 years after deglaciation and stabilise within another 30 to 90 years (Curry et al., 2006) then during these 80 to 140 years alpine glaciers will have retreated by 10^2 to 10^3 m, exposing areas in the order of 10^3 to 10^6 m², corresponding to average rates in the order of 1 to 10 m yr^{-1} of terminus retreat and 10^2 to 10^4 $m^2 yr^{-1}$ area exposed, respectively. That implies that, for a valley glacier with large LIA lateral moraines, only the most distal parts of kilometre-long reaches of the glacial valley would be approaching stability, while the majority of the proglacial system hillslopes would continue being heavily reworked.

A further implication of this trend in proglacial system evolution is that an insight into the future status of many proglacial systems could be gained by substituting space for time (Fig.17), and in particular examining those proglacial areas that have recently become entirely disconnected from a glacier, either because the glacier has disappeared entirely, or because an ice-marginal lake has formed.

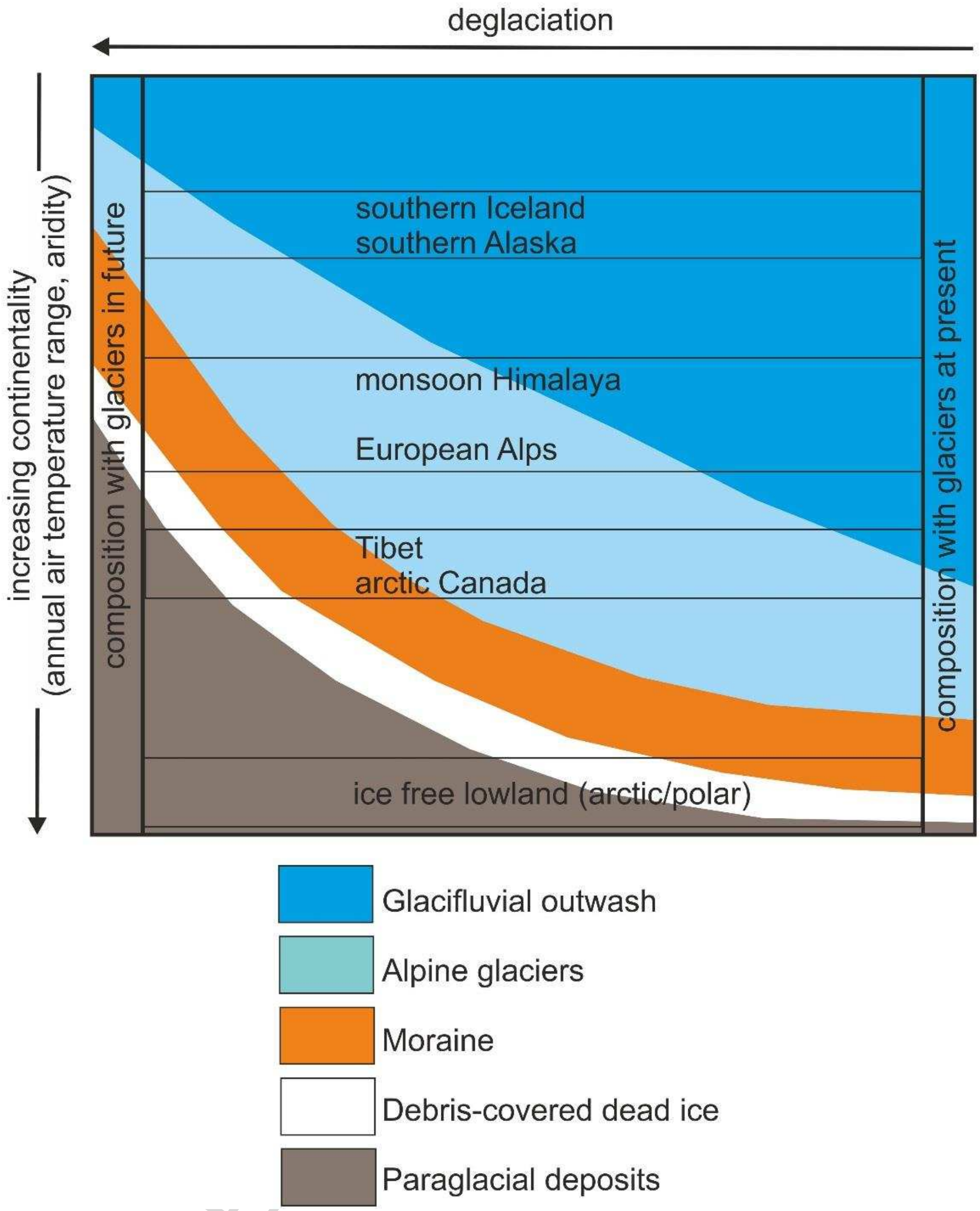

Figure 17. Model of typical spatio-temporal percentage landcover composition of proglacial systems, where horizontal bars denote present-day regional examples and where vertical bars denote degree of glaciation. Sediment transfer becomes more interrupted or disconnected or even decoupled as the dominance of glacial processes diminishes. No hierarchy is implied in the ordering or colouring of the components. This figure was inspired by that representing the glaciers of the Ben Ohau range in New Zealand, by Benn et al. (2003).

4.2.3 Connectivity

We have argued that change is inherent to proglacial systems in times of ongoing deglaciation; this automatically implies that sediment connectivity can itself also change. Given the rates and types of geomorphological activity within proglacial landscapes (section 3), we assume that the latter are prone to changes in system structure and hence sediment connectivity. Such changes can be the result of single events or developments that are driven by external forcing (e.g. the occurrence of a mass movement), or are not driven by external forcing (e.g. the generation and vanishing of proglacial lakes). Profound changes to the system structure can also result from system-internal dynamics such as the availability of new sediment sources (e.g. by permafrost degradation), the exhaustion of older sediment stores, or a change in the drainage pattern leading to the (de-/re-)coupling of sediment sources with the channel network (e.g. Heckmann et al., 2008). Additionally, functional connectivity would be expected to change with changes in external forcing (e.g. an increase or decrease in precipitation) and in the amount of discharge in glacial meltwater channels the increase of which is a transient response to global warming (e.g. Moore et al., 2009).

It is therefore rather unexpected that recent studies have shown that an increase of transport capacity due to enhanced meltwater runoff did not lead to a corresponding increase of sediment flux (Lane et al., 2016; Micheletti and Lane, 2016). These authors reasoned that a lack of connectivity between (new) hillslope sediment sources and the channel network had developed. These studies highlight that connectivity is related to a system's sensitivity to changes, specifically that it effectively moderates the downstream propagation of sediment pulses from upstream (and upslope) sediment sources. As a consequence, connectivity has to be taken into account when particular results related to sediment generation and transfer (section 3) need to be integrated. Connectivity could also be a key for a better upscaling of sediment budgets (Slaymaker, 2006).

Sediment connectivity assessment is still a challenge. There are indices that can be computed using a high-resolution DEM and, where applicable, flow impedance information derived from landcover maps. The index of connectivity (IC) published by Cavalli et al. (2013) represents an adaptation of an original index (developed by Borselli et al., 2008) to high mountain areas where high-resolution DEMs exist. This index has been applied to proglacial areas in a semi-quantitative fashion (Goldin et al., 2016; Micheletti and Lane, 2016; Cavalli et al., in review) for the appraisal of spatial patterns of high or low sediment connectivity. These studies emphasise disconnection by lateral moraines (Goldin et al., 2016; Lane et al., 2016) or rock glaciers (Micheletti and Lane, 2016), and the role of the width and topography of the proglacial area that governs the accommodation space available for sediment storage landforms (Cavalli et al., in review). A study by Vigiak et al. (2012), whilst not within a proglacial system, points out that the IC can be quantitatively related to specific sediment yield. Another index that can be computed from a high-resolution DEM has been proposed by Lane et al. (2016). It evaluates the changes in flow accumulation on a high-resolution DEM whose sinks are progressively being filled to increasing depths; where the resulting curve becomes asymptotic, a distinction between sinks that are due to DEM uncertainty and "real" sinks that disconnect sediment flux is inferred. Common to those indices is their static nature, i.e. they are computed on a DEM with no reference to forcings of different magnitude, for example.

Therefore, we propose two research avenues that should offer good opportunity to improve our understanding of connectivity and changes in connectivity in proglacial areas. Firstly, there should be more use of time series of historical aerial photos plus DEMs and geomorphological maps derived from the latter for diachronic comparisons of sediment connectivity (see also Heckmann et al., 2014). Secondly, there should be further investigation of the significance of static, DEM-based indices with respect to sediment transfer and delivery driven by different magnitudes of hydrometeorological or glacifluvial forcing.

5. SUMMARY, FUTURE OPPORTUNITIES AND CONCLUSIONS

5.1 Conceptual models, geomorphological maps and interpretations of DEMs

In summary, the definition of proglacial areas as 'the area between the LIA glacier extent and a contemporary glacier margin (Schiefer and Gilbert, 2007; Heckmann et al., 2012) has been championed because it has utility for defining a specific relation to glacier retreat and to subsequent paraglacial dynamics. This definition also has the advantage that it sets a physical boundary to a proglacial area, which in many cases is clearly visible on the ground and also within remotely-sensed data, which is of great importance to studies wishing to spatially-discriminate landforms and processes.

Prevailing conceptual models of proglacial systems have proven essential as a framework for designing field monitoring programs of water and sediment fluxes. However, they are generally too simple for understanding geomorphological activity because sediment sources, fluxes and sinks are not linear in space, e.g. with distance from glacier, or time, e.g. with time since deglaciation. Furthermore they are inter-dependent, connected, and with positive and negative feedbacks for each other, e.g. scour and fill at a point, e.g. source and sink zones of a single event. The balance between lateral fluxes from hillslope activity and longitudinal fluxes from glacial sediment sources and meltwater has yet to be explicitly considered but could be expected to be critical to developing the three-dimensional shape, or 'landscape evolution' of proglacial systems.

The paucity of geomorphological maps published for proglacial systems is a problem because the way in which these systems have developed since the LIA and the way in which they will respond to ongoing climate change depends on the extrinsic controls of geology and climate for affecting glacier mass balance and permafrost, and the intrinsic control of topography for its role in pre-conditioning paraglacial processes. Until the inter- and intra-catchment variability is quantified, proglacial landscape development will not be able to be modelled.

The increasing availability of high-resolution DEMs is likely to be a part of this mapping effort, and will additionally assist with transferring or 'regionalising' local field measurements, as well as for changing the scale of process of reference, or 'upscaling'. However, whilst analysis of elevation change is quantitative and fully spatially-distributed, the associated interpretation of landforms or of geomorphological processes, both *within* a proglacial area (e.g. Carrivick et al., 2013) and *between* deglaciated catchments (e.g. Strunden et al., 2015) is subjective and relies on interpreting mass transfer from vertical and horizontal displacements (e.g. Micheletti et al., 2015).

Caution must be exercised when interpreting elevation changes at-a-point or along transects because they can be associated with sediment erosion/deposition but they can also be produced by non- sediment transfer (and thus non-mass change) processes, such as permafrost degradation/aggradation. Elevation changes must also be scrutinised carefully because two different processes may overlap spatially. Ignoring preconceived landform boundaries is perhaps preferential approach for making full use of fully-distributed topographic survey data and for application in spatially-complex and temporally-dynamic proglacial systems. It is therefore exciting to note that there are emerging opportunities for automated pattern, spatial coherence or clustering, or neighbourhood analysis and that these need not be conceptually complicated nor computationally intensive.

5.2 Quantification of short-term changes within proglacial systems

Spatio-temporal variability in inter- and intra-catchment controls on proglacial runoff and consequent geomorphological response makes comparing rates of change measured at one site to those at another difficult. This does not diminish the importance of doing so. The variability is not simply a function of process and environment, and landform position, age and connectivity, but also due to survey time(s) and survey intervals. There is a growing recognition via geophysical investigations that the subsurface character, especially the presence of ground ice, exerts a strong control landform evolution and geomorphological activity.

Distributed topographic surveys have demonstrated coupling between upper and lower valley floor components of a proglacial fluvial system as a result of major glacier meltwater discharge events, and a spatial association between hillslope activity as responses to major snowmelt and precipitation events. The relative influence within a proglacial system of these longitudinal and lateral influences is perhaps a defining factor for considering proglacial landscape evolution.

5.3 Future challenges and opportunities

There is an urgent need for inventories of proglacial systems, to form a baseline from which changes could be detected. At present, LIA extents are only locally resolved:

for example for southern Baffin Island, arctic Canada (Svoboda and Paul, 2010), for southern Vatnajökull, Iceland (Hannesdóttir et al., 2015) for the Qilian Shan part of Tibet (Shiyin et al., 2003) and the Jotunheimen part of southern Norway (Baumann et al., 2009). Proglacial system inventories will most plausibly be derived via DEM analyses with elevation data derived from ALS. Subsequent re-surveys could be achieved with ALS, or with SfM methods. Emerging technologies of automatic surface extraction will also likely prove useful because they utilise high-resolution satellite imagery, which covers virtually the entire World and does so at multiple time frames.

Once inventories have been reconstructed there will be a need for objective and automatic classification of land cover and landforms. These classifications will inform interpretations of elevation changes and geomorphological activity, local stability assessments and hence ultimately modelling of proglacial system evolution.

There is a research gap for utilising historical aerial photographs and historical maps to reconstruct the development of proglacial systems. Such studies could usefully examine the questions raised in this review of path dependence or topographic inheritance, and (changing) connectivity. A longer-term perspective would increase the likelihood of observing extreme events, allow for a more profound estimation of magnitude-frequency relationships and trends, and increase the chance to observe the cumulative effect of slow, otherwise undetectable processes.

Comparison of records of water discharge and inferred sediment transport capacity with sediment yield should be made (cf. Micheletti and Lane, 2016) wherever possible to test the hypothesis that proglacial systems are largely transport-limited systems and to examine if such a discrepancy between sediment transport capacity and sediment yield is due to (changes in) proglacial system connectivity.

Emerging high-resolution and high-accuracy topographic surveying methods will allow future detection of differences in geomorphological activity between proglacial areas under different conditions: different glacial meltwater discharge, different age-since-deglaciation, different regional/local climate in otherwise similar areas, and thus help to answer the question as to whether (or, where) climate change intensifies paraglacial response.

5.4 Conclusions

Rates of change to landforms in proglacial systems have generally confirmed the paraglacial exhaustion model but the decline of the volume of remaining sediment also reflects stabilisation of the storage landform (Curry et al., 2006). Declining connectivity between rockwalls, hillslopes and the proglacial channel network has been invoked to explain an increase in meltwater yield and sediment transport capacity with warmer weather conditions but an (unexpected) disassociation of sediment yield can exist with these parameters (Micheletti and Lane, 2016). Some landforms have transitioned, e.g. from glacial to periglacial, in their geomorphological process types.

These intra-catchment conditions of inherited topography, sediment supply / exhaustion and (changes in) connectivity explain why the net efflux of sediment from proglacial systems, as represented in sediment yield, can be inversely proportional to drainage basin area as well as due to the factors of low vegetation cover, high surface erosion and mechanical denudation. Furthermore, where hillslope processes dominate there is greater opportunity for deposition of sediment with transport distance on slopes, as well as within channels and on floodplains. Thus specific sediment yield can decrease with increasing catchment area despite absolute load increasing.

Proglacial systems with thick moraine accumulations are especially common in alpine environments because cirques or shallow-gradient valley floors that are inherited from previous glacier activity do not support an efficient link between the glacial and the hydrological transport systems. Proglacial systems with strong glacier–permafrost interactions are especially common in arctic and high-altitude environments because polythermal glaciers can have indistinct and even intercalated boundaries with surrounding hillslopes and glacier forefields and because permafrost in hillslopes introduces ice into the sediment system.

There is a scarcity of quantitative data on short-term changes in proglacial systems. Field monitoring of water and sediment fluxes is intensive and so necessarily restricted to a handful of sites, most notably those part of the SEDIBUD network. To date topographic survey data sets are of short duration, too targeted (spatially-restricted), or biased in geographical coverage. There is little baseline data from which short-term changes within proglacial systems can be assessed and a shortage of the number and variety of sites that have capability to support quantitative assessments of short-term changes in a decadal and centennial context.

Forecasting changes in geomorphological activity within proglacial systems, and by inference of hazard occurrence, magnitude and frequency, has occurred to date via statistical upscaling models. Future advancements must therefore depend on process-based modelling. That process-based knowledge demands an improved understanding of geomorphological process-response systems and their impacts on human activity. The former of these should now be possible with recourse to both historical and emerging high-resolution topographic datasets covering proglacial systems.

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