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Decadal-scale changes of the Ödenwinkelkees, central Austria, suggest increasing control of topography and evolution towards steady state

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Abstract

Small mountain glaciers have short mass balance response times to climate change and are consequently very important for short-term contributions to sea level. However, a distinct research and knowledge gap exists between (i) wider regional studies that produce overview patterns and trends in glacier changes, and (ii) in situ local scale studies that emphasise spatial heterogeneity and complexity in glacier responses to climate. This study of a small glacier in central Austria presents a spatiotemporally detailed analysis of changes in glacier geometry and changes in glaciological behaviour. It integrates geomorphological surveys, historical maps, aerial photographs, airborne LiDAR data, ground-based dGPS surveys and Ground Penetrating Radar surveys to produce 3D glacier geometry at thirteen time increments spanning from 1850 to 2013. Glacier length, area and volume parameters all generally showed reductions with time. The glacier equilibrium line altitude increased by 90 m between 1850 and 2008. Calculations of the mean bed shear stress rapidly approaching less than 100 kPA, of the volume-area ratio fast approaching 1.458, and comparison of the geometric reconstructions with a 1D theoretical model could together be interpreted to suggest evolution of the glacier geometry towards steady state. If the present linear trend in declining ice volume continues, then the Ödenwinkelkees will disappear by the year 2040, but we conceptualise that non-linear effects of bed overdeepenings on ice dynamics, of supraglacial debris cover on the surface energy balance, and of local topographically-driven controls; namely wind-redistributed snow deposition, avalanching and solar shading, will become proportionally more important factors in the glacier net balance.

Keywords: mountain glacier; glacier geometry; response time; ELA; LiDAR

Introduction

Small mountain glaciers are extremely important indicators of climate change due to their fast response times (Raper and Braithwaite, 2006; Dyurgerov and Meier, 2000; Haeberli et al., 2007). On a global scale small mountain glaciers (i) are so numerous that they have a significant contribution to the world's total ice volume and may be a notable source of error if excluded (Bahr and Radic, 2012) and, (ii) are much more important for short term contributions to sea level than that from Greenland or Antarctica (Raper and Braithwaite, 2006; Glasser et al., 2011). On a local scale, mountain glacier ice mass variations and the consequent changes in meltwater runoff magnitude and timing is of concern for understanding proglacial geomorphology and sediment fluxes (e.g. Carrivick et al., 2013; Carrivick and Rushmer, 2009) and riparian ecology, and in application for power generation, irrigation, and tourism.

Understanding small mountain glacier mass balance, especially processes representing responses to climate, is most ideally achieved with local direct measurements; i.e. distributed surface elevation and ice and snow density measurement programs. Such field studies are (i) limited to a few tens of points at most on a glacier surface so require extrapolation to the whole glacier, (ii) are expensive in time and resources required and (iii) are consequently rare; in the European Alps there are ~ 25 glaciers now with these programs (Vincent et al., 2004; Zemp et al., 2008). Furthermore, direct mass balance measurements were not initiated until late 1940s and many of these records are intermittent.

Given the deficiencies in the coverage of direct mass balance data, there have therefore been a series of efforts to analyse changes in glacier terminus position, which are often reported as glacier length changes (in the European Alps; e.g. Hoelzle et al., 2003; Klok and Oerlemans, 2004; Oerlemans, 2005; Hormes et al., 2006; Zemp et al., 2008). Indeed Zemp et al. (2008) report on over 200 European Alps glaciers with > 18 terminus position measurements since 1850. However, prior to 1890 there were only ~10 European glaciers with measured terminus positions. Furthermore, there is a strong bias towards larger glaciers in the length measurement sample due to the remote location of most small glaciers. Perhaps most crucially, whilst glacier length records can correlate well with weather patterns, albeit with a lag time (or 'response time'), dynamic reactions of a glacier to climate change must be assessed with consideration of area and volume (Harrison et al., 2009 in Paul, 2010).

Analyses of changes in glacier area and ice surface elevation have relied on (i) field-based geomorphological surveys of moraine crests and contemporaneous ice limits, (ii) remote sensing; either photogrammetry for local-scale studies or image-processing of satellite data, for regional studies (Paul et al., 2002, 2007a; Paul and Haeberli, 2008), the latter which restricts studies to the 1970s onwards, or (iii) climate-driven ice dynamics models. One of the simpler remote sensing approaches is that of the repeated observations of glacier transient snow lines (Chinn, 1995), which at the end of the summer can approximate the ELA, and which can permit high resolution spatiotemporal analyses (e.g. Carrivick and Chase, 2011). Perhaps the ultimate expression of the remote sensing approach is represented by glacier inventories, such as the Austrian glacier inventories of 1969 (Gross, 1987) and 1998 (Lambrecht and Kuhn, 2007) and the recently revisited French glacier inventories (Gardent et al., 2014), for example. Whilst these inventory-based studies provide insight into spatial variability in glacier changes, they usually consider changes in glacier surface elevation between just two points in time (e.g. Kääb et al. 2002; Paul et al. 2007a; Hoezlze et al. 2007; Lambrecht and Kuhn, 2007) or for glaciers aggregated regionally (e.g. Paul, 2002; Stocker-Waldhuber et al., 2012).

There is therefore a lack of consideration of decadal-scale changes at individual glaciers, and a lack of information on absolute volume, which necessitates knowledge of ice thickness and bed topography (Paul et al, 2007b). Knowledge of ice thickness and hence of bed elevation are necessary not only for computing absolute changes in glacier volume, which is of necessity when projecting glaciological changes into the future (c.f. Huss, 2012), but also for determining ice flow mechanics, which is a crucial step in understanding processes that could explain reconstructed geometric changes; see the critical review on previous ice dynamics reconstructions by Carr and Colman (2007). However, the complexity of obtaining direct data in the field on ice thickness and bed elevation, most usually from Ground Penetrating Radar (GPR), means that even then considerable uncertainty can be inherent in the data and interpolation between survey lines is necessary (e.g. Binder et al., 2009). Thus studies of volume changes in mountain glaciers usually report relative volume changes (e.g. Bauder et al., 2007; Villa et al., 2007; Stocker-Waldhuber et al., 2012) without any knowledge of distributed ice thickness and hence of absolute volume. Indirect approaches to estimating absolute volume have been developed; namely (i) relatively simple volume-area scaling (Bahr et al., 1997), (ii) reconstructions using moraine crests, (iii) flow line ‘perfect-plasticity’ models (e.g. Li et al., 2012), and (iv) more complex physics-based models (e.g. Farinotti et al., 2009). Climate-driven ice-dynamics models that reconstruct ice volume are limited to very discrete periods of time, usually the Last Glacial Maximum (LGM) and the Little Ice Age (LIA), where geological evidence in the form of moraine ridges and trimlines can test the model (e.g. Golledge et al., 2012; Harrison et al., 2014).

Overall, whilst a few hybrid studies exist that have integrated multiple types of data (i.e. remote sensing data with mass balance data; e.g. Haeberli et al., 2007; Abermann et al., 2009), a distinct research and knowledge gap exists between (i) wider regional studies that emphasise spatial heterogeneity and complexity in glacier responses to climate and (ii) in situ local scale studies that develop process-understanding. Therefore, as Haeberli et al. (2007) and Linsbauer et al. (2012) both comment, there is a need for spatio-temporally detailed studies on individual glaciers. In order to achieve spatiotemporally detailed studies, it is possible to augment remote sensing and mass balance data with primary geomorphological surveys and compilation of secondary datasets such as historical maps and historical oblique field sketches. Use of historical maps for glacier reconstructions is exemplified by Hall et al. (2003), Villa et al. (2007), Zumbühl et al. (2008), and Knoll et al. (2009), for example. Use of historical oblique photographs and sketches was made by Steiner (2001) using mono-plotting for georeferencing to reconstruct the Rhone glacier from 1874 onwards.

The aim of this study is to create a spatiotemporally detailed analysis of changes in glacier geometry and to infer what these mean for glaciological behaviour. This study from the Glockner region in central Austria achieves this aim by integrating geomorphological surveys, historical maps, aerial photographs, airborne LiDAR (ALS) data gridded at 1 m resolution, ground-based dGPS surveys and Ground Penetrating Radar surveys to derive the glacier bed elevation and hence ice-thickness. Altogether these datasets produce measurements of three-dimensional glacier geometry at thirteen time increments spanning from 1850 to 2013. Whilst acknowledging the inherent uncertainty in these data sets, this study contributes to the very small number of studies achieving this spatiotemporal resolution of study of glacier changes.

Study area

The Ödenwinkelkees glacier (~47°7′00″ 12°30′00″E) is located within the Glockner mountain range within the Hohe Tauern National Park in the Salzburg province of central Austria (**Figure 1**). The Ödenwinkelkees was chosen for this study for four main reasons: (i) availability of (analogue) historical data pertaining to the glacier configuration at high spatial and temporal resolution (**Table 1**), (ii) long term monitoring of the terminus position of the Ödenwinkelkees providing dated moraine ridges (Slupetzky and Teufl, 1991), (iii) availability of airborne LiDAR (ALS) data (Carrivick et al., 2013) which provided excellent positional information for georeferencing the analogue data, (iv) two GPR surveys that together provided reasonable spatial coverage and from which an estimate of ice thickness and bed elevation could be made to derive absolute glacier volumes (**Table 1**). Access to the area is most convenient via the Rudolfshuette, which is the ex-Austrian Alpine Club headquarters, a site of continuous weather and snow monitoring by ZAMG (Austrian Meteorological Institute) and ÖBB (national train company) portal, and is now a mountain hotel.

Regional climate records for the Alps are available from the 'HISTALP' database (ZAMG, 2014). HISTALP data aggregates multiple weather station records and those from the NE region, which covers the Odenwinkelkees, show no change in precipitation and ~ 1 ^{o}C air temperature increase between 1760 and 2008. Local weather data from the Rudolfsheuette was obtained as monthly means, but it only covered the 1960s and 1970s, and comparison of this local dataset with the regional HISTALP means yielded very poor correlation in air temperature and no correlation in precipitation.

Georeferencing

All analogue historical maps (**Table 1**) and aerial photographs were georeferenced to the airborne LiDAR (ALS) data as projected in the Universal Transverse Mercator World Geodetic System (UTM WGS 1984 33N) and gridded at 1.0 m resolution, using a spline fit between identifiable points on the terrain that were assumed to be static; such as the Weizsee dams, buildings, roads, major gully heads and major stream confluences (c.f. Mennis and Fountain, 2001). All modern maps were georeferenced using their gridline intersections. Maintaining minimal distortion of the available maps required as wide a point coverage as possible, and some maps were at insufficient resolution or with insufficient geospatial reference to be georeferenced with any confidence so they were discarded from the quantitative analysis of this study; although they were still useful for gaining contextural knowledge of the glacier changes.

Horizontal errors in georeferencing analogue maps by different authors and aerial photographs are discussed at length by Hall et al. (2003) and Knoll et al., (2009) for example and in **Table 2**, but nonetheless remained extremely difficult to quantify absolutely in this study due to the restriction in georeferencing point pair synthesis, which is a product of both limited identifiable features from topographic maps and disagreements between maps as a function of varying resolution (**see Table 2**). We only used thirteen datasets (of all those listed in Table 1) to limit this study to using those datasets in which we had confidence in the georeferencing. Images of these georeferenced maps are available in the supplementary information. Overall error in reconstructed glacier length and area is reckoned to be ± 5 %, in elevation ± 10 m and in thickness ± 20 m (**Table 2**).

Reconstruction of glacier character

Ice extent and 3D ice surface elevation

The 1850 (last Holocene maximum; Little Ice Age) and 1890 moraine ridges were digitised from the maps by Slupetzky and Teufl (1991) and together with their associated trimlines are very visible in the ALS-derived 2008 DEM (**Fig. 1**) in aerial photographs and on the ground (**Fig. 2a and 2b**). Glacier outlines and contour lines for 1969 and 1998 were obtained in digital format and are from Lambrecht and Kuhn (2007). All other glacier outlines and ice surface contours were manually digitised on screen from each georeferenced topographic map (**Table 1; Fig. 3**), as by Linsbauer et al., (2012) and Paul (2010), for example. Ice surface contours had to be estimated for the 1850 and 1890 datasets and this was achieved in the following steps: (i) the glacier extent polyline was converted to points and elevation data attributed to it using the intersection of it with the 2008 DEM, (ii) contours (polylines) were drawn manually across the surface of the glacier extent to link points with identical elevation at ~ 20 m intervals, but this was not a straight line and the simplest way to estimate the palaeo-surface geometry was to (iii) match the planform curve of this contour to that of the 1929 dataset, which in the accumulation zone was concave and in the ablation zone was convex. There is unquantifiable error in the published topographic map ice surface elevation contours, and our reconstruction of the 1850 and 1890 ice surfaces is the most simple (and thus defensible) approach (**Table 2**). Nonetheless, this approach to reconstructing ice surfaces from lateral moraine crests has been used elsewhere, by Carrivick et al., (2012), for example. Furthermore, we note that Paul (2010) has shown that an accurate reconstruction of glacier (outline) extent (area) is much more important than that of the surface elevation for determination of former glacier mass balance. Interpolation between the ice surface contour lines using the ArcGIS 'Topo to Raster' (ANUDEM algorithm) tool converted the 20 m ice surface contour lines to a smooth surface. Finally, this reconstructed ice surface was clipped to the glacier outline (**Fig. 3**).

Ice thickness and 3D geometry

The area and length of each reconstructed glacier extent was calculated from the digitised glacier outlines. Ice surface long profiles were generated along the same centreline for each time period to ensure comparability between each ice surface. The same centreline was used when measuring the long profile of the bed surface to calculate ice thickness of each reconstructed glacier. Ice thickness data is crucial to gain a better understanding of how climate perturbs glacier changes, but is unknown for most glaciers.

The Ödenwinkelkees bed elevation was estimated primarily using ground penetrating radar (GPR) data gathered in 1998 by Span et al, (2005) and supplemented by our GPR surveys in 2010 (**Fig. 1**). Our April 2010 GPR surveys (**Fig. 2c**) covered a total of 2050 m horizontally and utilised a PulseEkko Pro system, following the protocol used by Rippin et al., (2011). In brief, GPR data were collected while moving continuously; the sledge was hauled manually (**Fig. 2c**), and at a centre frequency of 50 MHz. To accurately locate the GPR on the glacier, real-time kinematic (RTK) differential global positioning (dGPS) system data were simultaneously collected via a Leica GPS500. The rover unit was mounted on the sledge, whilst the base station was located at the Hinterer Schafbichl [47°08'04.21827" N and 12°37'41.73905" E, 2352 m a.s.l.] (**Fig. 1**) as positioned accurately (+/- 0.01 m) relative to Salzburg using a 8 hour static occupation of that point. A mean 3D positional accuracy of 0.2 m was achieved with the rover in this RTK mode. The GPR data was processed in the software packages ReflexW; filtered to remove low-frequency noise, resampled to a uniform 0.5 m step-size as recommended for 50 MHz antennas, migrated using an FK algorithm, bandpass filtered with a pass 30-80 MHz, topographically corrected using the dGPS data, conversion of two-way travel time to ice thickness assuming a radar wave velocity of 0.168 m ns^{-1} (cf. Rippin et al., 2011).

Abermann et al., (2009) discuss in detail the errors associated with GPR data, and our 2010 data was no different (**Table 2**). Our processed radargrams were tested for internal consistency by checking the surface and bed elevations at crossover points between intersecting GPR lines. Surface elevation crossover errors were very small with a mean error magnitude of just ± 0.2 m. Bed crossover errors had a mean magnitude of ± 6.7 m, but due to there just being 3 cross-over points (**Fig. 1**) we consider (GPR-derived) bed elevation uncertainty to be ± 10 m. Both the 1998 and the 2010 GPR-derived bed elevation were acquired on a series of transects and along a long profile (**Fig. 1**) and in this study that data was digitised; to georeferenced 3D points, and interpolated using the ArcGIS 'Topo to Raster' (ANUDEM algorithm) tool (c.f. Hutchinson, 1989; Abermann et al., 2009; Linsbauer et al, 2012). We acknowledge that given the spacing between the GPR survey transects due to steep ice, crevasses and rock falls, the necessary interpolation leads to an over simplification of reality, which can be crude (Linsbauer et al, 2012). Therefore, the uncertainty in reconstructed ice thickness is ± 20 m; i.e. approaching 50 % in the steep headwall area. An 'ice-free DEM' was created by merging the GPR-derived bed elevation for the present glacier extent with elevation data on the rest of the catchment from the 2008 DEM.

Ice thickness at each historical time interval was estimated by subtracting the ice-free DEM from the respective reconstructed ice surface. Volume of each reconstructed glacier was calculated using the ArcGIS 'cut and fill' tool, and 3D elevation changes between successive reconstructed glaciers were calculated using the ArcGIS 'raster calculator' tool. We had no choice but to assume that the contemporary valley floor as represented in the 2008 ALS data is the elevation as the bed beneath the reconstructed glacier; i.e. that there has been no valley in-fill (sedimentation), because we do not know the thickness of the Eisboden valley floor sediments. Whilst this thickness of sediments is likely to be at least an order of magnitude less than the reconstructed ice thickness, it must be considered that our reconstructed volumes are a minimum estimate. ArcGIS tool 'zonal statistics' produced information on mean, maximum and minimum ice surface elevation and ice thickness at each time interval. Spatially distributed rates of change in ice surface elevation and ice thickness were obtained for each time period by dividing the differenced surface (a DEM of difference: 'DoD') by the number of years within the time period.

To bring this study up to date, a survey of the glacier terminus position and ice surface elevation was conducted in July 2013; the latter comprising a long profile and 6 transects recorded using a Leica GPS500 differential GPS. The dGPS was used in Real Time Kinematic (RTK) mode as described above for the GPR surveys. Interpolation between the dGPS points on glacier terminus, on the long profile and along transects was carried out to produce an ice surface DEM, using the "Topo to raster" tool in ArcGIS. This 2013 survey and hence any subsequent analysis using the 2013 dataset were limited to the lower part of the glacier only due to accessibility.

Reconstruction of glacier behaviour

The utility of generating three-dimensional glacier geometries for successive time periods is that it permits quantitative assessment of the physical response of a glacier to climate change. Several key glaciological parameters that are largely dependent on ice geometry were analysed in this study; namely the conceptual parameter of glacier equilibrium line altitude (ELA), and the physical parameters of basal shear stress and theoretical ice surface velocity. A consideration of whether basal sliding and subsole deformation were likely was also made.

ELA

The ELA provides a good indicator of glacier response to climate change, acting as a climate proxy both for the past and future glacier extents (Bate, 2008). In this study the ELA was reconstructed for the Ödenwinkelkees for each time period using a number of (map based) methods, thereby recognising that each of these methods has strengths and weaknesses (Benn and Lehmkuhl, 2000; Leonard and Fountain, 2003; Porter, 2001). Therefore, an ELA sensitivity analysis was carried out by comparing the resultant ELA values of five of the most frequently used methods; namely the The Areax Altitude Balance Ratio (AABR), Accumulation Area Ratio (AAR), glacier Terminus-to-Headwall Altitude Ratio (THAR), Median elevation (Hmed), and contour inflection (kinematic ELA).

The AABR firstly weights the mass balance according to the distance above or below the ELA of that area. Secondly, the calculation is refined by considering different linear slopes of the mass balance/altitude curve above and below the ELA. Since many glaciers conform roughly to this specification (Osmaston, 2005), the AABR method-derived ELA serves as a useful first approximation for former glaciers, such as that analysed herein, for which there is no a priori knowledge about their mass balance. We used the freely-available AABR spreadsheet of Osmaston (2005) and simply entered in the contour interval, contour values and area of the glacier for each interval. We then applied balance ratios of 0.5 to 4, at 0.5 increments and noted the ELA for each. Osmaston (2005) suggests with his Table 4 to select the ELA with the lowest standard deviation as calculated with different BRs for different glaciers in a region. However, because we are only considering a single glacier we simply calculated the mean.

The AAR method uses the ratio of the accumulation area to the total glacier area (Carrivick and Brewer, 2004; Bate, 2008) to estimate the ELA. AAR values used in the literature are normally within the region of 0.5 to 0.8; representing steady state conditions (Bate, 2008; Carrivick and Brewer, 2004). In this study it was assumed that the accumulation area accounted for 50 % of the total glacier area. The area between 100 m interval contours was calculated automatically and fed into the equation:

$$ELA = \frac{\sum A_i \, h_i}{\sum A_i}$$

where A_i is the area between each 100 m contour interval and h_i is the mid-point altitude for each contour interval. This method includes the error associated with reconstruction of glacier extent and contours (**Table 2**). However, this error is usually quite minor and indeed unavoidable when using historic map data (Bate, 2008).

The THAR method uses a ratio between the maximum and minimum ice altitude to estimate the ELA. This method cannot consider ice surface geometry. Examples throughout the literature of a THAR ratio have consistently used a ratio of 0.35 to 0.4 for valley and cirque glaciers (Carrivick and Brewer, 2004; Bate, 2008). Therefore this study used the equation:

$$ELA = A_t + THAR\,(A_h - A_t)$$

where THAR is a set ratio of 0.4, A_h is maximum altitude (headwall crest altitude) and A_t is the minimum altitude (terminus altitude). Error is introduced when finding maximum and minimum altitudes of a glacier, as these altitudes depend on DEM grid resolution as well as vertical accuracy. In this study, altitudes were automatically identified across the margins of the headwall and along the terminus enabling a mean headwall crest elevation and a mean terminus elevation to be calculated.

The Hmed method simply uses the median altitude between the headwall and the terminus to estimate the ELA. Its advantages are in its simplicity and speed of calculation (Bate, 2008). The literature suggests that the Hmed method tends to overestimate the ELA, and it must be noted that it does not take into account valley topography, which has a strong control over the distribution of glacier area (Nesje, 1992; Porter, 2001; Carrivick and Brewer, 2004). The Hmed method uses the following equation:

$$ELA = (A_h - A_t)/2 + A_t$$

where Ah is maximum altitude (headwall crest altitude) and At is the minimum altitude (terminus altitude).

The contour inflection (kinematic ELA) method determines ELA via analysis of the shape of ice surface contours, which tend to change from concave to convex down the glacier profile and thus in the transitional zone (around the ELA) appear relatively flat (Leonard and Fountain, 2003). The contour inflection (kinematic ELA) method is simple conceptually and indeed is based on glaciological physics principles. However, error can be introduced due to ELA estimations when applying the contour inflection method due to the subjectivity of deciding where the shape of the contour changes (Bate, 2008; Leonard and Fountain, 2003).

Basal shear stress

Glacier ice can flow via internal deformation, basal sliding and subsole deformation (Piotrowski and Tulaczyk, 1999). The relative contribution of these flow types to the overall behaviour of a glacier is determined by the pressure exerted by the ice on the bed, i.e. the basal shear stress. Basal shear stress depends on ice thickness and ice surface slope, the resistance of the glacier boundaries and ice viscosity:

$$\tau_b = \rho \, g \, h \sin \alpha$$

where τ_b is basal shear stress, ρ is ice density (900 kg.m^{-3}), g is gravitational acceleration (9.81 m.s^{-2}), h is mean ice thickness at the glacier's centreline (m), and α is the glacier surface slope angle parallel to the ice flow direction; calculated in this study using the long profile 50 to 100 m either side of the ELA.

Theoretical ice surface velocity

The theoretical surface velocity at the glacier centreline due to ice deformation was calculated by:

$$U_s = \left(\frac{2 \, A \tau_b^n \, h}{n+1} \right) 31536000$$

where A is an ice-softness parameter of 5.3 x 10^{-15}, n is a constant (usually = 3, according to Glen's Flow Law, h is ice thickness (m), and 31,536,000 is the number of seconds per year (Kerschenner et al., 1999; Carr and Coleman, 2007). Theoretical velocity will differ from actual velocity due to variation in glacier depth. Furthermore, this equation assumes that the basal shear stress is driving movement in a direction parallel to the bed only (Benn and Evans, 2010).

Basal sliding and subsole deformation

The likelihood of basal sliding or of subsole deformation can be determined by considering if the reconstructed ice thickness was sufficient to melt the basal ice; i.e. if the bed of the glacier was temperate, at pressure melting point (Golledge, 2012). At the pressure melting point a film of meltwater between the ice and the bed surface reduces ice-bed friction and encourages sliding. Additionally, this pressure and sliding could warm any basal sediments and lead to flow due to deformation. Whether ice thickness at the ELA (h) is thick enough to cause basal sliding (Weertman, 1961) can be calculated by:

$$h \geq \left(K \frac{\Delta T}{Q_g + Q_s} \right) \Big/ 100$$

where K is the thermal conductivity of ice (1.7 x 10^5 cal.cm^{-1}.yr^{-1}.°C^{-1}), ΔT is the temperature difference between the melting point of ice and the mean annual temperature of upper surface of the glacier (°C), Q_g is the geothermal heat flux (47.3 cal.cm^{-2}.yr^{-1}) and Q_s is the heat of sliding friction (assumed to be 0).

Theoretical steady state geometry

An assessment of the degree to which the Ödenwinkelkees was out of equilibrium at each time step was made firstly by comparing the geometrically-derived ice surface with that suggested by a simple one-dimensional (1D) model applied along a centreline profile of the glacier. The model was that presented by Benn and Hulton (2010), which assumes a steady state ‘perfectly plastic’ ice rheology, where irrecoverable strain only occurs when the basal shear stress equals a specified yield stress. We used a valley-wide yield stress of 130 kPa, as suggested by Hoelzle et al., 2007 (from Haeberli and Hoelzle, 1995) as representative of European Alps glaciers. Note that we did not tune the model in any way; e.g. to best fit the height of lateral moraines; we wanted to consider the departure of the geometrically-reconstructed ice surface to that of a theoretical steady state. Secondly, we analysed the volume (V) - area (A) relationship given by $V = c.A^{\gamma}$ where c = 0.027 and γ = 1.458 for steady state conditions (Adhikari and Marshall, 2012).

Results: Changing geometry and behaviour of the Ödenwinkelkees

Glacier terminus retreat, glacier length reductions and glacier ice surface lowering and ice thickness reductions are depicted together in **Figure 5**. The Ödenwinkelkees terminus position has retreated by ~ 1400m during the 163 years of the study (1850 – 2013), and is now just 70 % of its 1850 length (~ 4600 m) (**Fig. 4**). A linear best-fit line represents the change in absolute glacier length with $r^2 = 0.97$. The rate of change in the terminus position reduced from -10 $m.yr^{-1}$ between 1850 and 1890 to -5 $m.yr^{-1}$ between 1890 and 1930, and then increased to – 25 $m.yr^{-1}$ by 1980. Since 1980 the rate of change in the terminus position has decreased from -25 $m.yr^{-1}$ to < -5 $m.yr^{-1}$.

The change in absolute glacier area (**Fig. 4**) has decreased by 40 % between 1850 to 2013 and can be represented with a linear best-fit line with $r^2 = 0.98$. The rate of change of glacier area was ~ -0.01 $km^2.yr^{-1}$ between 1850 to 1970, but between 1970 to 1985 was positive (i.e. enlarging) slightly. Between 1985 and 2013 the rate of change in glacier area has decreased from -0.01 $km^2.yr^{-1}$ to 0 $km^2.yr^{-1}$ (**Fig. 6**).

Mean ice thickness has reduced by 62 m and maximum ice thickness has reduced by 125 m between 1850 and 2008; i.e. by 60 % and 43 % respectively of the reconstructed 1850 Ice thickness. The rate of change in ice thickness between 1850 to 1977 varied between -0.5 m.yr-1 to 0.2 $m.yr^{-1}$; i.e. a transition from losses in ice surface elevation to a period of time of slight gain in ice surface elevation (**Fig. 6**). Ice thickness also increased at ~ 0.5 $m.yr^{-1}$ between 1985 and 1992. There were prominent periods of ice surface lowering between 1977 and 1985 and between 1992 and 1998 at -0.25 $m.yr^{-1}$ and -0.32 $m.yr^{-1}$, respectively (**Fig. 6**).

Glacier volume in 2008 was just 25 % of that reconstructed for the year 1850. The change in absolute glacier volume through time can be represented by a linear best-fit line with $r^2 = 0.94$ (**Fig. 6**). The rate of change of volume has generally decreased from -0.0025 $km^3.yr^{-1}$ between 1850 and 1890 to -0.001 $km^3.yr^{-1}$ between 1998 and 2008. However, there was faster rates of volume change both negative (~ -0.006 $km^3.yr^{-1}$) between 1977 and 1985 and between 1992 and 1998, and positive (~ 0.001 $km^3.yr^{-1}$) between 1988 and 1992 (**Fig. 6**).

There is a large variation between the five different methods of calculating ELA, with an average standard deviation of 63 m (**Fig. 6**). The AABR, AAR and Hmed methods consistently overestimated the mean, whilst the THAR and the contour inflection methods were consistently underestimating the mean. The glacier ELA, as reconstructed using the mean of the five different map-based or geometric methods, increased by 90 m between 1850 and 2008, and can be represented by a linear best-fit line with $r^2 = 0.94$ (**Fig. 6**). The rate of change of the ELA progressively increased between 1850 and 1950 (**Fig. 6**). However, the times around 1985 and 1998 saw a drop in ELA elevation (**Fig. 6**). The application of Weertman's (1961) equation suggests that ice thickness at the ELA has been consistently thick enough to cause basal sliding, i.e. the glacier has had a temperate bed at all times since 1850 to the present day.

Basal shear stress, as calculated using the mean ice thickness on the centre line and the ice surface gradient at this point, generally decreased from 260 kPa to 200 kPa between 1850 and 1992, as the glacier thinned (**Fig. 6**). However, in 1998 the mean basal shear stress had dropped to 192 kPa and in 2008 to just 112 kPa, due to a reduction in ice surface gradient. This rapid reduction in mean bed shear stress since the 1990s is interesting because Dreidger and Kennard (1986) noted that glaciers in the Cascades apparently reached a thickness sufficient to obtain a critical shear stress of 1 bar; i.e. 100 kPA. The rate of change of basal shear stress was very slight between 1850 and 1977 but since 1988 has progressively and rapidly decreased (**Fig. 6**).

Glacier velocity, as reconstructed using the mean ice thickness on the centre line and the ice surface gradient at this point, and hence the mean basal shear stress, has decreased from 230 $m.yr^{-1}$ to 10 $m.yr^{-1}$ between 1850 and 2008 (**Fig. 6**). However, two distinct periods of reductions in ice velocity are apparent; one between 1850 to 1932 and the other from 1969 to 2013 (**Fig. 6**).

The spatial pattern of ice elevation changes between successive reconstructions is represented in **Figure 7**. The time periods 1950 to 1890, 1890 to 1929, 1929 to 1931, 1931 to 1969, 1969 to 1977 and 1998 to 2008 all exhibit some increase in ice surface elevation at higher altitudes, and some ice surface lowering at lower altitudes (**Fig. 7**). Thereafter, the pattern of surface elevation changes becomes more complex. The maps for the time periods 1977 to 1985 and 1992 to 1998 evidence surface lowering across virtually the entire glacier (**Fig. 7**). The maps for the time periods 1985 to 1988 and 1988 to 1992 suggest ice surface elevation increases in the easterly sections of the glacier and in the lower and central parts of the glacier, respectively, but ice surface lowering at the higher altitude parts of the glacier (**Fig. 7**). Between 2008 and 2013 the lower part of the glacier lost elevation. Asymmetric surface elevation changes become clearer in more recent time periods; especially 1969 to 1977 and 1985 to 1988, where there is an east-west contrast in ice surface elevation changes (**Fig. 7**).

Comparison of the geometric profiles to that predicted by a theoretical steady state model is only depicted in **Figure 5** for selected time periods for clarity. A discrepancy between the geometric profiles and the model-predicted profiles suggests that the Ödenwinkelkees has not been in a steady state condition at any time. **Figure 5** arguably suggests a progression towards more equilibrium conditions with time towards the present day; one can compare the geometric-model difference for 2008 and 1977 difference, which are both slight, with the geometric-model difference of earlier time periods. The most obvious geometric-model differences were in the glacier snout region, where the model consistently suggests higher surface elevations than measured. The geometric and model surface elevation profiles tend to converge towards the mid-reaches of the glacier. However, there was poor agreement between the geometric and model centreline ice surface profiles in the upper steep head region of the glacier, but Benn and Hulton (2010) acknowledge that the model does not do well in steep headwall areas. Changing glacier 3D geometry is summarised in a series of screenshots and a long profile graph in the supplementary information.

Analysis of the volume-area relationship for the Ödenwinkelkees at each time step (**Fig. 8**) shows that the value of the exponent γ increased up until the 1970, suggesting a progressively greater disequilibrium, and then a rapid decrease. The value of the exponent γ is now very close to 1.458; that suggested by Adhakari and Marshall (2012) as representing steady state conditions.

Discussion

This study highlights that the choice of geometric dimension(s) of glaciers that are considered has a very big influence on the patterns, amounts and rates determined. Firstly, the decline in glacier length in 2013 was 30 % of the 1850 length, the 2013 surface area was 61 % of the 1850 surface area, the maximum and mean thickness in 2008 was 50 % and 30 %, respectively of those thicknesses in 1850, and the 2008 volume was only 24 % of the 1850 volume (**Fig. 6a**). Secondly, we found (small) increases in glacier volume that were not reflected in area or length changes, which is an attribute of glacier adjustments to climate recognised by Adhikari and Marshall (2012) with their volume-area analyses. Thirdly, whilst the 1D, 2D and 3D measurements of absolute length, area and volume, respectively, all have linear relationships with time (**Fig. 6a**), the effect of increasing complexity is best manifest in the equation of these lines being used to predict when the glacier will cease to exist; the length regression equation predicts no ice in the year 2354, that of area predicts no ice in the year 2228 and that of volume predicts no ice in the year 2040. We therefore support the widespread view that glacier length studies will only give an indication of glacier responses to climate, whereas studies that include thickness and volume calculations provide much more detailed and accurate knowledge of glacier dynamics processes (Bishop et al, 2004; Dyurgerov and Meier, 2005; Kääb et al, 2002; Zemp, 2006).

Evidence of small re-advances in the 1890s, 1920s and early 1980s were expected with consideration of the work by Knoll et al (2009). However, increases in ice mass were only suggested by the data in this study for area and volume in 1985 (**Fig. 6**) and also for thickness in the mid 1970s and mid 1980s. This lack of detection of an advance between 1850 to 1929 is due to the interval of measurements, which is not fine enough to capture evidence of re-advance (Hoelzle et al, 2007). The positive ice thickness change in the mid-1970s is unexpected but because: (i) the magnitude of the change is not outside of our level of uncertainty, and (ii) this time period is outside of the studies of glacier inventories and thus of most other alpine glacier geometry change studies.

Previous regional studies by Abermann (2011), Kääb et al. (2002), Bishop et al. (2004) and Zemp (2006), for example have shown accelerating loss of ice volume; 48 % between 1850 and 1975 (0.5 % a^{-1}), 25 % between 1975 and 2000 (1 % a^{-1}), and 10 to 15 % between 2000 and 2005 (3 % a^{-1}). The results of this study are in broad agreement for the same time periods; 48 % between 1850 and 1975 (0.5 % a^{-1}), 45 % between 1975 and 2000 (1.8 % a^{-1}) and 10 % between 2000 and 2005 (2 % a^{-1}) but this study has many more intervals that can be considered. With some caution because of our 20 m ice thickness uncertainty, the overall volume changes calculated in this study for the period 1975 and 2000 are notable for being twice the volume loss of ice reported in regionally-averaged studies.

The surface lowering calculated by this study for the Ödenwinkelkees (0.4 $m.a^{-1}$) is less than half that reported to be the average across Austria, which was 1 $m.a^{-1}$ (Fisher, 2009; Lambrecht and Kuhn, 2007). This discrepancy is not likely to be due to the glacier size, because Abermann et al.'s (2009) study of 81 different glaciers within the Otztal Alps in Austria, suggested that glaciers of a similar size to the Ödenwinkelkees (1 to 5 km^2) had an average thickness of approximately 60 m (from the periods 1969 to 2007) and this fits well with the thickness seen at the Ödenwinkelkees; where mean ice thickness reduced from 80 m to 42 m over the same time period. Therefore the unexpectedly slow surface lowering is probably due to more localised factors, namely supraglacial debris cover (**Fig. 2**) and topography. The datasets used in this study to not include information of the spatial coverage of supraglacial debris. However, qualitatively we consider that supraglacial debris cover extent has increased as a proportion of the glacier surface between 1920 and the present day, as based on our examination of historical postcards and paintings and photographs within the Rudolfshuette.

Topographic effects are evidenced by asymmetric (east-west) thinning rates as depicted in **Fig. 2**. Topography at the Ödenwinkelkees probably determines enhanced mass input from lee side wind-redistributed snow deposition and from avalanching, and reduced mass loss due to solar shading (**Fig. 2**). These topographic factors could have become more important as an increasing proportion of the glacier has close proximity to steep surrounding headwalls (**Fig. 2**). These topographic factors could also explain why the ELA has risen by 67 m (**Fig. 6**) rather than by 150 m as would be expected by a 1 ^{o}C air temperature rise that has been observed across the Austrian Alps (e.g. Paul et al., 2007b), notwithstanding the fact that an ELA value is theoretical and assumes equilibrium; a condition which is clearly not met. Indeed recognition of a departure by the Ödenwinkelkees from modelled equilibrium conditions (**Fig. 5**) is unsurprising, except for the LIA time period perhaps. However, it must be remembered that in the absence of mapped outlines or surface contours for 1850 and 1890, glacier surface profiles at those time periods were reconstructed using the position and elevation of lateral moraines. Thus the difference between the 1850 and 1890 geometric and modelled centreline surface profiles could provoke suggestions on the timing of lateral moraine emplacement; at maximum ice extent or during ice thinning. **Figure 5** also evidences spatiotemporally complex surface elevation (and hence mass) evolution. For example, ablation area gains in 1969 to 1977 and between 1988 and 1992 are unrelated to regional climate (warming) and thus could be indicative of a kinematic wave of mass moving down glacier; mass that was derived from precipitation, wind re-distributed snow or avalanching, for example.

Predictions of future relationships between length, area and volume with time, which historically have been linear (**Fig. 6**) could be complicated by two factors. Firstly, the estimated glacier bed surface indicates a number of undulations along its long profile, and these are very likely real because their inflexions line up with pronounced topographic ridges on either side of the glacier. These overdeepenings are likely sites for accumulation of meltwater, which with diminishing overburden pressure as the ice thins will promote ice melt and ice mass loss via calving and enhanced glacier terminus retreat (Benn et al., 2007; Kääb and Haeberli, 2001; Linsbauer et al, 2012; Carrivick and Tweed, 2013).

Secondly, the present ELA at 2560 m a.s.l. coincides with the elevation of a very steep bed surface and very thin ice; indeed there is bedrock protruding through the glacier ice surface at this point (**Fig. 2b**); in other words the accumulation area is becoming physically separated or decoupled from the ablation area. If this decoupling proceeds, then ice flow may reduce dramatically or indeed stop altogether, leading to glacier stagnation in the lower part (Small, 1995; Linsbauer et al, 2012). The lower part of the glacier would continue to melt, albeit slowly due to the supraglacial debris cover. The Ödenwinkelkees ice thickness trends support this stagnation theory where continued reductions in thickness and surface slope have led to basal stresses being reduced, most rapidly since the 1990s and towards the critical bed shear stress limit of 1 bar (100 kPa) suggested by Dreidger and Kennard (1986). Between 1998 and 2008, basal shear stress dropped by 43 % and the associated maximum velocity (at the ELA) by 82 % from 386 m a^{-1} to 68 m a^{-1} (c.f. Vacco et al, 2010).

Thirdly, the linear relationships between length, area and volume with time, could be disrupted by the glacier apparently approaching a steady state condition, as suggested by the V-A analysis (**Fig. 8**) and the progressive decline in the difference between a 1D steady state model and our measured surface profiles (**Fig. 5**). Overall, the suggestion of increasing rate of change in mountain glacier ice mass loss with time by Haeberli et al (2003), Zemp et al (2007) and Abermann (2011) is far less pronounced in the results of this study. In fact this study demonstrates decreased rates of change in glacier geometry; length, area and volume, and glaciological parameters; bed shear stress and surface velocity, since 1980s to the present.

Conclusions and indications of future glacier response

This study emphasises spatial heterogeneity and complexity in glacier responses to climate. It contributes to bridging a gap between glacier inventories that have a high spatial but comparably low temporal resolution, and length and mass balance records that have low spatial but high temporal resolution. It quantifies absolute amounts, rates and trends of changes in the 3D geometry of the Ödenwinkelkees glacier, central Austria, over thirteen time periods spanning from 1850 to 2013.

The Ödenwinkelkees terminus position retreated by ~ 1400 m since the Little Ice Age; the last Holocene maximum, and in 2013 was just 70 % of its former length. Glacier area decreased by 40 % between 1850 and 2013 and glacier volume in 2008 was just 25 % of that reconstructed for the year 1850. Regression equations of glacier length with time, area with time and volume with time have r^2 values of > 0.94, permitting prediction of no ice in the year 2354 if using length, no ice in the year 2228 if using area, and no ice in the year 2040 if using volume.

Volume-area scaling analyses and discrepancy between our reconstructed geometric profiles and those suggested by a 'perfect-plasticity' model, both suggest that the Ödenwinkelkees has not been in a steady state condition at any time in the last 160 years, but there has arguably been evolution of the Ödenwinkelkees towards 'equilibrium conditions' with time towards the present day. As another line of enquiry to see if the glacier was 'stabilising', we had wished to compare the 'climatic' ELA; that suggested by lapse rates, with the 'geometric' ELA; that suggested by measured glacier hypsometry, but the regional and local weather data that we were able to obtain proved unsuitable in temporal coverage. Further investigation of this idea of 'glacier stabilisation' could be made with hypsographic modelling (c.f. Paul et al., 2007c).

We found a tendency for greater asymmetry in rates of change of reconstructed surface elevation with more recent time periods, and we infer that this sheds light on processes that may become more important for the state of the Ödenwinkelkees and other small mountain glaciers in the future; namely mass inputs from avalanching and wind-blown snow, for example, and topographic shading. We suggest that future geometric changes of the Ödenwinkelkees and other small mountain glaciers will be controlled by ice thickness thinning and ice surface slope reductions both together acting to reduce ice velocity. Together with developing supraglacial debris cover and separation of the accumulation area from the ablation area due to thinning over a bedrock ridge, conditions close to stagnation will ensue. With the above conditions small mountain glaciers could become less sensitive to climate and have longer response times.

There is no reason to suppose that the Ödenwinkelkees is unrepresentative of very many small mountain glaciers and the approach used in this study is directly applicable to other mountain glaciers, so the findings of this study should stimulate similar work at other sites and comparisons. The information on volume changes is effectively net balance (Paterson, 1994, p. 29), and so provides a good opportunity to quantify and compare glacier changes locally, regionally (c.f. Bauder et al., 2007) and even globally. Where there is confidence in 3D reconstructions, future studies could calculate ice-flux through selected cross-sections and balance gradients along the glacier tongue (Kerschner et al., 1999). Carr and Coleman (2007) more than adequately demonstrate the potential of gaining glaciologically meaningful information from geometric reconstructions. Where future studies encounter unknown bed topography, which is very common, it should be possible to use the trends within thickness and area parameters as methods of modelling ice thickness (Linsbauer et al., 2012; Raper and Braithwaite, 2009); the Ödenwinkelkees data of this study supports these models, for example, mean ice thickness accounts for 30 % of maximum ice thickness (with an error of 0.04 %) and there is a strong correlation between mean ice thickness increase with area, with a log scale trend (Linsbauer et al., 2012). Future work should quantify the effect of bed overdeepenings on ice dynamics, of supraglacial debris cover on the surface energy balance, and of local topographically-driven factors; namely wind-redistributed snow deposition, avalanching and solar shading, on mass balance.

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References

Abermann, J., Lambrecht, A., Fischer, A., and Kuhn, M., 2009. Quantifying changes and trends in glacier area and volume in the Austrian Ötztal Alps (1969–1997–2006). The Cryosphere Discussions, 3, 415-441.

Adhikari, S., and Marshall, S. J., 2012. Glacier volume-area relation for high-order mechanics and transient glacier states. Geophysical Research Letters, 39, L16505.

Austrian Alpine Club, 1969. Map of the Grossglockner region, 1:25,000. Austrian Alpine Club.

Austrian Alpine Club, 1985. Map of the Grossglockner region, 1:25,000. Austrian Alpine Club.

Bahr, D. B., Meier, M. F., and Peckham, S. D., 1997. The physical basis of glacier volume-area scaling. Journal of Geophysical Research: Solid Earth, 102, 20355-20362.

Bahr, D. B., and Radić, V., 2012. Significant contribution to total mass from very small glaciers. The Cryosphere, 6, 763-770.

Bate, S., 2008. A Reconstruction of Equilibrium Line Altitudes of the Little Ice Age Glaciers in Linnédalen, Western Spitsbergen, Svalbard. MSc, University Centre in Svalbard

Bauder, A., Funk, M., and Huss, M., 2007. Ice-volume changes of selected glaciers in the Swiss Alps since the end of the 19th century. Annals of Glaciology, 46, 145-149.

Benn, D. I., and Hulton, N. R., 2010. An Excel spreadsheet program for reconstructing the surface profile of former mountain glaciers and ice caps. Computers & Geosciences, 36, 605-610.

Benn, D.I., and Lehmkuhl. F., 2000. Mass balance and equilibrium-line altitudes of glaciers in high mountain environments. Quaternary International, 65-66, 15-29.

Binder, D., Brückl, E., Roch, K. H., Behm, M., Schöner, W., and Hynek, B., 2009. Determination of total ice volume and ice-thickness distribution of two glaciers in the Hohe Tauern region, Eastern Alps, from GPR data. Annals of Glaciology, 50, 71-79.

Bishop, M.P, Haeberli, W., Kääb, A., Paul, F., Hall, D. K., Kargel, J. S., Molnia, B. F., Trabant, D. C., Wessels, R., Olsenholler, J. A., Shroder, J. F., Barry, R. G., Raup, B. H., Bush, A. B. G., Copland, L., Dwyer J. L. and Fountain, A. G., 2004. Global land ice measurements from space (GLIMS): Remote sensing and GIS investigations of the Earth's cryosphere. Geocarto International. 19, 57-84.

Carr, S., and Coleman, C., 2007. An improved technique for the reconstruction of former glacier mass-balance and dynamics. Geomorphology, 92, 76-90.

Carrivick, J. L. and Brewer, T. R., 2004. Improving local estimations and regional trends of glacier equilibrium line altitudes. Geografiska Annaler, 86A, 67-79

Carrivick, J. L., and Rushmer, E. L., 2009. Inter- and Intra-catchment variability in proglacial geomorphology: An example from Franz Josef Glacier and Fox Glacier, South Westland, New Zealand. Arctic, Antarctic and Alpine Research, 41, 18-36.

Carrivick, J. L., and Chase, S. E., 2011. Spatial and temporal variability in the net mass balance of glaciers in the Southern Alps, New Zealand. New Zealand Journal of Geography and Geophysics, 54, 415-429.

Carrivick, J.L., Tweed, F.S., 2013. Proglacial lakes: character, behaviour and geological importance. Quaternary Science Reviews, 78, 34–52.

Carrivick, J. L., Davies, B. J., Glasser, N. F., Nyvlt, D., and Hambrey, M. J., 2012. Late-Holocene changes in character and behaviour of land-terminating glaciers on James Ross Island, Antarctica. Journal of Glaciology, 58, 1476-1490.

Carrivick, J. L., Geilhausen, M., Warburton, J., Dickson, N. E., Carver, S.J., Evans, A. E., Brown, L. E., 2013. Contemporary geomorphological activity throughout the proglacial area of an alpine catchment. Geomorphology, 188, 83-95

Chinn, T. J. H., 1995. Glacier fluctuations in the Southern Alps of New Zealand determined from snowline elevations. Arctic and Alpine Research, 27, 187-198.

Cornelius and Clar, 1929. Geologische karte des Grosglockergebietes. Aufgenommen im Auftrage des Deutschen u. Österreischen Alpenvereins.

Driedger, C. L., and Kennard, P.M., 1986. Glacier volume estimation on Cascade volcanoes: an analysis and comparison with other methods. Annals of Glaciology, 8, 59-64.

Dyurgerov M. B., and Meier, M. F., 2005. Glaciers and the changing Earth system: a 2004 snapshot (Occasional Paper 58, Institute of Arctic and Alpine Research, Univ. of Colorado, Boulder, Colorado).

Farinotti, D., Huss, M., Bauder, A., Funk, M., and Truffer, M., 2009. A method to estimate the ice volume and ice-thickness distribution of alpine glaciers. Journal of Glaciology, 55, 422-430.

Federal Office of Metrology and Surveying, 1931. Grosglockner, 1:50000, Vienna: Federal Office of Metrology and Surveying.

Gardent, M., Rabatel, A., Dedieu, J. P., and Deline, P., 2014. Multitemporal glacier inventory of the French Alps from the late 1960s to the late 2000s. Global and Planetary Change. 120, 24–37.

Glasser, N. F., Harrison, S., Jansson, K. N., Anderson, K. and Cowley, A. 2011. Global sea-level contribution from the Patagonian Icefields since the Little Ice Age maximum. Nature Geoscience, 4(5), 303-307.

Golledge, N. A., Mackintosh, A. N., Anderson, B. M., Buckley, K. M., Doughty, A. M., Barrell, D. J. A., Denton, G. H., Vandergoes, M. J., Andersen, B. G. and Schaefer, J. M., 2012. Last Glacial Maximum climate in New Zealand inferred from a modelled Southern Alps icefield. Quaternary Science Reviews, 46, 30-45.

Gross, G., 1987: Der Flächenverlust der Gletscher in Österreich 1850–1920–1969. Zeitschrift für Gletscherkunde und Glazialgeologie, 23, 131-141.

Haeberli, W., and Hölzle, M., 1995. Application of inventory data for estimating characteristics of and regional climate-change effects on mountain glaciers: a pilot study with the European Alps. Annals of Glaciology, 21, 206-212.

Haeberli, W., Hoelzle, M., Paul, F., and Zemp, M., 2007. Integrated monitoring of mountain glaciers as key indicators of global climate change: the European Alps. Annals of Glaciology, 46, 150-160.

Hall, D. K., Bayr, K. J., Schöner, W., Bindschadler, R. A., & Chien, J. Y. 2003. Consideration of the errors inherent in mapping historical glacier positions in Austria from the ground and space (1893–2001). Remote Sensing of Environment, 86(4), 566-577.

Harrison,W. D., Cox, L. H., Hock, R., March, R. S., and Petit, E. C., 2009. Implications for the dynamic health of a glacier from comparison of conventional and reference-surface balances, Annals of Glaciology, 50, 25–30.

Harrison, S., Rowan, A.V., Glasser, N.F., Knight, J., Plummer, M.A. and Mills, S., 2014. Little Ice Age glaciers in Britain: glacier-climate modeling in the Cairngorms. The Holocene, 24(2), 135-140.

Hoelzle, M., Haeberli, W., Dischl, M., and Peschke, W., 2003. Secular glacier mass balances derived from cumulative glacier length changes. Global and Planetary Change, 36, 295-306.

Hoelzle, M., Chinn, T., Stumm, D., Paul, F., Zemp, M., and Haeberli, W., 2007. The application of glacier inventory data for estimating past climate change effects on mountain glaciers: A comparison between the European Alps and the Southern Alps of New Zealand. Global and Planetary Change, 56, 69-82.

Hormes, A., Beer, J., and Schlüchter, C., 2006. A geochronological approach to understanding the role of solar activity on Holocene glacier length variability in the Swiss Alps. Geografiska Annaler A, 88, 281-294.

Huss, M., 2012. Extrapolating glacier mass balance to the mountain range scale: the European Alps 1900–2100. The Cryosphere Discussions, 6, 1117-1156.

Humlum, O., 2002. Modelling late 20th-Century precipitation in Nordenskiöld Land, Svalbard, by geomorphic means. Norsk Geografisk Tidsskrift 56, 96-103.

Hutchinson. M. F., 1989. A new procedure for gridding elevation and stream line data with automatic removal of spurious pits. Journal of Hydrology, 106, 211-232.

Kääb, A., Paul, F., Maisch, M., Hoelzle, M., and Haeberli, W., 2002. The new remote-sensing-derived Swiss glacier inventory: II. First results. Annals of Glaciology, 34, 362-366.

Kerschner, H, S.Ochs and C. Schlüchter. 1999. Paleoclimatic interpretation of the early late-glacial glacier in the Gschnitz valley, Central Alps, Austria. Annals of Glaciology, 28, 135-140

Klok, E. J., and Oerlemans, J., 2004. Climate reconstructions derived from global glacier length records. Arctic, Antarctic, and Alpine Research, 36, 575-583.

Knoll, C., Kerschner, H. Heller, A. and Rastner, P., 2009. A GIS-based Reconstruction of Little Ice Age Glacier Maximum Extents for South Tyrol, Italy. Transactions in GIS. 13, 449-463.

Kuhn, M., Lambrecht, A., Abermann, J., Patzelt G., and Groß. G., 2009. Die österreichischen Gletscher 1998 und 1969, Flächen- und Volumenänderungen. Wien: Verlag der österreichischen Akademie der Wissenschaften.

Lambrecht, A., and Kuhn, M., 2007. Glacier changes in the Austrian Alps during the last three decades, derived from the new Austrian glacier inventory. Annals of Glaciology, 46, 177-184.

Leonard, K.C., and Fountain, A.G., 2003. Map-based methods for estimating glacier equilibrium-line altitudes. Journal of Glaciology, 49, 329-336.

Li, H., Ng, F., Li, Z., Qin, D., and Cheng, G., 2012. An extended "perfect-plasticity" method for estimating ice thickness along the flow line of mountain glaciers. Journal of Geophysical Research: Earth Surface, 117(F1).

Linsbauer, A., Paul, F., and Haeberli, W., 2012. Modeling glacier thickness distribution and bed topography over entire mountain ranges with GlabTop: Application of a fast and robust approach. Journal of Geophysical Research, 117(F3).

Mennis, J. L., and Fountain, A. G., 2001. A spatio-temporal GIS database for monitoring alpine glacier change. Photogrammetric engineering and remote sensing, 67, 967-974.

Nesje, A., 1992. Topographical Effects on the Equilibrium-Line Altitude on Glaciers. GeoJournal, 27, 383-391.

Oerlemans, J., 2005. Extracting a climate signal from 169 glacier records. Science, 308, 675-677.

Osmaston, H., 2005. Estimates of glacier equilibrium line altitudes by the Area× Altitude, the Area× Altitude Balance Ratio and the Area× Altitude Balance Index methods and their validation. Quaternary International, 138, 22-31.

Patzelt, G., 1980. The Austrian glacier inventory: status and first results. IAHS Publication (Riederalp Workshop 1978 – World Glacier Inventory), 126, 181-183.

Paul, F., 2002. Changes in glacier area in Tyrol, Austria, between 1969 and 1992 derived from Landsat 5 Thematic Mapper and Austrian Glacier Inventory data. International Journal of Remote Sensing, 23, 787-799.

Paul, F., 2010. The influence of changes in glacier extent and surface elevation on modeled mass balance. The Cryosphere Discussions, 4, 737-766.

Paul, F., and Haeberli, W., 2008. Spatial variability of glacier elevation changes in the Swiss Alps obtained from two digital elevation models. Geophysical Research Letters, 35, L21502.

Paul, F., Kääb, A., Haeberli, W., 2007a. Recent glacier changes in the Alps observed by satellite: Consequences for future monitoring strategies. Global and Planetary Change, 56, 11-122.

Paul, F., Maisch, M., Rothenbühler, C., Hoelzle M., and Haeberli, W., 2007b. Calculation and visualisation of future glacier extent in the Swiss Alps by means of hypsographic modelling. Global and Planetary Change 55, 343-357.

Paul, F., Maisch, M., Rothenbühler, C., Hoelzle, M., & Haeberli, W. (2007c). Calculation and visualisation of future glacier extent in the Swiss Alps by means of hypsographic modelling. Global and Planetary Change, 55(4), 343-357.

Piotrowski, J., and Tulaczyk. S., 1999. Subglacial conditions under the last ice sheet in northwest Germany: ice-bed separation and enhanced basal sliding? Quaternary Science Reviews, 18, 737-751.

Porter, S. C., 2001. Snowline Depression in the Tropics During the Last Glaciation'. Quaternary Science Reviews, 20, 1067-1091.

Raper, S and Braithwaite, R., 2009. Glacier volume response time and its links to climate and topography based on a conceptual model of glacier hypsometry. The Cryosphere, 3, 183-194.

Rippin, D. M., Carrivick, J. L., and Williams, C., 2011. Evidence towards a thermal lag in the response of small Arctic glaciers to climate change. Journal of Glaciology, 57, 895-903.

Slupetzky, H., and Teufl, J., 1991. Ödenwinkelkees Gletschervorfeld 1:5000. Institut für kartographie und reproduktionstechnik (IKR). Universität Salzburg und Oesterreichischer Alpenverein.

Steiner, L., 2011. Reconstruction of Glacier States from Geo-Referenced, Historical Postcards. Masters thesis. Eidgenossische Technische Hochschule Zurich. 54 pages plus appendices.

Stocker-Waldhuber, M., Wiesenegger, H., Abermann, J., Hynek, B., and Fischer, A.; 2012. A new glacier inventory of the province of Salzburg, Austria 2007/2009. Zeitschrift für Gletscherkunde und Glazialgeologie, 43, 121-128.

Vacco, D.A. R. Alley, D. Pollard and D. Reusch. 2010. Numerical modeling of valley glacier stagnation as a paleoclimatic indicator Quaternary Research, 73, 403–409

Villa, F., De Amicis, M., and Maggi, V., 2007. GIS analysis of Rutor Glacier (Aosta Valley, Italy) volume and terminus variations. Geografia fisica e dinamica quaternaria, 30, 87-95.

Vincent, C., Kappenberger, G., Valla, F., Bauder, A., Funk, M., and Le Meur, E., 2004. Ice ablation as evidence of climate change in the Alps over the 20th century. Journal of Geophysical Research: Atmospheres, 109(D10).

Weertman, J., 1961. Stability of Ice-Age Ice Sheets. Journal of Geophysical Research, 66, 3783-3792.

ZAMG, 2014. Historical Instrumental Climatological Surface Time Series Of The Greater Alpine Region. http://www.zamg.ac.at/histalp/ last visited 28/11/14.

Zemp M., 2006. Glaciers and climate change. Spatiotemporal analysis of glacier fluctuations in the European Alps after 1850. Zürich: Universität Zürich,

Zemp, M., Paul, F., Hoelzle, M., and Haeberli, W., 2008. Glacier Fluctuations in the European Alps, 1850-2000. Darkening Peaks: Glacier Retreat, Science, and Society, 152 pp.

Zumbühl, H. J., Steiner, D., and Nussbaumer, S. U., 2008. 19th century glacier representations and fluctuations in the central and western European Alps: An interdisciplinary approach. Global and Planetary Change, 60, 42-57.

Decadal-scale changes of the Ödenwinkelkees, central Austria, suggest increasing control of topography and evolution towards steady state

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Data type	Date of glacier extent	Source
Digitised outline of 1850 moraines. Thought to be the maximum extent during the Little Ice Age.	1850	Primary dataset, field geomorphological survey and digitised from catchment DEM from ALS data Carrivick et al. (2013). Moraine date from historical observations compiled by Slupetzky and Teufl (1991)
1:250,000 map of the Hohe Tauern Range	1880 - 88	Karte der Ost-Tiroler-Alpen, Tauern & Dolomiten (Karte der Ostalpen, Blatt V), Maßstab 1:250.000. Frankfurt, Ludwig Ravenstein Verlag (1889)
Digitised outline of 1890 extent from moraines.	1890	Primary dataset, field geomorphological survey and digitised from the catchment DEM (Carrivick et al, 2013). Moraine date from historical observations compiled by Slupetzky and Teufl (1991)
Austrian topographic map 1:25,000	1929 - 32	Federal Office of Metrology and Surveying (1931?)
Geological map of the Grossglockner region 1:25,000	1929(?)	Cornelius and Clar (1929)
Austrian Alpine Club map 1:25,000	1969	Austrian Alpine Club (1969)
Ice surface DEM from the Austrian Glacier Inventory.	1969	Gross (1987) and Patzelt (1980)
Austrian topographic map 1:25,000	1977	Unknown glacier extent date, therefore, the date was approximated using its position relative to other dated extents and supported by work by Kuhn et al (2009)
Orthophoto of the Ödenwinkelkees.	1979	Federal Office of Metrology and Surveying (1979)
Geological map of Austria 1:50,000 scale	1985	Höck and Pestal (1994)
Austrian Alpine Club map 1:25,000	1985 (B)	Austrian Alpine Club (1985)
Topographic map 1:50,000 scale	1988	Official Austrian topographic map (Federal Office of Metrology and Surveying, N.D). Glacier extent date unknown, therefore approximated using its position relative to other dated extents and supported by work by Kuhn et al. (2009) and Slupetzky and Teufl (1991)
Austrian topographic map 1:50,000 scale	1992	Federal Office of Metrology and Surveying (1992)
Ice surface DEM from the Austrian Glacier Inventory.	1998	Lambrecht and Kuhn (2007)
Catchment DEM	2008	Carrivick et al. (2013)
Ground Penetrating Radar data (ice thickness).	1998 2010	Span et al. (2005) Unpublished surveys made by Jonathan Carrivick, Christopher Williams and Daniel Carrivick in April 2010 and processed by David Rippin

Table 1 Datasets used in this study to reconstruct Ödenwinkelkees geometry and discrete time increments.

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Source of error	**Error acknowledgement and management**
Original map production	Error exists within the original mapping of ~ 5% error; we have calculated difference between three 1985 datasets. This error is deemed acceptable for the purpose of this study. We note that this method is used throughout the literature (due to its simplicity) and effectiveness in capturing ice extent, thereby permitting further analysis (e.g. Mennis and Fountain, 2001; Kääb et al, 2002; Hoelzle et al, 2007; Lambrecht and Kuhn, 2007; Paul and Andreassen, 2009).
Original map scale	Unquantifiable uncertainty due to data used to construct the map and also the map scales (1:25,000 and 1:50,000), which introduce error through cartographic generalisation.
Interpreting 1850 and 1890 extents	The interpretation of the lateral moraines is in places obvious, and in others less so due to some degradation since deposition. Reconstruction of the accumulation area outline is much more difficult and largely subjective by examining any supposed trimlines.
Estimating 1850 and 1890 ice surface contours	There is no available evidence of the shape and position of the contour lines for the interpreted 1850 and 1890 extents. The elevation of the outline was taken from the catchment DEM. It was clear from more recent datasets that the ice surface contours were concave in the accumulation zone and convex in the ablation zone (c.f. Leonard and Fountain, 2003). Therefore, the general shape of the 1929 contours were used to estimate the ice surfaces but they were ‘lifted’ and ‘extended across-valley’ to intersect with the 1850 and 1890 outlines.
Positional error	Georeferencing positional errors were minimised by positioning all layers relative to the 2008 LiDAR data, i.e. to the best positioned layer multiple control points (c.f. Mennis and Fountain, 2001).
Glacier extent digitisation	The digitisation processes inherently causes cartographic generalisation, the level of detail captured depends on the person digitising and the amount of time available, as it is a very labour intensive process. The highest possible level of detail was aimed for, trying to take into account every small “finger”. However, these fingers varied on the original maps, sometimes classed as part of the glacier, other times classed as disconnected. This could be reflective of the actual state of the glacier during certain times or it could be down to subjectivity of the map creator as to their classification (c.f. Mennis and Fountain, 2001).
Contour map digitisation	This incorporates errors from the original historical map and from the digitisation process. Leonard and Fountain (2003) suggest that contour error is typically half the contour interval, which in this case was 10 m. Where possible DEMs from remotely sensed data were used as the ice surface (for 1969, 1998 and 2008), which reduced this error.
Surface DEM creation	Interpolation between the 20 m interval contour lines varies depending on the method used (e.g. inverse distance weighting, spline, kriging etc.) but generally does not negatively affect the results, and is a good method of estimating previous ice surfaces (Mennis and Fountain, 2001).
ELA calculations	There has been wide discussion on the errors of the methods of ELA calculations (e.g. Benn and Lehmkuhl, 2000; Leonard and Fountain, 2003; Porter, 2001). Therefore, to avoid bias due to choosing one method, four methods were applied in this study (all appropriate to mountain glaciers) and the mean value was used for further calculations.

Table 2 Sources of error within the glacial reconstruction methodology and their management. These errors are to be expected when working with such old datasets (in addition to the modern datasets).

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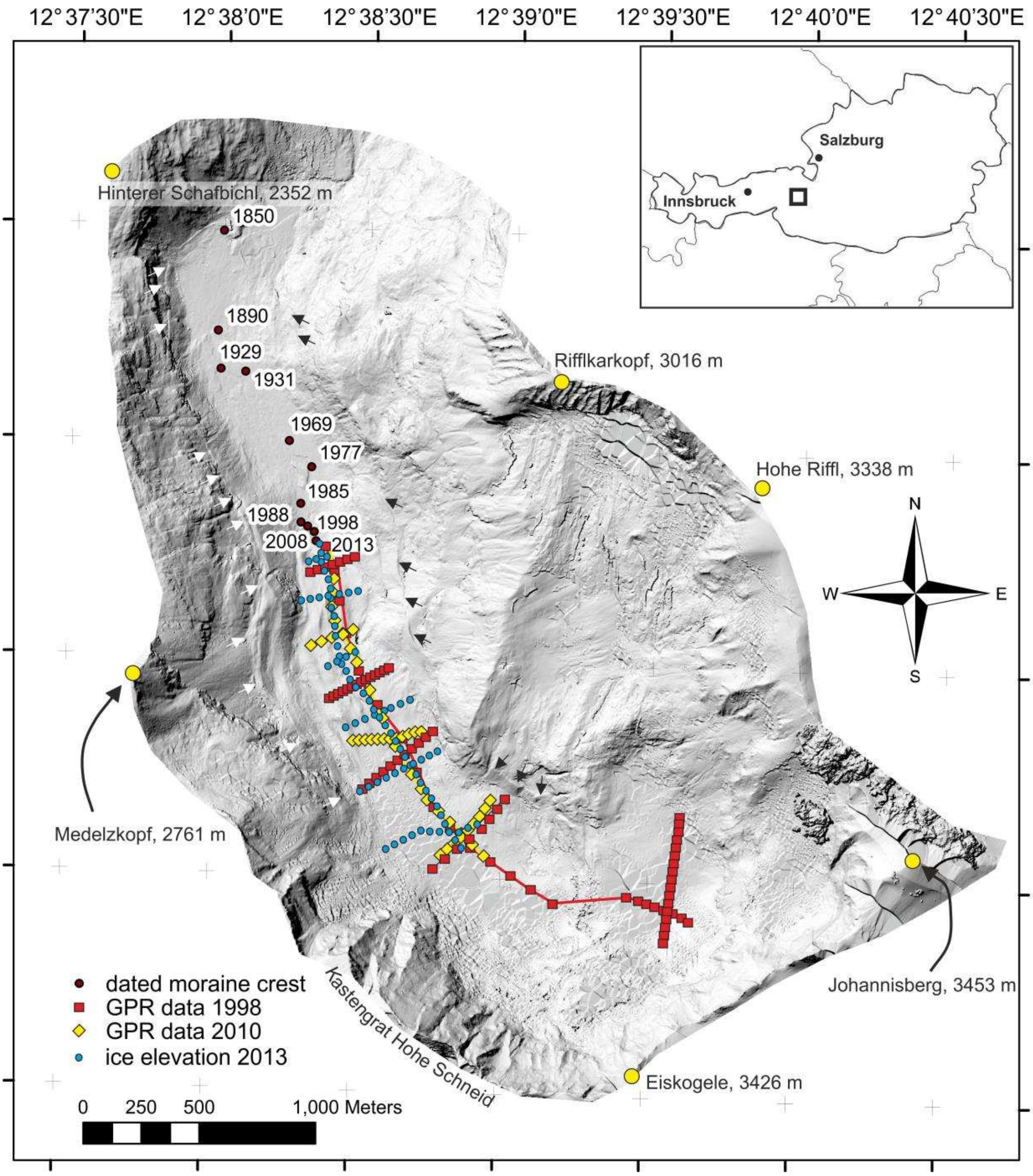

Figure 1 Study location and detail of the Ödenwinkelkees and its catchment, central Austria. The hillshaded elevation model (main image) is derived from airborne LiDAR (ALS) data gridded at 1 m resolution. The terminal moraines of the 1850, 1890 and 1929 extents as dated by Slupetzky and Teufl (1991) are visible in this terrain model, as are associated lateral moraines and trimlines, the upper limit of which is indicated by the white and black arrows on the west and east hillsides, respectively.

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Figure 2 Field photographs annotated with 1850 and 1890 glacier extents; white dashed lines in part a, and separation of accumulation area via ice thinning over bedrock steps; black arrows in parts a and b. Note east-west asymmetry in surface of glacier; perpendicular to valley axis, in b. Part c depicts April 2010 GPR data collection using 50 Hz antennae mounted on a manhauled sledge.

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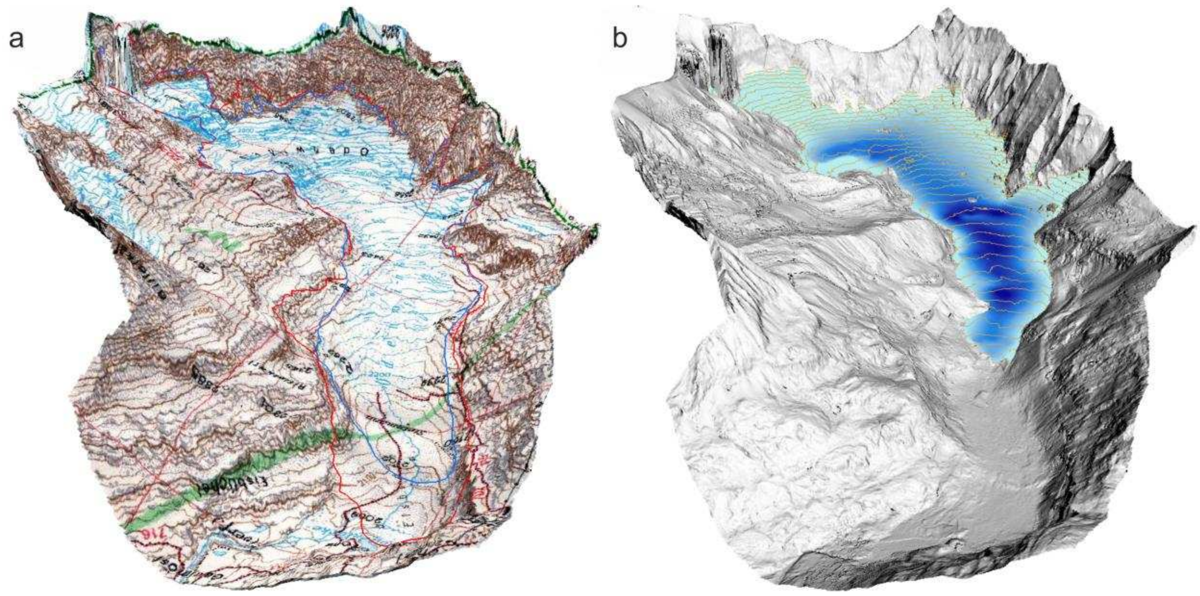

Figure 3 Indication of some of the data sets used in this study; digitisation of glacier outlines from historical maps (a) and from geomorphological evidence (b), coupled with ice thickness (part b; blue shades) derived from ground penetrating radar surveys.

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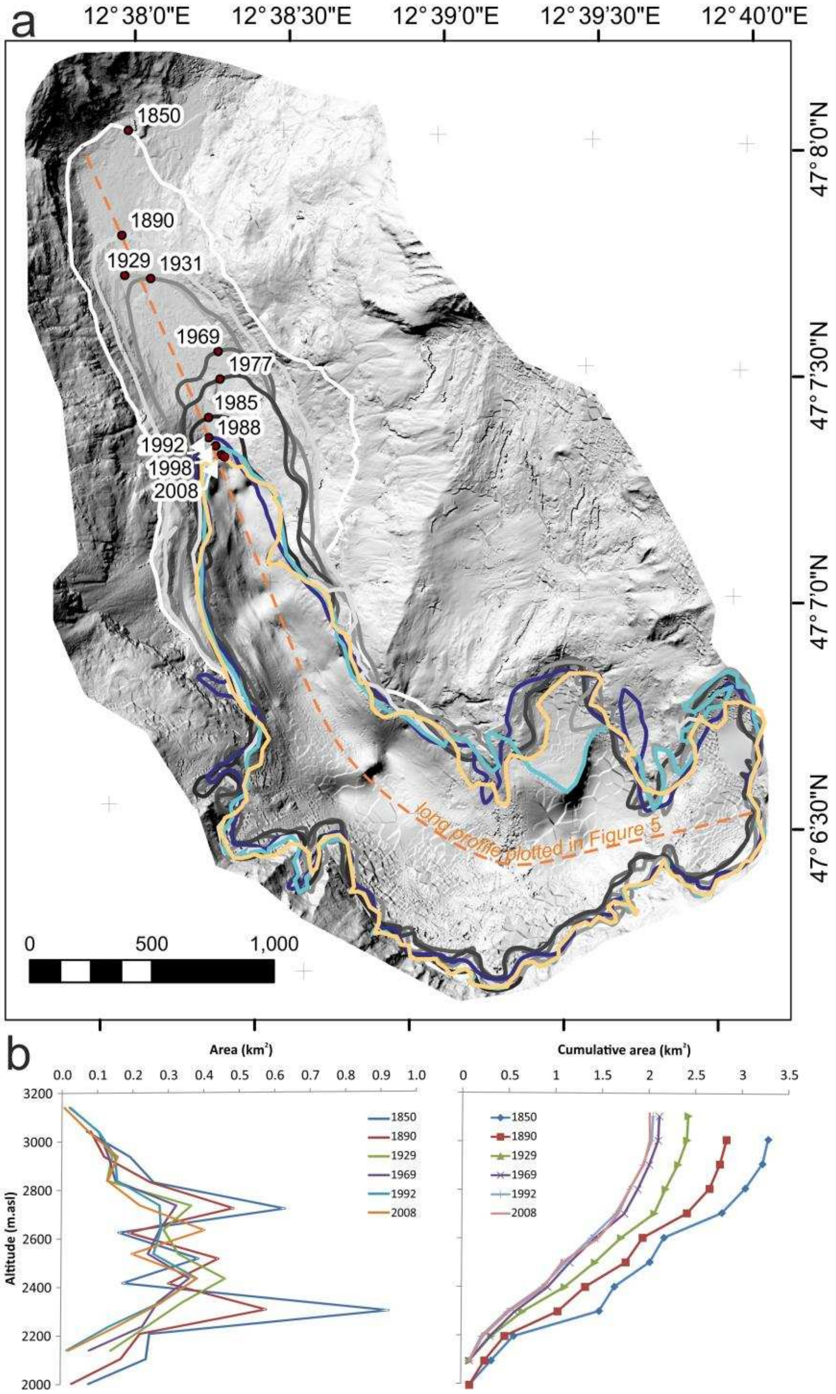

Figure 4 Reconstructed glacier outlines for 1850 and 1890, and mapped glacier outlines for 1929 to 2008 (a). The hillshaded DEM in part A includes the estimated glacier bed, i.e. with the contemporary ice thickness removed. The centre line as mapped in part a is that used to define the long profiles in Figure 5. Reconstructed glacier hypsometry is depicted in part b.

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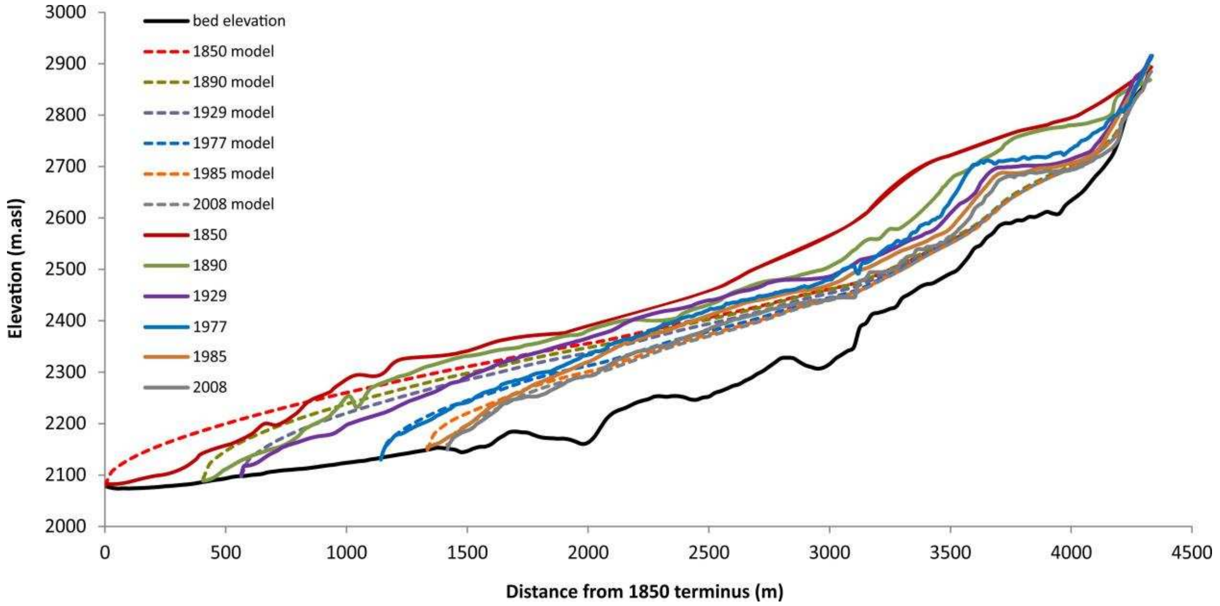

Figure 5 Reconstructed ice surface elevations along a centre-line long profile (solid lines) compared with those modelled (dashed lines) using a steady-state ‘perfect-plasticity’ 1D model.

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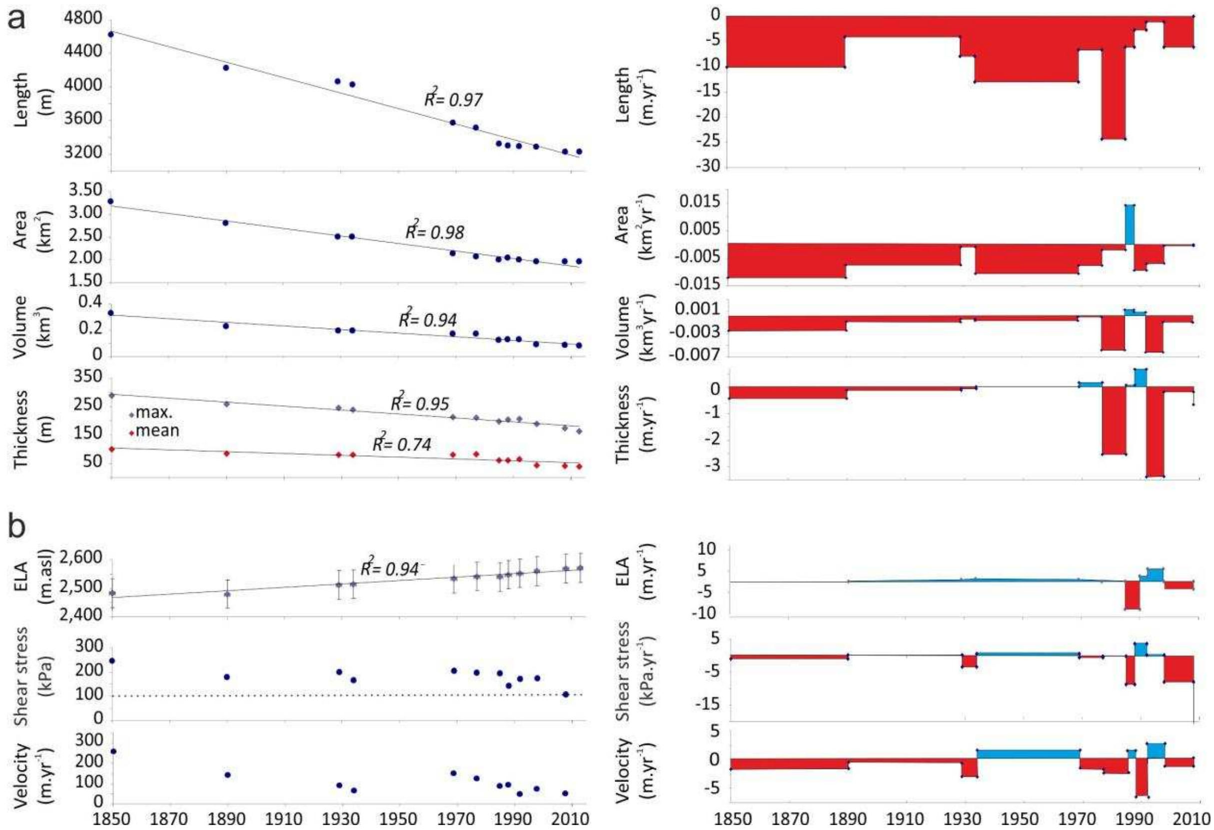

Figure 6 Reconstructed absolute measures of glacier geometry; length, area, volume and thickness (a), and glaciological properties; ELA, shear stress, velocity, both with corresponding rates of change (b). The dashed line in the absolute bed shear stress panel is the critical bed shear stress of 1 bar (100 kPA) suggested by Dreidger and Kennard (1986).

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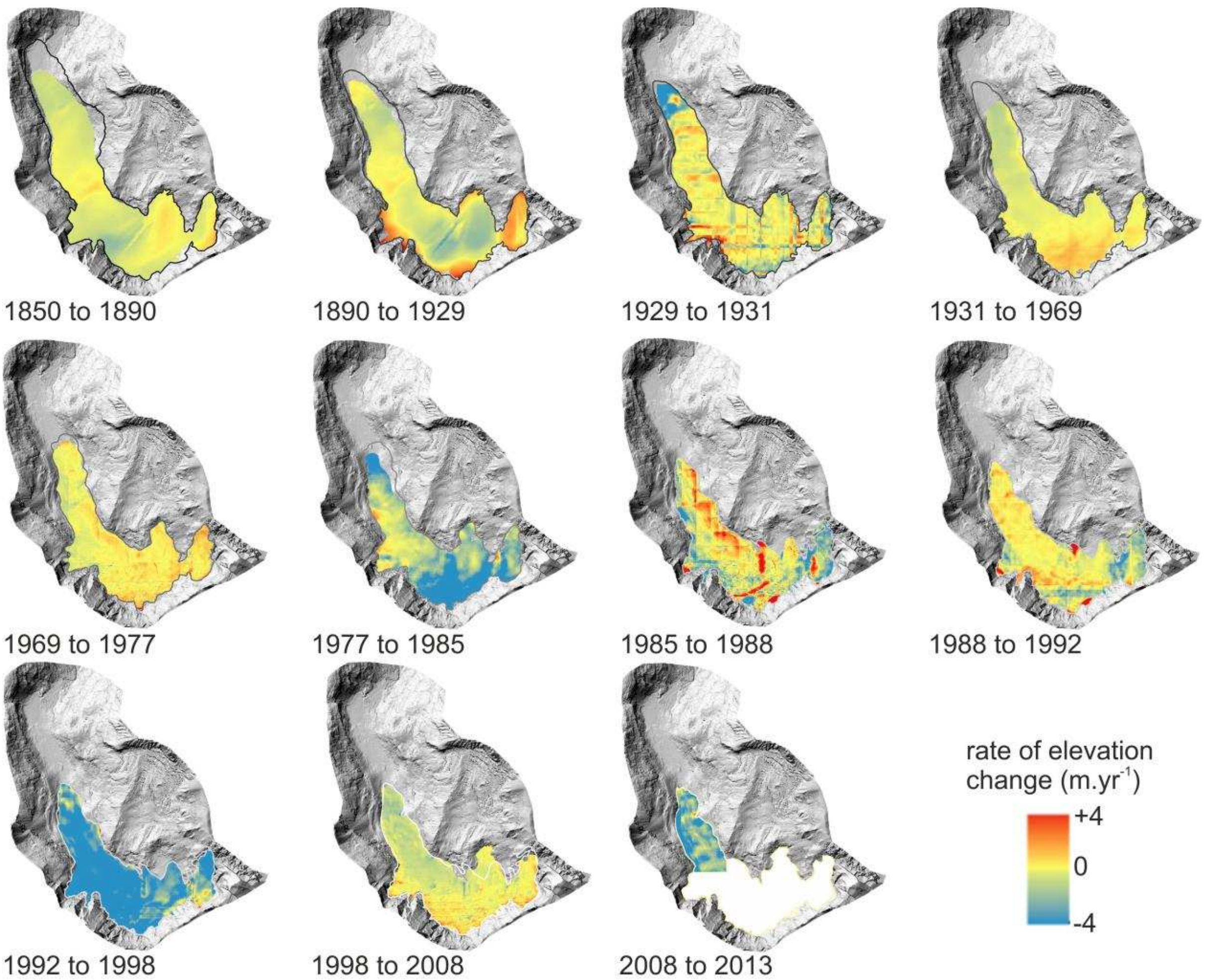

Figure 7 Spatiotemporal variability in the rate of elevation change of the Ödenwinkelkees.

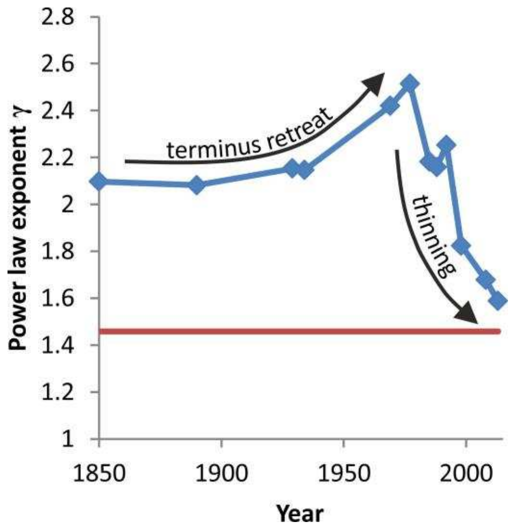

Figure 8 Suggestion of evolution of Ödenwinkelkees glacier geometry towards equilibrium as represented by Adhikari and Marshall's (2012) volume-area power law exponent γ. The red line with value 1.458 is that determined for mountain glaciers in steady state by Adhikari and Marshall (2012). Up until 1970 the dominant geometric change at Ödenwinkelkees was terminus retreat, whilst post-1970 geometric change was dominantly thinning.