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Post-LIA glacier changes along a latitudinal transect in the Central Italian Alps

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Abstract

The variability of glacier response to atmospheric temperature rise in different topo-climatic settings is still matter of debate. To address this question in the Central Italian Alps we compile a post-LIA (Little Ice Age) multitemporal glacier inventory (1860-1954-1990-2003-2007) along a latitudinal transect that originates north of the continental divide in the Livigno mountains, and extends south through the Disgrazia and Orobie ranges, encompassing continental-to-maritime climatic settings. In these sub-regions we examine area change of 111 glaciers. Overall, total glacierized area has declined from 34.1 to 10.1 km^2, with a substantial increase in the number of small glaciers due to fragmentation. Average annual decrease (AAD) in glacier area has risen of about an order of magnitude from 1860–1990 (Livigno: 0.45; Orobie: 0.42; and Disgrazia: 0.39 $\% a^{-1}$) to 1990–2007 (Livigno: 3.08; Orobie: 2.44; and Disgrazia: 2.27 $\% a^{-1}$). This ranking changes when considering glaciers $< 0.5 km^2$ only (i.e., we remove the confounding caused by large glaciers in Disgrazia), so that post-1990 AAD follows the latitudinal gradient and Orobie glaciers stand out (Livigno: 4.07; Disgrazia: 3.57; and Orobie: 2.47 $\% a^{-1}$). More recent (2007–2013) field-based mass balances in three selected small glaciers confirm post-1990 trends showing consistent highest retreat in continental Livigno and minimal area loss in maritime Orobie, with Disgrazia displaying a transitional behaviour. We argue that the recent resilience of glaciers in Orobie is a consequence of their decoupling from synoptic atmospheric temperature trends. A decoupling that arises from the combination of local topographic configuration (i.e., deep, north-facing cirques) and high winter precipitation, which ensures high snow-avalanche supply, as well as high summer shading and sheltering. Our hypothesis is further supported by the lack of correlations between glacier change and glacier attributes in Orobie, as well by the higher variability in ELA_0 positioning, post-LIA glacier change, and inter-annual mass balances, as we move southward along the transect.

1 Introduction

Mountain glaciers are prominent players in the hydrologic and geomorphic functioning of glacierized drainage basins. They are effective agents of landscape evolution (Montgomery, 2002; Brardinoni and Hassan, 2006) and modulate present hydrologic, sedimentary, and geochemical fluxes along the receiving fluvial systems. In consideration of the current generalized conditions of atmospheric temperature rise, despite the relatively small contribution of most of mid-latitude mountain glaciers to sea-level change (e.g., Zemp, 2006; Radic and Hock, 2011), a quantitative appraisal of their retreat and an improved understanding of the spatial variability in relation to different climatic settings hold critical implications for: (i) water supply to hydropower plants (e.g., Barnett et al., 2005; Schaefli et al., 2007; Huss, 2011), and to agricultural and civil compartments (e.g., Braun et al., 2000; Piao et al., 2010; Huss, 2011; Hagg et al., 2013), (ii) mountain tourism (e.g., Scott et al., 2007; Beniston, 2012); and (iii) the assessment of relevant natural hazards (e.g., Huggel et al., 2004; Frey et al., 2010).

Composite glacier sensitivity to recent and ongoing climate changes has been reported through models based on empirical glacier mass balances from selected case studies (Oerlemans and Fortuin, 1992). Accordingly, low-elevation glaciers under maritime conditions, with high accumulation and mass turnover, would display higher sensitivity to climate fluctuations compared to their counterparts located in drier, continental settings. Similar findings have been reported by Hoelzle et al. (2003), who reconstructed the mass balance of more than fifty glaciers around the world on the basis of front retreat information during the entire 20th century. More recently, results from remotely-sensed multitemporal (2 to 5 decades) glacier inventories conducted across maritime-to-continental climatic transects have proved this question to be still open. For example, while Pan et al. (2012), when comparing six mountain systems in China, ranging from monsoonal-temperate to extreme-continental climatic conditions, could not draw a conclusive picture on glacier response in relation to climate properties, other authors in the Canadian Cordillera have even shown that maritime glaciers in the

Coast Mountains retreat less than continental counterparts in the Rockies (De Beer and Sharp, 2007; Bolch et al., 2010).

Within a given climatic setting, glacier dynamics are typically size dependent, with large glaciers retreating, on average, at slower pace than smaller ones (e.g., Paul et al., 2004; Bolch et al., 2010; Diolaiuti et al., 2012a; Tennant et al., 2012; Scotti, 2013; Carturan et al., 2013b). The latter, in turn, display high variability of area change, a variability that has been attributed to the local topographic heterogeneity of the hosting landscape (e.g., Kuhn, 1995; Paul et al., 2004; Abermann et al., 2009; DeBeer and Sharp, 2009; Hagg et al., 2012; Tennant et al., 2012; Carturan et al., 2013b). In fact, region-wide inventories have been customarily conducted from Landsat imagery (30 m grid ~ 0.001 km^2) with automated procedures of detection, which, if on one side allow a rapid cover of entire mountain ranges, cannot capture the area variation of very small glaciers (e.g., < 0.01 km^2: Paul et al., 2004, 2011; Carturan et al., 2013b; and < 0.05 km^2: Bolch et al., 2010; Tennant et al., 2012), and most likely are less accurate than high-resolution aerial photographs (e.g., 0.5 m grid). This is a critical shortcoming since small glaciers (e.g., < 0.5 km^2) in the European Alps represent more than 80 % in number and 15 % in area of the whole glacier population (Paul et al., 2011), with much higher percentages in most sub-regions located south of the continental divide (e.g., Scotti, 2013 and this study).

In this physiographic context, there is a general lack of systematic studies tracking the area change of medium-to-small sized mountain glaciers from the Little Ice Age (LIA) to the beginning of the 21st century, a minimal temporal scale for constraining relevant interactions (coupling vs. decoupling) between climate and glacier fluctuations (Zemp et al., 2011). In fact, most of the relevant literature on the Italian Alps is of extremely difficult access (i.e., published in Italian, e.g., Caccianiga et al., 1994; Pelfini et al., 2002; Curtaz et al., 2013; Lucchesi et al., 2013), has examined post-LIA area change for single glaciers (Carturan et al., 2013a, 2013c), or for a limited number of case studies (e.g., seven, Federici and Pappalardo, 2010), or has considered much

TCD
8, 4075–4126, 2014

shorter time intervals (e.g., Maragno et al., 2009; Diolaiuti et al., 2011, 2012a, b; Carturan et al., 2013b).

In order to fill this research gap and improve our understanding of alpine glacier response to climatic forcing in relation to climate spatial heterogeneity, we conduct post-LIA multitemporal, high-resolution, glacier inventories in three sub-regions of the Central Italian Alps. These are home to medium-to-small glaciers, located along an idealized latitudinal transect that encompasses maritime, transitional, and continental glaciers, ranging in size from 0.002 to 2.3 km^2. Along this transect, we aim to: (i) characterize glacier properties, (ii) calculate changes in glacierized area and evaluate acceleration/deceleration trends, (iii) elucidate correlations between area changes and environmental properties including glacier and terrain topographic attributes, and precipitation; and (iv) evaluate the spatial variability of glacier response to climatic forcing.

2 Study area

We focus on the glaciers of the Livigno, Disgrazia and Orobie sub-regions, located along a north-to-south transect within the Central Italian Alps (Fig. 1). The Livigno sub-region sits in the northern side of the Alpine continental divide (Inn-Danube River basin) and reaches 3303 m a.s.l. at Piz Paradisin. The area is dominated by a SW–NE trending valley that is chiefly underlain by orthogneiss and paragneiss of the Austroalpine basement. The Disgrazia sub-region is placed south of the Alpine continental divide and feeds the Masino and Mallero River valleys (Adda-Po River basin). The largest glaciers flow down radially from the higher peak of Monte Disgrazia massif (3678 m a.s.l.) that is built by Malenco Metaophiolites (mainly serpentinites). The Orobie are an E–W trending mountain range representing the southernmost glacierized area within Lombardy. It is located in the Southalpine tectonic domain that consists of metamorphic lithologies (paragneiss, phyllites and micaschists) covered by thick sedimentary deposits (conglomerates, marls and limestones). The highest peak is Pizzo di Coca (3052 m a.s.l.) and only two other summits exceed 3000 m a.s.l.

The climate of the Central Italian Alps above 2000 m a.s.l. is classified as Tundra Climate (ET) according to the Köppen–Geiger scheme (e.g., Peel et al., 2007). In the three selected sub-regions precipitation (rainfall and snowfall) exhibits high spatial variability in terms of total annual values (Fig. 1b) and seasonal distribution (Ceriani and Carelli, 2000). In the northernmost mountain range (Livigno) annual precipitation ranges locally between 790 and 1200 mm with a winter minimum in February and a single summer maximum in August (e.g., Cancano weather station, 1950 m a.s.l.) (Fig. 2a). The opposite extreme can be observed in the southernmost mountain range (Orobie) where two precipitation peaks in June and October (Scais WS, 1500 m a.s.l.) contribute to annual precipitation values ranging between 1620 and 1770 mm (Figs. 1b and 2c). The Disgrazia region is located at an intermediate latitude, exhibits a transitional behavior in terms of total annual values (range 1210–1370 mm), and mimics the Orobie seasonal distribution (Alpe Gera WS, 2125 m a.s.l.) (Figs. 1b and 2b). The foregoing high spatial variability in total annual precipitation is confirmed by field data of glacier winter mass balances (Bonardi et al., 2014). Specifically, the Lupo glacier (Orobie) despite its 500 m lower elevation, shows more than three times (2.9 m.w.e.q.) the accumulation observed at the Campo Nord glacier (0.9 m.w.e.) (Livigno).

Mean Annual Air Temperature (MAAT) is 1.7 °C at Cancano (Livigno), 1.3 °C at Alpe Gera (Disgrazia) and 6.3 °C at Scais (Orobie). December and August are respectively the coldest and hottest months at Cancano and Scais while at Alpe Gera the monthly extremes occur in January and July.

The progressive climatic shift from oceanic (Orobie) to continental (Disgrazia and Livigno) was detected as the main cause of the lower treeline elevation observed in the Orobie range (2260 m a.s.l. for trees ≥ 3 m) compared to the Disgrazia (2420 m) and Livigno (2480 m) areas (Lucini, 2000; Caccianiga et al., 2008).

3 Data collection and methods

In order to constrain the recent trend of glacier retreat, we reconstructed the extent of glacier, glacierets and perennial snow fields (here all termed “glaciers”) starting from the last maximum advance associated with the Little Ice Age (LIA) and proceeding with those from 1954, 1990, 2003 and 2007 (Fig. 3). The detection of the LIA maximum was conducted by integrating: (i) field mapping of moraines and trim-lines, (ii) remotely-based interpretation of aerial photographs and DSM (digital surface models) shaded-relief rasters; and (iii) historical information including maps, paintings, photographs, reports and scientific literature. LIA moraine ridges in the region are usually well preserved but in some glaciers the interpretation is more challenging, therefore in order to quantify the planimetric accuracy of the mapping we assumed a conservative buffer of ±10 m around the digitized glacier boundaries.

The shape and position of LIA moraines in the study areas and surrounding regions resembles that of other regions in the Alps where examples of LIA glacier reconstructions exist (e.g., Gross, 1987; Maisch, 1992; Maisch et al., 2000). Moraine ages have been determined by means of dendrochronology (e.g., Pelfini, 1999), geopedology (e.g., Caccianiga et al., 1994; Trobio glacier in the Orobie), as well as lichenometry (e.g., Orombelli, 1987; Ventina glacier in the Disgrazia) and combination of these methods (e.g., Pelfini et al., 2002; Disgrazia/Sissone glaciers). These studies significantly improved the confidence of our reconstruction and helped setting the generic date of the last LIA maximum glacial advance in the Disgrazia, Livigno, and Orobie sub-regions to 1860 AD (Pelfini and Smiraglia, 1992). This constitutes our benchmark against which we have computed historical area fluctuations.

Glaciers’ limits in 1954 have been stereographically interpreted on paper copies of black and white aerial photographs (nominal scale 1 : 45 000) then manually drawn on digital orthophotos. In this context, a careful visual inspection of available terrestrial oblique pictures was carried out in order to improve mapping consistency and accuracy that was assessed to be ±5 m (e.g. Diolaiuti et al., 2011).

The glacial extent of the third time step (1990) relies on the Lombardy glaciers inventory (Galluccio and Catasta, 1992), a data set based on detailed field surveys conducted between 1988 and 1991. Since most fieldwork was conducted in 1990 we have decided to set this year as reference. To maximize consistency with the original data, the glacier limits, formerly on paper, have been digitized in GIS environment and slightly revised on the basis of terrestrial and aerial oblique photos. The planimetric uncertainty of this inventory (±2 m) is due to the reading error on the map used by the authors (scale 1 : 10 000) (Citterio et al., 2007; Diolaiuti et al., 2011, 2012a).

The most recent inventories of glacial extent have been reconstructed from 2003, 2007 and 2012 digital orthophotos. Despite the existence of a similar 2003 regional inventory (i.e., Diolaiuti et al., 2012a), in order to minimize the degree of subjectivity due to multiple interpreters, we decided to map independently all glaciers on 2003 orthophoto mosaic (0.5 m grid). This mosaic is characterized by minimal snow cover over the glaciers due to the extremely high temperatures recorded throughout that summer (i.e., García-Herrera et al., 2010). The 2007 inventory was compiled via manual delineation of glacier limits on a high-resolution (0.5 m pixel) orthophoto mosaic and a 2 m gridded Digital Surface Model (DSM, 2007). Thanks to the dry and hot accumulation season, snow cover is very limited in the 2007 images too (Scotti et al., 2013). Such conditions improved substantially our ability to identify glacier limits and constituted a hard stress test for the survival of glacierets and perennial snow fields previously detected during field surveys.

Despite the excellent quality of the orthophoto mosaics, in order to minimize problems related to the delimitation of debris-covered glaciers, we conducted complementary GPS field surveys on three sample glaciers that provided critical ground control for data extracted from remotely-based inspection. We consider the planimetric uncertainty of the digitized 2003 and 2007 glacier limits equal to ±1 m, that is the uncertainty associated with the orthophoto mosaic as specified by the manufacturer (e.g., Diolaiuti et al., 2012a).

The uncertainty associated with glacier area was evaluated for each glacier by setting a buffer of ±10 m (LIA), ±5 m (1954), ±2 m (1990) and ± 1 m (2003 and 2007) on the digitized glacier limits. Subsequently, to evaluate the uncertainty of estimated glacier change we used the root of the squared sum of buffer areas along the study time series (e.g., Xu et al., 2013; Tennant and Menounos, 2013).

In order to improve our understanding on the factors controlling the site specific variability of glacier retreat we have collected a number of environmental attributes for the 2007 dataset. These include glacier primary classification, contribution of snow avalanching to accumulation, surface area (A), maximum elevation (E_{max}), terminus elevation (E_{min}), elevation range (ΔE), theoretical Equilibrium Line Altitude (ELA_0), elevation of the ridgecrest upslope of the glacier (E_{rc}), mean slope gradient (S), main aspect (MA), summer clear-sky radiation (CSR) and annual precipitation on the glacier (MAP) (Fig. 1b and Table 1).

The glacier primary classification and the definition of the avalanche contribution to glacier accumulation are crucial to characterize glacier types in the three study areas. The former follows the Illustrated GLIMS Glacier Classification Manual (Rau et al., 2005); the latter, which we define as Avalanche Area Accumulation Basin Ratio (ABR), is the ratio between the area usually occupied by avalanche supply at the end of the accumulation season and the area of the accumulation basin (above the ELA_0). This classification scheme, which is based on decadal field observations, consists of three classes: low ($ABR \leq 0.33$), moderate ($> 0.33\ ABR \leq 0.66$) and high ($ABR > 0.66$). The main topographic attributes (i.e., E_{max}, E_{min}, ELA_0, E_{rc} and S) have been extracted from the 2 m gridded DSM using zonal statistics in ArcGIS v.9.3 (Paul et al., 2009). The terminus (E_{min}) and the maximum glacier elevation (E_{max}) are effective tools to define the lower and upper limit of the glacial domain and their fluctuations are usually related to surface and volume changes. The analysis of the elevation fluctuations was applied to a fixed sample of glaciers through time. This approach minimizes the errors caused by the increase (or decrease) in number of glaciers due to fragmentation (or extinction). The use of the entire dataset from each sequential photoset would have resulted in under or overestimation of the E_{min} and E_{max} change. The maximum difference we have found by comparing the two approaches is 45 % (e.g., underestimation of the E_{max} drop of Livigno glaciers from the LIA to 2007). The elevation range (ΔE) is the arithmetical difference between E_{max} and E_{min} and depends on glacier length and slope gradient (S).

The theoretical equilibrium line altitude (ELA_0), or balance budget ELA (Meier and Post, 1962; Cogley et al., 2011), is a widely used parameter in glacier and paleoclimatic reconstructions (e.g., Miller et al., 1975; Benn and Lehmkuhl, 2000). We delineate ELA_0 (also termed local-topography ELA_0) by considering a 0.67 balance budget Accumulation Area Ratio (AAR_0) (i.e., ratio of the accumulation zone to the area of the glacier with mass balance equal to zero) (Gross et al., 1978). This topography-based parameter, differs from the regional-climatic ELA (i.e., $_{rc}ELA_0$), which relies on synoptic climatic data and on mass balances of a limited number of selected glaciers (e.g., 14 glaciers for the European Alps, and only two belonging to the Italian portion, Zemp et al., 2007). The elevation of the ridgecrest upslope of the glacier (E_{rc}) is computed as the median elevation of the 10 m wide buffer drawn along the ridgecrest feeding the glacier accumulation basin. The elevation difference between the E_{rc} and the ELA_0 is considered to be correlated to both the degree of avalanching contribution to the glacier's mass balance and the shading effect of the rock walls upslope of the glacier. The main aspect of the glacier, divided in 8 classes, was manually defined along the direction of the main flow axis, or for snow fields, the dominant aspect of the mountain slope. The summer clear sky global radiation (June to September) was calculated with ArcGIS Spatial Analyst (Dubayah and Rich, 1995) using a 20 m resampled version of the DSM. This parameter is directly affected by glacier slope aspect and by the shading proprieties of the rock walls surrounding the glacier. Mean annual precipitation for each glacier is derived from a 250 m gridded precipitation map (Fig. 1b) and represents a proxy for snow accumulation on the glacier.

TCD
8, 4075–4126, 2014

4 Results

4.1 Glacier proprieties

In the presentation of the results we provide an overview of the glacier properties, as inventoried in 2007. We proceed from the northernmost Livigno sub-region, home to 16 glaciers (total glacier area = 1.1 km^2 ± 0.02), continue with the Disgrazia sub-region that hosts 37 glaciers (7.3 km^2 ± 0.09), and conclude with the Orobie sub-region in which we identify 44 glaciers (1.8 km^2 ± 0.05). Along this transect, we observe a remarkable increase in mean annual precipitation (MAP) as we move from the interior ranges (Livigno; 790–1200 mm) towards the outer ranges (Orobie; 1620–1770 mm) (Fig. 4). Concurrently, median ELA$_0$ (Fig. 4) and clear-sky radiation mirror the spatial variability of local relief in that they slightly increase from the interior, plateau-like topography of Livigno (2833 m a.s.l.; 176 W m^2), to the Disgrazia Massif (2890 m a.s.l.; 210 W m^2), and drop abruptly in the Orobie Range (2517 m a.s.l.; 145 W m^2). The altitudinal distribution of ELA$_0$ displays an increase in within-regional scatter with increasing MAP (i.e., moving from Livigno down south; Fig. 5a). This variability is imparted by the combination of two spatial patterns in which ELA$_0$ rises progressively: (i) from north- to south-facing glaciers, within the same mountain range (i.e., Disgrazia in Fig. 5b); and (ii) for a given aspect category (e.g., N and NW in Fig. 5b) moving from the peripheral Orobie range inland to the Livigno mountains.

In the Livigno mountains, glacierets and cirque glaciers are dominant typologies, and face mainly northwest to northeast (Fig. 5b and Table 2). Despite the presence of relatively high peaks across the entire sub-region, glaciers today survive almost only in the southernmost portion of the range (with one exception), where incidentally MAP is higher. Glacier size ranges between 0.003 and 0.37 km^2 (Val Nera Ovest glacier). Propensity to avalanche snow/ice supply (ABR) is high (11 cases) to moderate (4 cases), while slope (S) ranges between 19.6° and 33.0° (median 29.2°).

In the Disgrazia sub-region, besides the abundance of permanent snowfields, glacier types comprise in decreasing order of frequency: cirque, niche, and simple/compound basin valley glaciers (Table 2). Glaciers face preferentially northwest and southeast, but thanks to the radial structure of the massif, all aspects are well represented (Fig. 5b). Compared to the other study sub-regions, ice masses are evenly distributed across the N–S transect, they are relatively larger, and range from 0.002 to 2.31 km^2 (Disgrazia glacier). ABR is high, moderate, and low for respectively 24, 10, and 3 glaciers. Median slope is comparatively lower (27.1°), and we observe the largest slope variability (18.1–45.0°).

Glaciers in the Orobie are located exclusively within north-to-northwest facing cirques. They are clustered around a narrow latitudinal range, along the main ridge of the sub-region (Fig. 4), and are particularly small in size, ranging between 0.002 and 0.22 km^2 (Lupo glacier) (Fig. 5b). The peculiar morphometric setting made of high and steep rock walls, located immediately upslope of each glacier, is confirmed by the high elevation difference (259 m) recorded between ELA_0 and mean ridgecrest elevation (E_{rc}). Accordingly, all of Orobie glaciers exhibit a high ABR potential of avalanche snow supply. Slope range is similar to that observed in Disgrazia (18.8–42.2°), while median slope (29.1°) is higher and resembles that of Livigno.

4.2 Area changes

Since the LIA all of the 111 glaciers of the study sub-regions have gone extinct (14) or have experienced a strong net areal reduction (97) for a combined area loss of 24 km^2 (Fig. 6a–c). At the apex of LIA advance, the 15 glaciers of the Livigno cluster used to cover an area of 5.4 km^2 (Fig. 6a and Table 3). By 1954 a total of 21 glaciers (i.e., 3 of the initial 15 had fragmented into smaller ones) occupy 2.5 km^2 (52.6 ± 14.6 %) for an average annual decrease (AAD) of about $0.031 \pm 0.006\ km^2\ a^{-1}$ (Table 3). In the same period, the 27 LIA glaciers of the Disgrazia Mountains increased to 36 (Fig. 6b), but with an overall area loss of 43.6 ± 6.4 % (from 22.0 to 12.4 km^2) and an AAD of about $0.102 \pm 0.015\ km^2\ a^{-1}$ (Table 3). Finally, in the Orobie sub-region by 1954 we record a 52.6 ± 14.6 % of LIA surface reduction (from 6.7 to 3.2 km^2), which corresponds to an AAD of about $0.038 \pm 0.010\,km^2\,a^{-1}$ (Table 3). In this period, the fragmentation of 3 glaciers caused a minor increase in glacier count (from 45 to 49) (Fig. 6c and Table 3).

The 1990 inventory depicts a much slower rate of areal contraction with values small enough to fall within the envelope of uncertainty (Fig. 7). The glacierized area in the Livigno Mountains records the highest relative contraction (i.e., $9.5 \pm 8.3\,\%$) equal to $0.23\,km^2$ (AAD = $0.007 \pm 0.006\,km^2\,a^{-1}$) (Table 3). Glaciers in the Disgrazia lost $3.5 \pm 5.1\,\%$, which corresponds to a net loss of $0.43\,km^2$ (AAD = $0.012 \pm 0.017\,km^2\,a^{-1}$) (Table 3). Similarly, in the Orobie we observe a $3.5 \pm 10.4\,\%$ loss, corresponding to a net loss of $0.11\,km^2$ (AAD = $0.003 \pm 0.009\,km^2\,a^{-1}$).

In the 1990–2003 period, glaciers exhibit consistent fast retreat throughout the three study areas (Fig. 7). In increasing order, Disgrazia glaciers witness a decrease of $3.5\,km^2$ (from 12.0 to $8.4\,km^2$) that corresponds to a $29.5 \pm 2.0\,\%$ reduction (AAD = $0.271 \pm 0.018\,km^2\,a^{-1}$); Orobie exhibit a $1.2\,km^2$ decrease (from 3.1 to $2.0\,km^2$), which amounts to a $35.0 \pm 4.2\,\%$ contraction (AAD = $0.083 \pm 0.010\,km^2\,a^{-1}$); and Livigno glaciers lost $1\,km^2$ (from 2.3 to $1.3\,km^2$), equal to a $42.7 \pm 3.3\,\%$ loss of the 1990 glacierized area (AAD = $0.075 \pm 0.006\,km^2\,a^{-1}$) (Table 3). During the 2003–2007 interval we observe for the first time that glacier area loss increases northward, with Livigno displaying highest retreat ($16.9 \pm 2.5\,\%$, from 1.3 to $1.1\,km^2$) (AAD = $0.063 \pm 0.009\,km^2\,a^{-1}$), followed by Disgrazia ($12.8 \pm 1.6\,\%$, from 8.4 to $7.4\,km^2$) (AAD = $0.309 \pm 0.037\,km^2\,a^{-1}$), and Orobie ($10 \pm 3.6\,\%$, from 2.0 to $1.8\,km^2$) (AAD = $0.057 \pm 0.020\,km^2\,a^{-1}$) (Table 3). Overall, considering the entire study period (1860–2007), glaciers of the Livigno sub-region display the largest retreat recorded amongst the three study areas, losing a total of $4.4 \pm 0.5\,km^2$ ($80.1 \pm 9.8\,\%$ of the initial 1860 extension). Glaciers in the Disgrazia cluster lost a total of $14.6 \pm 1.3\,km^2$, ($66.5 \pm 5.9\,\%$) and in the Orobie range they lost $4.9 \pm 0.9\,km^2$ ($73.2 \pm 13.8\,\%$).

Data stratification into size classes reveals that most of the Disgrazia glaciers at the LIA maximum used to belong to the 0.1-to-0.5 km^2 class and that most of the total glacierized surface in this sub-region fell within the 2-to-5 and 5-to-10 km^2 classes (Table S2). Interestingly, we record a progressive reduction both in area and number of glaciers in all sizes except the ≤ 0.1 km^2 class, which increases in number due to glacier fragmentation from 6 (total area = 0.4 km^2) (LIA) to 28 (total area = 1 km^2) (2003), and then declines slightly to 26 (total area = 0.6 km^2) (2007) due to glacier extinction.

In the Orobie sub-region, after the disaggregation of the Trobio glacier, the largest one (1.1 km^2) at the LIA apex, and the reduction of the Scais glacier (0.6 km^2), only the 2 low-magnitude classes are present. By 1954 we observe a sharp decrease of glacier count and area in the 0.1-to-0.5 km^2, which translates into an increase of smaller glaciers (≤ 0.1 km^2) both in terms of number and area. Area contraction continues across the 1954–2007 period but glacier distribution in the 2 classes remains substantially unchanged.

At the LIA maximum the Livigno Mountains host the Mine glacier, a relatively larger ice body (1.5 km^2). By 1954, its disaggregation had generated 7 distinct glaciers. As a consequence of glacier fragmentation and progressive contraction, similarly to what observed in the Orobie mountains, by 2007 the distribution of glaciers across sizes displays the survival of the 2 smallest classes only. The main difference, in comparison to the Orobie cluster, is the presence of glaciers in the 0.5-to-1 km^2 class up until 1990, and the higher abundance of 0.1-to-0.5 km^2 ice bodies compared to the ≤ 0.1 km^2 category in every time interval.

Examination of AAD across size classes shows that relative change rate in glacier area in the 1860–1954 period has been fairly low (0.46 % a^{-1} in the Disgrazia cluster, 0.56 % a^{-1} in Orobie and 0.57 % a^{-1} in Livigno) and complementary among small- and large-size classes of the study sub-regions (Table 4). Subsequently (1954–1990), the < 0.1 km^2 class displays the lowest reduction (Livigno: 0.02; Disgrazia: 0.16 % a^{-1}), and in the Orobie case even a modest increase (−0.09 % a^{-1}). In Disgrazia and Livigno the largest retreat rates are observed in the intermediate classes (0.5-to-1 km^2 and 0.1-to-0.5 km^2 respectively), whereas larger glaciers exhibit a slight area increase (Disgrazia: −0.22 % a^{-1} for the 2-to-5 km^2; Livigno: −0.04 % a^{-1} for the 0.5-to-1 km^2) (Table 4).

The strong glacier shrinkage recorded in the two more recent periods (1990–2003 and 2003–2007) has affected especially small glaciers (i.e., $< 0.1\,\text{km}^2$ and 0.1-to-$0.5\,\text{km}^2$) and we observe progressively slower retreat rates within the larger size classes (i.e., 0.5-to-1, 1-to-2 and 2-to-$5\,\text{km}^2$) (Table 4). In particular, the 2003–2007 period denotes high retreat variability both across size classes and among the different sub-regions. In Disgrazia small glaciers ($< 0.1\,\text{km}^2$) exhibit the highest retreat rate of the whole study period ($11.11\,\%\,\text{a}^{-1}$), 5 times higher than the 2-to-$5\,\text{km}^2$ class. A similar behavior, even though less pronounced, is observed in Livigno ($8.73\,\%\,\text{a}^{-1}$) for the $< 0.1\,\text{km}^2$ class; by contrast, in Orobie this size class shows much slower retreat ($3.77\,\%\,\text{a}^{-1}$) (Table 4).

4.3 Elevation changes

The area changes detailed above correspond to changes in glacier ice elevation, both in terms of E_{min} and E_{max}. The median E_{min} of the 111 glaciers detected at the LIA maximum lies at 2480 m a.s.l. and rises progressively throughout the 20th century to a maximum of 2628 m in 2007, which translates to an average annual gain of $1.0\,\text{m}\,\text{a}^{-1}$. In the same period, median E_{max} drops from 2893 to 2810 m a.s.l. ($-0.6\,\text{m}\,\text{a}^{-1}$). Data stratification into sub-regional domains reveals a considerable spatial variability in E_{min} and E_{max} fluctuations. Both glacier attributes in the Livigno cluster are characterized by a markedly lower variability compared to the Orobie and Disgrazia (Fig. 8). The 1860–2007 overall rise in E_{min} is lowest in Livigno ($0.7\,\text{m}\,\text{a}^{-1}$), intermediate in Orobie ($1.0\,\text{m}\,\text{a}^{-1}$), and highest ($1.9\,\text{m}\,\text{a}^{-1}$) in the Disgrazia sub-region, where we note a sharp increase between 1860 and 1954 (Fig. 8). Conversely, Disgrazia exhibits the lowest drop in E_{max} ($-0.6\,\text{m}\,\text{a}^{-1}$), followed by Livigno ($-0.7\,\text{m}\,\text{a}^{-1}$), and Orobie ($-1.1\,\text{m}\,\text{a}^{-1}$), with the last characterized by two large drops in 1860–1954 and 1990–2003 (cf. median lines in Fig. 8).

Simultaneous analysis of elevation (E_{min} and E_{max}) and area changes through time is instructive in that it allows inferring qualitatively characteristic trends of volumetric glacier shrinkage (Fig. 9). Up until 1990 we observe a general decline in average annual decrease and a general convergence of the E_{min} and E_{max} trend lines in Livigno and Orobie clusters, while in the Disgrazia both E_{min} and E_{max} rise slightly (Fig. 9). This latter trend suggests that, on average, glacier ice lost at the terminus was nearly completely replaced (i.e., at least in terms of area) by the increase in elevation of the accumulation basin (Fig. 9b). From 1990 we start observing a progressive divergence of the E_{min} and E_{max} trend lines (Fig. 9), an indication of net, generalized, glacier volume loss. While such trend continues to the end of the study period in Livigno and Disgrazia, in the Orobie we observe an opposite behaviour between 2003 and 2007, with E_{min} and E_{max} overlapping around a null elevation change rate (Fig. 9c), an indication of about volumetric stationarity.

4.4 Area change with glacier attributes

Analysis of changes in glacier area within the same sub-region allows to detect, and possibly rank, the main environmental attributes driving glacier retreat. To this purpose, we analyze the mutual correlations among the "1860–2007 area change" in relation to glacier size (GS), main aspect (MA), mean slope gradient (S), minimum elevation (E_{min}), maximum elevation (E_{max}), glacier relative relief (ΔE), mean annual precipitation (MAP), ridgecrest elevation (E_{rc}), and clear-sky radiation (CSR) (Tables S3–S5).

Relative area change (AC %) in Livigno exhibits strong direct correlation with E_{rc} (r = 0.77), E_{max} (r = 0.72) and ΔE (r = 0.65), and moderate correlation with E_{min} (inverse, r = −0.46), former glacier size (GS, r = 0.43), and clear-sky radiation (CSR, r = 0.43) (Table S3). These correlations with relative area change weaken progressively moving south to Disgrazia (i.e., E_{rc} (r = 0.35), E_{max} (r = 0.45), ΔE (r = 0.47), and glacier size (GS, r = 0.42)) (Table S4), and virtually disappear in the Orobie (i.e., E_{rc} (r = −0.03); E_{max} (r = −0.20); ΔE (r = 0.20); and E_{min} (r = −0.40)) (Table S5).

Despite the moderate glacier size-retreat correlations previously identified in the Livigno and Disgrazia sub-regions, representing relative area changes as a function of former glacier size does not aid constraining an empirical envelope of variability (Fig. 10).

Interestingly, in the Orobie glacier size (GS) not only is completely unrelated to retreat, but also shows a weak inverse correlation with E_{min} ($r = -0.32$) (Fig. 10).

In order to gain further insights on the elevation–retreat correlations identified above, we have represented relative area change as a function of E_{rc} (i.e., the elevation of the ridgecrest located upslope of the glacier) (Fig. 11). We hypothesize this variable to be a useful proxy of the local climatic conditions (e.g., snowfall available for subsequent avalanche inputs, shading effect and wind shielding) that characterize a glacier's source basin. Although we reckon that E_{rc} is tightly related to other glacier elevation attributes i.e., E_{max} and ΔE (Tables S3-S5), unlike these, E_{rc} does not change with time, and as such would constitute a more reliable reference across changing climate conditions. In addition, E_{rc} is a more statistically sound attribute, as it is not based on a single datum of elevation (i.e., E_{min} and E_{max}).

The representation presented in Fig. 11 shows that E_{rc} declines progressively along our north-to-south transect. In the Livigno and Disgrazia sub-regions relative area change (AAD) varies inversely with E_{rc}, and this relation is well-constrained for AAD up to 80 %. Beyond this threshold the degree of scatter increases. Stratification of glaciers according to south- and north-facing categories allows constraining two distinct retreat-elevation envelopes, with the former glaciers plotting about 300 m higher. In this context, the northern facing Disgrazia-Sissone and Ventina glaciers display a smaller relative retreat (56 and 45 % respectively), compared to the south facing counterparts of Predarossa (69 %) and Cassandra (83 %) that are similar in size and that flow down from the same summits (Figs. 6b and 11b). Finally, in the Orobie mountains we see that the wide range of retreat rates is completely unrelated to E_{rc} (Fig. 11c) and glacier size (Fig. 6c), suggesting that different mechanisms must control contemporary glacier dynamics in this physiographic setting.

5 Discussion

5.1 Equilibrium line altitude

The equilibrium line of a glacier is a climate-dependent attribute that, when estimated at the regional scale using climatic data and a limited set of glacier mass balances ($_{rc}ELA_0$; e.g., Ohmura et al., 1992; Zemp et al., 2007), can mask the intrinsic spatial heterogeneity modulated by glacier aspect and other local topographic variables (Dahl and Nesje, 1992). Such topographic effects can be evaluated by comparing the local topography ELA_0($_{lt}ELA_0$) (i.e., the ELA_0 considered in this study) with the regional climatic one ($_{rc}ELA_0$) (Dahl and Nesje, 1992; Lie et al., 2003; Zemp et al., 2007). In this respect, the distributed $_{rc}ELA_0$ map of the Central European Alps presented by Anders et al. (2010) (i.e., based on equations by Ohmura et al. 1992 and Zemp et al. 2007) reports values that are about 70, 150, and 400 m higher than the actual topography-based analogues for the Disgrazia, Livigno, and Orobie respectively, suggesting that local topography, on average, has a different weight in each sub-region.

Since the $_{rc}ELA_0$ approach typically tends to respectively underestimate and overestimate southerly and northerly aspects (Zemp et al., 2007), the relatively small "climate-topography" mismatch in the Disgrazia cluster should not surprise, given that in this area glaciers are distributed on all aspect categories (Fig. 5b) and so aspect effects tend to cancel out. Following this logic, from a synoptic climatic standpoint Orobie glaciers should not exist, as the $_{rc}ELA_0$ in this sub-region ($\sim$ 2900 m a.s.l.) plots some 180 m above the median ridgecrest, hence confirming the characteristic topo-climatic adjustment of these glaciers (on average). In this context, the comparison between Orobie and Livigno (both characterized by dominantly north-facing glaciers) is instructive, as it removes any potential confounding associated with slope aspect. In the Orobie, we observe a four-fold increase in ELA_0 variability (> 800 m) compared to Livigno ($\sim$ 300 m) (Fig. 5b), a variability that reinforces prior hints (Sect. 4.4) on the potential decoupling between Orobie glaciers and synoptic climatic conditions, and that we interpret as the effect of local morphometric properties of the hosting cirques and niches.

At these locations, peculiar conditions of snow avalanching, shading and wind accumulation would be able to sustain glaciers but not significant ice flow, as this latter would imply the existence of larger glaciers, characterized by higher elevation ranges (ΔE).

5.2 Area change of small glaciers

Considering the characteristic limited size of our study glaciers, the high sensitivity of small glaciers to climate change (i.e., Haeberli and Beniston, 1998; Paul et al., 2004; Jiskoot and Mueller, 2012; Tennant et al., 2012), and the relatively low elevation of the study terrain (Fig. 4), it is not surprising that, at first glance, post-LIA Annual Average Decrease (AAD) in Livigno (0.55 % a^{-1}), Disgrazia (0.45 % a^{-1}), and Orobie (0.50 % a^{-1}) plot well above the estimated average of 0.33 % a^{-1} for the European Alps (1850–2000, Zemp et al., 2008). However, since this regional estimate relies chiefly on satellite imagery, it is likely to carry high uncertainties on the area change of small glaciers, and therefore a direct comparison with our sub-regional glacier inventories seems inappropriate. Comparisons with other sub-regions within the Alps characterized by larger glacier and higher mountains, and where inventories of comparable temporal and spatial resolution are available, highlight lower retreat rates in: (i) Les Ecrins (AAD = 0.45 % a^{-1}; MAP ~ 1200–1400 mm a^{-1}), the Mont Blanc (AAD = 0.25 % a^{-1}; MAP ~ 1400–2000 mm a^{-1}), and the Vanoise (AAD = 0.20 % a^{-1}; MAP ~ 900–1400 mm a^{-1}) (1820/50–2006/09, Gardet and Deline, 2013), (ii) Val d'Aosta (0.39 % a^{-1}; MAP ~ 800–2000 mm a^{-1}) (1820/50–2005, Curtaz et al., 2012), and (iii) the Swiss Alps (AAD = 0.26 % a^{-1}; MAP ~ 600–2600 mm a^{-1}) (1850–2000, Zemp et al., 2008). Elsewhere, post-LIA retreat rates are higher (0.78 % a^{-1}) in the Spanish Pyrenees (MAP ~ 1600–2000 mm a^{1}) (1894–2001, Gonzales Trueba et al., 2008), about the same (0.50 % a^{-1}) in the Canadian Rocky Mountains (MAP ~ 730–1970 mm a^{-1}) (1919–2006, Tennant et al., 2012), and substantially lower (0.13 % a^{-1}) in the Jotunheimen (Southern Norway, MAP ~ 1300–1650 mm a^{-1}) (1750–2003, Baumann et al., 2009).

TCD
8, 4075–4126, 2014

In order to remove the possible confounding exerted by glacier size and conduct a more appropriate evaluation of glacier area change at the local (i.e., comparison among three subregions) and regional (e.g., against the alpine average) scales, we now consider the two smaller glacier size classes only i.e., < 0.1 and 0.1–0.5 km^2 (DeBeer and Sharp, 2007). This adjustment yields a 1860–2007 AAD that decreases progressively moving southward, from Livigno (0.62 % a^{-1}) to Disgrazia (0.58 % a^{-1}) to Orobie (0.48 % a^{-1}). These retreat rates are similar to: (i) data by Lucchesi et al. (2013), who report an average AAD (1860–2006) of 0.50 % a^{-1} for the Western Italian Alps, starting from LIA glaciers of 0.5 km^2 (average size), a value similar to the combined average size of our study glaciers (i.e., 0.4 km^2); and (ii) the estimated average of the European Alps (1850–2000, 0.51 % a^{-1}) for the same size class (Zemp et al., 2008). It is worth highlighting that his latter figure would have risen significantly if post-2000 data were to be added, given that the 2001–2007 period was characterized by intense glacier retreat (WGMS, 2009).

5.3 Glacier retreat and temporal variability

The examination of 4 different periods (1860–1954–1990–2003–2007) in 3 sub-regions allows us to detect the temporal and spatial variability of glacier change. Glaciers in the study area underwent a low relative retreat in the 1860–1954 period, remained almost stable up until 1990, and then started retreating at progressively faster rates in the 1990–2003 and 2003–2007 intervals (Fig. 7), with greater retreat acceleration of the very small glaciers (≤ 0.1 km^2). In this temporal context, the Orobie sub-region represents the exception, in that the retreat rate across 1990–2003 and 2003–2007 stays constant with, in the latter period, an AAD value for glaciers ≤ 0.1 km^2 that is much lower than in Livigno and Disgrazia sub-regions (Table 4). The gradual increase with time of the spread of the relative change in glacier area (Fig. 7b) is a warning that these results need to be used with caution since the study intervals differ significantly in length. In particular, potential decadal fluctuations in glacier area within the 1860–1954 and 1954–1990 periods would have gone undetected (i.e., the re-advance phase of alpine glaciers in the 1970s and 1980s, Patzelt, 1985; Hoelzle et al., 2003; Citterio et al., 2007).

In order to partly solve this issue and conduct a more sound comparison of our results with other inventories, we consider the AAD values associated with the 1860–1990 and 1990–2007 periods. One of the most striking results is the significant increase in AAD that one observes after 1990. In particular, post-1990 AAD in Livigno, Disgrazia and Orobie is respectively 7.2, 6.6, and 6.1 times faster than before. These values are gradually decreasing along our latitudinal transect, indicating that glaciers in the most continental sub-region (Livigno) not only depict a higher total post-LIA retreat, but also that such retreat has been much faster in recent years compared to more maritime environments (i.e., Orobie mountains). Similar rates (i.e., 7.1) have been reported only in the Spanish Pyrenees between 1894–1991 and 1991–2001 (Gonzales Trueba et al., 2008), whereas in many other alpine regions the acceleration is still detectable but less intense (i.e., 2.2 times in France between LIA and the 70's to 2006–2009 (Gardet and Deline, 2013), and 2.9 times in Swiss Alps between LIA and 1973 to 1999, (Paul et al., 2004).

The previously disclosed differences in glacier retreat pattern along our latitudinal transect are even more apparent when increasing the temporal resolution to an inter-annual basis. To this end, we present unpublished data from multiple GPS field surveys and glaciological mass balance campaigns (2007–2013) on three sample glaciers: Campo Nord (GS = 0.30 km^2; Livigno), Vazzeda (GS = 0.23 km^2; Disgrazia), and Lupo (GS = 0.22 km^2; Orobie) glaciers (Table 5 and Figs. 1b and 3). Mass balances are combined with glacier limits updated to summer 2012 (delineated on a 0.5 m grid orthophoto mosaic; planimetric uncertainty ±1 m) (Table 5 and Figs. 1b, 3, 12 and 13). In particular, the relevant winter and summer specific mass balances, measured across the ELA_0 (Figs. 3, 12 and 13), even though referred to three glaciers only, are useful to infer the mechanisms responsible for the differences in glacier retreat observed along our transect (Table 4 and Fig. 7). Since 2007, Campo Nord glacier depicts an uninterrupted series of negative net balances for a total loss of 12.9 m.w.eq. and an area loss of 0.02 km^2. Lower mass losses are recorded at Vazzeda and Lupo glaciers (6.3 and 5.6 m.w.eq.), with the former losing 0.03 km^2 and the latter showing no significant changes in glacier area (Figs. 12 and 13). Despite the small latitudinal difference from Campo Nord to Lupo glacier (about 40 km), the mass balance turnover increases dramatically along the transect. At Lupo, years with high winter accumulation are able to compensate for more consistent rates summer ablation throughout the 2007–2013 period. This trend suggests a higher sensitivity of Orobie glaciers to winter precipitation, as 2009, 2010, and 2011 were characterized by both above-average winter precipitation and summer temperatures, which resulted in negative mass balances across most of the European Alps (WGMS, 2011, 2013).

5.4 Small, avalanche-dominated glaciers

The tendency of small, avalanche-dominated glaciers to be poorly coupled to synoptic temperature changes has been reported in different studies. Kuhn (1995) discusses a conceptual model to explain the mass balance of "very small" glaciers (i.e., glacier area < 1 ha, or 0.01 km^2), suggesting that snow drifted by wind and accumulated by avalanching activity would be crucial to sustain glaciers below the $_{rc}ELA_0$. Furthermore, he suggests that glaciers in small cirques are partly de-coupled from precipitation as in winters with heavy snow falls once the cirque is completely filled with snow, this surplus would be conveyed below the glacier terminus via avalanching and thus lost to accumulation. More recently, DeBeer and Sharp (2009) have shown that a sample of very small glaciers (< 0.4 km^2) in the Monashee Mountains (British Columbia) displayed no observable change in area during the 1951–2004 period, while the neighboring larger glaciers suffered generalized retreat. Accordingly, these small glaciers after an initial post-LIA retreat are now placed in locations that would favor their survival (i.e., in sheltered sites surrounded by high and steep rock walls). The authors suggest that the enhanced mass inputs at these particular sites can compensate for the decline in winter precipitation observed in the region.

TCD
8, 4075–4126, 2014

Dahl and Nesje (1992), while reconstructing the paleo-ELA of a small glacier in western Norway, attribute the resilience of small avalanche-dominated glaciers to patterns of winter precipitation, as opposed to summer temperature. More recently, Carturan et al. (2013a) provide empirical data supporting this explanation for the Montasio glacier (GS = 0.07 km^2; E_{median} = 1903 m a.s.l.), in the Eastern Italian Alps. Accordingly, during the 2009–2011 period years with heavy winter snow-falls (and related high snow avalanche inputs) would be able to generate a positive mass balance sufficient to compensate one or more subsequent negative years. This interpretation is further supported by the limited post-LIA area loss, which the authors estimate to be about 30 %.

Even though most of the glaciers in our study sub-regions are small and avalanche fed (Table S1), only those of the Orobie cluster appear to be poorly coupled to the contemporary synoptic climatic conditions and deviate from the other two (Fig. 7), hence from the average alpine trend (Zemp et al., 2008). In consideration of the progressively lower decoupling inferred moving northward along the study transect, we hypothesize that snow avalanching activity is efficiently increasing glacier accumulation, hence dampening glacier retreat, only where precipitation is relatively high, as in the Orobie case. In other words, we propose that the dynamics of these glaciers are (snow) *supply-limited*, rather than limited by summer ablation.

Despite the lack of reliable long-term climatic series for each sub-region, the progressive north-to-south decoupling of glacier change is supported by the southward increase in variability of ELA_0 (Fig. 5a), post-LIA glacier change (Fig. 7), and inter-annual mass balances of the monitored sample glaciers (Figs. 12 and 13). Further to this, the below alpine average post-LIA retreat (for the same glacier size) and the lack of relations between glacier change and glacier attributes found in the Orobie sub-region (Fig. 11c and Table S5) are evidences of enhanced glacier-climate decoupling.

It should be highlighted, however, that such decoupling exhibits a high degree of variability, as exemplified by post-LIA area losses of the initial Orobie 45 ice bodies: ranging from as little as 33 % (Aga glacier, comparable to the area shrinkage reported in Montasio), including respectively 6 and 12 glaciers that have recorded an area loss lower than 50 and 60 %, and up to 5 cases that have reached extinction. It follows that generalizations and extrapolations on small, avalanche-fed glaciers to other regions, based on a single glacier mass balance, should be conducted and evaluated with caution. Further work in the Orobie is presently ongoing to investigate causal linkages between climatic forcing, landscape (i.e., hosting cirques and niches) structure, and glacier dynamics to better constrain the environmental conditions and the feedback mechanisms promoting glacier survival in temperate, maritime, mountain settings.

6 Summary and conclusion

With a multitemporal, airphoto-based glacier inventory, combined with inter-annual, field-based mass balances of selected small glaciers we can link glacier and terrain morphometric attributes, climatic characteristics, and glacier response to climatic forcing. In particular, we examine post-LIA glacier area and elevation changes, along a latitudinal transect, and across a 150 year time window. Within a latitudinal distance of less than 60 km we move from small continental-like glaciers surviving between 2800–3200 m a.s.l. with as little precipitation as 790 mm a^{-1} (Livigno sub-region) to maritime ones located between 2100–2500 m a.s.l. with as much as 1770 mm a^{-1} (Orobie sub-region). As one moves southward, this physiographic set up corresponds to: (i) a progressive depression of ELA_0 values with a concurrent increase (doubling) of ELA_0 within-subregional variability; and (ii) a weakening and/or disappearance of correlations between basic altitudinal glacier attributes and 1860–2007 glacier area change.

We further show that post-1990 glacier area change is about an order of magnitude faster than before, and that this trend accelerates even more in Livigno and Disgrazia between 2003–2007, in line with the European Alps trend. By contrast, Orobie glaciers, which have been retreating comparatively less since 1990, are basically stationary in the post-2003 period. This behaviour is further confirmed and extended through 2013 by an overall (2007–2013) equilibrium mass balance at Lupo glacier (Orobie), as op-

TCD
8, 4075–4126, 2014
Post-LIA glacier changes along a latitudinal transect
R. Scotti et al.

posed to persistent net deficits observed in Campo Nord (Livigno) and Vazzeda (Disgrazia) glaciers. This equilibrium is achieved thanks to heavy accumulation seasons that, during the seven years of monitoring, have been able to compensate for consistent summer ablation losses and relevant dry winters. Therefore, we argue that the dynamics of Orobie glaciers are currently supply-limited (i.e., their survival depends on the magnitude-frequency of winter accumulations) rather than controlled by ablation. In other words, we hypothesize that the recent resilience of glaciers in Orobie is a consequence of their decoupling from synoptic atmospheric temperature trends (i.e., rise). A decoupling that originates from local topographic conditions (i.e., deep, north-facing cirques), but most importantly from high winter precipitation, which represents the distinctive attribute of the Orobie cluster. This combination of topo-climatic conditions ensures high snow-avalanche supply, as well as high summer shading and sheltering. In this context, we introduce the parameter E_{rc} (i.e., the elevation of the ridgecrest located upslope of a given study glacier), which, when represented as a function of relative glacier area change, proves to be an efficient proxy for discriminating climatically-coupled from decoupled settings.

The case of the Orobie, in which for the first time we identify a population of maritime, climatically-decoupled small glaciers (i.e., beyond the documentation of a single glacier behaving as an outlier), is in contrast with empirically-based mass balance models and comparative studies according to which low-elevation glaciers under maritime conditions, with high accumulation and mass turnover, would display higher sensitivity to climate fluctuations compared to their counterparts located in drier, continental settings (e.g., Oerlemans and Fortuin, 1992; Hoelzle et al., 2003; Benn and Evans, 2010). Interestingly, since winter precipitation is expected to rise by 15 to 30 % in the future decades across the Central European Alps (e.g., CH2011, 2011; Beniston, 2012), Orobie glaciers may continue to find favourable conditions for surviving much longer than previously thought.

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TCD
8, 4075–4126, 2014

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Table 1. Glacier variables considered.

Glacier variable	String	Unit
Size	GS	km^2
Maximum elevation	E_{max}	m a.s.l.
Minimum elevation	E_{min}	m a.s.l.
Theoretical Equilibrium Line Altitude	ELA_0	m a.s.l.
Ridgecrest elevation	E_{rc}	m a.s.l.
Glacier relative relief	ΔE	m
Mean slope gradient	S	degrees
Main Aspect	MA	na
Clear-Sky Radiation (Jun–Sep)	CSR	$W m^2$
Mean Annual Precipitation	MAP	$mm a^{-1}$
Avalanche Area Accumulation Basin Ratio	ABR	na

Table 2. Glacier characteristics in the study sub-regions as inventoried in 2007.

Classification		Sub-region		
Primary	Secondary	Livigno	Disgrazia	Orobie
Valley Glacier	Simple basin	–	1	–
	Compound basins	–	1	–
Mountain Glacier	Cirque	3	13	24
	Niche	–	2	–
	Compound basins	–	2	–
Glacieret	Cirque	4	4	9
	Niche	1	1	–
Permanent snowfield		8	13	11
Total sample		16	37	44
Area (km^2)		1.1 (±0.02)	7.3 (±0.09)	1.8 (±0.05)

TCD
8, 4075–4126, 2014

R. Scotti et al.

Table 3. Variation of glacier count and glacierized area through time in the study sub-regions.

Sub-region	1860		1954		1990		2003		2007	
	Count	Area (km^2)	Count	Area (km^2)	Count	Area (km^2)	Count	Area (km^2)	Count	Area (km^2)
Livigno	15	5.4 ± 0.53	21	2.5 ± 0.20	22	2.3 ± 0.07	21	1.3 ± 0.03	16	1.1 ± 0.02
Disgrazia	27	22.0 ± 1.28	36	12.4 ± 0.59	38	11.9 ± 0.22	39	8.4 ± 0.10	37	7.3 ± 0.09
Orobie	45	6.7 ± 0.93	49	3.2 ± 0.31	49	3.1 ± 0.12	48	2.0 ± 0.06	44	1.8 ± 0.05

TCD
8, 4075–4126, 2014

Table 4. Relative change rate in glacier area, expressed as average annual decrease (AAD), across glacier size classes.

Size Classes km^2	AAD (% a^{-1}) 1860–1954	1954–1990	1990–2003	2003–2007
		Livigno		
< 0.1	0.63	0.02	5.20	8.73
0.1–0.5	0.68	0.62	2.88	3.17
0.5–1	0.41	−0.04	2.31	–
1.0–2.0	0.60	–	–	–
Total AAD	0.57 ± 0.11	0.26 ± 0.23	3.28 ± 0.25	4.82 ± 0.70
Median AAD	0.58	−0.04	3.92	9.52
		Disgrazia		
< 0.1	0.41	0.16	3.54	11.11
0.1–0.5	0.63	0.36	2.71	3.31
0.5–1	0.63	0.43	3.14	3.74
1.0–2.0	0.47	0.18	2.82	–
2.0–5.0	0.34	−0.22	1.52	2.17
5.0–10.0	0.43	–	–	–
Total AAD	0.46 ± 0.07	0.10 ± 0.14	2.27 ± 0.15	3.67 ± 0.44
Median AAD	0.47	0.20	3.06	7.14
		Orobie		
< 0.1	0.55	−0.09	3.27	3.77
0.1–0.5	0.55	0.25	2.21	2.04
0.5–1	0.56	–	–	–
1.0–2.0	0.60	–	–	–
Total AAD	0.56 ± 0.15	0.10 ± 0.29	2.69 ± 0.32	2.87 ± 1.02
Median AAD	0.55	−0.11	2.72	2.64

TCD
8, 4075–4126, 2014

TCD
8, 4075–4126, 2014

Post-LIA glacier changes along a latitudinal transect

R. Scotti et al.

Table 5. Topo-climatic attributes of the glaciers selected for inter-annual mass balance analysis.

Glacier[a]	Sub-region	LIA Area (km^2)	2012 Area (km^2)	MA	ABR	S (°)	CSR ($w\,m^2$)	MAP ($mm\,a^{-1}$)	E_{rc} (m a.s.l.)	E_{min} E_{max} (m a.s.l.)	ELA_0 (m a.s.l.)	Ablation stakes (m a.s.l.)
Campo Nord	Livigno	0.84	0.30	NW	moderate	19.1	134	1140	3137	2837–3178	2977	2970–2972
Vazzeda	Disgrazia	1.09	0.23	NE	low	25.3	133	1350	2978	2732–3081	2898	2908–2914
Lupo	Orobie	0.42	0.22	N	high	25.1	96	1680	2844	2435–2760	2545	2555–2564

[a] Glacier attributes are referred to year 2007. The location of the glaciers is reported in Fig. 1b.

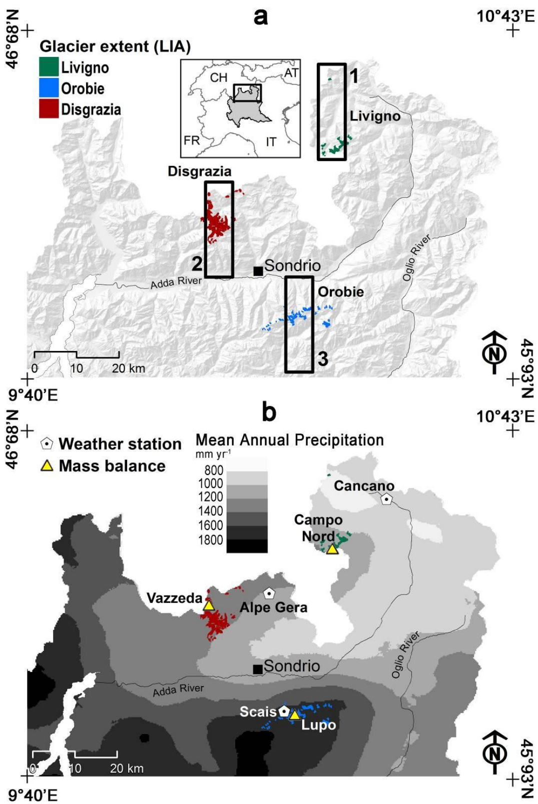


Figure 1. Maps of northern Lombardy showing **(a)** the three sub-region location and the transects used to create the swath profiles (see Fig. 6) and **(b)** spatial distribution of mean annual precipitation with sample weather stations and mass balance measured glaciers (see text for further details). Mean annual precipitation was interpolated by using ordinary co-kriging with 374 rainfall stations (1981–1990) (Ceriani and Carelli, 2000) and 50 000 elevation points randomly distributed within the Region.

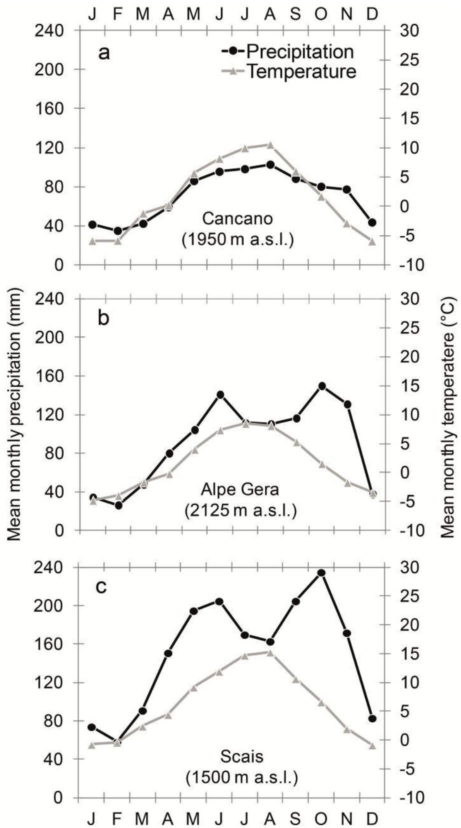


Figure 2. Climographs for Cancano (Livigno sub-region), Alpe Gera (Disgrazia sub-region) and Scais (Orobie sub-region) weather stations. Time series: temperature (1990–2000); precipitation (1951–2000 Cancano, 1990–2000 Alpe Gera and 1958–2000 Scais,). Data sources: Servizio Idrografico e Mareografico Nazionale, Consorzio dell'Adda, ARPA Lombardia, Database OLL – Regione Lombardia D.G.S.P.U.

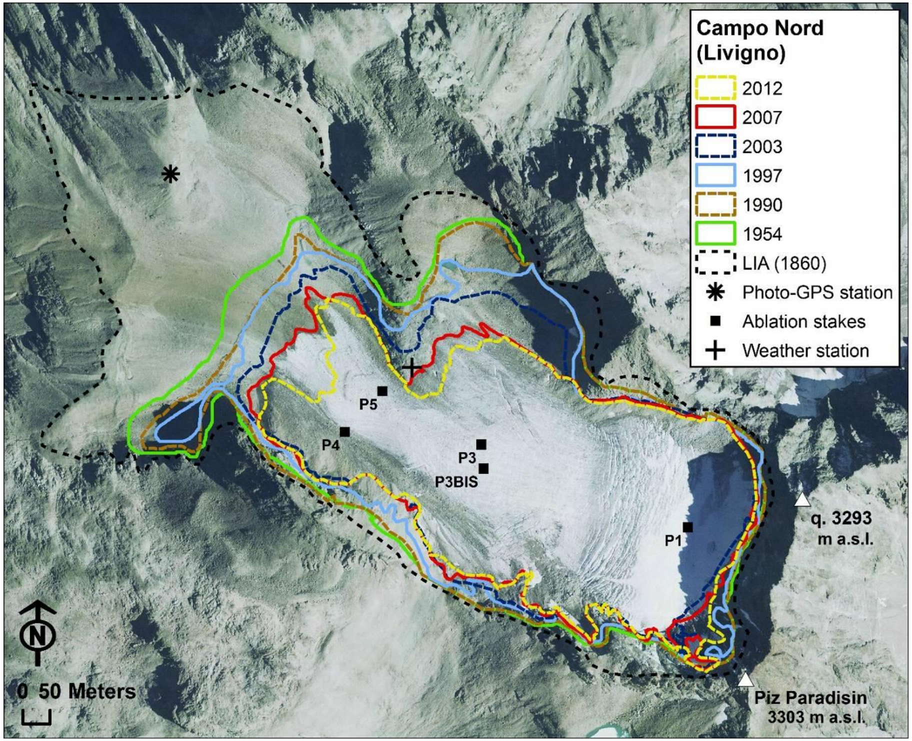


Figure 3. Example of multitemporal glacier delineation i.e., Campo Nord glacier (Livigno sub-region) with 2007 orthophoto in the background.

TCD
8, 4075–4126, 2014

Post-LIA glacier changes along a latitudinal transect

R. Scotti et al.

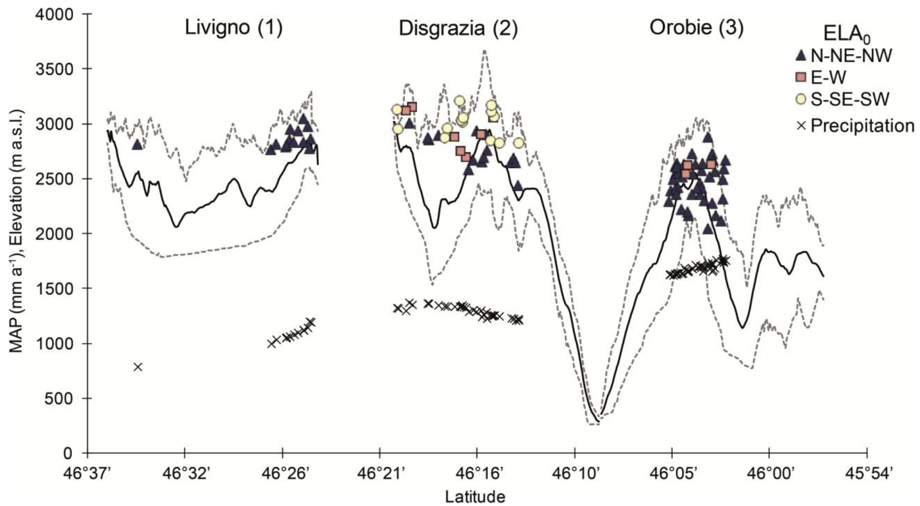


Figure 4. Latitudinal transect across Livigno, Disgrazia, and Orobie sub-regions. Dashed lines indicate minimum and maximum elevation, solid line indicate mean elevation. Filled symbols and crosses refer respectively to ELA_0 (stratified by dominant slope aspect) and Mean Annual Precipitation (MAP) values associated to each study glacier.

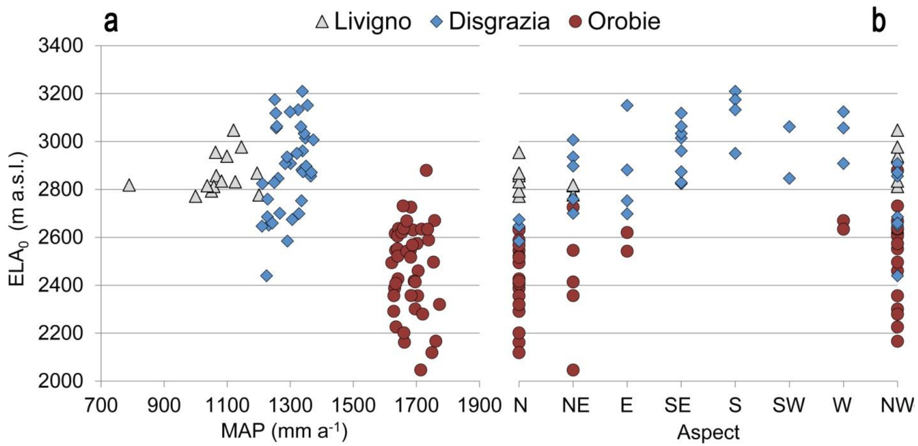
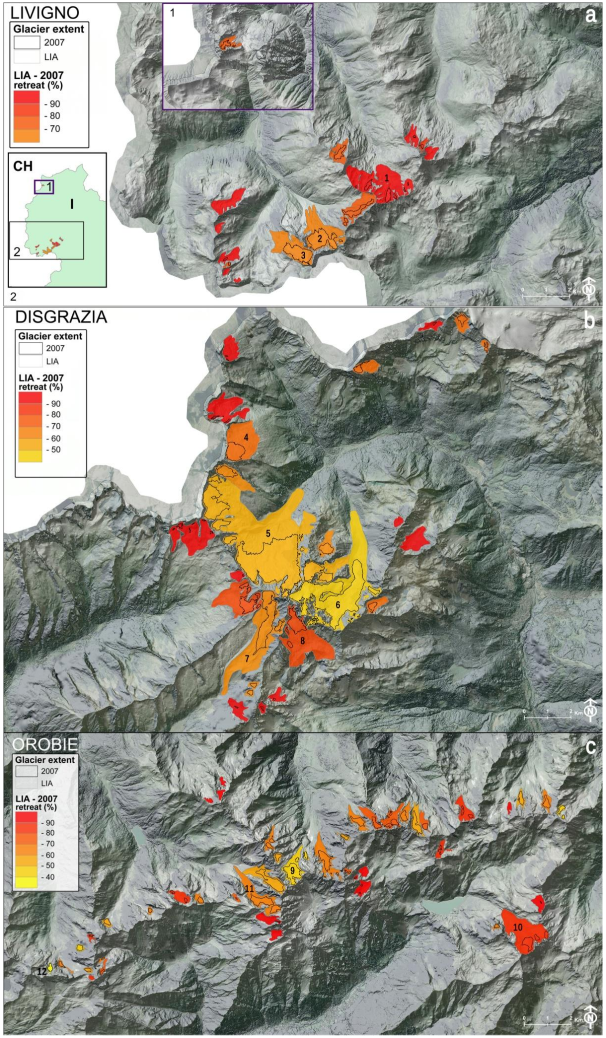


Figure 5. Theoretical equilibrium line altitude (ELA_0) as a function of: **(a)** mean annual precipitation (MAP); and **(b)** slope aspect.



TCD

8, 4075–4126, 2014

Post-LIA glacier changes along a latitudinal transect

R. Scotti et al.

Figure 6. Maps showing the glacier extent in 1860 (LIA) and 2007, and the spatial distribution of the relative change in glacier area in: **(a)** Livigno; **(b)** Disgrazia; and **(c)** Orobie. Numbers refer to glacier cited in the text. 1: Mine, 2: Campo Nord, 3: Val Nera Ovest, 4: Vazzeda, 5: Disgrazia/Sissone, 6: Ventina, 7: Predarossa, 8: Cassandra, 9: Lupo, 10: Trobio, 11: Scais, 12: Aga.

TCD
8, 4075–4126, 2014

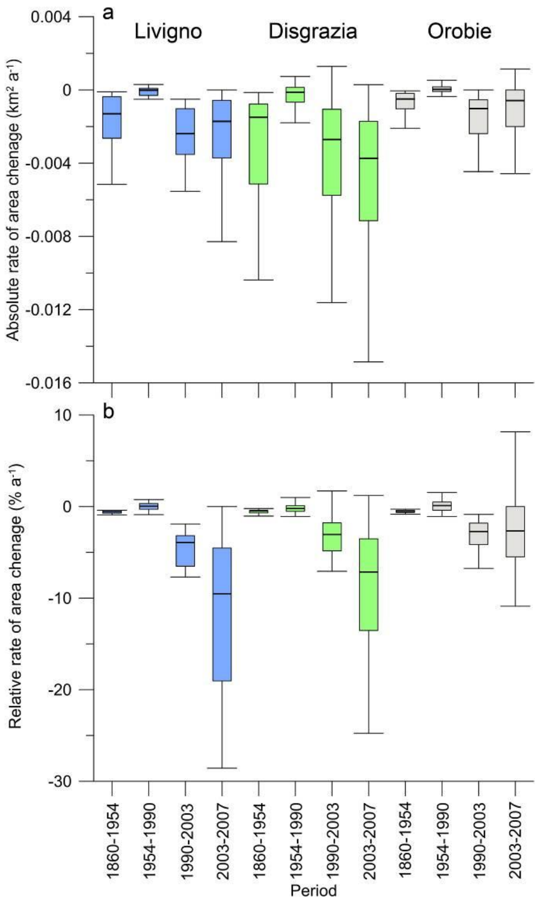


Figure 7. Box-plots showing: **(a)** absolute rate of glacier area change; and **(b)** relative rate of glacier area change. Horizontal lines indicate median values, boxes constrain 25th and 75th percentiles, and whiskers mark 10th and 90th percentiles. Outliers are not presented due to scale constraints.

TCD
8, 4075–4126, 2014

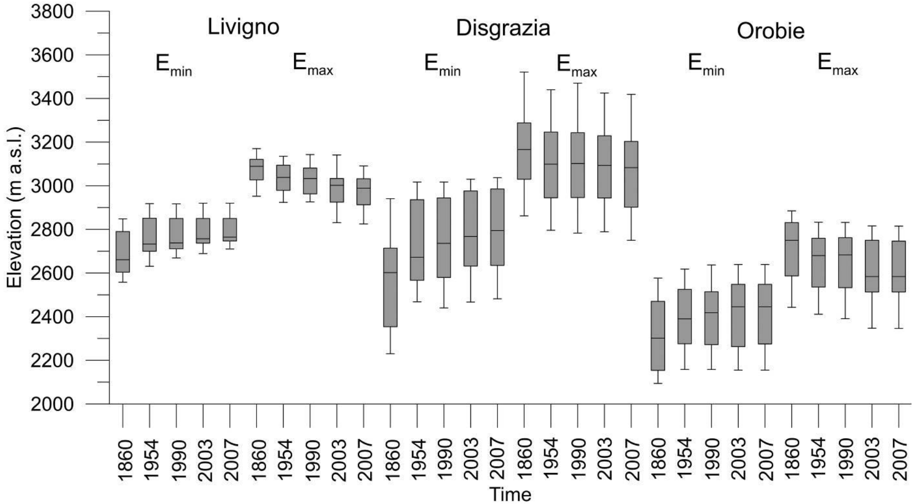


Figure 8. Change in glacier maximum maximum (E_{max}) and minimum (E_{min}) elevation across the 4 study intervals. Horizontal lines indicate median values, boxes constrain 25th and 75th percentiles, and whiskers mark 10th and 90th percentiles. Outliers are not presented due to scale constraints.

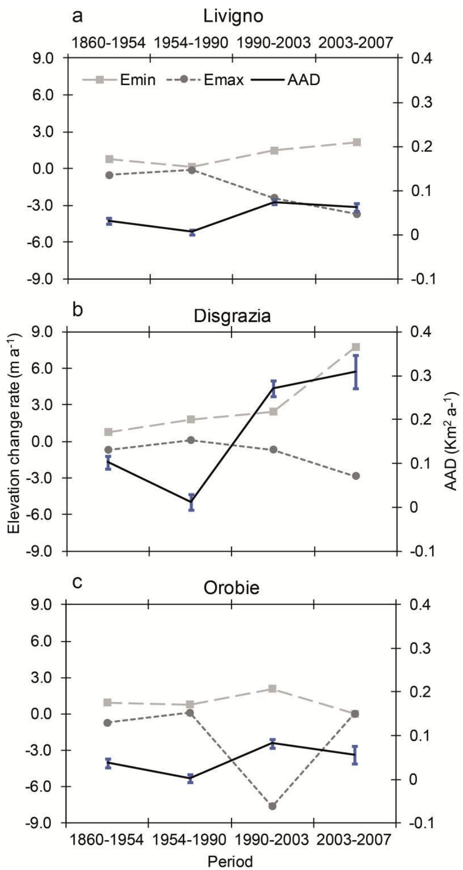


Figure 9. Mean annual elevation change rate ($m\,a^{-1}$) and average annual decrease (AAD) in glacier area ($km^2\,a^{-1}$) in: **(a)** Livigno; **(b)** Disgrazia; and **(c)** Orobie. Bars indicate uncertainty in glacier area delinetation.

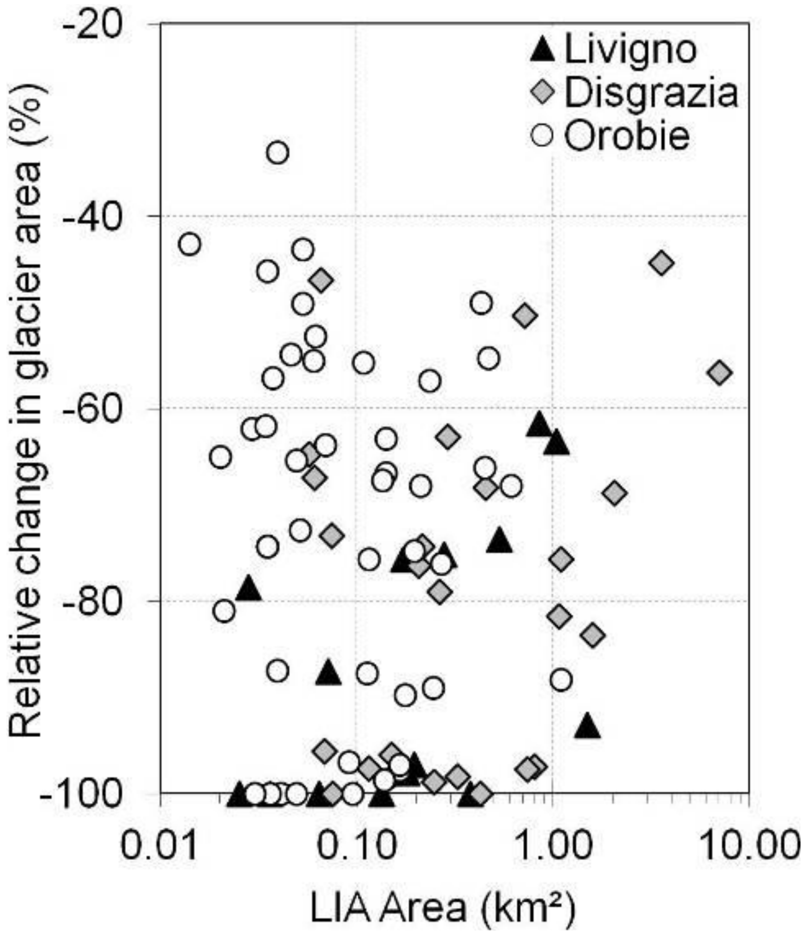


Figure 10. Relative change in glacier area (1860–2007) as a function of former glacier size.

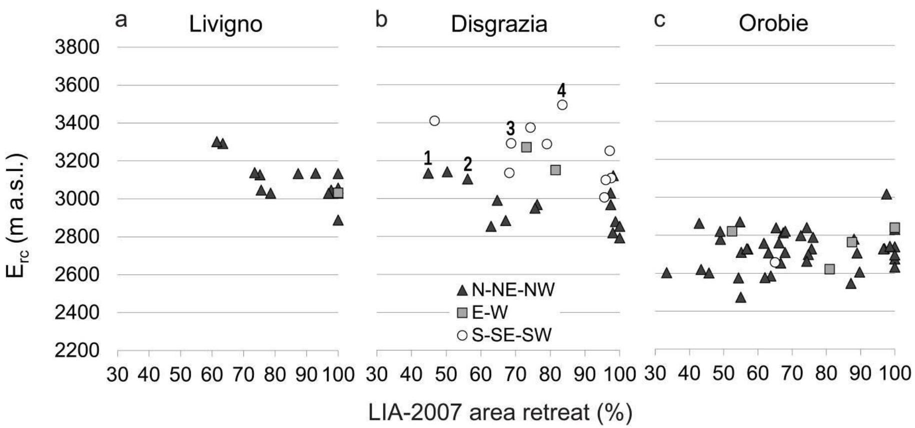


Figure 11. Relative area retreat in (1860–2007) as a function of E_{rc} (ridgecrest elevation upslope of the glacier) in: **(a)** Livigno; **(b)** Disgrazia; and **(c)** Orobie. Glaciers are stratified by dominant slope aspect (note different symbols). Numbers refer to glacier cited in text; 1: Ventina, 2: Disgrazia/Sissone, 3: Predarossa, 4: Cassandra.

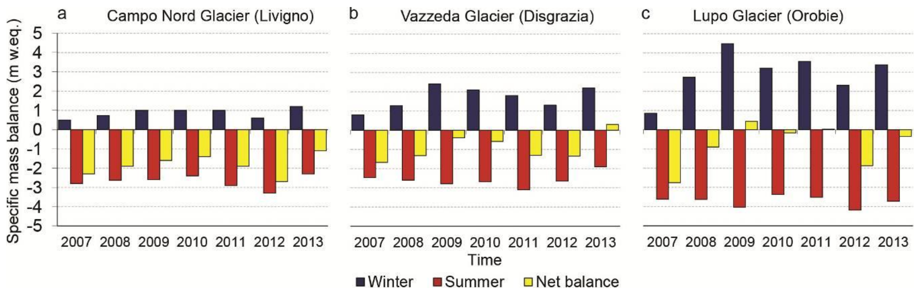


Figure 12. Histogram showing winter, summer, and specific net mass balance at: **(a)** Campo Nord glacier (Livigno); **(b)** Vazzeda glacier (Disgrazia); and **(c)** Lupo glacier (Orobie) from 2007 to 2013. Specific mass balance data are measured with two ablation stakes placed across the ELA_0 of each glacier (see Table 5 for further details).

TCD
8, 4075–4126, 2014

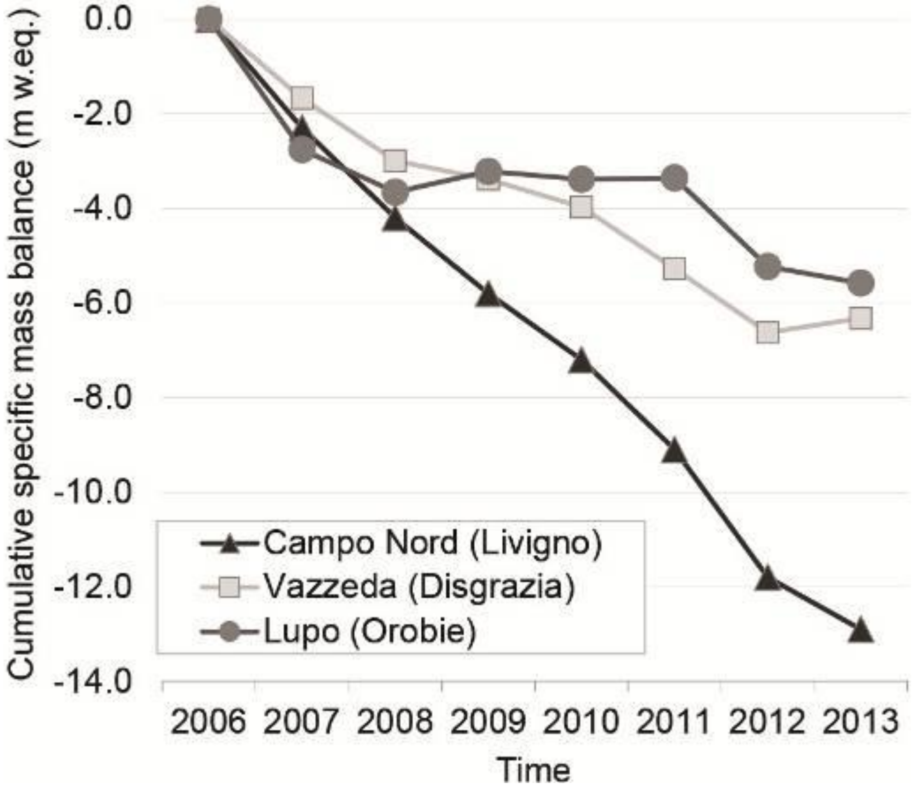


Figure 13. Cumulative specific net mass balance in Campo Nord (Livigno), Vazzeda (Disgrazia), and Lupo (Orobie) glaciers from 2007 to 2013 (see Table 5 for further details).

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