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SPATIAL AND TEMPORAL VARIABILITY OF GLACIERS AND ROCK GLACIERS IN THE CENTRAL ITALIAN ALPS (LOMBARDY REGION)

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Ciclo XXV°

Contents

1. INTRODUCTION

1.1 MOTIVATION

The term "cryosphere" is derived from the Greek word 'kruos' for frost. It is defined as the portion of the Earth's system where water is in solid form, and consisting in ice masses and snow deposits, including ice sheets, ice caps, glaciers, intact rock glaciers, sea ice, snow cover, lake and river ice, seasonally frozen ground and permafrost. The cryosphere is an integral part of the global climate system with important linkages and feedbacks generated through its influence on surface energy and moisture fluxes, clouds, precipitation, hydrology, and atmospheric and oceanic circulation. Through these feedback mechanisms, the cryosphere plays a significant role in regulating global climate and in climate model response to global change (Goodison et al., 1999).

Perennially frozen ground (permafrost) controls soil water content and vegetation over vast regions and is one of the most sensitive components of the cryosphere to atmospheric warming trends. As permafrost thaws, organic material stored in permafrost may release greenhouse gases into the atmosphere and increase the rate of global warming. On the other hand glaciers react quickly to climate effects, influencing ecosystems and human activities on a local scale and act as reliable indicators of climate change.

Glaciers, ice caps and continental ice sheets cover almost 10% (or 14.5×10^6 km^2) of the Earth’s land surface (Table 1.1). This corresponds to about 75% of the Earth’s freshwater (Lemke et al., 2007).

Fig. 1.1 The Cryosphere – world map (UNEP/GRID-Arendal)

A complete melt of continental ice has been estimated to produce a sea level rise of about 64 m, with the Antarctic Ice Sheet contributing for 56.6 m, the Greenland Ice Sheet for about 7.3 m, and all the other glaciers and ice caps for about 0.15 – 0.37 m (Bamber et al., 2001; Lythe et al., 2001; Ohmura, 2004; Dyurgerov and Meier, 2005). Despite the relatively small contribution of mountain glaciers to sea level rise, their role has been recently re-considered from a more composite standpoint (Radic and Hock, 2011). For example, if on one hand glaciers of the European Alps with their 2,909 km^2 (1970s estimate) would contribute to a rise of just one millimeter or less (Zemp, 2006); on the other, they are prominent sources of fresh water for agriculture and industry (e.g., Braun et al. 1999; BUWAL et al. 2004), an important economic component in terms of tourism (e.g., Bonardi, 2012) and hydro-power production (e.g., UNEP, 1992; Barnett et al., 2005), and a potential source of serious natural hazards (e.g., Huggel et al., 2004, Salzmann et al., 2004; Fisher et al., 2006; Frey et al., 2010).

Table 1.1. Area, volume and sea level equivalent (SLE) of cryospheric components. Indicated are the annual minimum and maximum for snow, sea ice and seasonally frozen ground, and the annual mean for the other components. The sea ice area is represented by the extent (area enclosed by the sea ice edge). The values for glaciers and ice caps denote the smallest and largest estimates excluding glaciers and ice caps surrounding Greenland and Antarctica (Lemke et al., 2007).

Cryospheric component		**Area** **(10^6 km^2)**	**Ice Volume** **(10^6 km^3)**	**Potential Sea Level Rise (SLE) (m)**[g]
Snow on land (Northern Hemisphere) Annual (min ~ max)		1.9 ~ 45.2	0.0005 ~ 0.005	0.001 ~ 0.01
Sea ice Annual (min ~ max)		19 ~ 27	0.019 ~ 0.025	0
Glaciers and ice caps (smallest[a] and [largest[b]] estimate)		0.51 - [0.54]	0.05 - [0.13]	0.15 - [0.37]
Ice shelves[c]		1.5	0.7	0
Ice sheets	Total	14.0	27.6	63.9
	Greenland[d]	1.7	2.9	7.3
	Antarctica[c]	12.3	24.7	56.6
Seasonally frozen ground[e] (Northern Hemisphere) Annual (min ~ max)		5.9 ~ 48.1	0.006 ~ 0.065	0
Permafrost[f] (Northern Hemisphere) Annual (min ~ max)		22.8	0.011-0.037	0.03-0.10

Notes:

[a] Ohmura (2004); glaciers and ice caps surrounding Greenland and Antarctica are excluded.

[b] Dyurgerov and Meier (2005); glaciers and ice caps surrounding Greenland and Antarctica are excluded.

[c] Lythe et al. (2001).

[d] Bamber et al. (2001).

[e] Zhang et al. (2003).

[f] Zhang et al. (1999) , excluding permafrost under ocean, ice sheets and glaciers.

[g] Assuming an oceanic area of 3.62×10^8 km^2, an ice density of 917 kg m^{-3}, a seawater density of 1,028 kg m^{-3}, and seawater replacing grounded ice below sea level.

Despite the wealth of studies that have investigated the relationship between glaciers (e.g., areal and volumetric fluctuations) and climate change in many regions of the planet (e.g., Fitzharris et al., 1992; Greene et al., 1999; Oerlemans, 2001; Braithwaite and Zhang, 2000; Hoelze et al., 2003; Bonardi, 2008; Zemp et al., 2008; Abermann et al., 2009; Tennant et al., 2012), the recent rapid downwasting of Alpine glaciers calls for further work in this topic

(Paul et al., 2004, 2007). As advocated by the World Glacier Monitoring Service (WGMS) in order to better understand the coupling between climate and glacier fluctuations, it is critical to extend as far as the Little Ice Age the reconstruction of glacier extent (Zemp et al., 2011). In the central Italian Alps (Lombardy region) (Fig. 1.2) this task has been pursued only for shorter time windows (Citterio et al., 2007; Diolaiuti et al., 2011, 2012) or if so, for a limited number of case studies only (Pelfini, 1988, 1994; Diolaiuti and Smiraglia, 2010): a post-LIA regional appraisal of glacier fluctuations is long overdue. Part of the present doctoral thesis aims to fill this gap.

In the European Alps research on discontinuous permafrost is a young discipline (see review by Haeberli et al., 2011) compared to glaciological studies (e.g., Agassiz, 1840; Marcou, 1886; Dana, 1886). The rapid increase of air temperature recorded in the last 30 years and the relevant abrupt changes in the hydro-geomorphic regime of high mountain environments have produced a growing interest on Alpine permafrost. The spatial distribution of permafrost has been shown to be highly sensitive to climate change (Barsch, 1996; Haeberli et al., 2011), and its degradation is regarded as one of the main sources of environmental risk at high altitude causing a possible increase of rockfalls and debris flows (e.g., Haeberli and Beniston, 1998; Haeberli, 2005; Harris et al., 2009; Fisher et al., 2006, 2012; Deline et al., 2011).

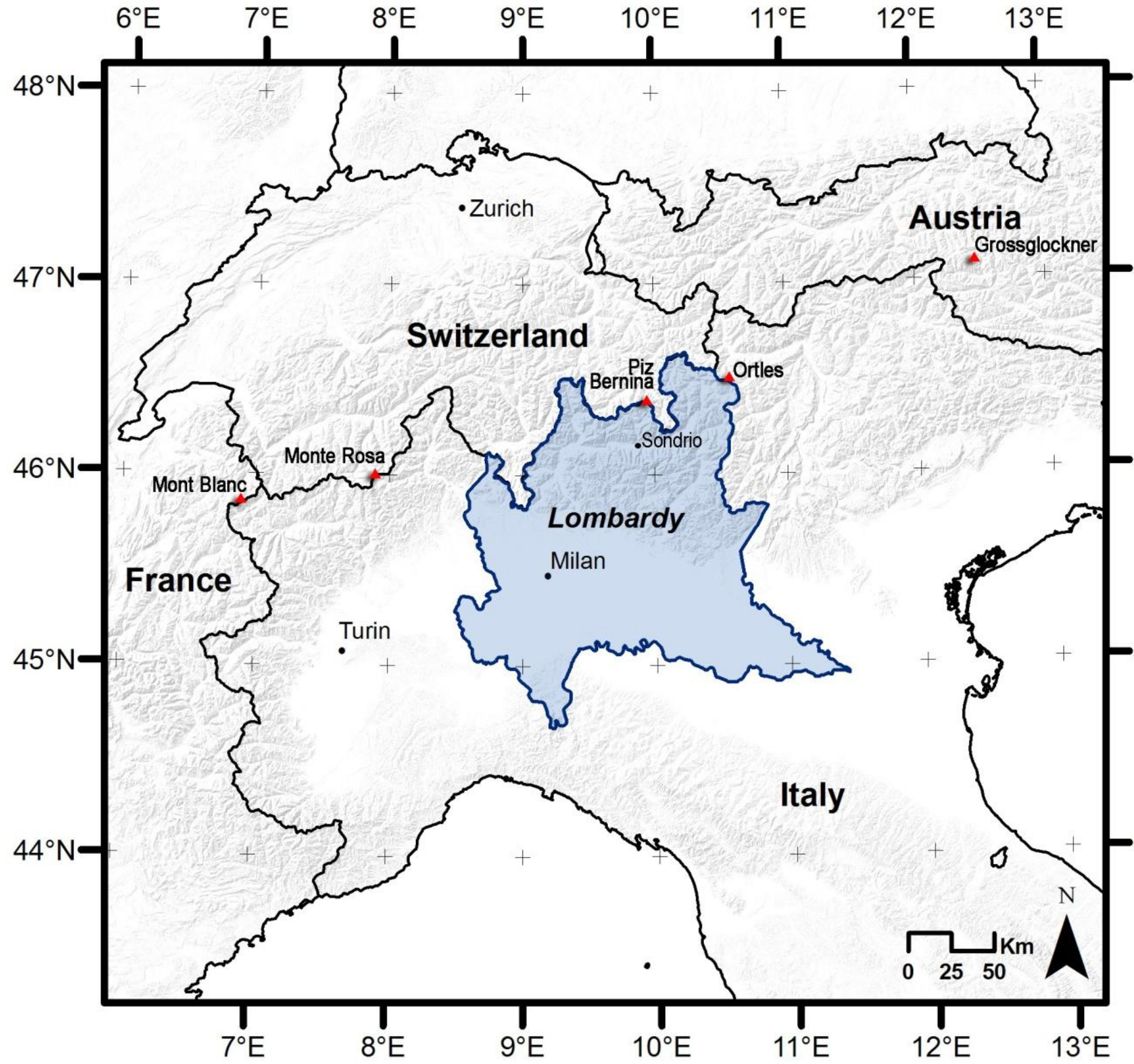

Fig. 1.2. The Alps with the main summits and Lombardy region. Background: ASTER Digital Elevation Map (METI-NASA). Country boundaries provided by ESRI.

In this context, documenting the spatial distribution of rock glaciers and understanding the relevant controlling factors holds implications and useful applications for modelling the past and/or present extent of discontinuous permafrost (e.g., Imhof, 1996; Frauenfelder et al., 2001, Lambiel and Reynard, 2001; Etzelmüller et al., 2003; Boeckli et al., 2012), as well as for assessing geohazards related to permafrost degradation (Haeberli and Burn, 2002).

To date, no quantitative and systematic recognition of such landforms has been made for the Central Italian Alps. The inventory of rock glaciers and protalus ramparts here presented, which represents a necessary preliminary step for modelling the spatial distribution of discontinuous permafrost at the regional scale, fills a critical geographic gap in the context of ongoing permafrost research in the European Alps (e.g., PermaNET).

1.2 AIMS

This study aims to investigate the cryosphere of the Lombardy region (Central Italian Alps) by compiling and analysing glacier, rock glacier, and protalus rampart inventories. To this purpose, I have employed remote sensing techniques and in-situ measurements for delineating the contemporary and historical extent of glaciers and contemporary extent of rock glaciers and protalus ramparts. In order to better identify and highlight the response of the regional cryosphere to climate change, the LIA maximum extent has been reconstructed for nine selected glaciers and their fluctuations since then have been discussed and analyzed in relation to temperature and precipitation recorded at ground weather stations.

The main objectives of this doctoral dissertation are to:

- compile glacier, rock glacier and protalus rampart regional inventory for the Central Italian Alps (Lombardy region) exploitable for investigations about permafrost distribution;
- elucidate the linkages between the occurrence of periglacial landforms (rock glaciers and protalus ramparts) and local litho-topographic attributes;
- examine the variability of periglacial activity in relation to terrain elevation and mean annual precipitation;
- gain new insights about the impact of the Pleistocene-Holocene climatic transition on the periglacial domain within the study area;
- examine the linkages between glaciers location, glaciers attributes (e.g., size, aspect, ELA_0) and mean annual precipitation;
- analyze the relations between glaciers location, size, aspect, ELA_0 and areal variations since 1991;
- analyze post-LIA (1860) glacier fluctuations for nine selected glaciers from different sub-regions of the central Italian Alps;
- investigate glacier sensitivity to climate change (i.e., precipitation and temperature).

2. A REGIONAL INVENTORY OF ROCK GLACIERS AND PROTALUS RAMPARTS IN THE CENTRAL ITALIAN ALPS (LOMBARDY REGION)

This chapter is based on the eponymous manuscript accepted for publication on Geomorphology December 13, 2012 and published online January 4, 2013.

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2.1 ABSTRACT

We present a regional inventory of rock glaciers (*n* =1514) and protalus ramparts (228) from the central Italian Alps, Lombardy region. To identify and classify the landforms we inspect three sequential air-orthophoto mosaics and a 2m-DSM, and conduct confirmatory field work. The inventory forms an empirical basis to analyze: (i) the relative contribution of hillslope (i.e., talus slopes) and glacial (i.e., moraines) sediment stores to rock glacier sediment supply; (ii) linkages between the landforms inventoried and local topographic attributes; (iii) the spatial variability of periglacial activity in relation to a parsimonious set of environmental variables (i.e., elevation, precipitation, and lithology); and (iv) the effects of the Pleistocene-Holocene climatic transition on the distribution of intact and relict landforms. This analysis reveals that the elevation of rock glacier termini can vary over 200 m as a function of slope aspect. In turn, the distribution of rock glaciers among aspect categories is controlled by the structure of the valley network that promotes NW and SE exposures. Talus rock glaciers prevail numerically over the glacier-related typology, even though the latter population appears to have increased during the Holocene. Relict and intact rock glaciers have distinct spatial patterns in that the former display, on average, a 400-m elevation drop and a less clustered distribution towards northern aspects, suggesting that they have developed in more "permafrost-prone" climatic conditions. Analyzing the study region through a 27.5 km-grid has been instrumental for showing that rock glacier specific area and terminus elevation are: (i) positively correlated with terrain elevation; and (ii) negatively correlated with mean annual precipitation. As a consequence, in relation to Holocene generalized atmospheric temperature rise, intact rock glaciers have progressively disappeared from the wetter and milder portions of the central Italian Alps. Analysis of rock glacier occurrence across litho-tectonic sectors does not provide conclusive dependences and requires further analysis. This inventory, which represents a necessary preliminary step for modelling the spatial distribution of discontinuous permafrost at the regional scale, fills a critical geographic gap in the context of ongoing permafrost research in the European Alps (e.g., PermaNET).

Highlights:

1. We compile a regional inventory of rock glaciers and protalus ramparts.
2. We analyze landform location in relation to topography, precipitation, and lithology.
3. We show strong interactions between landform aspect and elevation.
4. Rock glacier elevation and specific area correlate with elevation and precipitation.
5. Relict and intact rock glaciers display climate-driven spatial distributions.

2.2 INTRODUCTION

Rock glaciers are depositional features that characterize the periglacial geomorphic domain in high mountain areas, such as the European Alps (Haeberli et al., 2006). They usually form as a result of downslope creep of perennially frozen ice-rich debris (Harris et al., 2009), although there are examples showing that glacial/former glacial activity can represent an alternative formative mechanism (e.g., Whalley and Martin, 1992; Krainer and Mostler, 2000; Whalley and Azizi, 2003). The development of active (i.e., mobile) rock glaciers typically requires: (i) a climate that warrants ground thermal conditions appropriate for the development of discontinuous permafrost which is dry enough to prevent the formation of glaciers; (ii) sufficient sediment supply deriving from the degradation of upslope/upstream source areas such as rock walls and moraines; and (iii) topographic gradient that can support creep of the rock-ice mixture (Barsch, 1996; Harris et al., 2009).

Documenting the spatial distribution of rock glaciers and understanding the relevant controlling factors holds implications and useful applications for modelling the past and/or present extent of discontinuous permafrost (e.g., Imhof, 1996; Frauenfelder et al., 2001, Lambiel and Reynard, 2001; Etzelmüller et al., 2003; Boeckli et al., 2012), as well as for assessing geohazards related to permafrost degradation (Haeberli and Burn, 2002). From an exclusively geomorphic standpoint, rock glaciers constitute prominent sedimentary linkages within the alpine environment, and as such modulate large segments of the coarse sediment cascade (Caine, 1974).

A number of rock glacier studies have considered sub-regions, single lithological units, or valleys (e.g., Luckman and Crockett, 1978; Imhof, 1996; Hoelzle, 1998; Lambiel and

Reynard, 2001; Baroni et al., 2004; Janke and Frauenfelder, 2007; Scapozza and Fontana, 2009; Lilleøren and Etzelmüller, 2011); others have encompassed large portions of entire orogens (e.g., Barsch, 1977; Höllermann, 1983; Guglielmin and Smiraglia, 1997). While both approaches have contributed substantially to improve our understanding of the alpine environment, in the former case the limited number of landforms inventoried (e.g., from a few tens to a maximum of about 300) and the limited extent of the study area restrict the reliability of relevant statistical analyses, and the ability to assess the impact of environmental variables (e.g., precipitation), which tend to strongly change over a broad spectrum of spatial scales. In the latter case, since these inventories were compiled before the advent of high-resolution digital elevation models, and in some cases of GIS, the evaluation of terrain attributes suffered of insufficient accuracy and/or spatial resolution.

The present work, which focuses on geomorphic aspects of the alpine periglacial environment, aims at addressing the foregoing research gap. In particular, access to an historical distributed record of precipitation data (Ceriani and Carelli, 2000) and to high-resolution topography and photography enables us to search for dependences between rock glacier activity and a parsimonious set of environmental variables (i.e., elevation, precipitation, and lithology). By compiling an inventory of rock glaciers and protalus ramparts at the regional scale we seek to: (i) elucidate linkages between the occurrence of periglacial landforms and local litho-topographic attributes; (ii) examine the variability of periglacial activity in relation to terrain elevation and mean annual precipitation; and (iii) gain new insights about the impact of the Pleistocene-Holocene climatic transition on the periglacial domain across the central Italian Alps.

2.3 STUDY AREA

The study area includes the Alpine sector of the Lombardy Region (northern Italy), with elevation ranging from 180 m at Iseo Lake to 4049 m a.s.l. at Pizzo Bernina (Fig. 2.1a). Precipitation displays high regional variability in terms of total annual values (Fig. 2.1b) and seasonal distribution (Fig. 2.2), so that climate acquires an increasingly more continental character moving towards north and east. Extremes are found in the southern (Orobic Alps) and western mountain sectors (Val Chiavenna) in which mean annual precipitation reaches 2000 mm and more, with seasonal maxima in late spring and in autumn; as well as in the northeastern part of the region (Livigno Valley), where precipitation drops below 700 mm, and exhibits a summer maximum (Ceriani and Carelli, 2000). For reference we report monthly precipitations at Scais (1500 m a.s.l., Northern Orobic Alps) and Cancano (1950 m a.s.l., Saliente-Braulio) (Fig. 2.2). The former shows two precipitation peaks in June and October, the latter in August. Mean Annual Air Temperature (MAAT) is 6.3°C at Scais and 1.7°C at Cancano. In both stations December and August are respectively the coldest and hottest months; mean temperatures stay below 0°C for five (Cancano) and three (Scais) months (Fig. 2.2).

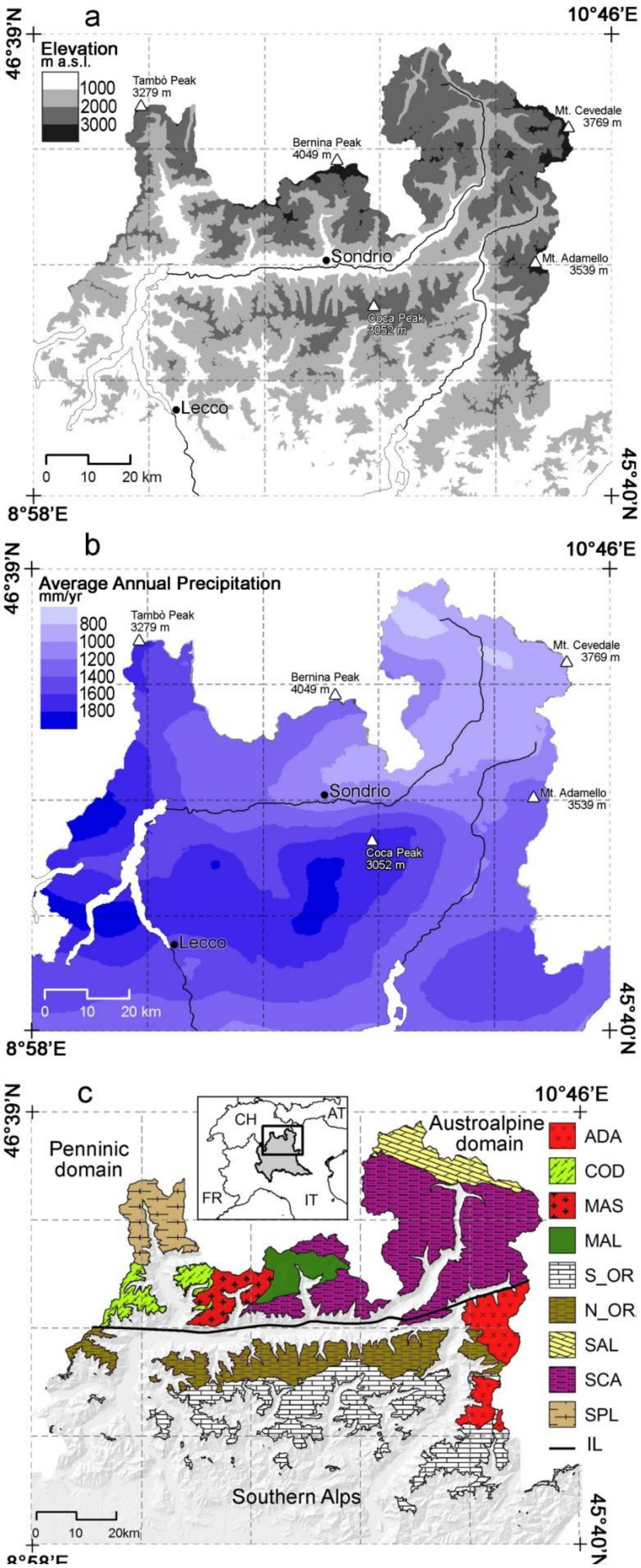

Fig. 2.1
Maps of northern Lombardy showing the spatial distribution of
(a) elevation, and (b) mean annual precipitation. Mean annual precipitation was interpolated by using ordinary co-kriging with 374 rainfall stations (1891-1990) and 50,000 elevation points randomly distributed within the Region. (c) Litho-tectonic sectors: ADA = Adamello, COD = Berlinghera-Codera, MAS = Masino, MAL = Malenco, S_OR = Southern Orobic, N_OR = Northern Orobic, SAL = Saliente-Braulio, SAC = Scalino-Cevedale and SPL = Spluga. Gridlines enclose squares 27.5-m wide (see text for further details).

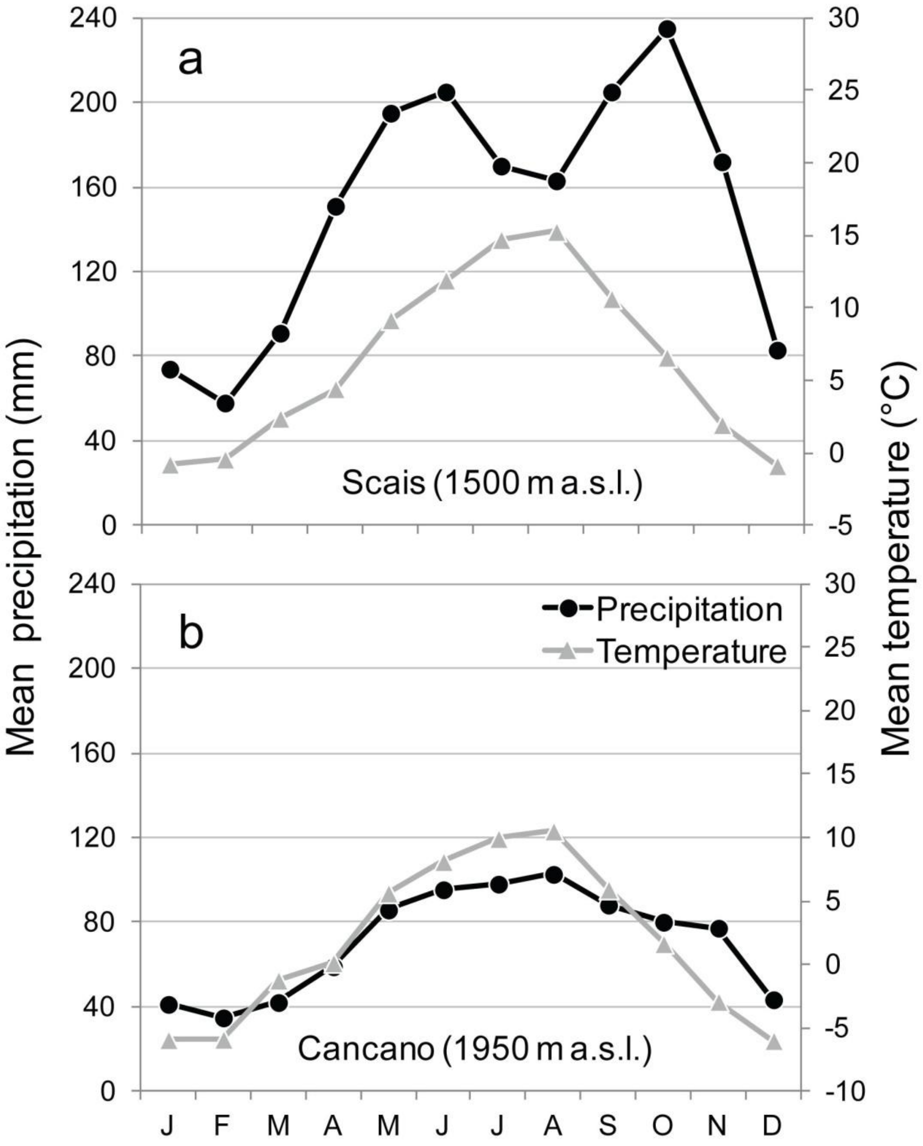

Fig. 2.2. Climographs for Scais (1500 m a.s.l., 46°45’21’’N – 9°55’34’’E) and Cancano (1950 m a.s.l. 46°30’52’’N – 10°19’01’’E) weather stations. Time series: temperature (1990-2000); precipitation (1958-2000 Scais, 1951-2000 Cancano). Data sources: Servizio Idrografico e Mareografico Nazionale, Consorzio dell’Adda, ARPA Lombardia, Database OLL – Regione Lombardia D.G.S.P.U.

The structural setting of the region comprises three main units: the Southern Alps, the Penninic unit, and the Austroalpine basement. These units are separated by the Insubric line, representing a steeply north dipping and east-west trending fault zone (Fig. 2.1c). The Southern Alps represent the youngest part where rocks can be subdivided into basement (metasedimentary rocks) and sedimentary covers (terrigenous and carbonate rocks). The units located north of the Insubric line consist of the Austroalpine nappes, to the east, and the Penninic nappes, to the west. Austroalpine units, although of similar paleogeographic provenance as the Southern Alps, consist of a completely rootless metamorphic basement (metasedimentary and metaigneous rocks) and sedimentary cover (carbonate rocks) which were detached from their lithosphere as early as the Cretaceous orogenesis (Froitzheim et al., 1994). The Penninic units are of extremely heterogeneous paleogeographic provenance, including remnants of oceanic lithosphere (Malenco-Forno Unit, metaigneous rocks) as well as basement of the European margin (Adula, Tambò and Suretta Units, metasedimentary rocks). Two important Tertiary tonalitic to granodioritic batholiths occupy large portions of Valchiavenna (Masino-Bregaglia) and Valcamonica (Adamello). In order to analyze the relationship between the landforms inventoried and geology, the study area was divided in 9 litho-tectonic sectors (Table 2.1 and Fig. 2.1c) characterized by distinct dominant lithology and tectonic history. This subdivision is based on the 1:250.000 regional geology map (Montrasio, 1990).

The foregoing structural context has been repeatedly overridden by Pleistocene glaciations (the Last Glacial Maximum occurred around 21 ka cal. BP (Keller and Krayss, 1993)), and relevant paraglacial responses along the major valley network (Hinderer, 2001) and on the slopes (Ballantyne, 2002). As a consequence, glacial and glacio-fluvial deposits blanket large portions of the mountain landscape. Today, in the region survive 203 active glaciers and glacierets that cover a total of 90.5 km² (Galluccio and Scotti, 2012).

Table 2.1. Litho-tectonic sectors of the Lombardy region

Litho-tectonic sector	Area (km²) [a]	Dominant lithology	Tectonic realm
Codera-Berlinghera (COD)	135.3	Granites & granodiorite gneisses	Lepontine basement Nappes
Spluga (SPL)	192.5	Paragneiss	Tambò-Suretta Nappes
Masino (MAS)	141.8	Quartzdiorites & tonalites	Bregaglia Tertiary Intrusive
Malenco (MAL)	134.1	Serpentinites, prasinites & metabasalts	Malenco Metaophiolites
Scalino-Cevedale (SCA)	1145.7	Paragneiss, phyllites & micaschists	Austroalpine basements
Saliente-Braulio (SAL)	179.7	Dolostones & limestones	Austroalpine covers
Adamello (ADA)	238.1	Quartzdiorites & tonalites	Adamello Tertiary Intrusive
Northern Orobic (N_OR)	538.4	Paragneiss, phyllites & micaschists	Southalpine basements
Southern Orobic (S_OR)	846.0	Conglomerates, marls & limestones	Southalpine covers

a. Not including glaciers, lakes, rivers, main alluvial fans and terrain below 1460 m a.s.

2.4 METHODS

2.4.1 COMPILATION OF THE ROCK GLACIER AND PROTALUS RAMPART INVENTORY

The regional inventory of rock glaciers and protalus ramparts was compiled via inspection of high-resolution (0.5-m pixel) sequential orthophotos (flown in 2000, 2003, and 2007) and a 2-m gridded Digital Surface Model (DSM, 2007). To assess the reliability of the identification and classification procedures we conducted complementary field surveys. Observations and measurements made during fieldwork provided critical ground control for data extracted from remotely-based analysis (e.g., Fig. 2.3).

The availability of the foregoing multiple and multi-temporal sources of imagery has been critical for reducing uncertainty due to vegetation, clouds, snow cover, and long shadows on steep north-facing slopes.

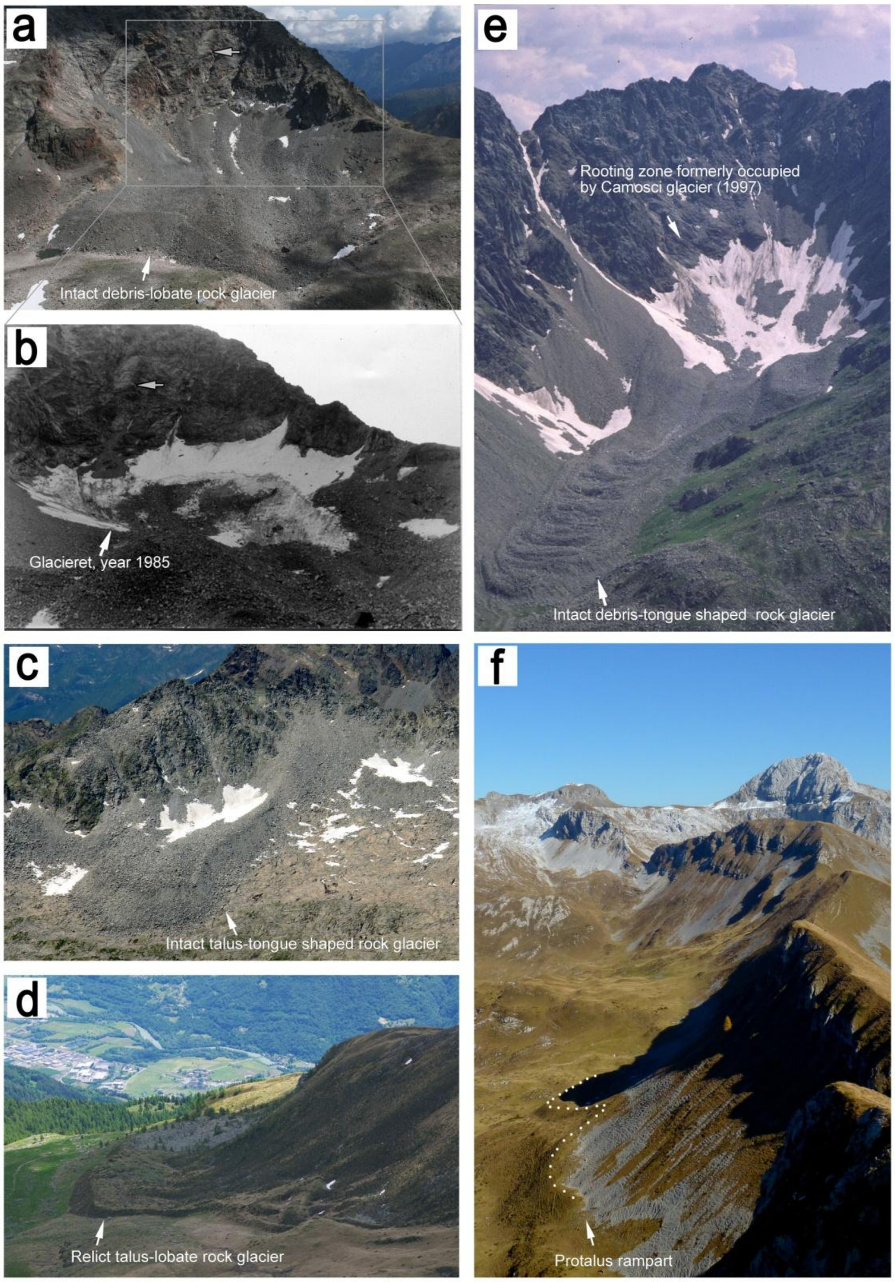

Fig.2.3. Rock-glacier classification examples: (a) intact debris-lobate rock glacier (46°22’53’’N – 10°11’07’’E) (year 2008); (b) the rock glacier rooting zone was occupied by the Passo Dosdè glacieret until 2006 (year 1985); (c) intact talus-tongue shaped rock glacier (46°09’45’’N – 10°27’24’’E) (year 2010); (d) relict talus-lobate rock glacier (46°12’54’’N – 9°56’37’’E) (year 2012); (e) intact debris-tongue shaped rock glacier (46°14’18’’N – 9°56’43’’E) with the rooting zone formerly occupied by the Camosci glacier until 1996 (year 1997); (f) protalus rampart (45°58’07’’N – 9°59’55’’E) in the lower portion of the picture (year 2012). Photos by R. Scotti (a, c, d), M. Lojacono (b), M. Butti (e) and L. Vezzoni (f)

For example, the 2003 ortho-mosaic allowed to improve substantially our capability to identify a number of landforms that were partially covered with snow in the 2000 and 2007 photo sets. In addition, the pseudo-3D vision of the DSM hillshade was extremely useful to interpret the topographic curvature of landforms, hence to delineate rock glacier boundaries and lobes.

The inventory adopts the morphological classification by Barsch (1996), and follows the specifics detailed by Seppi et al. (2005), Scapozza and Mari (2010), and Cremonese et al. (2011). Landform attributes include, geographic coordinates (i.e., centroid), mountain sector, surface area, elevation (minimum (i.e., terminus), maximum, and mean), slope gradient, slope aspect (manually defined along the direction of the main flow axis), vegetation at the front, and upstream presence/absence of a glacier or perennial snow field. The latter attribute is based on field checks and relies on the regional glacier inventory (Bonardi et al., 2012). In addition, we adopt a classification scheme that combines genetic (talus vs. debris), geometric (tongue-shaped vs. lobate), and dynamic (active/inactive vs. relict) attributes. Landform topographic attributes (i.e., elevation, slope gradient, and slope aspect) have been extracted from the 2-m gridded DSM.

Depending on the source of the sedimentary material transported downslope we classify rock glaciers into talus and debris ones. A *talus rock glacier* (Fig. 2.3c and 2.3d) is typically located at the base of a talus slope in which there is no visible ice upslope of the rock glacier body, and that, as such, transports mainly frost-shattered rock fragments derived from adjacent rock walls (e.g., Barsch, 1988, 1996; Lilleøren and Etzelmüller, 2011). A *debris rock glacier* (Fig. 2.3a and 2.3e) forms downslope of end moraines of (mostly) small glaciers

(Fig. 2.3b) and transports mainly reworked glacial debris (till) (Barsch, 1996). In this context, a clear distinction must be made with Humlum’s (1982, 1988) similar classification of *glacier derived-rock glaciers*. The term *debris rock glacier*, adopted in this inventory, exclusively relates to the sedimentary material and does not infer anything about the ice content, which cannot be determined from visual inspection alone. In summary, our classification scheme is basically concerned with the sedimentary connectivity between rock glaciers and: (i) talus slopes (i.e., talus rock glaciers) and (ii) moraines (i.e., debris rock glaciers).

The accelerated trend of glacier retreat and extinction of the last three decades (Paul et al., 2004; Galluccio and Scotti, 2012; Diolaiuti et al., 2012) has promoted fluvial overprinting of morainal deposits, hence increasing the degree of uncertainty for assuredly classifying the genetic origin of rock glaciers’ debris. In this work, we consider our remotely-based classification reliable for debris rock glaciers that have formed from the Little Ice Age (LIA) and younger moraines. Often, the degree of dissection associated with older glacial depositional landforms does not allow to disentangle talus from debris rock glaciers. As a result, it is likely that the number of the latter typology has been somewhat underestimated.

We further subdivide our inventory on the basis of rock glacier plan geometry (i.e., length to width ratio). Accordingly, rock glaciers are termed *tongue-shaped* (length/width >1) (Figs. 2.3c and 2.3e) and *lobate* (length/width <1) (Figs. 2.3a and 2.3d) (Wahrhaftig and Cox, 1959; Luckman and Crockett, 1977; Barsch, 1996; Imhof, 1996; Guglielmin and Smiraglia, 1997; Baroni et al., 2004; Nyenhuis et al., 2005).

The degree of activity (e.g., mobility) is probably the most important rock glacier attribute, both in scientific and applied terms, as it holds critical information for inferring the altitudinal distribution of discontinuous permafrost (current and past) (Barsch, 1996). In our inventory we distinguish between intact (Haeberli, 1985; Barsch, 1996; Cremonese et al., 2011; Lilleoren and Etzelmuller, 2011) (e.g., Figs. 2.3a, c, e and 2.4e) and relict rock glaciers (e.g., Fig. 2.3d). The former category includes active and inactive rock glaciers. Active rock glaciers have steep fronts and side slopes, mostly steeper than the angle of repose of the material. The upper surface is normally covered by boulders with a micro-relief of furrows and ridges, surface expression of decelerating viscous or plastic flow due to the presence of abundant ice (Barsch, 1996). Inactive rock glaciers do also contain ice, but are no longer mobile either due to melting of most of the upper layers within the front slope (climatically inactive) or topographic constrains and/or lack of material supply from the surrounding landscape components (dynamically inactive), while the frozen core of the rock glacier is protected from melting by the sediment cover (Barsch, 1996; Lilleøren and Etzelmüller, 2011).

Relict rock glaciers (e.g., Fig. 2.3d) are defined as formerly active landforms in which ice is vanished. They are characterized by collapsed structures at their surface, and their surface relief is much more subdued than on intact ones (Barsch, 1996). Normally they sit around or below the current tree line, have extensive vegetation cover and a less steep front compared to the intact ones (Seppi et al., 2005; Scapozza and Mari, 2010; Lilleøren and Etzelmüller, 2011).

Based on a number of prior studies conducted in the Alps, including a limited number of direct dates (e.g., Mortara et al., 1992; Calderoni et al., 1998; Haeberli et al., 1999b; Dramis et al., 2003), intact rock glaciers are thought to have formed between the end of the Subboreal and the Subatlantic (between 3100 BP and today), but some of them can be older and date back as far as the Atlantic (between 8000 and 5000 years BP). In fact, much older forms might exist in places where micro-climate conditions have warranted persistence of permafrost and glaciations did not interrupt rock glacier creep. As for relict rock glaciers, other investigations relying on the integration of present rock glacier velocity fields through time (e.g., Frauenfelder and Kaab, 2000; Frauenfelder et al., 2001), on their morphological relations with moraine ridges (Scapozza et al., 2009; Seppi et al., 2010), and direct dates (e.g., Böhlert et al., 2011) indicate that these landforms mainly developed during the Alpine Late Glacial (most probably in the Younger Dryas) and that they decayed between the end of the Alpine Late Glacial and the beginning of the Holocene.

For each rock glacier we have mapped the whole landform surface from the rooting zone to the foot of the front slope (Barsch, 1996). The upper boundary of rock glaciers was the most critical part in the mapping procedure (e.g., Roer and Nyenhuis, 2007). In particular, the presence of glaciers and glacierets during the LIA have caused in many situations the deposition of lateral and terminal moraines overlapping with the rock glacier rooting zone. In those cases where a debris rock glacier is gradually developing from a moraine, a clear distinction between the two landforms cannot be set and we delineated the whole body (i.e., moraine plus rock glacier, note circle in Fig. 2.4b).

In the case of multiple rock glaciers coalescing into one body, the delineation of single polygons can become subjective and as such it requires the use of general rules. In our inventory, when the frontal lobes of two (or more) rock glaciers originating from distinct source basins join downslope, we consider the two components as separate bodies (e.g., Figs. 2.4, RG 1 and 2). Where the limits between lobes are unclear and the lobes share other morphological characteristics (e.g., degree of activity, vegetation cover), we classify the whole system as an unique rock glacier (e.g., Fig. 2.4, RG 4). Lobes originating from the same source area developing along the same flow line are considered belonging to distinct rock glaciers only if we can clearly relate them to different pulses/cohorts (Fig. 2.4).

Our inventory includes also protalus (pronival) ramparts, a landform belonging to the alpine domain and morphologically similar to rock glaciers. Their identification and delineation is an intriguing yet challenging question (i.e., Whalley, 2009). According to Shakesby (1997; 2004), a pronival rampart is a ridge, a series of ridges, or a ramp of debris formed at the downslope margin of a perennial or semi-perennial snow bed, which is typically located near the base of a steep bedrock slope in an alpine environment. Other authors (Barsch, 1977; Haeberli, 1985; Scapozza at al., 2011) see protalus ramparts as embryonic rock glaciers, and as such hypothesize that permafrost creep plays an important role in their formation. These two explanations are closer than what it may appear if one considers ground thermal conditions. In fact, perennial or long-lasting snow induces permafrost conditions, because on if one hand it cuts the summer heat, on the other it does not prevent winter cold from penetrating into the ground.

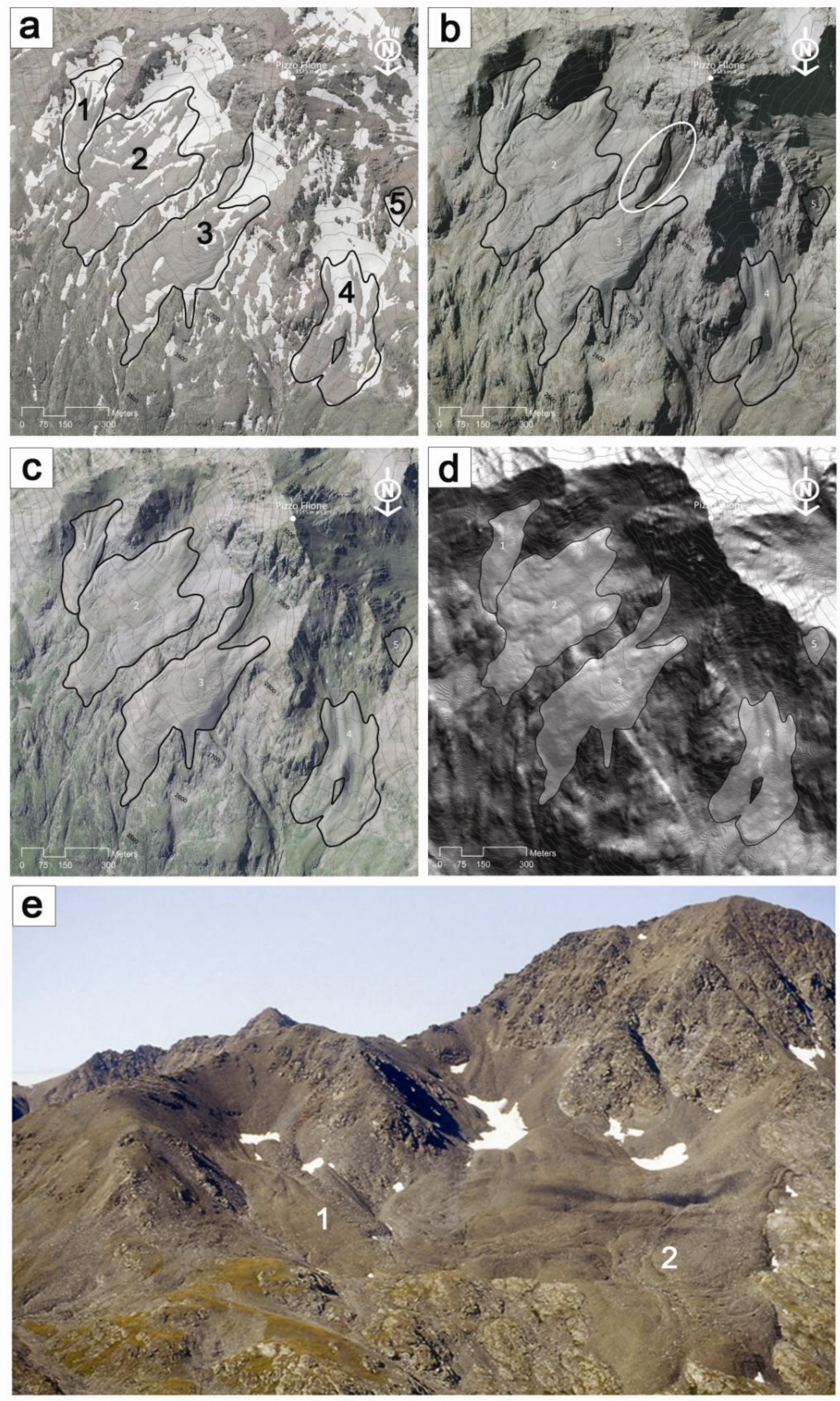

Fig. 2.4. Rock-glacier mapping in the upper Valtellina (Pizzo Filone 46°27’23’’N – 10°9’50’’E) perfomed via interpretation of aerial orthophotos flown in (a) 2000; (b) 2003; (c) 2007; and (d) a shaded relief derived from the 2007 2m-DSM. (e) Oblique terrestrial view of rock glaciers 1 and 2 (Photo by M. Marzorati).

Nevertheless, the debate on the genesis and the classification of these landforms remains still open (e.g., Hedding, 2011) and a clear genetic distinction is beyond the scope of our work. In this paper, we decided to map protalus ramparts using the general morphological definition proposed by Shakesby (2004). Due to the difficulties in resolving the degree of activity by means of aerial photo interpretation, this attribute was not defined for protalus ramparts. Finally, to account for the subjectivity associated with the identification and classification of the landforms inventoried, we have detailed a degree of "uncertainty". Uncertainties, listed in order of occurrence, include: (i) the degree of activity, which affects 12.5 % of the mapped landforms; (ii) the external boundary of complex landforms (10.8 %), that is, due to complex surface expressions (i.e., flow lines) it is difficult to discern whether a rock glacier body belongs to a single lobe or to different sequential pulses; (iii) uncertain boundary between the rooting zone and the rock glacier area (5.1 %), and (iv) landform type (2.8 %).

2.4.2 DATA ANALYSIS

In order to pursue the objectives detailed in section 1 we first analyze the whole database by evaluating: (i) how landform typology and degree of activity influence landform size, its local elevation, surface roughness (Fig. 2.6), and slope aspect (Fig. 2.7); and (ii) interactions between slope aspect and terminus elevation, rock glacier size, and total rock glacier area (Fig. 2.8) (section 2.4.1). Boxplots report median values (as opposed to means) since relevant variables are not normally distributed (e.g., Fig. 2.6a). Subsequently, we examine the within-regional variability of periglacial activity in light of possible interactions between elevation, precipitation, and dominant lithology (section 2.4.2). We use specific landform area (i.e., hectares of rock glaciers or protalus ramparts per unit (squared kilometer) terrain area; ha km^{-2}) as a proxy for present and past periglacial activity (or propensity to rock glacier formation) (Figs. 2.10 and 2.12). In particular, we calculate specific landform area by considering terrain area above 1460 m a.s.l. only, which is the minimum terminus elevation inventoried in the region. Specific landform area is finally evaluated across a regular grid (19 squares, each 27.5 km wide) overlaid on the study area (Figs. 2.1 and 2.12), and across litho-tectonic sectors (Figs. 2.1c and 2.13, and Table 2.3). Grid texture and positioning have been selected in a way that maximizes separation among litho-tectonic sectors and resolves the main sub-regional patterns of mean annual precipitation.

2.5 RESULTS

2.5.1 REGION-WIDE ANALYSIS

We inventory a total of 1742 landforms, including 1514 rock glaciers and 228 protalus ramparts, which cover an area of about 82 km² (Fig. 2.5). Numerically, talus rock glaciers dominate (1378; 91%) over debris rock glaciers (136) (Table 2.2). In particular, the most common rock glacier typology is the talus lobate (948; 63%), followed by the talus-tongue shaped (430; 28%), the debris-tongue shaped (91; 6%), and the debris lobate (45; 3%) (Table 2.2). These percentages agree with figures recorded from studies conducted in the neighboring Ticino Alps (Scapozza and Fontana, 2009; Scapozza and Mari, 2010) as well as with those reported from an investigation combining two study areas in southern and northern Norway (Lilleøren and Etzelmüller, 2011). Rock glaciers of the debris type are markedly larger than the others and occupy a significant portion of the alpine landscape. Specifically, the average rock glacier of the debris type is about one-third larger (7.6 ha) than a talus-derived one (4.9 ha), which in turn is about twice the size of the average protalus rampart (2.0 ha) (Fig. 2.6a). Cumulatively, talus and debris rock glaciers cover respectively an area of 67.1 and 10.3 km^2.

Table 2.2 Landform counts by type and activity

Landform type	Intact	Relict	Total
Talus - lobate	355	593	948
Talus - tongue shaped	171	259	430
Debris - lobate	34	11	45
Debris - tongue shaped	79	12	91
Total rock glaciers	639	875	1514
Protalus Ramparts	32	196	228
Total landforms	671	1071	1742

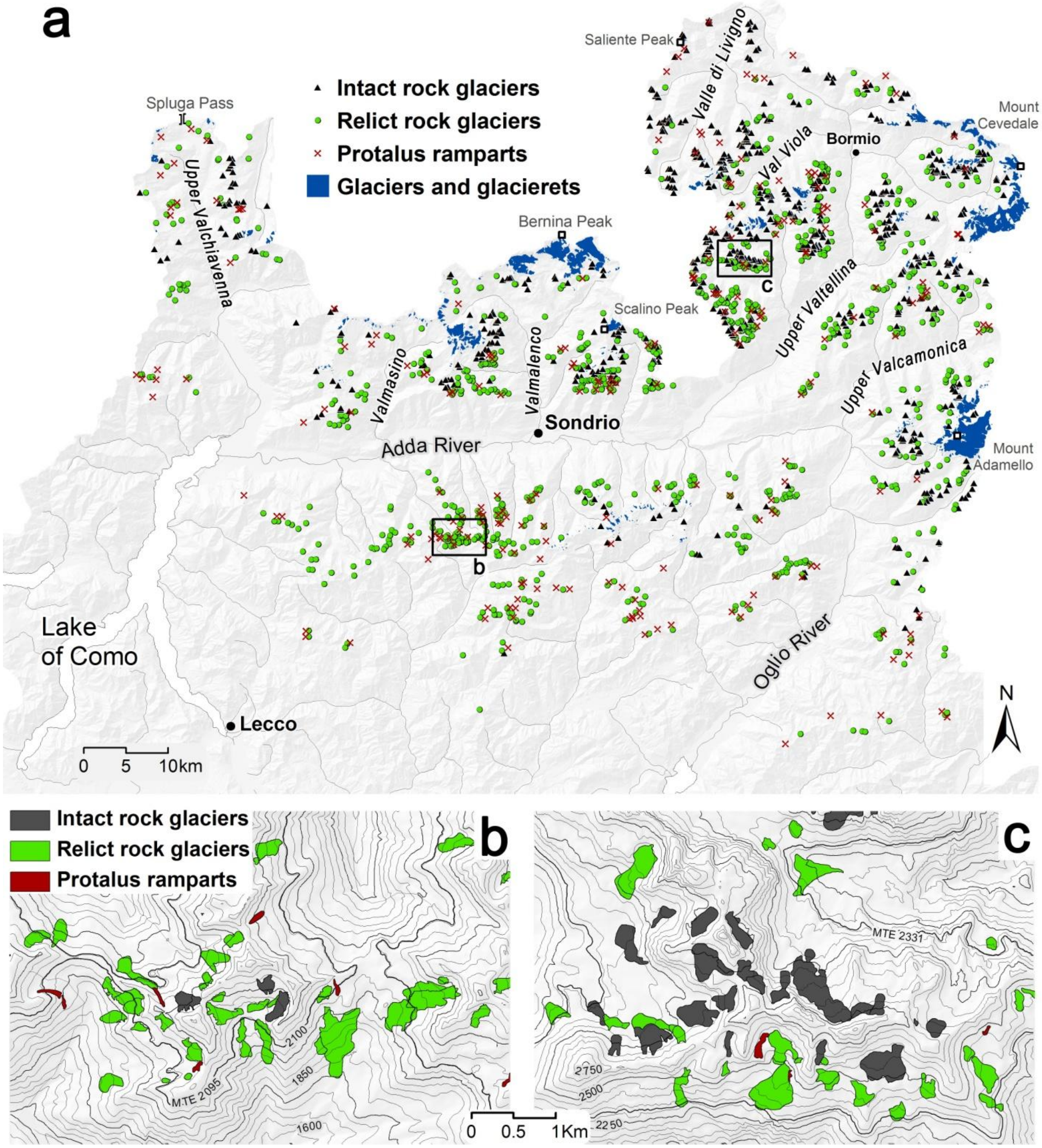

Fig. 2.5. Distribution of intact and relict rock glaciers (RG),protalus ramparts (PR), and glaciers and glacierets (i.e., extension in 2007) within the Lombardy region (a). For visual reference we report MTE (mean terminus elevation) of relict rock glaciers in the Northern Orobic Alps (b) and the Scalino-Cevedale sector (c).

In terms of activity, the inventory encompasses 639 (42%) intact and 875 (58%) relict rock glaciers (Table 2.2). In particular, debris rock glaciers are dominantly intact (lobate: 76%; tongue: 87%) whereas talus-derived ones are mainly relict (lobate: 64%; tongue: 60%) (Table 2.2). While similar intact/relict ratios have been reported by prior studies conducted in other sectors of the Alps (Baroni et al., 2004; Scapozza and Mari, 2010; Morra di Cella et al., 2011), we were not able to identify in the literature any work detailing information stratified by morphology, genesis, and activity. As a result, we cannot make any comparative statement of our findings.

Despite the general limitations associated with any remotely-based classification of rock glacier activity, boxplots of slope variability (i.e., slope range) across rock glacier typologies seem to provide an indirect quantitative confirmation of the reliability of our inventory. Specifically, slope range for intact landforms is consistently higher than for relict ones (Fig. 2.6b). The pattern is particularly evident for debris rock glaciers and less pronounced for the talus typologies. This outcome agrees with the notion that active rock glaciers exhibit a rougher surface expression (i.e., high variability of local slope) due to ongoing downslope creep of a mixture of coarse debris and supersaturated ice (Barsch, 1996). Considering the definition of intact rock glaciers, we can assume that less than half of the landforms inventoried could still incorporate ice.

The majority of the landforms identified is located within the 2000-2500 m a.s.l. (53%) and the 2500-3000 m (39%) elevation belts. In this context, relict landforms cluster chiefly between 2000 and 2600 m, intact landforms are distributed between 2400 m and 2900 m (Fig. 2.6c). The regional median elevation of rock glacier termini, typically considered a valuable morphological proxy of discontinuous permafrost lower limit (e.g., Barsch, 1978), sets respectively at 2590 and 2205 m for intact and relict types. This relict/intact altitudinal mismatch, which emerges distinctively across landform typology (Fig. 2.6c), ranges between 200 and 300 m in the case of debris rock glaciers (debris-lobate: 215 m; debris-tongue: 295 m), and increases for talus rock glaciers (talus-lobate: 400 m; talus-tongue: 360 m). While such elevations for intact and relict rock glaciers match the regional average for the entire Italian Alps (Guglielmin and Smiraglia, 1997), in the the Adamello-Presanella Massif (Baroni et al., 2004) the average relict elevation sets at about 2050 m so that the altitudinal discrepancy between relict and intact landforms approaches 470 m.

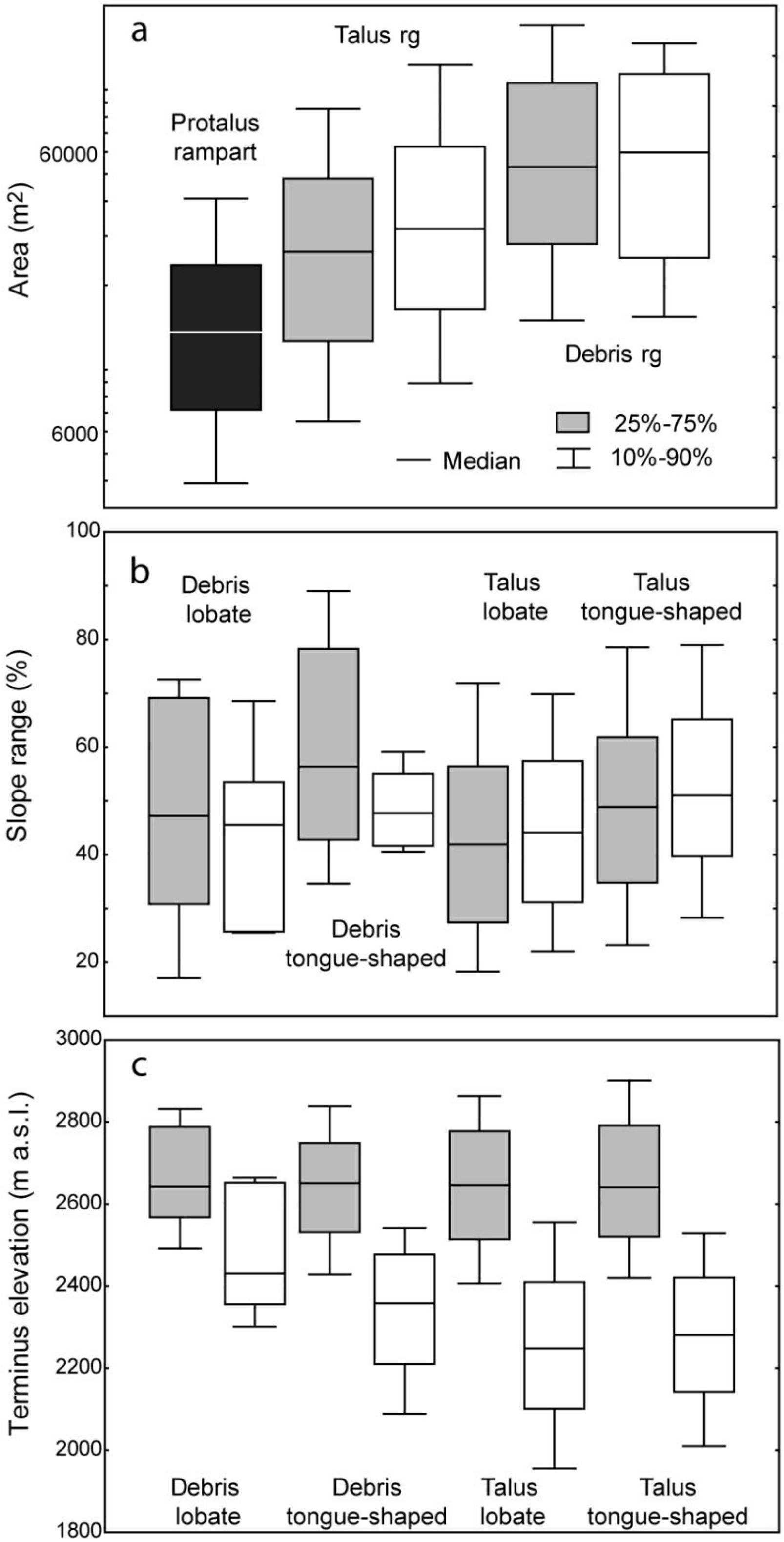

Fig. 2.6.
Boxplots showing the distribution of: (a) rock glacier and protalus ramparts size; (b) rock glacier slope range; and (c) rock glacier terminus elevation. Landforms are stratified by typology and degree of activity. Black box = protalus ramparts; grey boxes = intact rock glaciers; white boxes = relict rock glaciers.

Taken as a whole, the regional inventory exhibits a dominant northwesterly aspect (>25%), followed by northern (16%) and southeastern (16%) ones (Fig. 2.7a). The foregoing aspect configuration differs from prior alpine studies in which rock glaciers resulted more densely concentrated within the northern quadrant, with each remaining component representing less than 10% of the landforms inventoried (e.g., Barsch, 1996; Guglielmin and Smiraglia, 1997; Baroni et al., 2004; Scapozza and Mari, 2010). Stratification by degree of activity clarifies that all relict rock glaciers are moderately evenly distributed across aspect categories. In comparison, for intact counterparts we see an increase of northwest-facing landforms at the expense of southeastern ones (Fig. 2.7a). Further stratification into debris and talus rock glaciers allows to identify distinctive patterns. The former type exhibits high percentages in the entire northerly-facing quadrant (i.e., >20%), with a further increase in northern aspects for intact rock glaciers (Fig. 2.7b). In comparison intact talus rock glaciers, which used to flow along a more uniformly distributed set of directions (note relict dominance in NW and SE aspects, Fig. 2.7c), today display prevalent northern aspects. Finally, the pattern of protalus ramparts (Fig. 2.7d) appears to mimic that of talus rock glaciers. The representations of slope aspect as a function of terminus elevation (i.e., median value of the distribution) of relict and intact rock glaciers, as well as protalus ramparts, document analogous patterns (Fig. 2.8). Similarly to most reports from the Alps (e.g., Barsch, 1996; Imhof, 1996; Scapozza and Mari, 2010; Seppi et al., 2010), in all landform typologies we observe that north-facing landforms tend to plot at consistently lower elevations than south-facing ones, with an altitudinal mismatch ranging between 350 and 400 m.

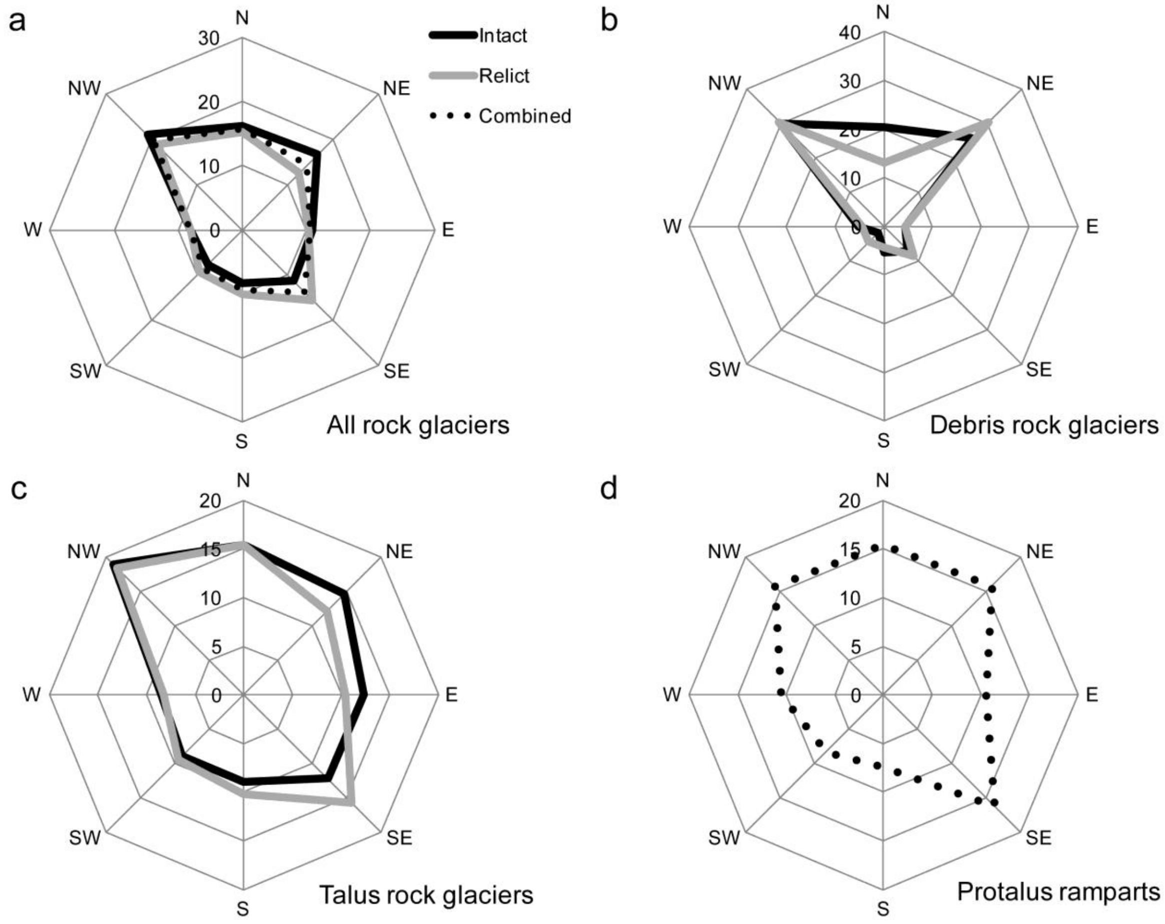

Fig. 2.7. Spider-web charts detailing the relative abundance (percentage) of rock glaciers across slope aspects. Landforms are stratified by typology and degree of activity.

This pattern is an expression of aspect-dependent surface and subsurface thermal conditions (e.g., rock glaciers appear at higher elevations at aspects where potential solar radiation input is highest). Relict rock glaciers and protalus ramparts, whose median values virtually overlap, denote higher sensitivity to aspect change (Fig. 2.8a).

Slope aspect appears to affect also rock glacier size (Fig. 2.8b) and its spatial distribution (Fig. 2.8c). It turns out that relevant aspect correlations with relict and intact rock glacier size are markedly different (Fig. 2.8b). The median size of relict landforms exhibits a monotonic increase from north-facing landforms (minimum), through southerly aspects, to westerly ones (maximum). In contrast, the median size of intact types shows a prominent peak at northwestern aspects. With respect to rock glacier areal distribution, relict landforms display a rather homogeneous pattern across aspects, with two peaks along NW and SE directions; the latter peak shifts towards NE aspects in intact types (Fig. 2.8c).

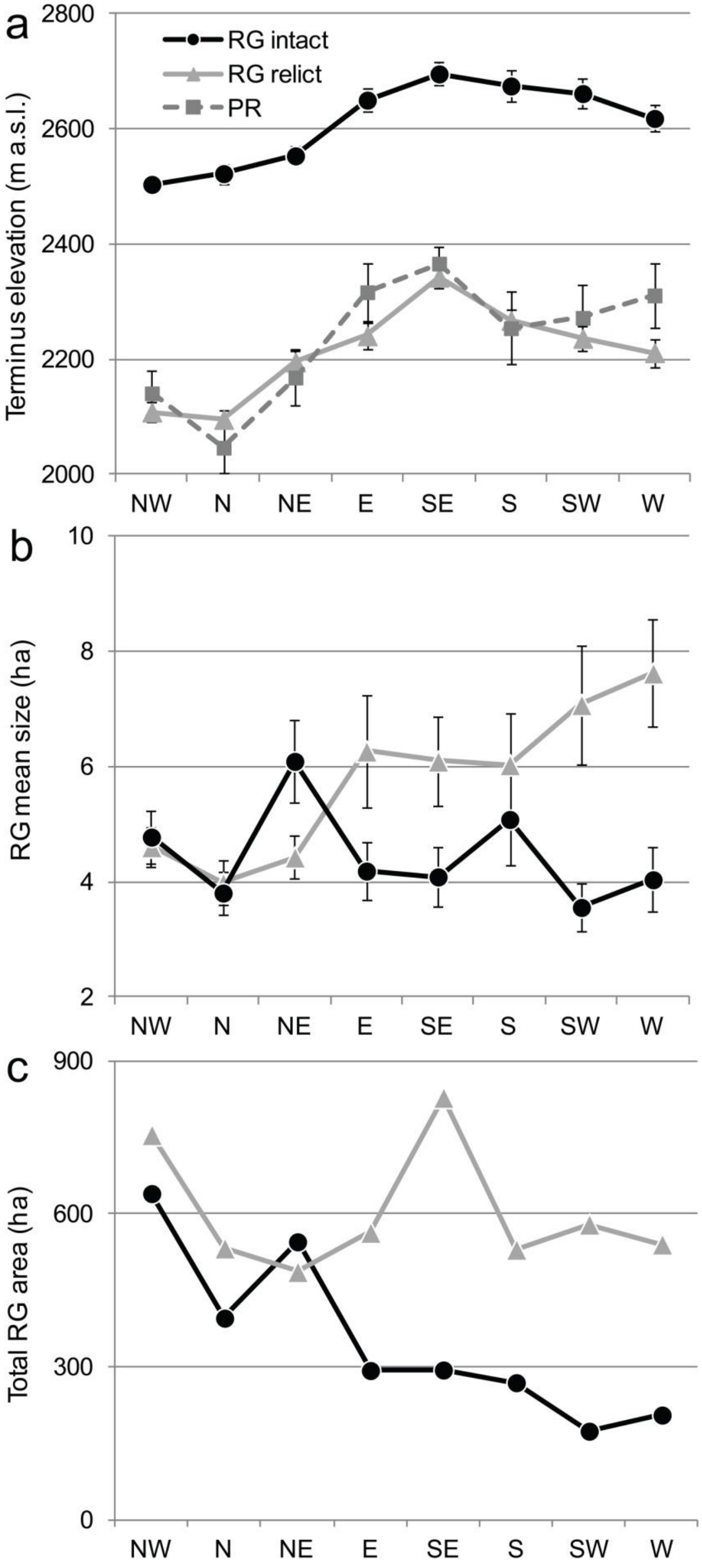

Fig. 2.8. The influence of slope aspect on: (a) mean terminus elevation; (b) mean rock glacier size; and (c) total rock glacier cover. Bars indicate relevant standard errors of the mean.

2.5.2 REGIONAL VARIABILITY

In this section we examine the regional variability of rock-glacier activity. We start by considering how the elevation of rock glacier terminus (Fig. 2.9) and rock glacier specific area (Fig. 2.10) vary along south-to-north and west-to-east transects in relation to the relevant variability in terrain elevation (a proxy for mean annual air temperature (MAAT)) and mean annual precipitation (MAP) (Fig. 2.11). We then examine the pattern of rock glacier and protalus ramparts' specific area within the study region (Fig. 2.12).

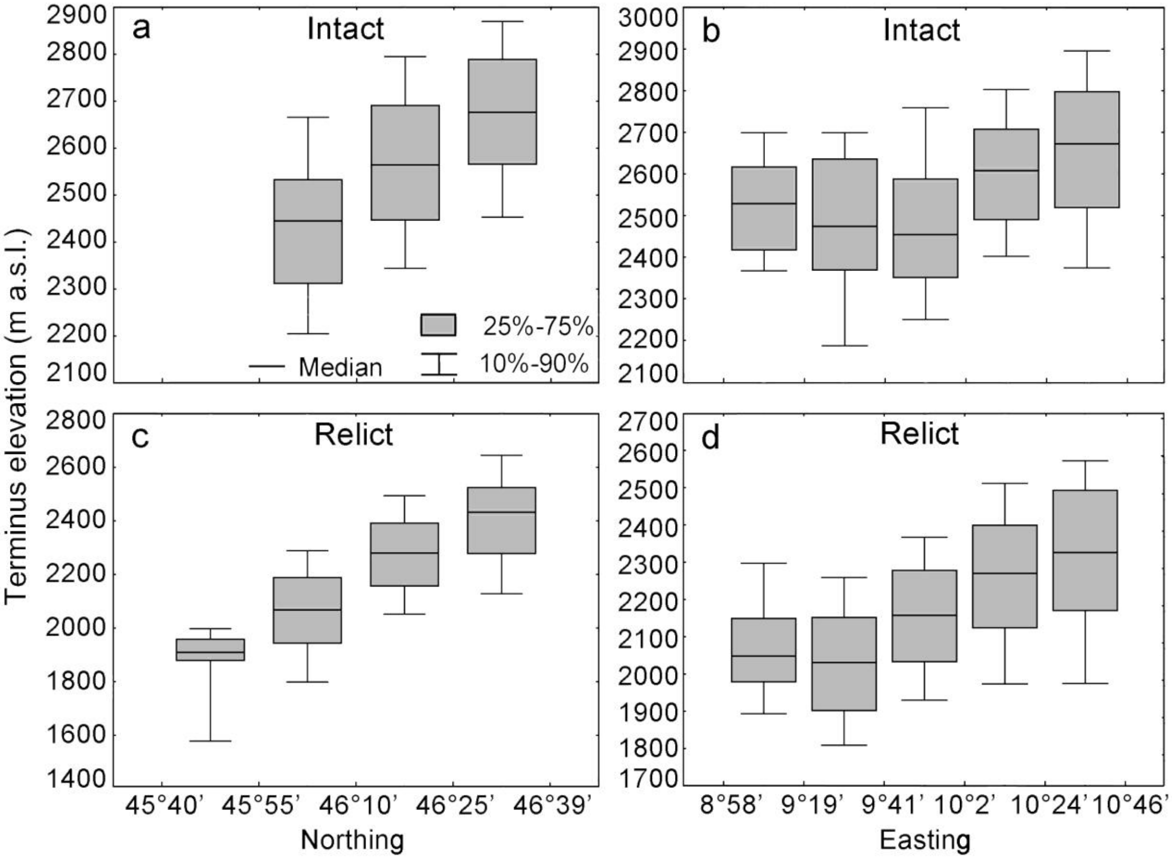

Fig. 2.9. Altitudinal distribution of rock glacier termini along latitudinal (a, c) and longitudinal (b, d) transects drawn across the 27.5-km grid. See text for details.

The elevation of rock glacier terminus increases consistently proceeding from south (median relict: 1920 m; median intact: 2420 m) to north (median relict: 2430 m; median intact: 2680 m; Figs. 2.9a and 2.9c). A more complex pattern is observed in the west-east direction in that we see rock glacier elevation remaining constant between 8°58' E and 9°41' E then increasing progressively towards east (Figs. 2.9b and 2.9d).

In terms of rock glacier specific area (ha km^{-2}), the regional portion covered by relict landforms is consistently greater than that associated with intact ones (note exception in Fig. 2.10a). Similarly to what was just described for terminus elevation, rock glacier specific area increases steadily from south (intact: 0; relict: 0.1 ha km^{-2}) to north (intact: 1.4; relict: 1.9 ha km^{-2}) and from east (intact: 0.1; relict: 0.3 ha km^{-2}) to west (intact: 1; relict: 1.5 ha km^{-2}), hence indicating that the northernmost and easternmost portions of the study region are the most suitable for sustaining permafrost. Specific area increases more according to latitude than longitude change. While this pattern is evident for intact types, relict specific area behaves slightly different, in that we observe a maximum at 46°18' N, as opposed to 46° 33' N (Fig. 2.10a).

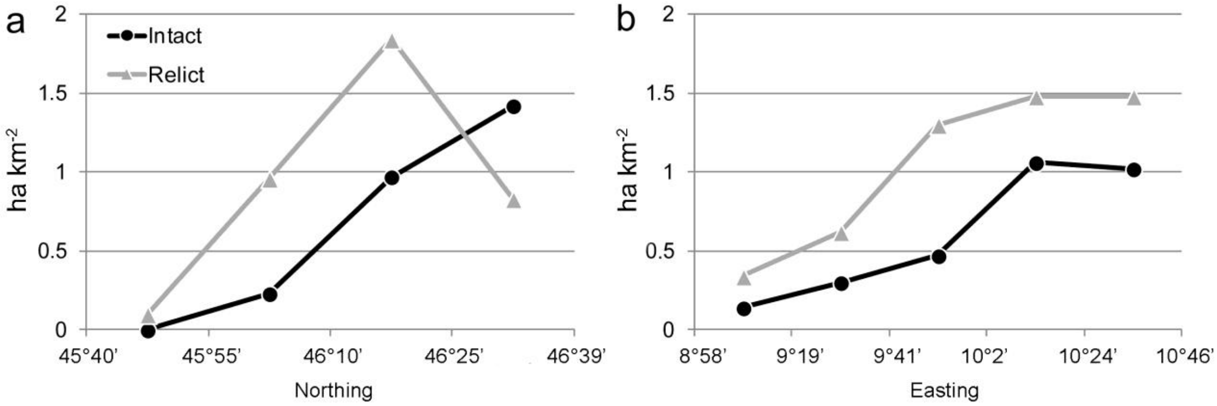

Fig. 2.10. Specific rock glacier area as a function of (a) latitude and (b) longitude.

In this context, inspection of the variability of terrain elevation and mean annual precipitation along south-to-north and west-to-east transects is insightful (Fig. 2.11). Concurrently to the increase of specific rock glacier area towards the northern and eastern parts of Lombardy, we observe a progressive increase in elevation (median S-N: from 1750 m to 2250 m; median W-E: from 2200 m to 2600 m) (hence MAAT) and a decline in precipitation (median S-N: from 1400 mm to 900 mm; median W-E: from 1600 mm to 1200 mm; Fig. 2.11). These outcomes suggest that the combination of higher elevation (cooler) and drier climate promotes the development of broader rock glacier surfaces, hence the persistence of discontinuous permafrost (Barsch; 1978; Haeberli, 1983).

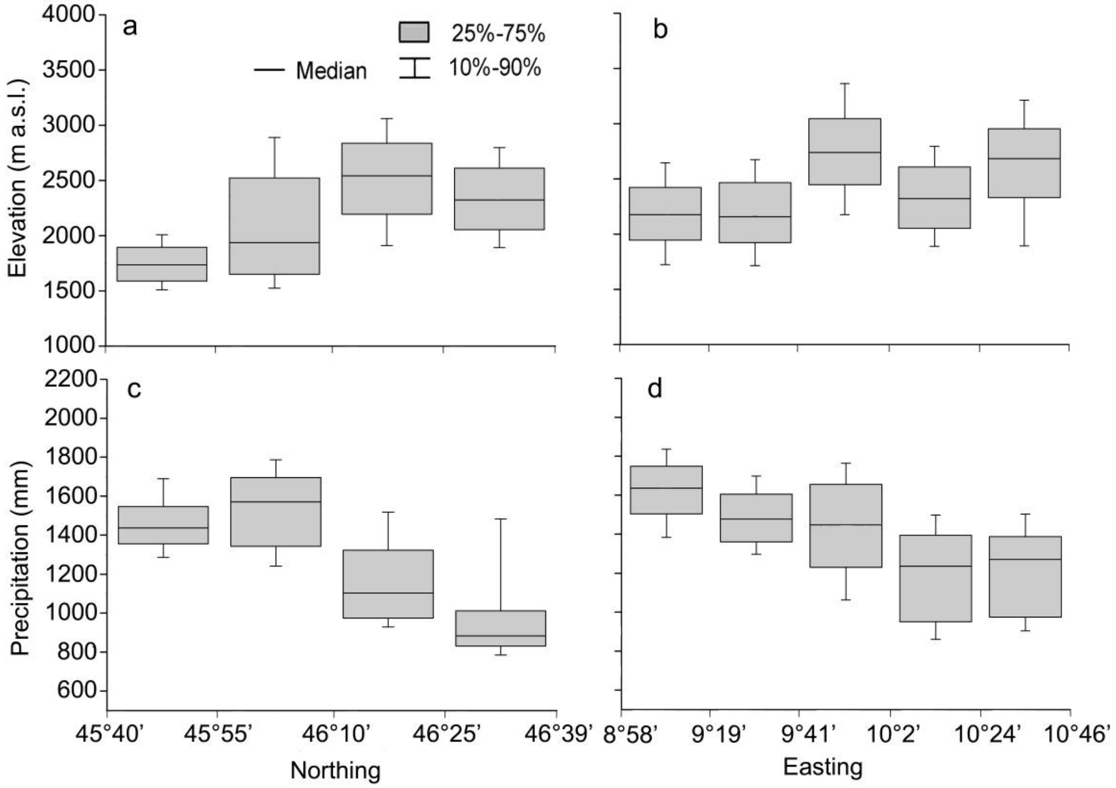

Fig. 2.11. Spatial variability of (a, b) terrain elevation, and (c, d) precipitation, along latitudinal and longitudinal transects drawn across the 27.5-km grid. See text for details.

A clearer picture of the past and present distribution of rock glaciers and protalus ramparts in alpine Lombardy is achieved by subdividing the regional terrain into 19 squares (each 27.5 km wide), hence calculate relevant specific areas (Fig. 2.12; see section 2.3.2 for details). Inspection of relict and intact specific areas clarifies that the former type displays generally higher values (e.g., max: 3.20 ha km^{-2}in Fig. 2.12b) than intact landforms (e.g., max: 1.76 ha km^{-2}in Fig. 2.12a) across the region. In addition, the spatial variability of the two categories differs substantially (cf. Figs. 2.12a and 2.12b). Relict rock glaciers tend to occupy the whole study area, except for its southwestern portion that is characterized by low-relief terrain (Fig. 2.1a). They cluster chiefly in the southern part of the Scalino-Cevedale sector (SCA in Fig. 2.1c), in Valmalenco (MAL) and along the Adamello batholith (ADA) (Fig. 2.12b), all areas associated with high relief and moderate-to-low mean annual precipitation. In comparison, the distribution of intact rock glaciers seems to have migrated northward (Fig. 2.12a). Specifically: (i) we did not identify any intact rock glacier south of 45°55'N (i.e., southern-most row in Fig. 2.12a); and (ii) highest specific areas are typically located north of 46°25'N (i.e., northern-most row in Fig. 2.12a). As we will discuss further in section 5, these findings point to the existence of a periglacial signature that records the spatial distribution of climate change at the Lateglacial-Holocene transition in the region (e.g., Frauenfelder et al., 2001). Finally, the distribution of protalus ramparts (Fig. 2.12c), characterized by typically lower specific areas in comparison with rock glacier figures, behaves in a distinct way. In particular, we observe somewhat high values not only in relatively dry areas like the southern portion of the Scalino-Cevedale (SCA in Fig. 2.1c; 0.22 ha km^{-2}) but also in rather wet terrain, such as the western part of the Spluga sector (SPL; 0.35 ha/km^2) and the central part of the Orobic Alps (OR; 0.15 ha km^{-2}).

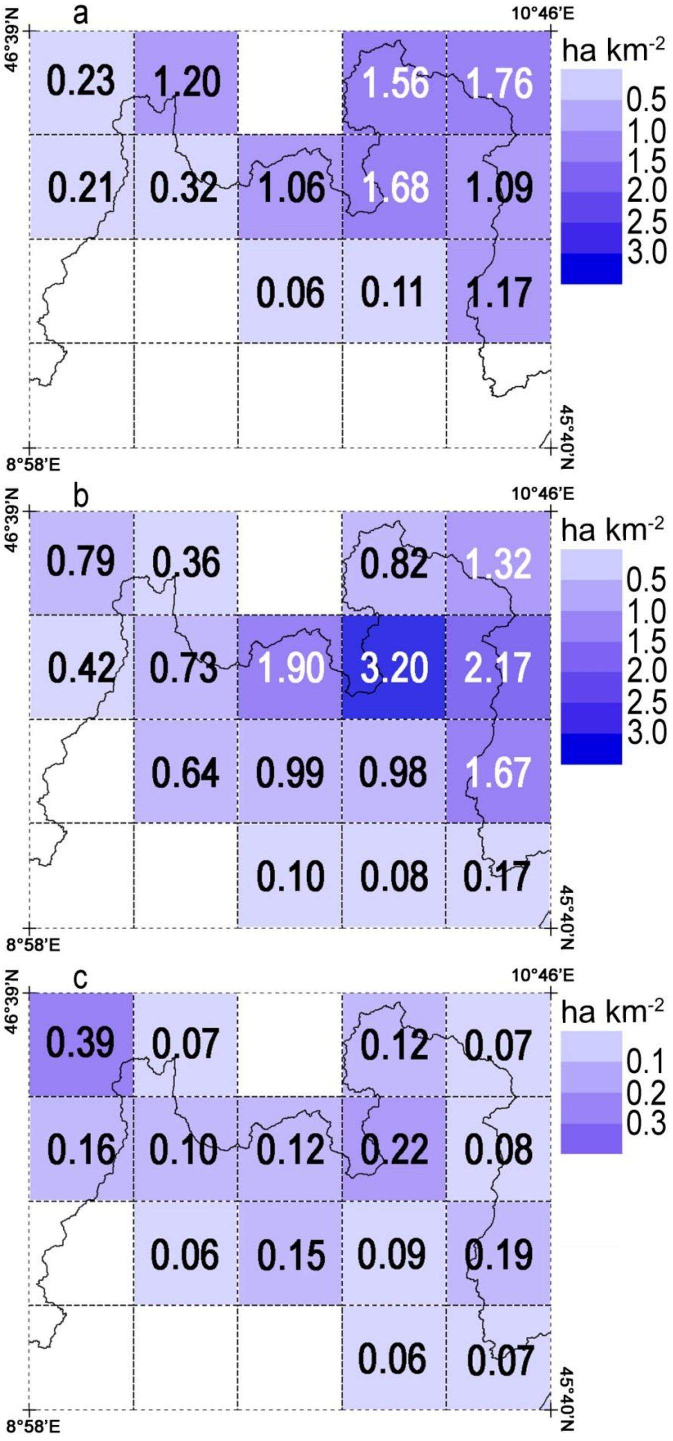

Fig. 2.12. Regional variability of specific landform area across the 27.5-km grid for: (a) intact rock glaciers; (b) relict rock glaciers; and (c) protalus ramparts.

2.6 DISCUSSION

This study characterizes the linkages between rock glacier occurrence, terrain topographic attributes, and precipitation regime in the central Italian Alps (i.e., Lombardy region). In so doing it complements prior smaller-scale studies conducted in other areas of the Alps (e.g., Frauenfelder and Kaab, 2000; Lambiel and Reynard, 2001; Baroni et al., 2004; Nyenhuis et al., 2005, Scapozza and Mari, 2010) and elsewhere (e.g., Luckman and Crockett, 1978; Humlum, 2000; Azocar and Brenning, 2010) in that it tackles the question of rock glacier and protalus rampart distribution at the regional scale. To our knowledge this work represents the largest published rock glacier and protalus rampart inventory that combines qualitative and quantitative attributes for assessing the present and former extent of the periglacial environment at the regional scale.

Unfortunately, when trying to compare our findings with results from other studies it became apparent that the number of inventories published in international journals is extremely limited. In addition, we faced a problem of data heterogeneity, mainly due to the lack of general consensus that until recently has characterized the international discussion on the classification of rock glaciers and protalus ramparts (e.g., Hamilton and Whalley, 1995). As a result, we found that most of the inventories were missing some "key attributes" in relation to the objectives of our work. The forthcoming discussion clearly suffers of this limitation. For example, we could identify only two studies (i.e., Scapozza and Mari, 2010; Lilleøren and Etzelmüller, 2011) that classify rock glaciers according to both degree of activity (intact vs. relict) and genesis (debris vs. talus), and that report all relevant descriptive statistics. This is a major shortcoming as it limits our ability to: (i) link present and former (i.e., intact/relict) climate conditions to the geomorphic activity of glacial and colluvial landforms (i.e., moraines/talus slopes) that have supplied material to rock glaciers; (ii) evaluate possible interactions between morphogenetic and topographic attributes; and (iii) characterize the spatial distribution of rock glaciers across different physiographic regions. In this regard, the PermaNET initiative is particularly timely (Cremonese et al., 2011). Among a series of guidelines for collecting permafrost evidences, Cremonese et al. lay out a strategy for standardizing existing and forthcoming rock glacier datasets that eventually will improve our ability to contrast strategic study areas, hence gain relevant insights.

Rock glaciers within the mountain terrain of Lombardy are dominantly developed on talus deposits, as opposed to glacigenic materials (Table 2.2), suggesting that slow, postglacial mass-wasting processes have been playing a prominent role in supplying sediment to the regional periglacial environment. Inspection of the relict and intact rock glacier populations reveals a peculiar evolutionary trend. If one accepts the assumption that relict and intact rock glaciers belong to distinct generations of landforms (e.g., Lateglacial vs. Holocene), -- even though such assumption rests on a limited set of information (e.g., Van Husen, 1997; Sailer and Kerschner, 1999; Frauenfelder and Kääb, 2000; Dramis et al., 2003 Böhlert et al., 2011) -- it follows that the regional surface occupied by active/inactive rock glaciers has decreased over time by about 60% (from a relict (formerly intact) surface of 48.2 km^2 to a presently intact surface of 29.2 km^2). In particular, talus- and debris-related populations display opposite dynamics. While the talus typology, which used to creep across 46.5 km^2 (96% of the total relict surface), today covers only 20.5 km^2 (70% of the total intact surface); debris rock glaciers record a 5-fold increase in surface: from just 1.7 km^2 (relict) to about 9 km^2 (intact).

We read this latter tendency as the effect of late Holocene (e.g., from LIA onward) generalized climatically-driven trend of glacier retreat and extinction that has promoted the formation/expansion of debris rock glaciers. Following this logic, the narrow spread around northerly aspects of the glacially-related rock glaciers (Fig. 2.7b), as opposed to the talus counterparts (Fig. 2.7c), could be related to age. Accordingly, the formation and persistence of youngest landforms, presumably those debris rock glaciers associated with LIA moraines, are likely to be more heavily controlled by aspect than older landforms that originated under conditions more favorable to permafrost development (e.g., the cooler and drier stages of the Lateglacial). In the latter situation, aspect was undoubtedly less of a limiting factor for permafrost persistence. A similar interpretation has been proposed for explaining the aspect distribution of rock glaciers across Norway (Lilleøren and Etzelmüller, 2011).

The foregoing reasoning alone does not explain the peak in south-easterly exposures of relict talus rock glaciers (Fig. 2.7c) and protalus ramparts (Fig. 2.7d). Inspection of the landscape structure reveals a peculiar organization of the valley network (Fig. 2.5a) that, in turn, dictates the prevalent exposure of rock glaciers and protalus ramparts. In sectors with high rock glacier density (Fig. 2.5), the axes of lateral tributary valleys, which are preferentially oriented along the SW-NE direction, tend to impose valley walls dominated by NW and SE (depending on the valley side) exposures. Examples include most tributary valleys located along Val Chiavenna, Valmasino, Valmalenco, and Upper Valcamonica, as well as the mainstem of Upper Valmalenco, Val Viola, Val di Livigno and Upper Valtellina (see Fig. 2.5). NW and SE maxima both in terms of percent number of landforms (Figs. 2.7c and 2.7d) and total relict rock glacier area (Fig. 2.8c) reinforces our interpretation.

The within-regional variability associated with the landforms inventoried has been evaluated by means of a 27.5-km grid (Fig. 2.1a). This analytical approach has been critical to: (i) derive south-to-north and west-to-east transects of rock glacier altitudinal (Fig. 2.9) and areal (Fig. 2.10) distributions, as well of terrain elevation and precipitation (Fig. 2.11); (ii) disentangle along the same transects the effects of elevation and precipitation change on the abundance of relict and intact rock glaciers; and (iii) obtain snapshots of the former and present rock glacier activity across sub-regions (Fig. 2.12).

The foregoing transects document that: (i) the altitudinal belt of intact rock glaciers (i.e., discontinuous permafrost boundary) lowers with increasingly continental climates (i.e., moving towards north and toward east); and (ii) rock glacier specific area displays respectively a positive and negative correlation with elevation and precipitation, hence indicating that cooler and drier conditions favour rock glacier activity. The former finding agrees with the classic work by Barsch (1978), Haeberli (1983), and King (1986) but contrasts with recent reports from the Chilean Andes (Azocar and Brenning, 2010) and with statistical modelling efforts conducted in the Alps (Boeckli et al., 2012). The latter finding has to be evaluated in relation to dominant lithology, a critical factor for debris supply to rock glaciers (e.g., Haeberli et al., 2006) hence for determining relevant landform size.

The spatial distributions of relict and intact rock glaciers across the regional grid exhibit substantial differences and tell us that discontinuous permafrost today covers a reduced portion of the region (Fig. 2.12a), as opposed to the past (Fig. 2.12b). As a result, intact rock glaciers have virtually disappeared from the Orobic Alps (1600 < MAP < 2000 mm; Fig. 1b) and record a substantial reduction in Valmasino (MAS), Valmalenco (MAL), and Adamello

(ADA), all areas at present characterized by mild-to-wet climates (1200 < MAP < 1400 mm). Keeping in mind the general assumption on the likely age of intact (younger) and relict (older) rock glaciers, this spatial evolutionary trend supports the hypothesis that presently relict landforms would have developed during the Younger Dryas in considerably drier and cooler conditions (Frauenfelder et al., 2001; Lambiel and Reynard, 2001). Climate conditions that could support permafrost persistence at substantially lower elevations than today (i.e., relict landforms display on average an elevation drop of about 400m that roughly corresponds to a difference in mean annual air temperature of about 2 to 3 °C (Frauenfelder et al., 2001), Figs. 2.6c and 2.8a). Notwithstanding the foregoing differences, both relict and intact landforms tend to cluster in the northeastern part of the region (Figs. 2.12a and 2.12b).

Examination of rock glacier specific area across litho-tectonic sectors reveals complex patterns (Table 2.3 and Fig. 2.13). The most striking case appears to be the Malenco valley, which although characterized by MAP and mean terrain elevation comparable with those in Masino and Adamello sectors (Table 2.3), displays substantially greater specific rock glacier area (2 to 4 times higher). We ascribe this discrepancy to the Malenco serpentinites and their tendency to weather (i.e., fracture) faster than quartz-diorites and tonalites. However, despite identical lithology, in Adamello we observe a specific area about 1.5 times larger than in Masino, a behaviour that we cannot explain with available information.

According to prior studies (Wahrhaftig and Cox, 1959; Evin, 1987) and reviews (e.g., Barsch, 1996; Imhof, 1996; Haeberli et al., 2006) rock glaciers would be comparatively rare in weak rocks that disintegrate into fine and platy debris. Our data seems to disagree with the foregoing reports in that the Codera Mountains (granites and granodiorite gneisses) exhibit specific areas about 4.5 times smaller than the Northern Orobic Alps (paragneiss, phyllites, and micaschists) (Table 2.3). Similarly, it is difficult to explain how lithologic effects alone can explain the discrepancies between Saliente-Braulio (dolostones and limestones) and Scalino-Cevedale (paragneiss, phyllites, and micaschists) (Table 2.3).

In order to improve our understanding of the factors controlling rock glacier activity, hence predict their areal distribution more accurately, we think we need: (i) access to higher resolution mapping, both in terms of bedrock geology and Quaternary deposits; (ii) to calculate and evaluate the importance of the source rock wall length and characteristics, (iii) to evaluate the incidence of glacially-derived rock glaciers on the sub-regional specific area values, (iv) to perform more detailed topography-based analysis (e.g., valley-wide hypsometric curves). These tasks, which are beyond the scope of the present paper, will be the subject of future research in selected areas of the Lombardy region.

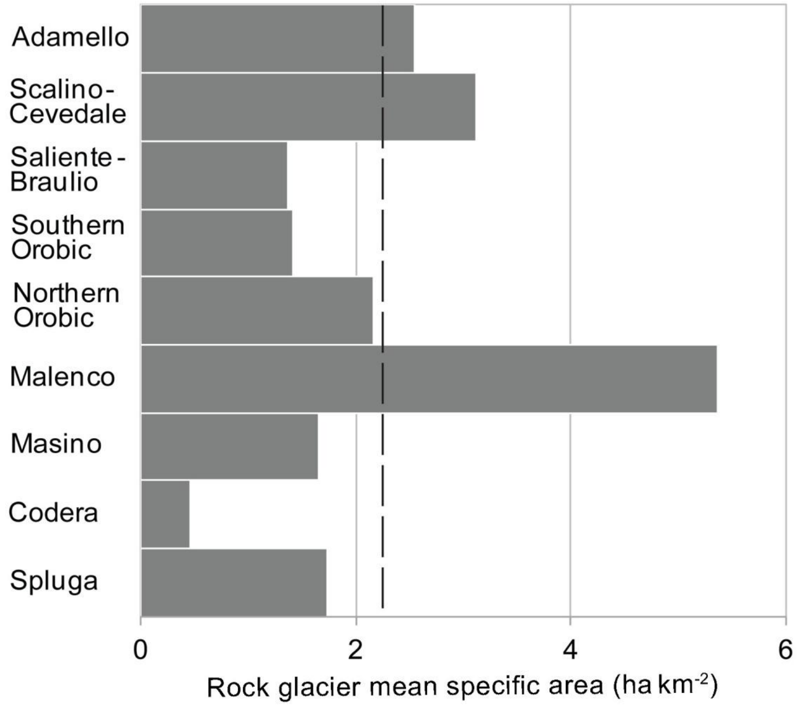

Fig. 2.13. Rock glacier mean specific area in the litho-tectonic sectors. The dashed line indicates the regional average.

Finally, the distribution of protalus ramparts (Fig. 2.12c), characterized by typically lower specific areas in comparison with rock glacier figures, behaves in a distinct way. In particular, we observe somewhat high values not only in relatively dry areas like the southern portion of the Scalino-Cevedale (SCA in Fig. 2.1c; 0.22 ha km^{-2}) but also in rather wet terrain, such as the western part of the Spluga sector (SPL; 0.35 ha/km^{2}) and the central part of the Orobic Alps (OR; 0.15 ha km^{-2}).

Even though their origin and functioning is difficult (if not unreliable) to investigate from largely remotely-based information, the distributions of protalus ramparts in terms of size (i.e., smaller than talus rock glaciers: Fig. 2.6a); aspect (i.e., same as talus rock glaciers: Fig. 2.7d), and elevation (i.e., same as relict talus rock glaciers: Fig. 2.8a) suggest they might be cohorts of relict embryonic talus rock glaciers that never attained complete development (Barsch, 1977; Haeberli, 1985; Scapozza et al., 2011). Their occurrence both in relatively dry and in rather wet terrain (Fig. 2.12c), provided that most protalus ramparts are relict, further supports this view.

Table 2.3. Specific rock glacier area, mean terrain elevation, and mean annual precipitation (MAP) range across litho-tectonic sectors

Litho-tectonic sector		Spluga	Codera	Masino	Malenco	Northern Orobic	Southern Orobic	Saliente-Braulio	Scalino-Cevedale	Adamello
Specific area (ha km^{-2})	Intact	0.98	0.16	0.33	1.28	0.57	0.27	0.73	1.51	0.75
	Relict	0.76	0.29	1.34	4.09	1.59	1.16	0.65	1.61	1.79
	Combined	1.74	0.46	1.66	5.37	2.16	1.42	1.37	3.12	2.54
Mean terrain elevation ± SD (m a.s.l.)		2180 (±358)	1900 (±278)	2240 (±397)	2265 (±404)	1925 (±296)	1835 (±269)	2400 (±382)	2310 (±447)	2275 (±424)
MAP range (mm)		1400-1800	1400-1800	1200-1400	1000-1400	1400-1600	1600-1800	600-1000	600-1000	1200-1400

2.7 CONCLUSIONS

In formerly glaciated mountain environments of Lombardy, central Italian Alps, local terrain topographic attributes (i.e., elevation and aspect), precipitation, lithology, landscape structure, and interactions among them profoundly affect the spatial distribution of rock glaciers and protalus ramparts. In addition, the degree of activity of these landforms (i.e., relict and intact), as modulated by Quaternary climate fluctuations, adds further variability to their positioning.

Our results indicate that between the end of the Last Glacial Maximum and the beginning of the Holocene hillslopes have prevailed over glacial systems in providing chronic sediment supply to rock glaciers. More recently, at least since the Little Ice Age, the number of glacially-derived rock glaciers has increased substantially. We further show that both rock glacier elevation and specific area (a proxy for rock glacier activity) plot respectively as positive functions of terrain elevation (here regarded as a proxy for MAAT) and negative functions of mean annual precipitation, thus indicating that cooler and drier conditions promote discontinuous permafrost formation and persistence.

In agreement with the foregoing correlations and with available palaeoclimatic reconstructions for the European Alps, today intact rock glaciers chiefly cluster around the innermost and dry ranges, as opposed to the outer and wetter mountain sectors (e.g., the Orobic Alps), where they have all virtually turned into relict landforms. The resolution of the geology mapping available for the entire region does not allow to isolate conclusive lithologic dependences. In order to address the question of sediment supply to rock glaciers, future work will seek to enrich the inventory with additional morphometric and lithologic attributes of the relevant source areas for selected litho-tectonic sectors.

2.8 ACKNOWLEDGMENTS

We are grateful to Massimo Ceriani (Regione Lombardia) for granting access to cartographic information and imagery. Richard Shakesby provided useful suggestions on protalus rampart recognition. Alberto Resentini kindly helped with the delineation of the litho-tectonic sectors. The paper has benefited from constructive reviews by Wilfried Haeberli and Brian Whalley.

3 RECENT AREA VARIATIONS OF GLACIERS IN THE CENTRAL ITALIAN ALPS (LOMBARDY REGION) AND THEIR INTERACTIONS WITH CLIMATE CHANGE

The dataset presented and analyzed in this chapter was used as data core of the volume:

I ghiacciai della Lombardia; evoluzione e attualità. (2012)

Servizio Glaciologico Lombardo, HOEPLI ed., Milano. 328 pp.

Edited by: Bonardi, L., Rovelli, E., Scotti, R., Toffaletti, A., Urso, M., Villa, F.

3.3 ABSTRACT

We present evidence of climate change impact on recent variations of glaciers in the Central Italian Alps (Lombardy region). To achieve that we have analyzed the area evolution of a sample of 304 glaciers in 1991, 2003 and 2007. The two most recent inventories (2003 and 2007), based on the manual delineation of glacier limits on orthophoto mosaics, and 2m-DSM, were used to

(i) examine the linkages between glacier location, glacier attributes (e.g., size, aspect, ELA_0) and mean annual precipitation; (ii) analyze the relations between those glaciers attributes and areal variations since 1991. In order to extend back in time the investigation of glacier fluctuations and to investigate glacier sensitivity to post-Little Ice Age (LIA) climate change we have reconstructed the 1997, 1954 and LIA maximum extent for nine selected glaciers (Pizzo Ferrè, Rasica Est, Casandra, Fellaria, Dosdè Ovest, Campo Nord, Zebrù, Pisgana and Lupo). We have analyzed temperature and precipitation anomalies from a long-term climate

station (Sils/Maria – Switzerland) to explain relevant glaciers' fluctuations from 1864 onward.

The analysis on the 2007 inventory reveals a great variability in the ELA_0 in the nine sub-regions considered (ranging from 3013 to 2481 m a.s.l.) and an inverse relation of ELA_0 with mean annual precipitation. Total glaciers' extent was 118.3 km^2 in 1991, 96.5 km^2 in 2003 and 89.4 km^2 in 2007 for a reduction of 28.8 km^2 (– 24.4 %) from 1991. The higher relative retreat was detected in the sub-regions with the smaller mean glacier size, here the area has almost halved from 1991. We have observed that glacier recession increased in speed from – 1.81 km^2 yr^{-1} (1991-2003) to – 2.02 km^2 yr^{-1}(2003-2007) as a consequence of an increase in summer temperature of 0.16 °C and a decrease in precipitation of 138 mm (– 23.9 %) between the two intervals considered.

The analysis of glacier fluctuations since LIA revealed that total relative area change to 2007 ranges from – 37% of the largest (Fellaria) to – 90.3 % of the smallest (Rasica Est). As precipitation during the accumulation season does not show any obvious trend, most of this shrinkage appears to be related to the increase in summer temperature (1864-2012 average of 0.0057 °C yr^{-1}). Furthermore, in comparison with the average retreat rate observed between LIA and 1990, the selected glaciers drastically increased (up to 5.7 times) their retreat rate from 1990 onward. Hence a large portion of the total relative loss since LIA (24-41 %) can be associated with the 1990-2007 interval which accounts for only the 12 % of the total time post-LIA time window.

3.4 INTRODUCTION

In the last few decades mountain glaciers worldwide have been undergoing a general mass (hence area loss) (e.g., Haeberli and Beniston, 1998; Haeberli et al., 1999a; Dyurgerov and Meier, 2000; Lemke et al., 2007; Bonardi, 2008); a behaviour that is considered consequence of the generalized atmospheric temperature rise driven by anthropogenic greenhouse gases (Oerlemans, 2005; IPCC, 2007). In particular, glaciers of the European Alps are experiencing one of the most severe warming trends of the planet (Böhm et al., 2001; Beniston et al., 2003; Rebetez and Reinhard, 2008) and as such, they have been shrinking rapidly since the 1980's (e.g., Kääb et al., 2002; Zemp et al., 2006; Paul et al., 2004; Citterio et al., 2007; Abermann et al., 2009).

The importance of alpine glaciers in the mountain hydrological cycle (e.g., river flow conditions for riparian habitat conservation, river discharge and hydroelectric power production) and as tourist attraction has long been acknowledged (e.g., Bonardi, 2012).

In order to capture the ongoing rapid trend of glacier reduction and predict glacier surface and/or mass dynamics in response to future climatic fluctuations, one has to undertake a frequent temporal update of glacier inventories using a consistent spatial resolution. Recently, this task has been feasible thanks to substantial advances in remote sensing (e.g., availability of multiple sensors, enhanced resolution and frequency of aerial and satellite photographs) (Paul et al., 2007).

In order to document and quantify the historical withdrawal of glaciers across the entire Lombardy Region we have to refer to the first available inventory conducted by Porro (1925) and to subsequent relevant update (Porro and Labus, 1927), which consisted exclusively on field-based surveys. The first regional inventory was published in 1961 (CGI-CNR) and included field campaigns conducted between 1927 and 1957. Subsequently, the glacier database was updated by the local section of the World Glacier Inventory (Serandrei-Barbero and Zanon, 1993) with data surveyed between 1981 and 1984. More recently, the Servizio Glaciologico Lombardo (SGL) published two inventories reporting data surveyed respectively in 1988-1991 (Galluccio and Catasta, 1992), and 1998-2001 (Galluccio, 2002, unpublished report). Finally, two additional inventories, which were mainly conducted through interpretation of aerial photographs, cover the 1997-1999 period (Citterio et al., 2007) and 2003 (Diolaiuti et al., 2012).

In this work we present an updated regional inventory based on the manual delineation of glacier limits on 2003 and 2007 orthophotos, and on a 2007 DSM. This inventory differs from the previous ones for including a larger dataset that considers a series of glacierets and permanent snow-fields mapped and surveyed during 16 years of detailed fieldwork by SGL (1992-2007).

Absolute and relative area variations from 1991 to 2007 are investigated and correlated with some basic glacier attributes including mountain sub-region, glacier size, and dominant aspect. Mean annual area change in the 1991-2003 and 2003-2007 periods are used to better constrain glaciers’ dynamics in terms of area and volume loss/gain. Furthermore, in order to extend back in time the investigation of glaciers’ fluctuations (Zemp et al., 2011), we have reconstructed the Little Ice Age maximum (LIA) extent for nine selected glaciers of different size classes and from different sub-regions (Fig. 3.1a). In order to better identify and highlight the strong trend of ice loss shrinkage occurred in the last two decades (i.e., from 1991), for these glaciers we have also conducted the 1954 and 1997 areal reconstructions derived from aerial photos. Temperature and precipitation anomalies from a long-term climate station (Sils/Maria – Graubunden – Switzerland ; Fig. 1b) were analyzed to explain relevant glacier fluctuations from 1864 onward.

3.5 STUDY AREA

The Central Italian Alps within the Lombardy Region (northern Italy), hosts 333 glaciers, glacierets, and perennial snowfields (Fig. 3.1a), including the two largest Italian glaciers (Adamello, 16.6 km², and Forni, 11.4 km²). The study area can be classified into nine sub-regions, including, from west to east: Spluga-Lei, Codera-Masino, Disgrazia-Mallero, Bernina-Scalino, Orobie, Livigno, Dosdè-Piazzi, Adamello and Ortles-Cevedale (Fig. 3.1b). The higher peak of the region, Punta Perrucchetti 4020 m a.s.l. (Bernina), is the only one above 4000 m, with about 270 peaks exceeding 3000 m a.s.l.

The climate of this sector of the Alps above 2000 m a.s.l. can be classified as Tundra Climate (ET) according to the Köppen-Geiger scheme (e.g., Peel et al., 2007; Toffaletti, 2012). Precipitation (rainfall and snowfall) exhibits high spatial variability in terms of total annual values (Fig. 3.1b) and seasonal distribution. Extremes are found in the southwestern sub-regions (Orobie and Spluga-Lei) where two seasonal maxima in late spring and autumn contribute to reach a mean annual precipitation of 2150 mm. In the northeastern part of the region (Ortles-Cevedale and northern Livigno sub-region) annual precipitation locally drops below 700 mm, and reveals a single summer maximum (Ceriani and Carelli, 2000). The foregoing high spatial variability in total annual precipitation is confirmed by field data of two glacier winter mass balances. Specifically, the Lupo glacier (Orobie Alps) despite its 500-m lower elevation, shows more than twice (448 cm) the snow accumulation observed at the Campo Nord glacier (186 cm) (Livigno Alps) (Toffaletti, 2012) (Fig. 3.1a).

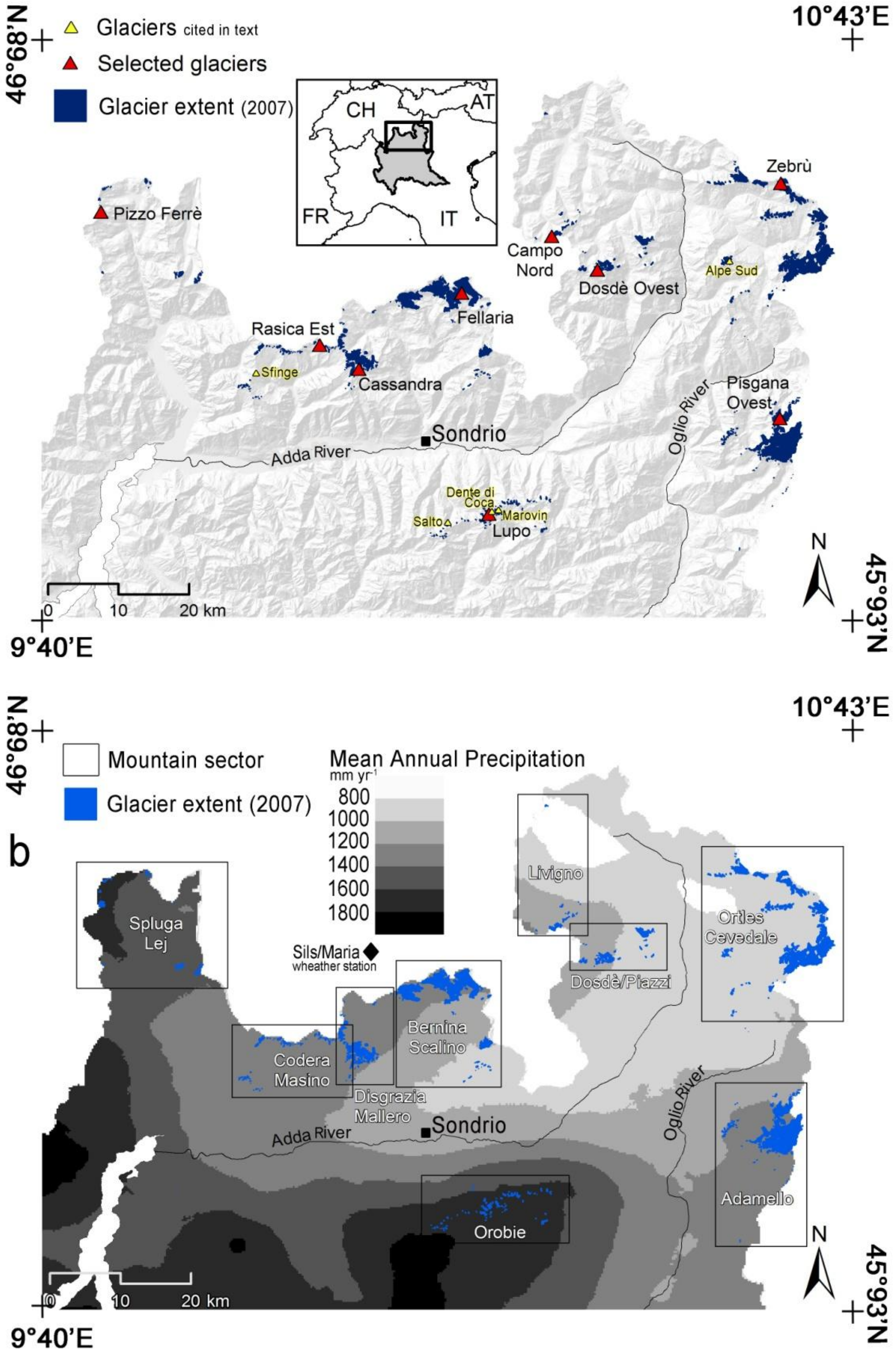

Fig. 3.1. Maps showing the distribution of glaciers in 2007. Selected glaciers (a) and the nine mountain sub-regions (b) (see text for further explanations) are localized. Mean annual precipitation (b) is interpolated by using ordinary co-kriging with 374 rainfall stations (1881-1990) and 50,000 elevation points randomly distributed within the Region.

3.6 DATA AND METHODS

3.6.1 CONTEMPORARY GALCIER MAPPING

The new inventory of glaciers, glacierets and perennial snow fields was compiled via manual delineation the glacier limits on a high-resolution (0.5-m pixel) orthophotos flown in 2007, and a 2-m gridded Digital Surface Model (DSM, 2007). In order to minimize problems related to the delimitation of debris-covered glaciers, we conducted complementary field surveys. In addition, GPS surveys, and the general supervision by SGL experienced glaciologists, provided critical ground control for data extracted from remotely-based inspection. Despite the existence of a similar inventory (i.e., Diolaiuti et al., 2012), to minimize the degree of subjectivity due to multiple interpreters, we decided to map independently all glaciers at the regional scale on a 2003 flight (1-m orthophoto mosaic). This photo mosaic is characterized by minimal snow cover over the glaciers and surrounding areas due to the extremely high temperatures recorded throughout that summer (i.e., Fink et al., 2004; García-Herrera et al., 2010). Thanks to the dry and hot accumulation season, snow cover is very limited in the 2007 images too (Scotti et al., 2008). 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.

The reconstruction of recent glacier variation involves the difficult tasks of data fusion and subsequent comparative analysis of various inventories. Different classifications and different time intervals used to assemble the data forced us to consider only the SGL 1992 inventory as a reliable benchmark for the 2003 and 2007 datasets.

Glacier attributes include, identification number, geographic coordinates, sub-region, river basin, surface area, elevation (terminus, max, and mean), theoretical Equilibrium Line Altitude (ELA_0), maximum elevation of accumulation basin, mean slope gradient, slope aspect (manually defined along the direction of the main flow axis, or for snow fields, the general aspect of the mountain slope) and mean annual precipitation on the glacier (Fig. 3.1b). The glacier primary classification and the definition of the main source of nourishment follows the Illustrated GLIMS Glacier Classification Manual (Rau et al., 2005). The main topographic attributes (i.e., elevation, theoretical $ELA_{0,}$ slope gradient, and slope aspect) have been extracted from the 2-m gridded DSM. Mean annual precipitation for each glacier is derived from a 250-m gridded precipitation map (Fig. 3.1b). The theoretical equilibrium line altitude (ELA_0), or balance budget ELA (Cogley et al., 2011), is defined here 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).

When quantifying the mean annual area change, one has to consider that the 2007 photo mosaic was flown in July, and as such it include a portion of the contingent ablation season. For this reason we have shortened the 1991-2007 and the 2003-2007 intervals respectively to 15.5 and to 3.5 years.

The delineation of different homogeneous sub-regions within the study area helped us to analyze the spatial distribution of some key attributes such as mean annual precipitation and the ELA_0 at the regional scale.

3.6.2 LITTLE ICE AGE RECONSTRUCTION

To expand the study interval, hence to better constrain the recent trend of glacier retreat, we reconstructed the glacier extent during the last maximum advance associated with the Little Ice Age (LIA) for nine selected glaciers of the region (Fig. 3.1a). This sample covers every sub-region, various slope aspects, different size classes from 0.03 to 9.32 km², and different elevations (ELA_0 from 2545 to 3118 m a.s.l.) (Table 3.3). The shape and position of LIA moraines in central Italian Alps and surrounding regions resembles that of the rest of the Alps where examples of regional LIA glacier inventories exist (e.g., Gross 1983, 1987; Maisch, 1992; Maisch et al., 2000). Moraine age determination have been carried out with different approaches including dendrochronology and geomorphology (e.g., Pelfini, 1999; Pelfini et al., 2009), lichenometry (e.g., Orombelli, 1987) and geopedology (e.g., Caccianiga et al., 1994). The foregoing studies improved the confidence of our reconstruction and helped setting to 1860 (Pelfini and Smiraglia, 1992) the date of LIA maximum glacial advance. This constitutes our benchmark against which we have computed historical area fluctuations. The detection of the LIA maximum extent over the nine selected glaciers was carried out merging different approaches including: (i) field mapping of moraines and trim-lines; (ii) remotely-based interpretation of aerial orthophotos and DSM hillshades; and (iii) integration of historical documentation (cartography, iconography, and literature). For the nine selected glaciers, in addition to the LIA, 1991, 2003 and 2007 areal mapping, data from 1954 and 1997 have been used to improve the temporal resolution of glacier fluctuations. The 1954 and 1997 glacier limits have been drawn on the basis of black and white orthophotos at a nominal scale of about 1:20,000 and 1:10,000 respectively.

3.6.3 SILS/MARIA CLIMATIC DATA

In our investigation, the Sils/Maria (46° 26.3' N / 9° 45.9' E) climatic station constitutes the main reference for climatic variability over the study period. This site is one of the 12 climatic stations of the MeteoSwiss network (Fig. 3.1b) with homogenized data (Berget et al., 2005). Considering its relative high elevation (1798 m a.s.l.) and the long and continuous historical record (1864-2012), it can be considered one of the most important stations in the entire Central Alps. Monthly precipitation and temperature were acquired and Standardized Anomaly Index (SAI) was used for comparing precipitation (October-May) and temperature (June-September) anomalies. The Standardized Anomaly Index is computed as:

$$SAI = (Xa - Xm) / Std\ Dev$$

where Xa is the mean (or total value) of the variable considered, Xm and Std Dev, represent respectively the mean and the standard deviation across the study period (i.e., Katz and Glantz, 1986). Even though the station is not located on the southern side of the Alps, the Sils/Maria historical series mimics closely that recorded at S. Antonio Valfurva (46° 27.5' N / 10° 25.7' E) between 1988 and 2006, a station located in the Ortles-Cevedale sub-region (Scotti, 2009). In particular, we note high SAI coefficients of determination for precipitation ($R^2 = 0.95$) and temperature ($R^2 = 0.73$). Precipitation at Sils/Maria follows the typical *intra-alpine* pattern, with a single monthly maximum in August (119 mm) and a single minimum in February (42 mm). The mean annual precipitation is 982 mm. The thermal regime shows the average minimum temperature in January (– 7.5 °C) and the maximum one in July (+ 10.6 °C), with a mean annual temperature of + 1.6 °C.

3.7 RESULTS

3.7.1 GLACIERS DISTRIBUTION

The number of glaciers, glacierets and snowfields in the study region, has been extremely variable and appears to be unrelated to the total extent (90.4 km² in 2007). In fact, despite a continuous decrease in area, from 1991 to 2003 the inventoried glaciers increased in number from 304 to 365 due to 87 new discoveries (i.e., previously unmapped), 29 extinctions, 4 fragmentations and 1 glacier previously considered split in 2 distinct bodies. Then, 33 extinctions and 1 more fragmentation from 2003 to 2007 decreased the total number to 333 (Table 3.1). Glacierets and perennial snowfields are most frequent (192) while the 141 glaciers are mostly of the mountain type (93 %), of the remaining 8 are valley glaciers and 2 ice fields (Fellaria/Palü and Adamello). By adding the Forni valley glacier to the two ice fields, we reach 37.2 km² or the 41 % of the glaciated area. On the other hand a great number of glaciers (257) are smaller than 0.1 km², representing 83 % in number and only 6.4 % of the regional surface. The area distribution in the different mountain sub-regions is heavily affected by the presence of the 3 largest glaciers. In fact, the 3 sub-regions hosting them (Bernina/Scalino, Adamello and Ortles/Cevedale) represent more than 80 % of the regional glaciarized area. The high relief energy typical of the "young" mountain chains as the Alps, leads the dominance of *avalanche fed* glaciers (i.e., > 90%). Nerveless it is worth noting that the two ice fields (Adamello and Fellaria) are mainly *snow* and *drift snow fed*.

Table 3.1. Glacier count, area, area change and mean annual area change in different mountain sub-regions with climatic data from Sils/Maria.

Sub-region	Glaciers, glacierets and snowfields (extinct/discovered/fragmented/merged)			Area (km^2)			Area change (km^2)			Area change (%)		Mean annual area change (km^2 yr^{-1})	
				Year 1991 reference sample (304 glaciers)									
	1991	**2003**	**2007**	**1991**	**2003**	**2007**	1991-2003	2003-2007	1991-2007	1991-2003	1991-2007	1991-2003	2003-2007
Spluga/Lei	17	17 (2/2)	17	3.5	2.2	1.9	-1.3	-0.3	-1.5	-36.9	**-44.6**	-0.11	-0.08
Codera/Masino	36	52 (2/17/1)	51 (1)	2.9	1.9	1.5	-1.1	-0.4	-1.4	-35.9	**-48.9**	-0.09	-0.11
Disgrazia/Mallero	32	34 (2/4)	32 (2)	11.1	8.1	7.1	-3.0	-1.0	-4.0	-26.9	**-36.0**	-0.25	-0.29
Bernina/Scalino	27	28 (3/3/1)	26 (2)	26.3	22.7	21.2	-3.5	-1.6	-5.1	-13.4	**-19.5**	-0.29	-0.45
Dosdè/Piazzi	25	28 (0/4/1/1)	25 (4/0/1)	5.6	4.0	3.4	-1.6	-0.5	-2.1	-28.8	**-38.4**	-0.13	-0.15
Livigno	20	22 (2/4)	16 (6)	2.1	1.3	1.1	-0.8	-0.2	-1.0	-38.7	**-48.6**	-0.07	-0.06
Ortles/Cevedale	57	65 (5/12/1)	62 (3)	38.6	31.6	29.7	-7.0	-1.9	-8.9	-18.2	**-23.1**	-0.59	-0.54
Adamello	46	57 (7/18)	48 (9)	25.8	23.1	22.1	-2.7	-1.0	-3.7	-10.5	**-14.5**	-0.23	-0.30
Orobie	44	62 (5/23)	56 (6)	2.4	1.7	1.5	-0.8	-0.2	-1.0	-31.9	**-39.2**	-0.06	-0.05
Lombardy	304	365 (29/87/4/1)	333 (33/0/1)	**118.3**	**96.5**	**89.4**	**-21.8**	**-7.1**	**-28.8**	**-18.4**	**-24.4**	**-1.81**	**-2.02**
Temperature (J-S) (°C)												**9.99**	**10.15**
Precipitation (O-M) (mm)												**576.5**	**438.5**

The regional ELA_0 is set at 2810 m a.s.l. with great differences in the 9 mountain sub-regions (Figs. 3.2, 3.3b). This value rises to 2895 m a.s.l. considering only the glaciers larger than 0.1 km². As the ELA_0 is a climatic related attribute, albeit dependent by other morphological factors, an inverse relationship with mean annual precipitation is found. The two extreme sub-regions are Ortles/Cevedale and Orobie: ELA_0 lowers from 3013 to 2481 m a.s.l. with a precipitation increase from 923 to 1674 mm, respectively (Fig. 3.2). The other sub-regions show relatively smaller differences following the recognized inverse relationship. Orobie and Codera/Masino exhibit the higher variance in ELA_0 elevation while the smaller is typical for the less represented sub-regions: Livigno and Spluga/Lei (Fig. 3.3a).

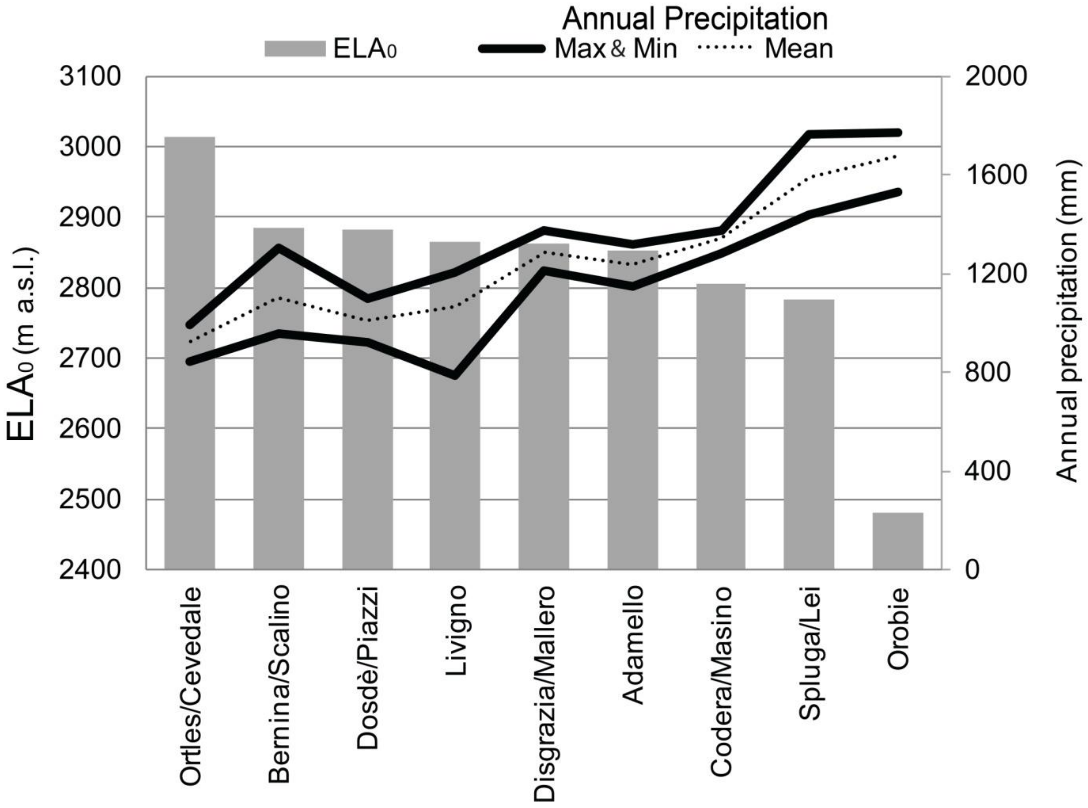

Fig. 3.2. Chart showing the mean elevation of the ELA_0 and the annual precipitation in the different mountain sub-regions.

Lombardy glaciers are mostly exposed to the north facing quadrant (67 %) with NW (25 %) as the most represented aspect, followed by N (23 %) and NE (19 %). Surprisingly the 4th most represented aspect is SW (10 %) followed by W (7 %) and SE (6 %), the remaining 10% being equally distributed between E and S (Fig. 3.3b). This general behavior can be easily explained with the evident advantages of the north facing slopes in terms of protection from incoming solar radiation. The elevation distribution of glaciers, related to their aspect, is of great interest as the results clearly show a lowering in the mean ELA_0 for the north facing glaciers compared to the south facing ones. Furthermore, the ELA_0 variance is much greater for the northern quadrant and it gradually decrease rotating to the southern quadrant (Fig. 3.3b). This behavior clearly shows the climatic limitation of the southern slopes to host glaciers at mid-low elevations due to the potentially higher incoming solar radiation. In fact, the lower ELA_0 of a south facing glacier is 2629 m a.s.l. (Sfinge glacieret, Codera/Masino, SE aspect), on the other side (NE aspect) the Salto glacier (Orobie) presents an ELA_0 at 2046 m a.s.l. The mean terminus elevation of Lombardy glaciers (whole sample) is 2730 m a.s.l., and if considering only glaciers ≥ 0.1 km², the elevation decreases to 2690 m a.s.l. The lower terminus of a south facing glacier is 2546 m a.s.l. for the Fellaria glacier, while Marovin glacier (N aspect) still flows down to 2030 m a.s.l., the lowest terminus of the region.

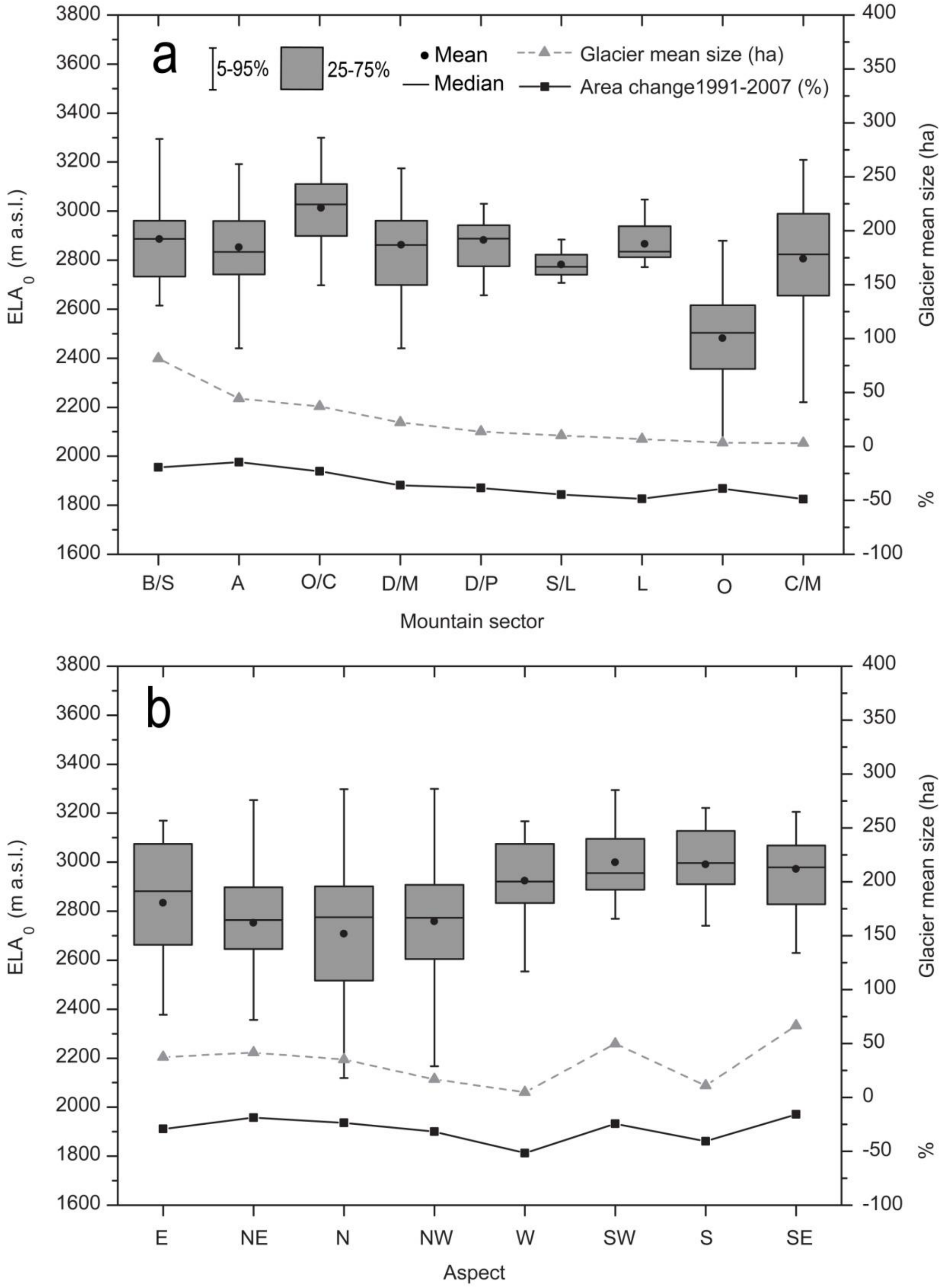

Fig. 3.3. Boxplot showing the variance of ELA$_0$ in different mountain sub-regions (a) and different main glacier aspects classes (b). Glacier mean size (ha) and relative area change (%) are insert for comparison. Mountain sub-regions: B/S = Bernina Scalino, A = Adamello, O/C = Ortles/Cevedale, D/M = Disgrazia/Mallero, D/P = Dosdè/Piazzi, S/L = Spluga/Lei, L = Livigno and O = Orobie.

3.7.2 RECENT AREA VARIATIONS (1991-2003-2007)

The 304 glaciers mapped in the 1991 inventory covered an area of 118.3 km². The same sample of glaciers decreased to 96.5 km² in 2003 (– 18.4 %) and to 89.4 km² in 2007, for a total area loss of 28.8 km² equal to the 24.4% of the 1991 surface (Table 3.1). The 1991-2007 mean annual area change is - 1.86 km² yr^{-1}. The rate of glacier shrinkage increased in speed during the recent years from – 1.81 (1991-2003) to – 2.02 km² yr^{-1} (2003-2007).

Diolaiuti et al. (2012) considering a smaller sample of glaciers (*n*=249), reported a higher retreat rate for the 1991-2003 interval (– 2.08 km² yr^{-1}), consequence of a higher value of general retreat (– 21.3 %, from 117.4 (1991) to 92.4 km² (2003)). The significant higher glacier extent we have found on our 2003 sample (even if larger in number) is probably due to the extensive field checks that have enabled an accurate mapping of the debris covered ice, often difficult to investigate by remote sensing.

Considering the different sub-regions, the shrinkage increasing trend is confirmed in Codera/Masino, Disgrazia/Mallero, Bernina/Scalino, Dosdè/Piazzi and Adamello while a slight decrease in rate is observed in Spluga/Lei, Livigno, Ortles/Cevedale and Orobie (Table 3.1). The sub-region with the worst performance in terms of acceleration of the areal retreat is the Bernina /Scalino where the mean annual change in area increased from - 0.29 to - 0.45 km² yr^{-1}. As this sub-region represents the 3rd largest (former 2nd), it heavily influences the regional value.

The 1991-2007 areal retreat is clearly related (R² = 0.827) to the former glacier size by a power law function (Fig. 3.4a). Things become much more complex if relative area retreat (in %, compared to the 1991 area) is considered. The interpolation of the complete pooled sample shows a weak coefficient of determination ($R^2 = 0.222$) between the relative area change (%) and the former area of each glacier (Fig. 4b; Table 3.2). In order to improve the analysis of this relation we divided the sample according to the glacier size (≤ 0.5 km^2 and > 0.5 km^2). The first sample is larger than the latter (267 and 37 glaciers, respectively) and it exhibits a non significant statistical relation ($R^2= 0.052$) while the 37 largest glaciers react in a more predictable way showing a higher, but yet not significant, coefficient of determination ($R^2= 0.344$) (Fig. 3.4b; Table 3.2). The 0.5 km^2 boundary is useful to highlight the behavior of the different sub-regions because all of them present a comparable number of glaciers in the ≤ 0.5 km^2 class. Moreover, Orobie and Codera/Masino are not represented in the second class (i.e. all their glaciers are smaller than 0.5 km^2). The comparison between the different sub-regions in the first category (≤ 0.5 km^2) is useful to highlight significant differences even in neighboring sub-regions. Notwithstanding, some of them (e.g., Spluga/Lei and Dosdè/Piazzi) exhibit high coefficients of determination even in this category, where the distribution of small glaciers is very scattered (i.e. many small glaciers were strongly reduced or completely disappeared in the considered time interval, while others showed a minor retreat, comparable with larger glaciers)(Figs. 3.5a, b; Table 3.2).

The best example of scattered distribution is represented by the Orobie sub-region where the relation between relative retreat and former area is totally absent ($R^2 = 0.003$) (Fig. 3.5d; Table 3.2). Qualitative photographic comparisons suggest that the areal decrease underestimate the real volume loss of small glacierets. Low elevation landforms, like the Dente di Coca glacieret (0.02 km^2) (Fig. 3.1a), gain or lose mass over the entire surface due to small size and characteristic avalanche feeding (Figs. 3.5d, e, f). Therefore, a decrease in volume is not clearly detectable by the areal change analysis until the final stage of the glacieret life when they completely disappear in a single ablation season.

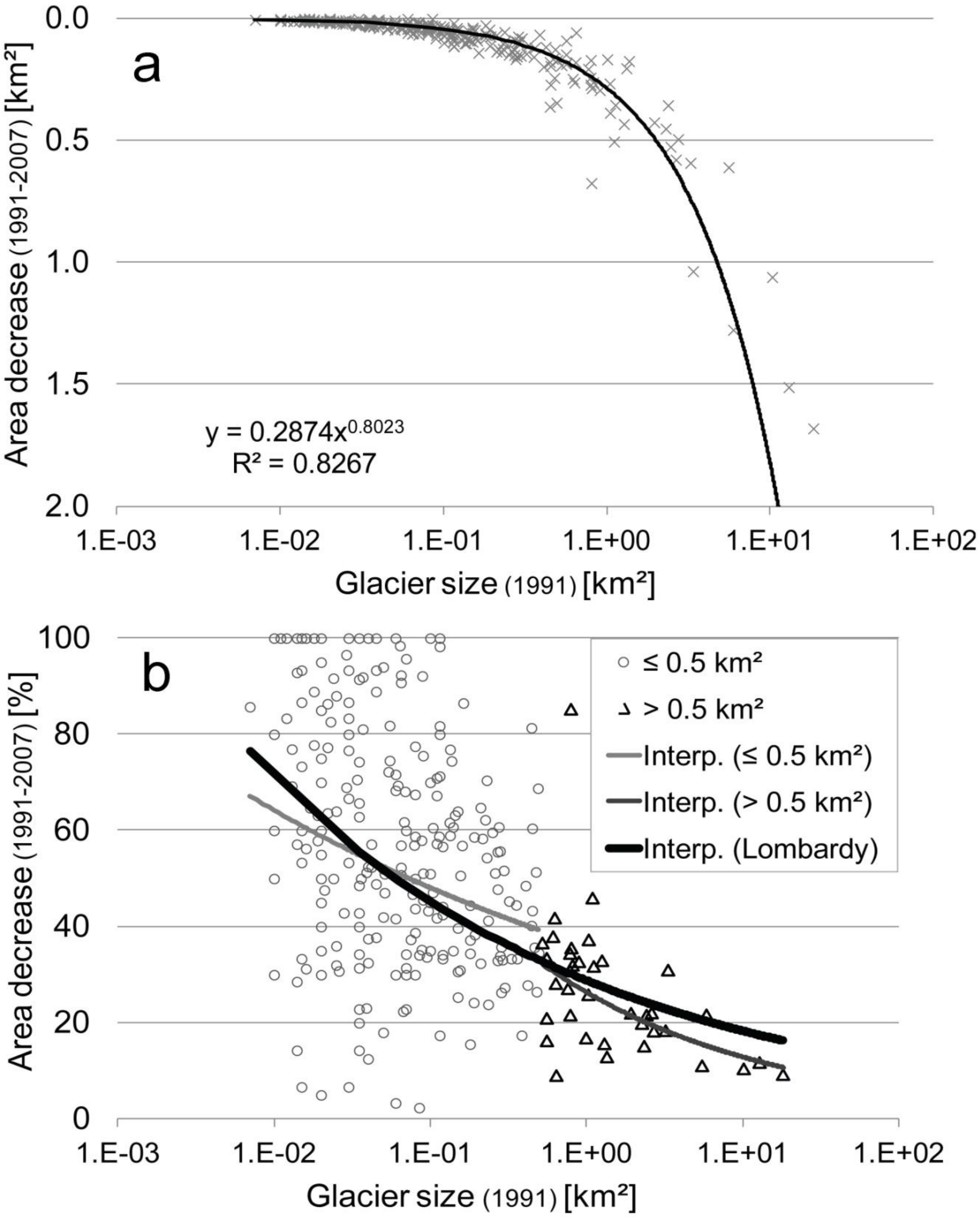

Fig. 3.4. Relation between area decrease (km², a) and relative area decrease (%, b) with glacier size (km²). Power law function is used.

The same behavior has been suggested for large glacier tongues (Abermann et al., 2009) where glacier downwasting is the main process and areal decrease underestimates the real decrease in thickness (i.e., volume). On the other hand, glaciers placed on flat slopes without avalanche feeding and orographic protection, like Alpe Sud (0.83 km²) (Figs. 3.5b, c) exhibit the higher relative shrinkage values and area losses are much more indicative of the real thickness decrease (Figs. 3.5a, b, c). We assume that volume relative changes should present higher values of coefficients of determination compared to the areal relative changes here available. However, some sub-regions with equal mean glacier size show substantial differences in relative area changes, so the position and the peculiar characteristics of the different mountain sub-regions are playing a role as well in this relation (Fig. 3.3a).

Table 3.2. Coefficient of determination between 1991/2007 area change and glacier size in different sub-regions.

Sub-region	Coefficient of determination R² (glacier count) 1991-2007 area change (%) / Area 1991		
	Glaciers ≤0.5 km²	Glaciers >0.5 km²	Total sample
Spluga/Lei	**0.484 (15)**	- (2)	**0.538 (17)**
Codera/Masino	0.041 (36)	-	0.041(36)
Disgrazia/Mallero	0.049 (28)	**0.402 (4)**	**0.130 (32)**
Bernina/Scalino	0.069 (21)	**0.707 (6)**	**0.634 (27)**
Dosdè/Piazzi	**0.510 (22)**	- (3)	**0.613 (25)**
Livigno	**0.203 (19)**	- (1)	**0.289 (20)**
Ortles/Cevedale	**0.142 (41)**	**0.151 (16)**	**0.467 (57)**
Adamello	0.098 (41)	**0.628 (5)**	**0.212 (46)**
Orobie	0.003 (44)	-	0.003 (44)
Lombardy Interpolation (Power law)	0.052 (267) $y=36.964x^{-0.197}$	**0.344 (37)** $y=26.283x^{-0.311}$	**0.222 (304)** $y=35.964x^{-0.126}$

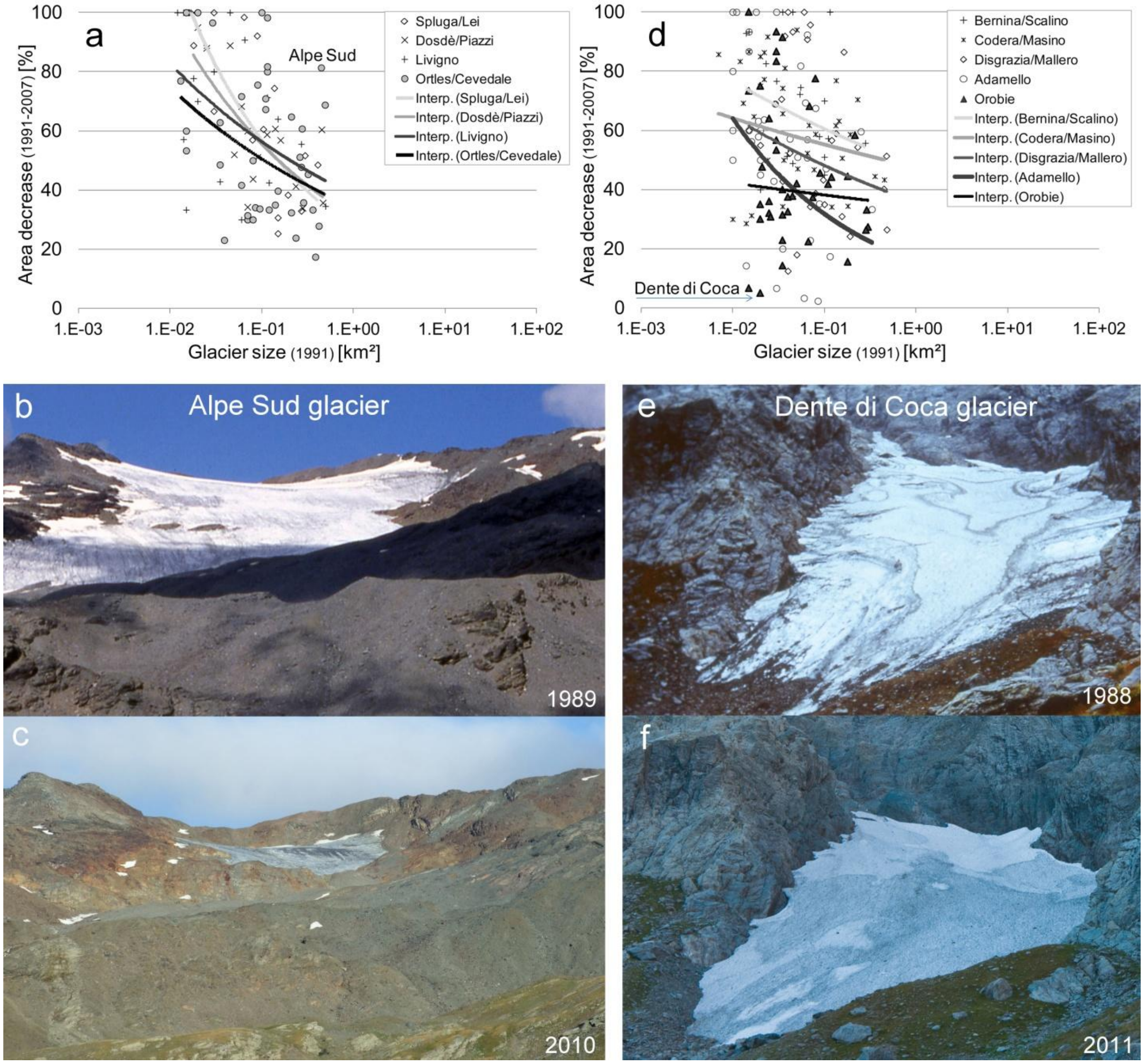

Fig. 3.5. Relation between relative area decrease (%, a) and glacier size (km²) for glaciers below 0.5 km² in mountain sub-regions with (a) and without (d) significant determination coefficients. Power law function is used. Photographs of two glaciers plotted in fig. a and fig. d are added: Alpe Sud, 46°23’30’’N-10°26’02’’E (b, c) and Dente di Coca 46°04’49’’N-9°59’43’’E (e, f). (b) photo A. Galluccio; (c) D. Colombarolli; (e) M. Butti and (f) R. Scotti. See text for details

Otherwise, the general rule is that largest areal losses are recorded in the sub-regions with the smaller mean glacier size (Fig. 3.3a). In fact, the sub-region showing the lower areal retreat (- 14.5 %, Adamello) is the one with the second mean glacier size (44.5 ha) while Codera/Masino mean glacier size (3.0 ha), exhibits the largest areal retreat (- 48.9 %). Bernina/Scalino and Orobie behave differently. The former, despite holding the largest mean glacier size (81.4 ha), shows a higher relative retreat (- 19.0 %) compared to Adamello.

The latter, despite its low mean glacier size (3.3 ha), records a relative low areal retreat (- 39.2 %). These two anomalies can be explained by the presence of many small size glaciers in the Monte Painale sub-sector of Bernina/Scalino witch probably increases the relative retreat value but is not affecting too much the mean glacier size of the whole sub-region.

On the other hand, the Orobie sub-region, benefits of the higher precipitation values in the region combined with high headwalls behind glaciers causing efficient avalanche feeding and orographic protection from solar radiation. These characteristics are affecting the ELA_0 of this sub-region which is by far the lowest of the region, even if ELA_0 in the other sub-regions, and by considering the different aspects, seems to be disjoined by relative areal retreat (Figs. 3.3a, b).

It is common sense that southerly sided glaciers retreat faster than others due to unfavorable climatic conditions. Nevertheless, if we consider the simple aspect parameter, this is stable over time so, southerly sided glaciers plot at higher elevations (i.e. lower temperature and better climatic conditions to sustain glaciers) compared to the others thus compensating the higher potential incoming solar radiation. This consideration is supported by the relative area variation analyzed by different aspect classes. W and S aspect classes show the higher areal decrease in the 1991-2007 interval, but SW and SE classes exhibit lower retreat values, comparable to, or even lower, than NE and N aspect classes. Also in this case the mean glacier size controls this behavior as shown in Fig. 3.3b where it is evident that the aspect class does not cause different retreat rates like mean glacier size.

3.7.3 CHANGES SINCE THE LITTLE ICE AGE

From LIA to the present, the sample of nine selected glaciers recorded area losses generally related with their size (Fig. 3.6; Appendix 1-9). Rasica Est, the smallest glacier here considered, has lost 90.3 % of its area to become a glacieret very close to its extinction (Appendix 2). On the other hand, Fellaria glacier, the largest in this sample, and the 3rd largest in Lombardy, has lost the 37.7 % reducing from 14.96 to 9.32 km² (Fig. 3.6; Table 3.3). Mid size glaciers as Campo Nord, Pizzo Ferrè and Dosdè Ovest, exhibit relative retreats between 60 and 70 % with two extreme cases: Lupo (Orobie) and Cassandra glacier (Disgrazia/Mallero) (Figs. 3.8a, b, c). The former, coherently with the anomalous low recent areal decrease in the Orobie sub-region, suffered a reduction from 0.43 to 0.22 km² (- 49.2 %), while the latter has lost 83.1 % of the former area, decreasing from 1.55 to 0.26 km².

Such a large reduction can be explained with the complex morphology of Cassandra glacier. At the LIA maximum it was flowing from a small accumulation basin dividing the flux downstream in 2 lobes that used to outflank the 2873 unnamed top (Figs. 3.8a, b, c). In order to sustain this extent, considering the tiny accumulation basin, a intense avalanche feeding from the side-walls was required. So, the increase of the ELA during the last century caused a sudden decrease of snow fall on the side walls (i.e. less avalanches) and an imbalance of most of the glacier surface below the current terminus position. This behavior is documented in the historical surveys completed during the last century and it is highlighted by the 1954 and 1990 glacier morphology where the lower part of the glacier is hardly dynamically linked with the upper portion and it is characterized by disaggregate in-situ dead ice lenses (Figs. 3.8a, b, c).

This behavior is mainly caused by the rapid rise of the ELA and the complex bedrock morphology, and occurs frequently in many other glaciers of the region. For this reason the classical investigation of glacier fluctuations by field measurements of the terminus is becoming less representative in this glacier downwasting phase.

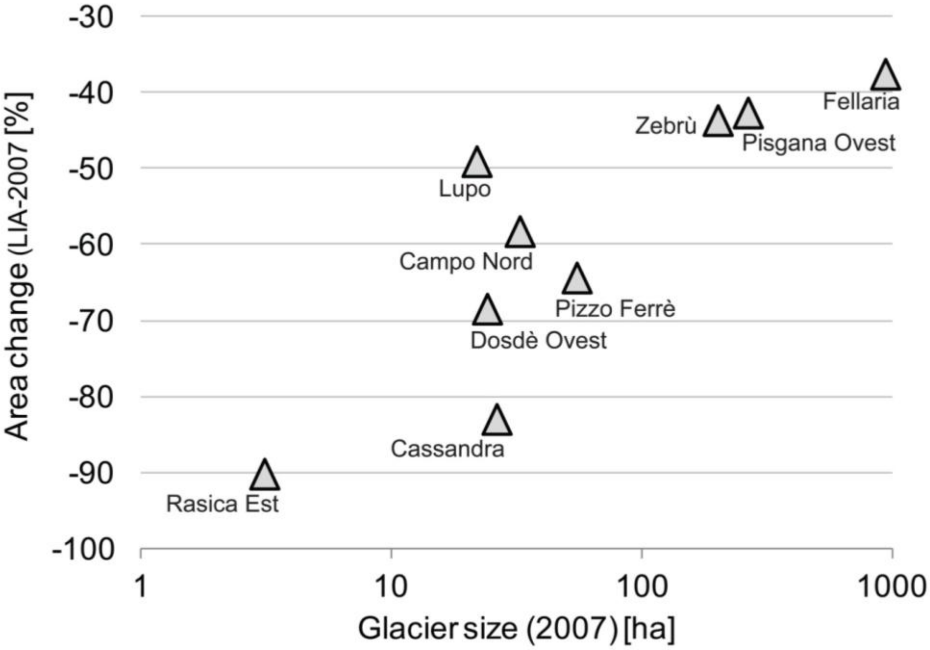

Fig. 3.6. Relation between relative area variation from LIA to 2007 and glacier size (ha) for nine selected glaciers.

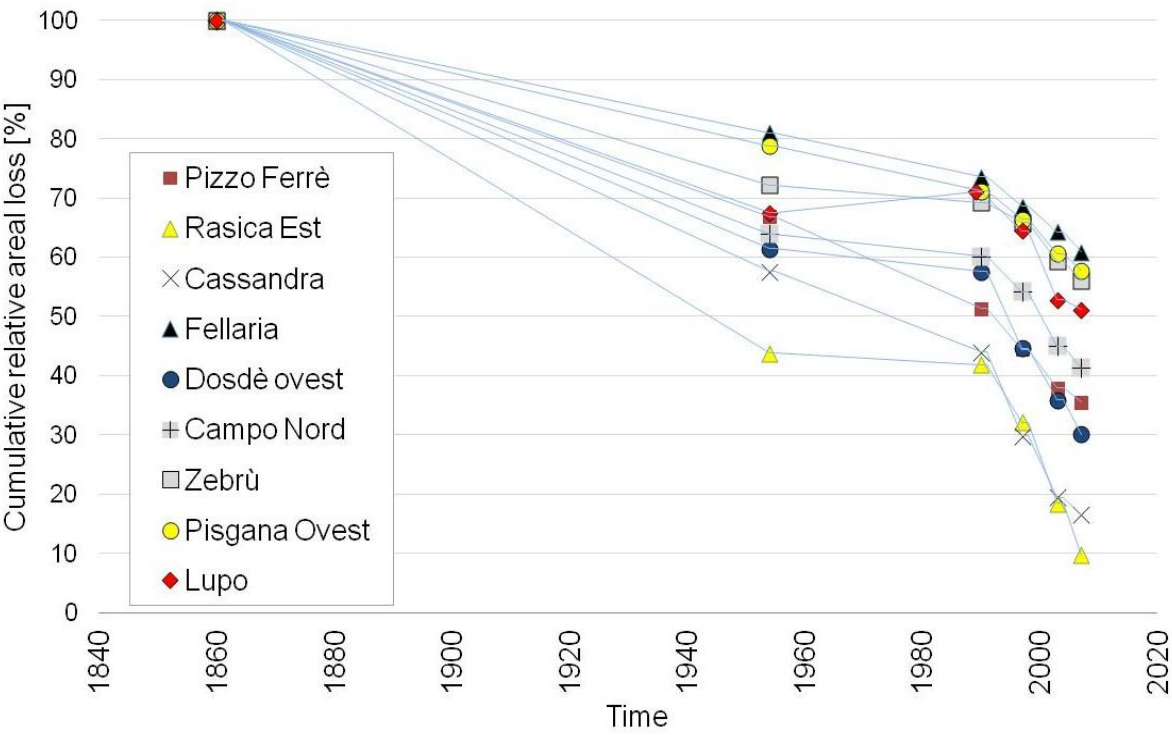

Fig. 3.7. Cumulative relative area loss from LIA (1860) to 2007 divided in 5 intervals (1860-1954, 1954-1990, 1990-1997, 1997-2003 and 2003-2007.

In a through-time prospective the mean annual retreat since LIA of the selected glaciers highlights the recent areal collapse. In the 1954-1990 interval a general decrease in the retreat rates, compared to the LIA-1954 interval, is observed (Table 3.4) (Fig. 3.7). The largest glaciers (i.e. Fellaria, Pisgana and Cassandra) maintained the same retreat rate of the previous interval. Rasica Est, Dosdè Ovest and Zebrù glaciers drastically decreased the retreat rate, and Lupo glacier slightly increased its extent (Table 3.3) (Fig. 3.7). A clear turning point is evident in 1990 as the subsequent intervals shows, even if very short, higher retreat rates. Even though the differences in retreat rate in the three more recent intervals are small, the maximum retreat rate is reached during the 1997-2003 interval by Pizzo Ferrè, Campo Nord, Zebrù, Pisgana Ovest and Lupo glacier. Cassandra and Dosdè Ovest glaciers reached the peak in the 1990-1997 interval, while Rasica Est and Fellaria in the 2003-2007 interval (Table 3.4) (Fig. 3.7).

Table 3.3. Area, and relative areal change since LIA maximum (1860) to 2007.

Glacier	Area (km^2)						Area change (%)
	LIA	1954	1990	1997	2003	2007	LIA-2007
Pizzo Ferrè	1.54	1.03	0.79	0.68	0.58	0.55	**-64.5**
Rasica Est	0.32	0.14	0.13	0.10	0.06	0.03	**-90.3**
Cassandra	1.59	0.91	0.70	0.47	0.31	0.26	**-83.1**
Fellaria	15.30	12.40	11.26	10.50	9.84	9.32	**-37.7**
Dosdè Ovest	0.80	0.49	0.46	0.36	0.29	0.24	**-68.6**
Campo Nord	0.78	0.50	0.47	0.42	0.35	0.32	**-58.4**
Zebrù	3.56	2.57	2.46	2.34	2.11	2.00	**-43.8**
Pisgana Ovest	4.61	3.63	3.28	3.06	2.79	2.66	**-42.8**
Lupo	0.43	0.29	0.30	0.28	0.23	0.22	**-49.2**

Every selected glacier exhibits a LIA-1990 retreat rate largely smaller than the one in the subsequent intervals (Table 3.4). In particular, if we consider the 1990-2007 interval, the retreat rate increased from 2.3 times (Zebrù glacier) up to 5.7 times (Dosdè ovest) (Table 3.4). The acceleration of the glacier shrinkage was so intense that even splitting the 1990-2007 interval by the 1997 and 2003 survey, it is not possible to find any retreat rate comparable with those of the LIA-1990 interval (Table 3.4).

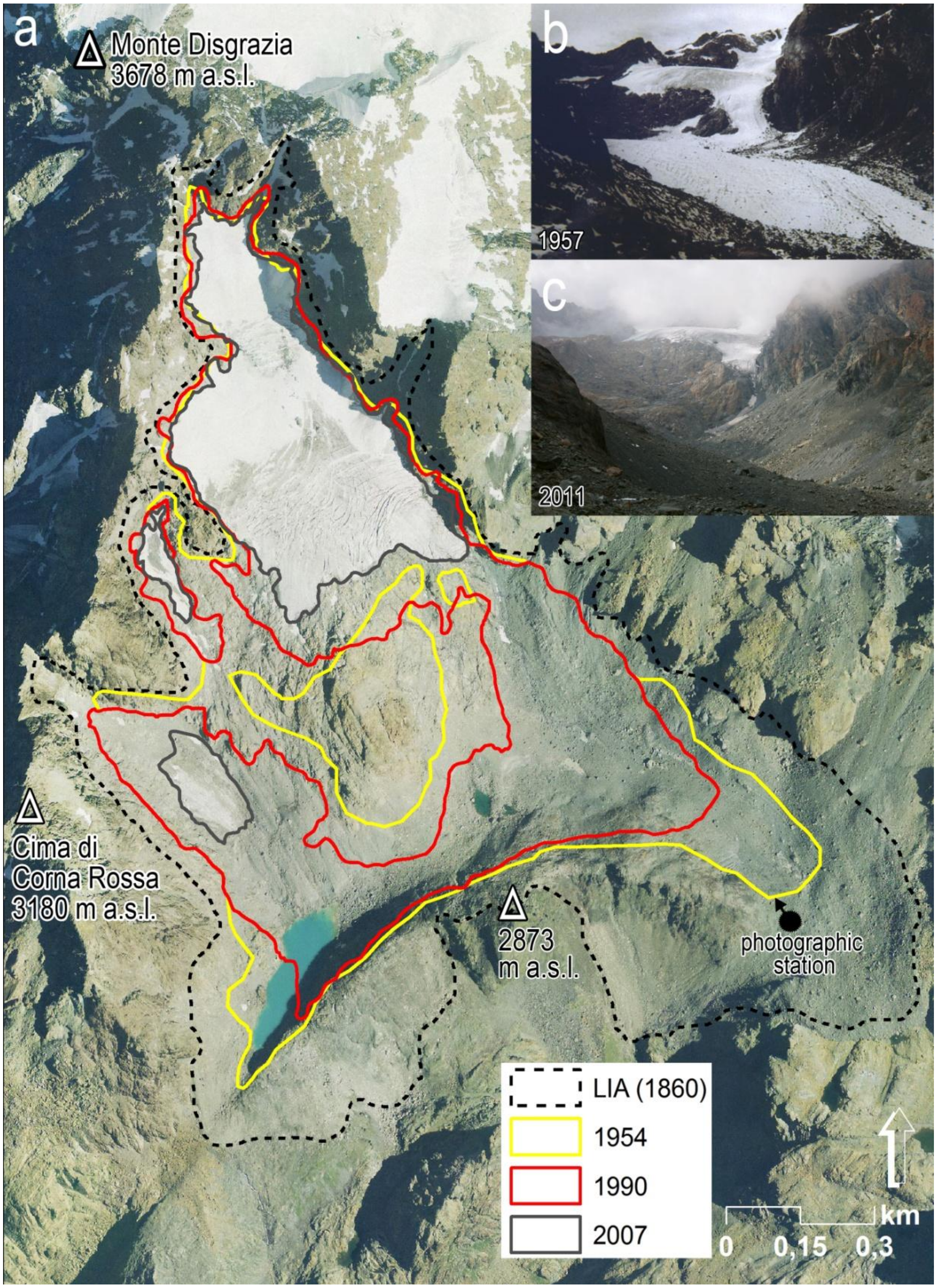

Fig. 3.8. Cassandra glacier (46°15'62''N-9°45'17''E; Disgrazia/Mallero) extent during LIA maximum, 1954, 1990 and 2007. The 2007 orthophoto is set as background image. Approx scale 1:10,000 (a). The east lobe of the glacier in 1957 (photo G. Nangeroni; b) and 2011 (photo R. Scotti; c).

Table 3.4. Mean annual areal change and ratio between the mean annual areal change LIA-1990 and 1990-2007 for nine selected glaciers. *For Lupo glacier the two intervals are LIA-1989 and 1989-2007

	Mean annual areal change (km^2 yr^{-1})							LIA-1990/ 1990-2007
Glacier	**LIA-1954**	**1954-1990**	**1990-1997**	**1997-2003**	**2003-2007**	**LIA-1990**	**1990-2007**	
Pizzo Ferrè	-0,005	-0,007	-0,015	-0,017	-0,011	-0,006	-0,014	**2.5**
Rasica Est	-0,002	0,000	-0,004	-0,007	-0,008	-0,001	-0,006	**4.2**
Cassandra	-0,007	-0,006	-0,032	-0,028	-0,013	-0,007	-0,026	**3.4**
Fellaria	-0,031	-0,033	-0,109	-0,109	-0,149	-0,031	-0,114	**3.7**
Dosdè Ovest	-0,003	-0,001	-0,015	-0,012	-0,013	-0,003	-0,013	**5.7**
Campo Nord	-0,003	-0,001	-0,007	-0,012	-0,008	-0,002	-0,009	**4.6**
Zebrù	-0,011	-0,003	-0,020	-0,039	-0,033	-0,008	-0,029	**2.3**
Pisgana Ovest	-0,010	-0,010	-0,031	-0,045	-0,038	-0,010	-0,036	**3.3**
Lupo*	-0,001	0,000	-0,004	-0,008	-0,002	-0,001	-0,005	**4.6**
Temperature (J-S) (°C)	9,14	9,07	9,82	10,19	10,15	9,12	10,03	**+ 0,91 °C**
Precipitation (O-M) (mm)	554	549	496	670	439	552	544	**- 8 mm (1,4%)**

3.8 DISCUSSION

3.8.1 CLIMATIC FLUCTUATIONS SINCE 1864

Analyzing the Sils/Maria climatic data, the long term atmospheric temperature trend is clear, with the steady increase of the temperature since the early 1980s. The warming trend for this period is so pronounced that only one ablation season (1984) out of the last 31 is reported below the average (1864-2012). The warmest summers are concentrated in the last 10 years: 2003 (June-September average: + 12.1 °C), 2012 (+ 11.1 °C) and 2009 (+ 11.0 °C) while the coolest ones are found within the cold phases of the 1920s and 1970s: 1912 (+ 7.3 °C), 1913 (7.5 °C) and 1972 (+ 7.6 °C) (Fig. 3.9a).

The overall trend of precipitation in the October – May interval can be assumed as steady. The 20-year dry period at the end of the XIX century is followed by almost 30 years of sustained precipitations. Wet years also occur in the 1970s and 1980s. The most abundant accumulation season is the 2000-2001 (1167.3 mm, twice the average value). Other wet accumulation seasons are those related to 1976-77 (985.7 mm) and 1982-83 (926.6 mm), whereas the driest is the 1941-42 with 218.3 mm. Other dry accumulation seasons are the 1943-44 (251.1 mm) and the 1879-80 (273.0 mm) (Fig. 3.9b).

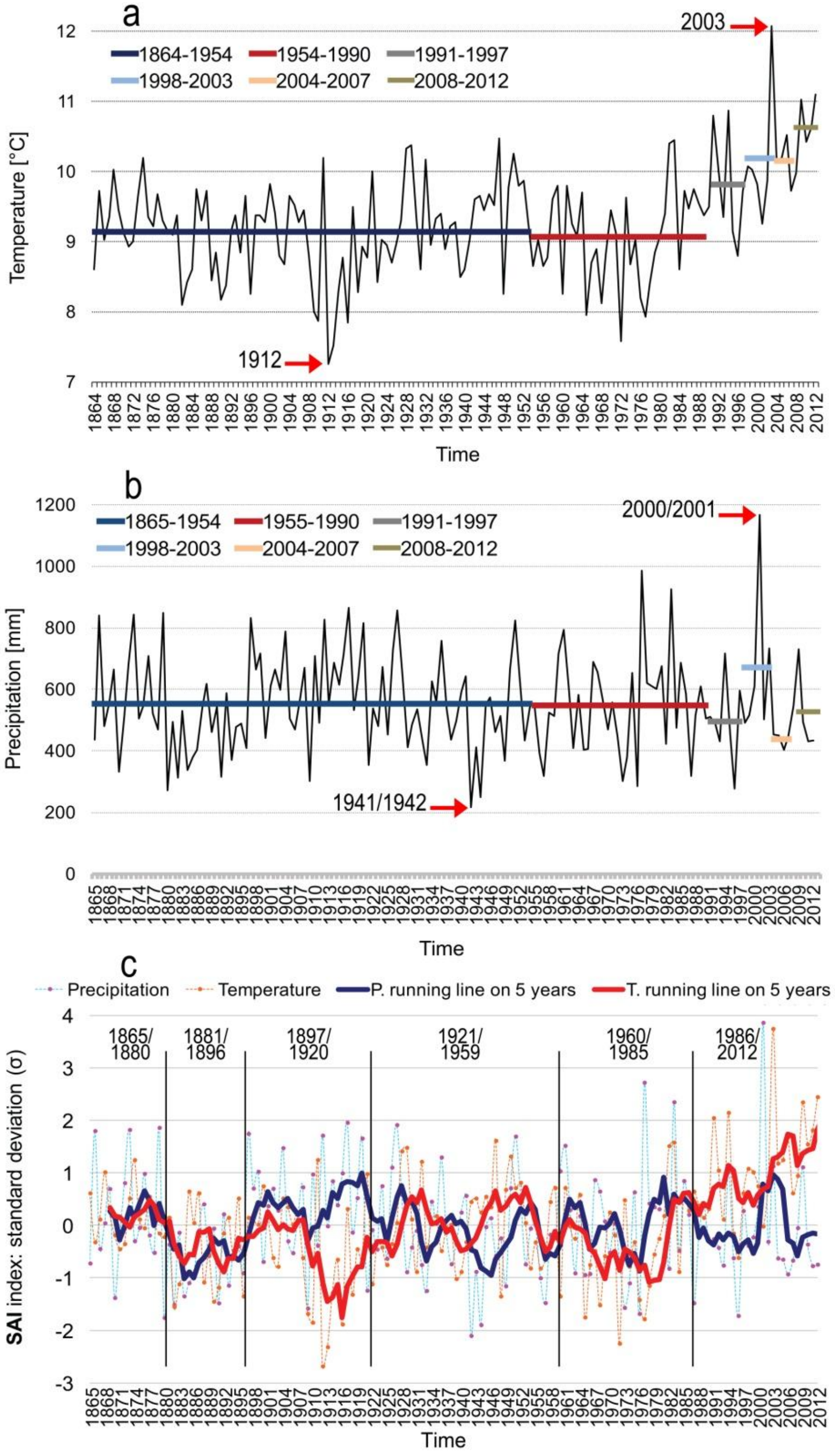

Fig. 3.9.
(a) Temperature (June-September) and (b) precipitation (October-May). See text for details. (c) Standardized Anomaly Index (SAI) for temperature (June-September) and precipitation (October-May) in Sils/Maria. Periods and sub-periods are defined on the basis of their likely effect on glaciers (time interval 1864/65-2011/12).

The SAI index was used for comparing precipitation anomalies in the accumulation season with temperature of the subsequent ablation season, thus to understand the climatic trend in relation to the glacier fluctuations. Analysis of both parameters (fig. 3.9c) points out six homogeneous climatic intervals. The first one runs from the beginning of the series to 1880 (15 years): this phase shows alternating trends for both variables (precipitation and temperature) unable to maintain in a steady state the glacial tongues that had reached their maximum extent a few years before.

In the following interval (1881-1896) temperature drops by about 0.5 °C (- 0.5 σ the mean anomaly of the period), while precipitations are the lowest of the whole time series (- 0.6 σ). This phase is neutral or even slightly positive for the Lombardy glaciers. From 1897 onward, both precipitation and temperature increases. Until 1909, positive (1902) and negative (1906) seasons make the assessment of their effect on glaciers rather difficult to define. Such problem does not arise for the subsequent 11- year period: after 1910 precipitation strongly increases while, at the same time, temperature shows a significant reduction. In fact, 1912 and 1913 are the most positive years for Lombardy glaciers since the end of the LIA. After the exceptional snowfall of 1916-17 (Galluccio and Cola, 2001), this positive pulsation ends in 1921 (e.g., Citterio et al, 2007). Considering the whole 1897-1920 interval, temperature maintains an anomaly of - 0.4 σ but precipitations increases at + 0.3 σ (the highest value of the entire series).

For the following years, both temperature and precipitations resume values close to the long term average (+ 0.1 and – 0.2 σ in the 1921-1959 interval, respectively). Once again, such mean value is not enough to maintain glaciation in a steady state; glaciers experience pronounced recessions despite the presence of a few favorable years (1927-1936).

From 1942 onward, ablation seasons turn out to be unusually warm while winter snowfalls become scarce. This period sees some of the worst years of the whole XX century for the glaciers (e.g., 1942, 1947 and 1949). In the 1950s the anomalies gradually recede, though not enough to stop the ongoing withdrawal of glaciers. It may be useful to highlight 1951 as a positive year for the glaciers but a tragic one for the Alpine population due to catastrophic avalanches causing hundreds of deaths (e.g., SLF, 1952, Roch, 1980).

The second wettest year of our historical sequence is 1960 marking the beginning of a new phase. Two consecutive abundant accumulation seasons (1960 and 1961) coupled with cool summers reconstitute residual snow-fields in the accumulation basins triggering the starting of glaciers expansion. From 1974 the slightly positive phase turns to significantly positive. Two exceptional seasons should be mentioned: 1977 and 1980. Between 1960 and 1985 glaciers advanced over the whole Alpine chain. This phase was triggered by a temperature decrease of about 0.3 °C (compared to the previous negative 1921-1959 period) and by an increase of precipitation of 51.8 mm (+ 9 %).

After 1977, summers become gradually warmer until the floods of 1987 which ended this series of positive years for the Lombardy glaciers. The anomalies of the whole period (1960-1985) are $- 0.3\ \sigma$ and $+ 0.1\ \sigma$ for temperature and precipitation, respectively.

The last 30 years show an unprecedented thermal increase. Summer temperatures during the ‘90s constantly exceed the long time average, apart from a small drop in 1996 and 1997. The XXI century seems to introduce another change, thanks to the strong snowfalls of the 2000-

2001 hydrologic year (precipitation anomaly: + 3.9 σ) witch, even if partly offset by a warm summer, remains a single episode, as the temperature continues to increase while the winter precipitation cannot compensate such negative anomaly. The thermal increase peaks with the 2003 season (+ 3.7 σ) followed by other extremely warm summers (e.g., 2009 and 2012). In the 1987-2012 interval, temperature anomaly reaches + 1.0 σ while precipitation is only slightly below the average (– 0.1 σ) (Fig. 3.9c).

3.8.2 GLACIER AND CLIMATIC FLUCTUATIONS

An organic reconstruction of glacier maximum extension in Little Ice Age (LIA) in Lombardy Alps is missing. Pelfini (1994) studied the Ortles/Cevedale glaciers and reported a reduction in area of 50 % since LIA. In the last twenty years of the XX century, the reduction rates increased (Ajassa et al., 1997), up to the last few years (2002-2012) characterized by a very severe de-glaciation. In this study we report areal relative retreats from 38 % to 90 % since LIA (Table 3.3; Fig. 6).

The retreat occurred in the 1990-2007 interval ranges from 24 % (Pizzo Ferrè glacier) to 41 % (Lupo glacier) of the total retreat since LIA. Such values are remarkable as the 1990-2007 interval account only for the 12 % of the total time span from LIA (1860-2007).

The increase in mean annual retreat rate in the 1990-2007 interval compared to the mean retreat rate from LIA to 1990, is remarkable in all the nine selected glaciers. Even if topographical and geographical considerations are difficult, the Dosdè Ovest and Campo Nord glaciers, placed in the neighboring Livigno and Dosdè/Piazzi sub-regions, increased the mean annual retreat rate up to 5.7 and 4.6 times respectively, showing the worst performance among the selected glaciers. This suggests an acceleration of glaciers recession in the north-eastern sector of Lombardy Alps (Table 3.4).

The lower value expressed by the Zebrù glacier (2.3), located at the NE corner of the study area, is not inconsistent with this supposition. In fact, this glacier drastically reduced the areal retreat after the 2004 Thuwieser rock avalanche (e.g., Cola, 2005; Sosio et al., 2008) which covered the glacier snout reducing the ablation (i.e. partly disjoining the glacier fluctuations from the climate forcing) (Figs. 3.10a, b) (Greggio, 2012).

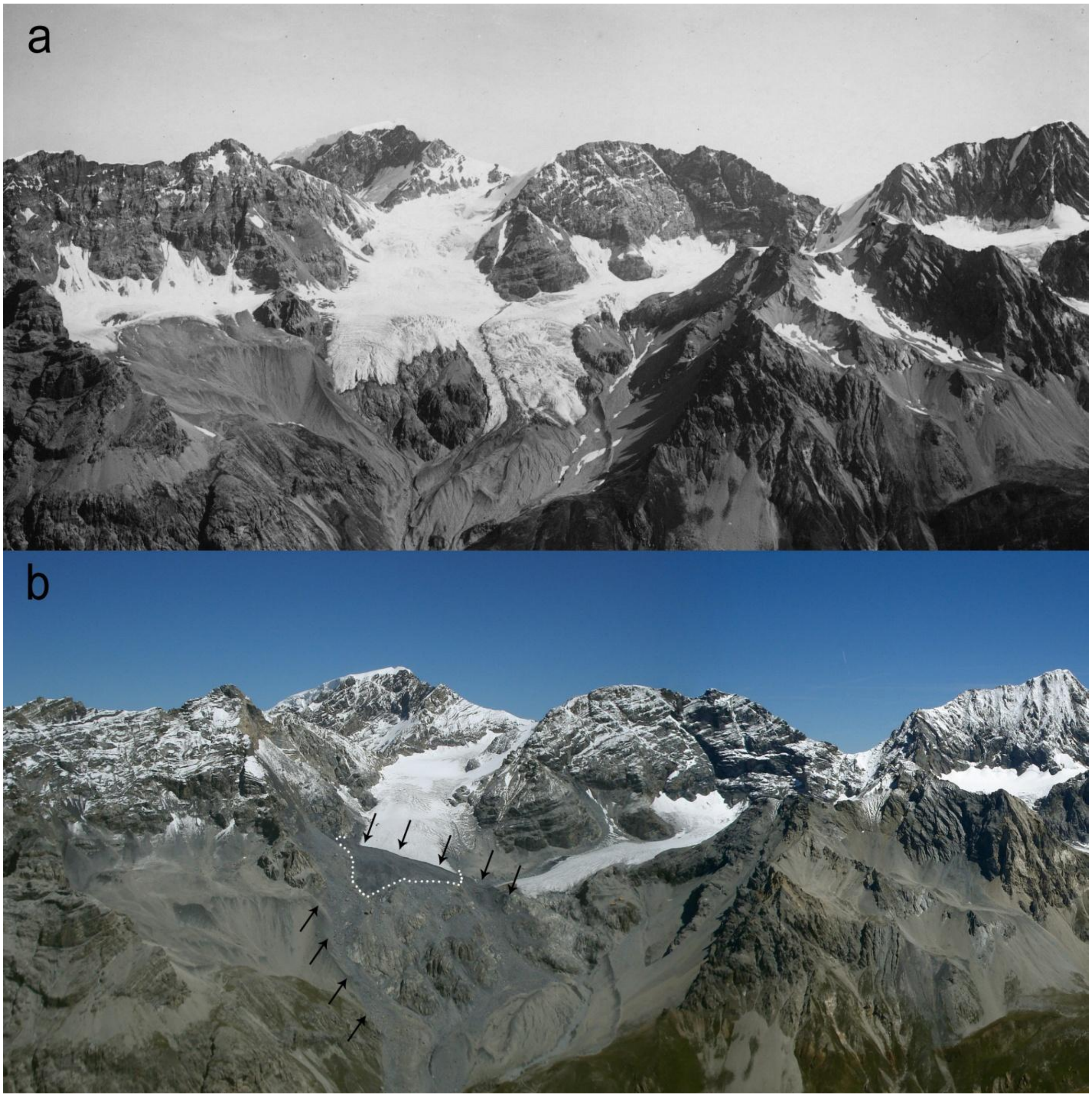

Fig. 3.10. Zebrù glacier (46°29’35’’N–10°32’08’’E) in 1907 (photo A. Corti – Arch. Sez. CAI Valtellinese; a) and in 2007 (photo V. Sciaresa – SGL; b). In (b) is evident (arrows) the debris accumulation derived from the Punta Thurwieser rock avalanche (18 September 2004) that covered the glacier snout (dotted line).

Looking closely to the climatic data in the different time intervals, it is obvious how a minimal decrease in temperature from 1864-1954 to 1954-1990 (– 0,07 °C), even if accompanied by a decrease in precipitation (– 5 mm, from 554 to 549 mm), was able, to significantly decrease the retreat rates or even maintain the glaciers stable (e.g., Rasica Est and Lupo glaciers) (Table 3.4). The subsequent intervals strengthen the linkage between glacier retreat rates and summer temperatures. The 1990-1997 time interval reveals an increase of 0.75 °C and a decrease in precipitation (– 53 mm) compared to 1954-1990. Glaciers reacted to this change drastically increasing the mean annual retreat rates with no exceptions in any of the nine selected glaciers (Fig. 3.7). The Dosdè Ovest glacier represents the extreme case, increasing the mean annual retreat rate from – 0,001 to – 0,015 km² yr^{-1} (Table 3.4). The following interval (1997-2003) is probably the most interesting as summer temperature rises further up to 10.19 °C (+ 0.37 °C) but precipitation in the accumulation season significantly rises too, shifting from 496 to 670 mm (+ 174 mm) compared with 1990-1997. We could expect that the relatively small step in temperature would be compensated by this increase in precipitation but glaciers reacted differently. In many cases they further increased the retreat rates reaching the higher values in the studied intervals (e.g., Pizzo Ferrè, Campo Nord, Zebrù, Pisgana Ovest and Lupo glaciers). This is highlighted in another study (Diolaiuti et al., 2012) where a sample of 249 glaciers in Lombardy increased the retreat rate from – 1.58 to – 3.08 km² yr^{-1} between similar time intervals (1991-1999 and 1999-2003).

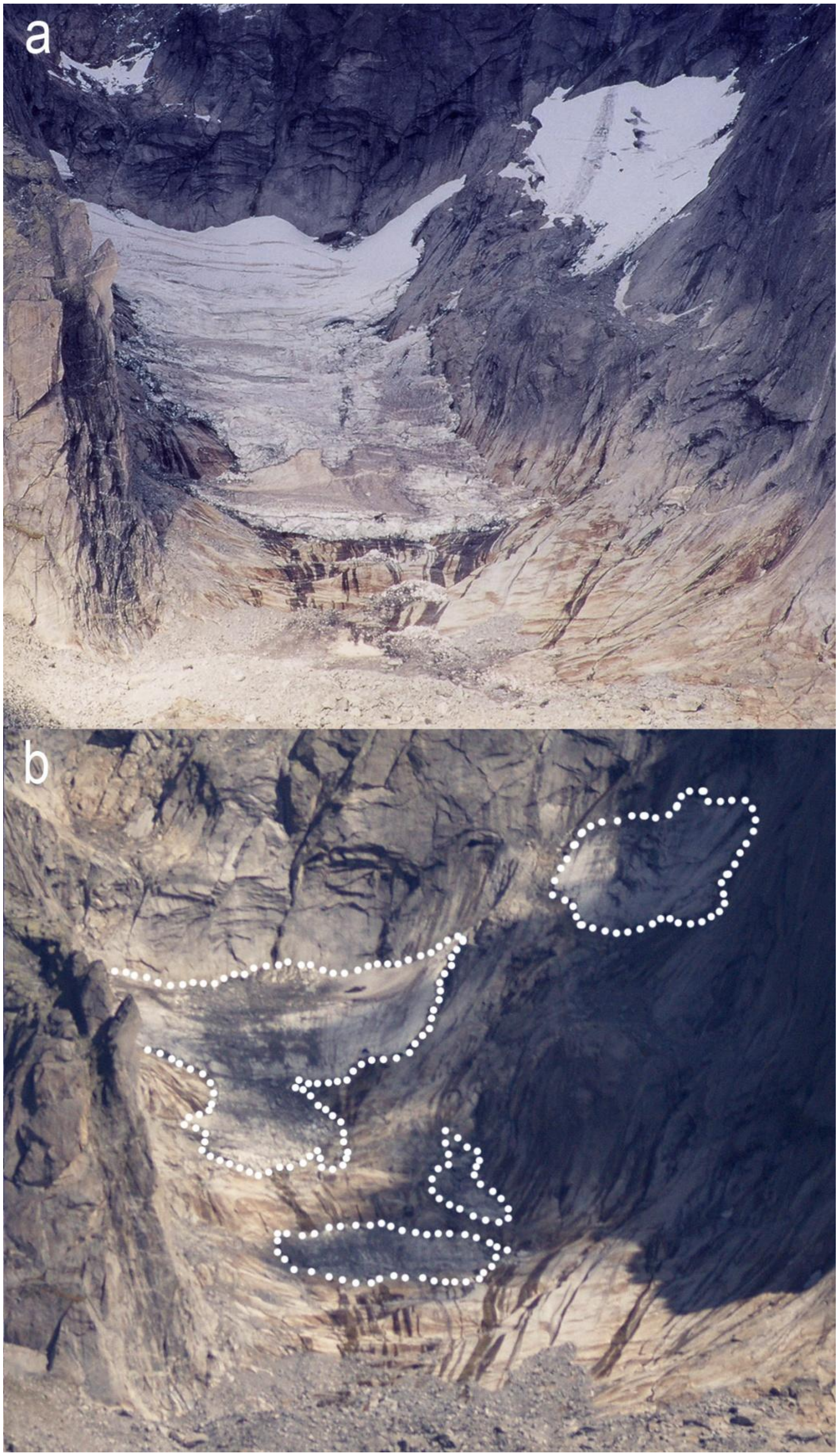

Fig. 3.11.
Rasica Est glacier (46°17’05’’N-9°40’77’’E) in 2003 (photo G.Di Gallo – SGL; a) and 2006 (photo A.Barilli – SGL; b).
Both pictures are taken at the end of the ablation season. The comparison highlights the downwasting of the glacier that lost 50% of its area in the 2003-2007.

During the last interval considered (2003-2007) temperature does not change significantly (– 0.04 °C) while precipitation decreases to 439 mm (i.e., the lowest value here considered) compared to the 1997-2003 interval. The retreat rates of the selected glaciers reaches the higher values at Rasica Est (Figs. 3.11a, b) and Fellaria glaciers and slightly decreases in the others. Considering the limited time span of the last two intervals, anomalies of a single year can play an important role. For instance, the 1997-2003 high precipitation average value was mainly caused by the 2000/2001 accumulation season alone. The extreme snowfall of that autumn-winter was able to drive a widespread positive glacier mass balance in Lombardy alps (i.e. blocking the areal retreat) (Fig. 3.9b). That season was soon followed by the extreme 2003 summer that cancelled the 2001 firn deposits and caused most of the areal retreat recorded in the whole interval (1997-2003). This particular behavior suggests the important role of summer heat waves and, in a broader sense, summer temperatures, driving glacier mass balance and consequently, areal variations. This relation is strengthen by the investigation of the long term climate data. While precipitation did not show any sign of a change from 1864 and temperature rose at a rate of 0.0057 °C yr^{-1}, Fellaria glacier, the largest here investigated, shrank at a rate of 0.041 km^2 yr^{-1}. Considering the wider and more representative sample of 304 glaciers analyzed since 1991, they show an increase in mean annual area change rate from – 1.81 (1991-2003) to – 2.02 km^2 yr^{-1} (2003-2007). Summer temperatures can still explain this increase as they rose of 0.16 °C, even if precipitation decreased from 576.5 to 438.5 (– 138 mm; – 23 %) between the two time intervals (Table 3.1).

Fig. 3.12. Pizzo Varuna glacier (46°20’09’’N-9°59’60’’E; Bernina/Scalino) in 1998 (photo M. Butti – SGL; a) and, completely disappeared in 2012 (photo R. Scotti; b). Pizzo Varuna glacier was part of the Fellaria glacier until 1926.

On the other hand qualitative analysis based on field surveys and preliminary results of ongoing monitoring projects highlight a not negligible role of precipitation anomalies on some glaciers (e.g., small avalanche fed glaciers and glacierets in the Orobie sub-region). Precipitation increased as well, even if only realigning to the long time average (i.e., the 2007-2012 interval was preceded by a drought 2003-2007 period).

In this time span (2007-2012) the Alpe Sud glacier (Ortles/Cevedale sub-region) showed a series of negative net mass balances, Pizzo Varuna glacier (Bernina/Scalino sub-region) completely disappeared (Figs. 3.12a, b), while the Lupo glacier (Orobie sub-region) reacted differently (Figs. 3.13a, b). Thanks to the extremely high mean accumulation, more than four times larger than Alpe Sud, the Lupo glacier (the largest glacier in Orobie sub-region), experienced a positive balance in 2009 and a substantial equilibrium balance for the following two years (Figs. 3.13a, b). These balances are not accurately explained by Sils/Maria data as only in 2009 precipitations were above the average. The Sils/Maria climatic station remains an exceptional milestone for glaciological studies in the Lombardy Alps and the general relation of glacier fluctuations with summer temperature remains strong but further investigations are required in order to better understand the different climatic trends (especially precipitation) in different sub-regions and in shorter intervals. This behavior is evident in the 2007-2012 interval, not yet covered by a complete areal analysis, as the average temperature (10.63 °C, + 0.48 °C from the 2003-2007 interval), is the highest of the time-series.

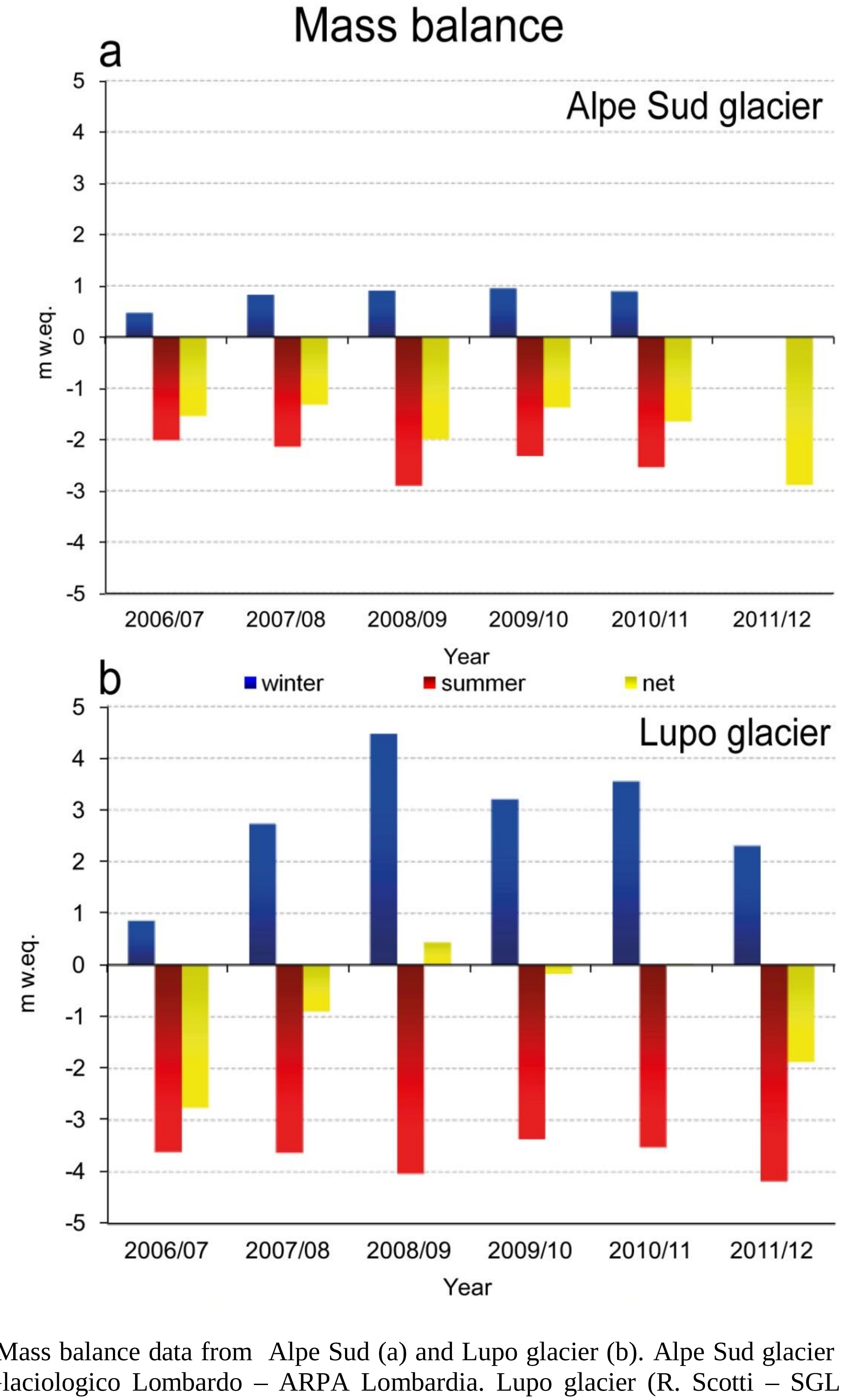

Fig. 3.13. Mass balance data from Alpe Sud (a) and Lupo glacier (b). Alpe Sud glacier data source: Servizio Glaciologico Lombardo – ARPA Lombardia. Lupo glacier (R. Scotti – SGL). Alpe Sud glacier winter and summer balances in 2012 are missing.

3.9 SUMMARY

Actually (2007) the 333 glaciers, glacierets and perennial snow fields detected in the study region cover an area of 90.4 km². To study the recent areal variations within nine sub-regions, a smaller fixed sample of 304 glaciers detected in 1991 (118.3 km^2) was used. Total glacier extent displays a progressive decrease through 2003 (96.5 km²) and 2007 (89.4 km²), for a total loss of 28.8 km², that is, 24.4 % of the 1991 surface. Considering the single glaciers, the relative areal retreat from 1991 seem to be unrelated to most of the variables examined such as, elevation, main aspect, ELA_0 (i.e., snow accumulation) and glacier type. A general relation with the mean glacier size was found, even if only the sample of 37 glaciers larger than 0.5 km² show a higher, but yet not significant ($R^2 = 0.344$). Supporting this relation, the higher relative retreat was detected in the sub-regions with the smaller mean glacier size (e.g., Livigno, Spluga/Lei and Codera/Masino) where the area has almost halved from 1991. Glacier recession increased in speed from – 1.81 km² yr^{-1} (1991-2003) to – 2.02 km² yr^{-1}(2003-2007) as a consequence of an increase in summer temperature of 0.16 °C and a decrease in precipitation of 23.9 % between the two intervals considered. Glacier fluctuation analysis was extended back to LIA maximum on nine selected glaciers. Most of them display an areal contraction of more than 50 % since LIA (ranging from 37.7 % of the largest (Fellaria) to 90.3 % of the smallest (Rasica Est)). Such values are consistent with prior studies conducted in the Ortles/Cevedale sub-area (Pelfini, 1994), in the Rutor glacier (Western Alps) (Orombelli, 2005; Villa et al., 2008), and with an alpine-wide reconstruction reporting an overall retreat of 50% between 1850 and 2000 (Zemp et al., 2008). The selected glaciers drastically increased (up to 5.7 times) their retreat rate from 1990 onward, in comparison with the average rate observed between LIA and 1990. Therefore, a large proportion of the total loss from LIA (24 – 41 %) can be associated with the 1990-2007 period, which accounts for only the 12 % of the total time post-LIA time window. Sils/Maria climatic data from 1864 were useful to explain glacier fluctuations especially in the most recent times. Most of the glacier shrinkage from LIA appears to be related to the increase in summer (J-S) temperature (1864-2012 average of 0.0057 °C yr^{-1}) (i.e., precipitation during the accumulation season (O-M) does not show any obvious trend). Since 1990, summer temperature increased by 0.91 °C but precipitation decreased only by 8 mm with respect to the 1864-1990 average. This substantial temperature rise appears to explain the recent acceleration of glacier retreat observed in the Central Italian Alps, even though the potential confounding exerted by local conditions deserves further investigations as highlighted by recent mass balance data (Figs. 3.12 a, b).

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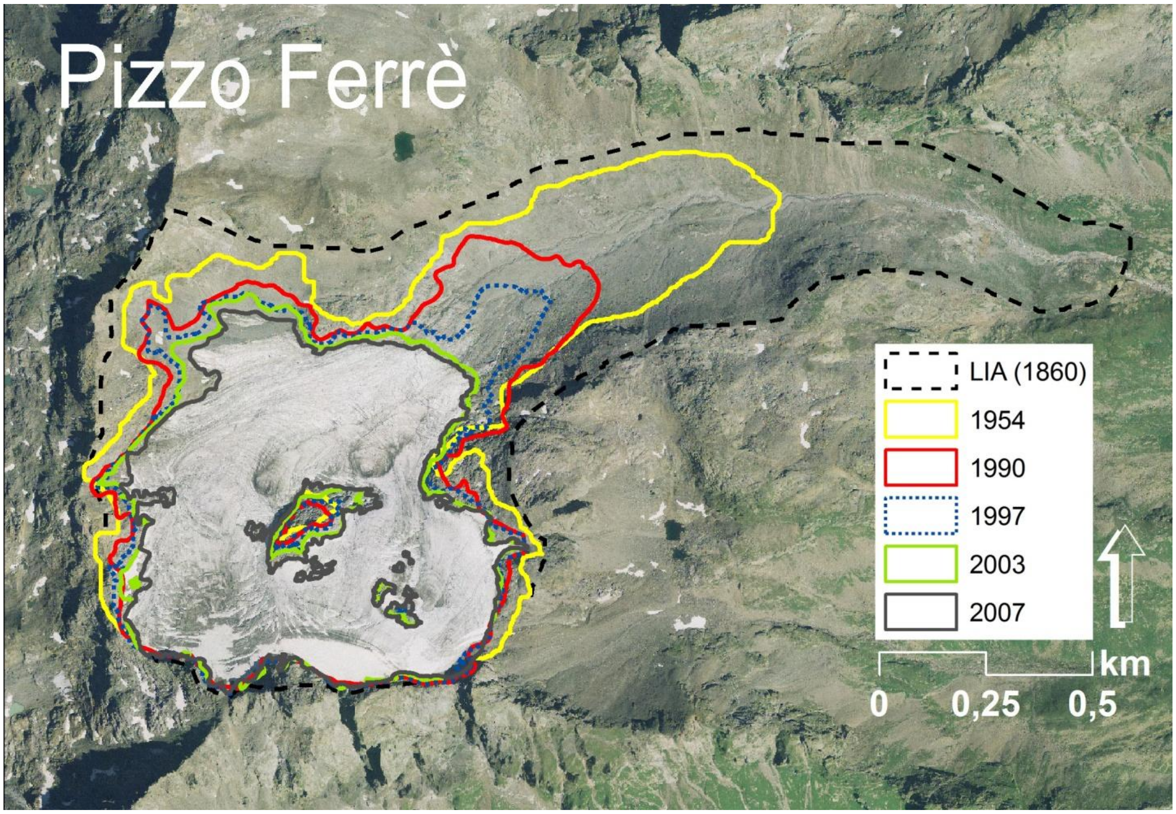
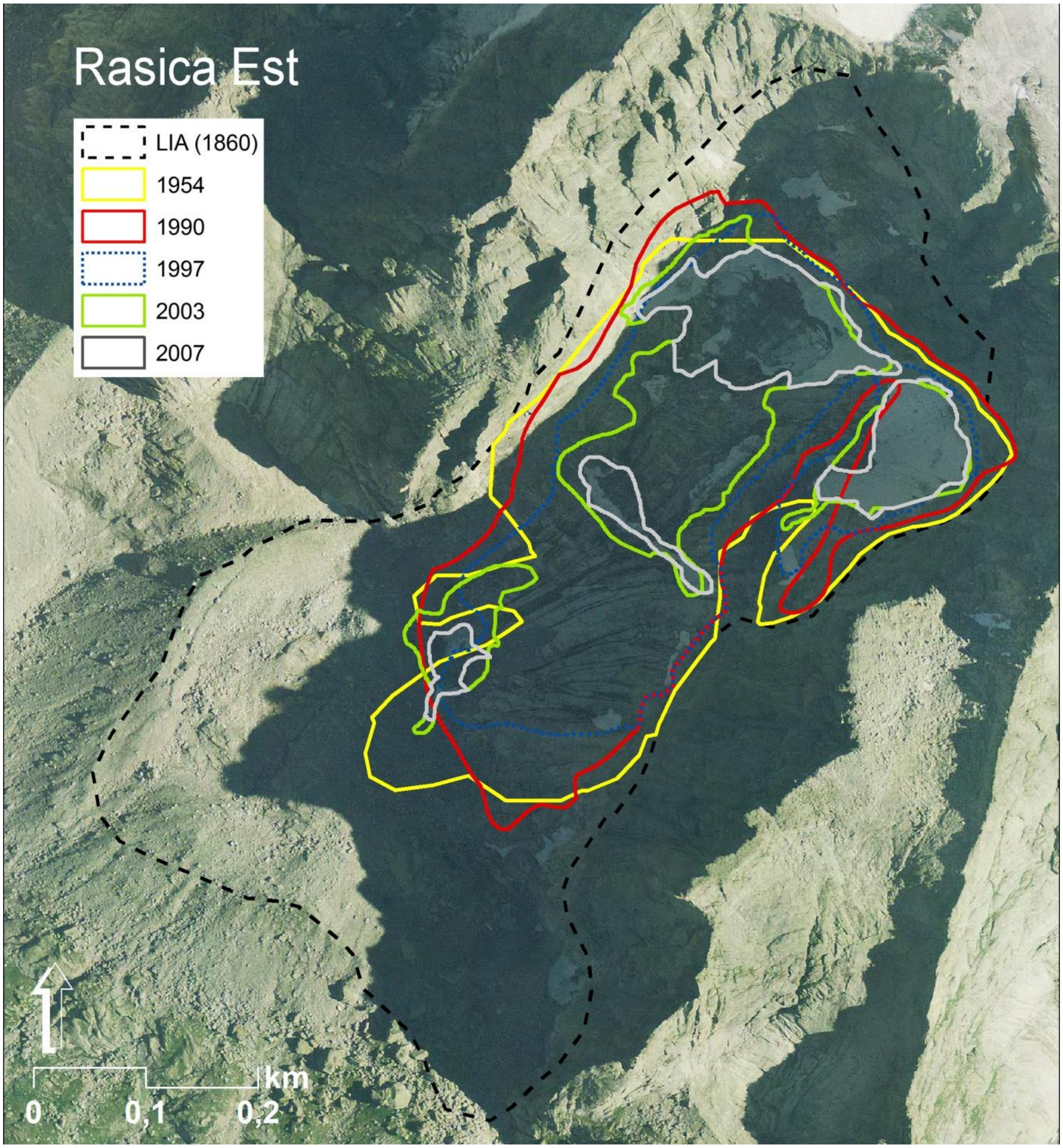
Appendix 1

Pizzo Ferrè glacier (Spluga/Lei) extent during LIA maximum, 1954, 1990, 1997,2003 and 2007. The 2007 orthophoto is set as background image. Approx. scale 1:15,000



Appendix 2

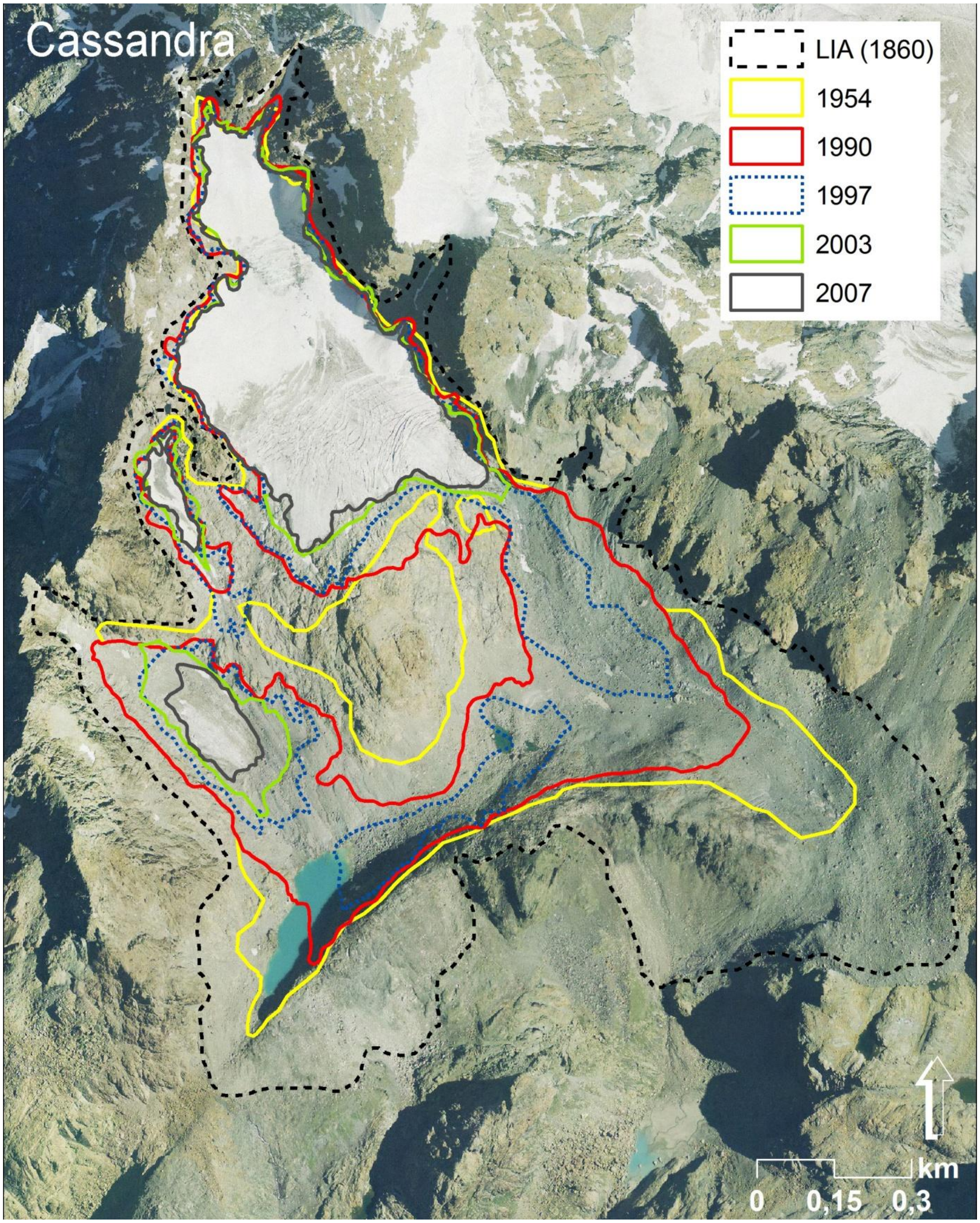
Rasica Est glacier (Codera/Masino) extent during LIA maximum, 1954, 1990, 1997,2003 and 2007. The 2007 orthophoto is set as background image. Approx. scale 1:5,000



Appendix 3

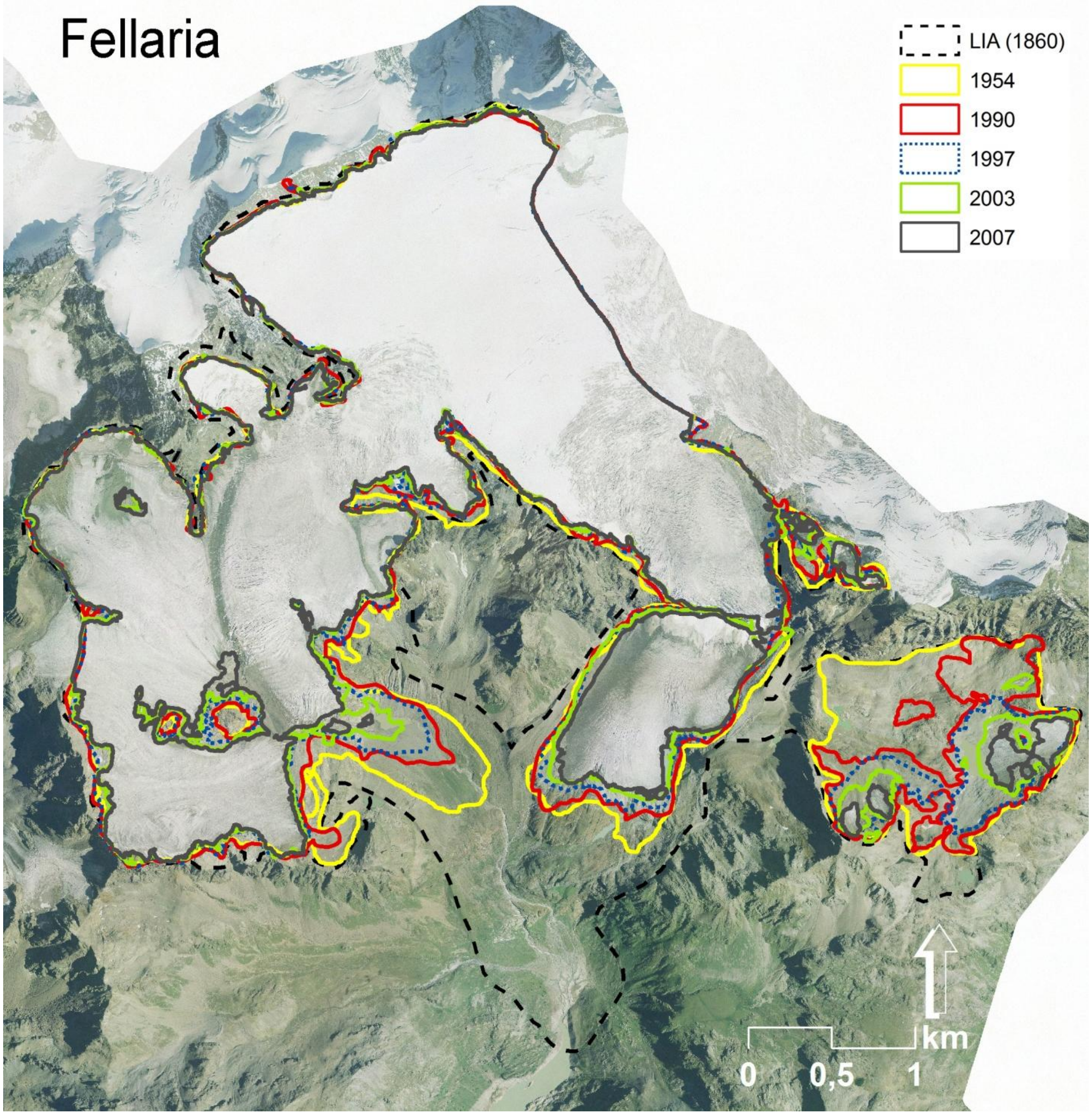
Cassandra glacier (Disgrazia/Mallero) extent during LIA maximum, 1954, 1990, 1997,2003 and 2007.

The 2007 orthophoto is set as background image. Approx. scale 1:10,000



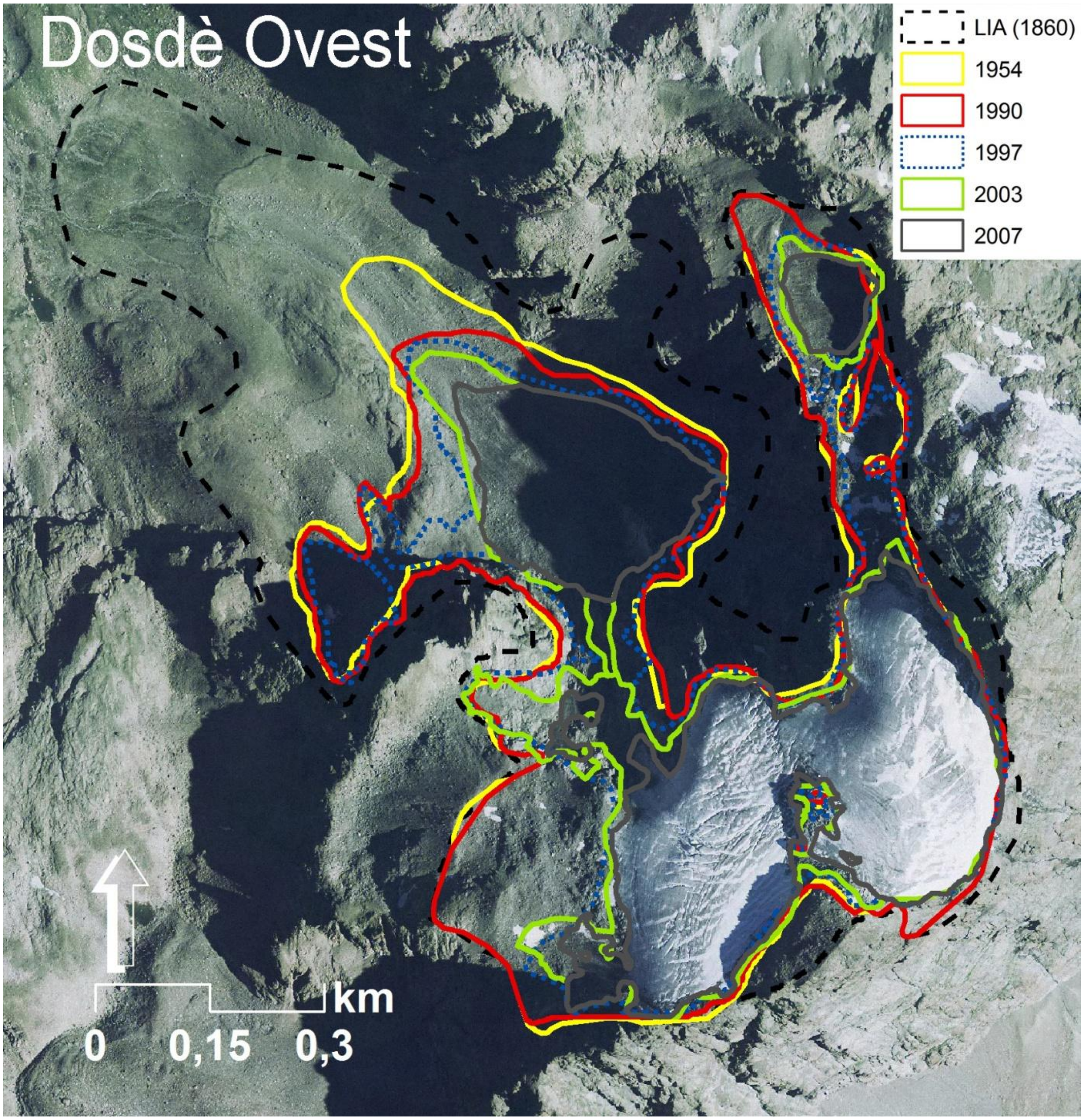
Appendix 4

Fellaria glacier (Bernina/Scalino) extent during LIA maximum, 1954, 1990, 1997,2003 and 2007. The 2007 orthophoto is set as background image. Approx. scale 1:35,000



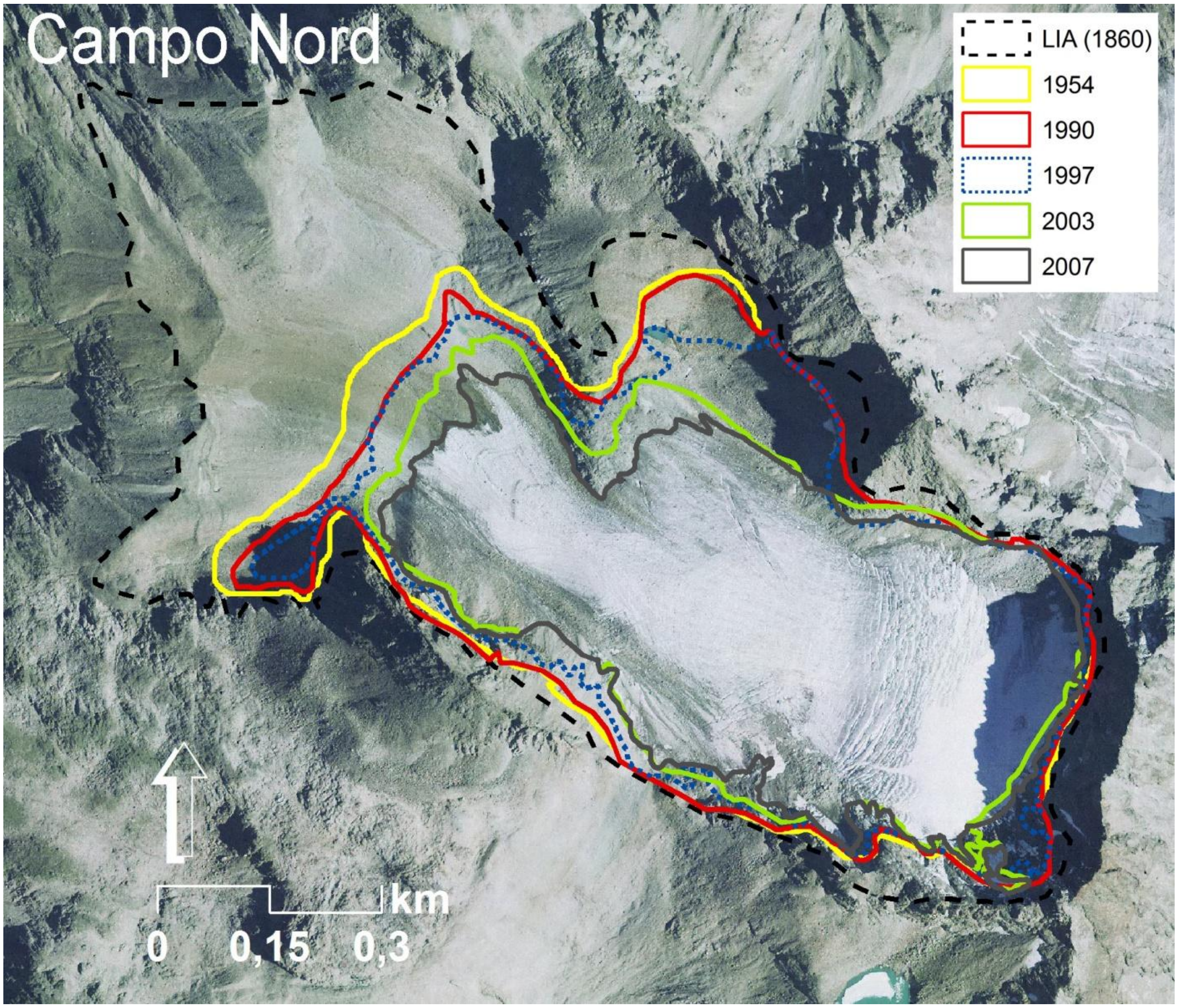
Appendix 5

Dosdè Ovest glacier (Dosdè/Piazzi) extent during LIA maximum, 1954, 1990, 1997,2003 and 2007. The 2007 orthophoto is set as background image. Approx. scale 1:10,000



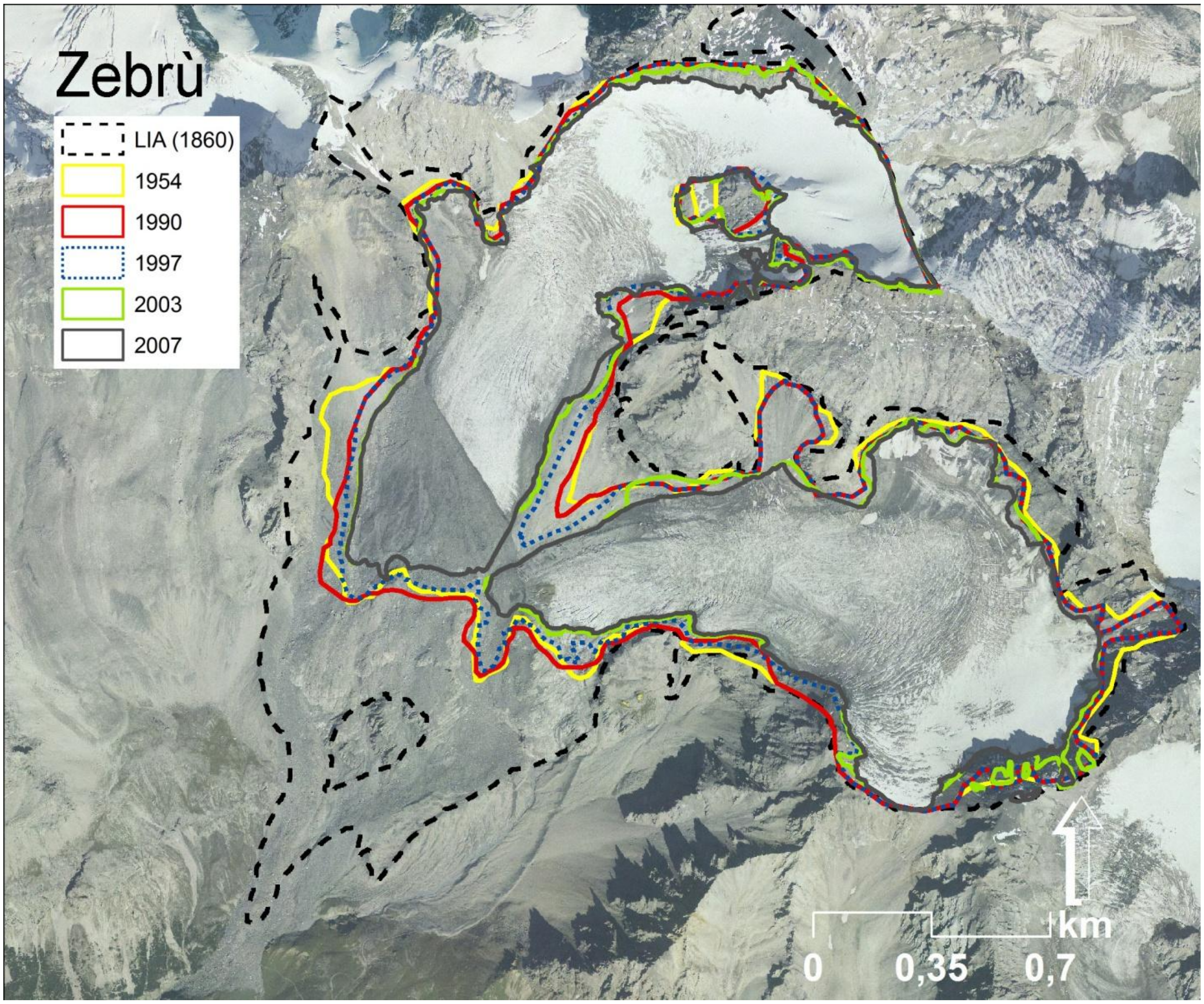
Appendix 6

Campo Nord glacier (Livigno) extent during LIA maximum, 1954, 1990, 1997,2003 and 2007. The 2007 orthophoto is set as background image. Approx. scale 1:10,000



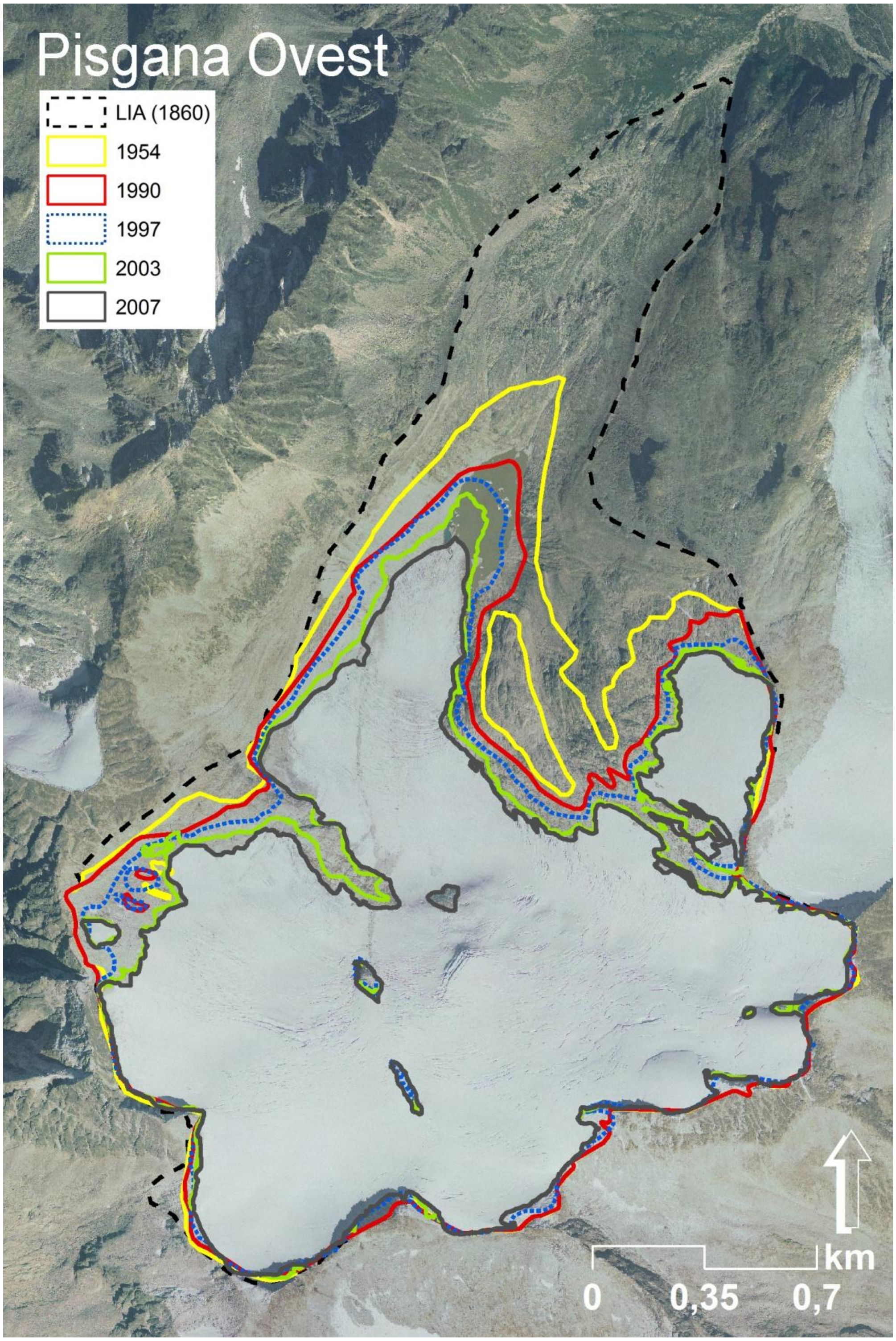
Appendix 7

Zebrù glacier (Ortles/Cevedale) extent during LIA maximum, 1954, 1990, 1997,2003 and 2007. The 2007 orthophoto is set as background image. Approx. scale 1:20,000



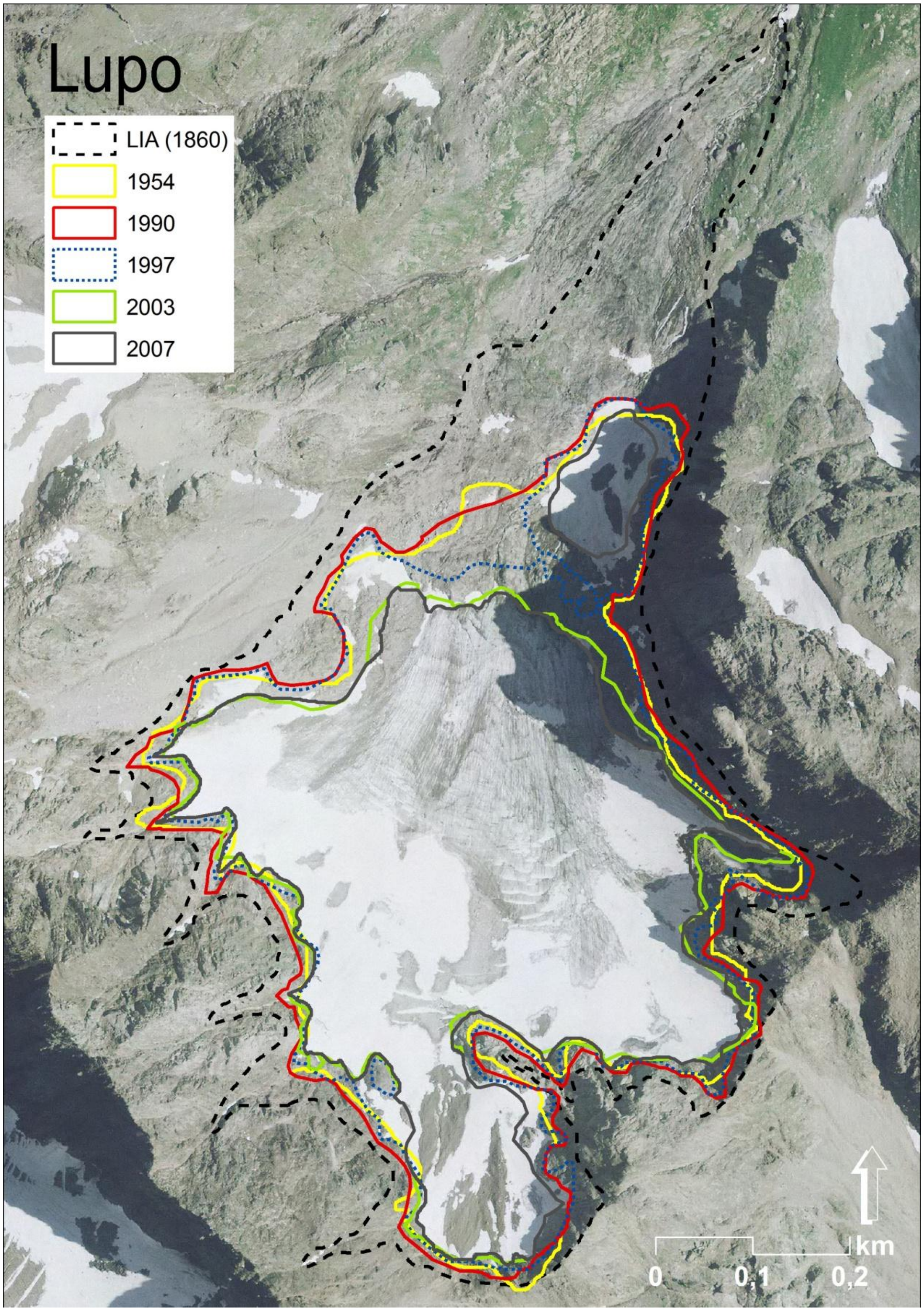
Appendix 8

Pisgana Ovest glacier (Adamello) extent during LIA maximum, 1954, 1990, 1997,2003 and 2007. The 2007 orthophoto is set as background image. Approx. scale 1:20,000



Appendix 9

Lupo glacier (Orobie) extent during LIA maximum, 1954, 1990, 1997,2003 and 2007. The 2007 orthophoto is set as background image. Approx. scale 1:5,000



Acknowledgments

My sincere thanks go to:

Giovanni Crosta, for believing in me and for the encouragement to always find new and interesting ways in research;

Francesco Brardinoni, for keeping me on the rope and for taking care of me even in the most complicated moments. With his great availability he was always an exceptional discussion partner when I had to decide which direction to take and even witch verb to use;

Fabio Villa, for encouraging me to undertake this adventure;

Paolo Frattini, for the essential help in GIS analysis and interesting scientific discussions;

Luca Bonardi, as my spiritual guru and all the **friends of Servizio Glaciologico Lombardo** for giving me pampered and scientifically brought up in the last 10 years;

Massimo Ceriani (Regione Lombardia) for granting access to cartographic information and imagery;

The colleagues of the "dinamica dei versanti" office and department: **Stephen, Serena, el Cobra, Marco, Alberto, Amal, Herr Fredy**, **Elena, Valentina, Nicoletta, Angela, Cinzia, Federico** for helping me in any way and for the great moments lived together;

Stefano Alberti and **Luca Greggio**. You plugged some of my informatics holes and you kept me in good spirit with your enthusiasm;

Roberto Renato Colucci and all the exceptional friends met at the University Centre in Svalbard in those 3 months that changed my life. During that experience I've found excellent professors that hugely improved my skills in glaciology and permafrost: **Doug Benn, Ole Humlum, Anne Christiansen, Tavi Murray, Lindsey Nicholson, Nick Hulton, Ian Rutt, Colin Ballantyne and Hans Hubberten**;

Prof. Kenneth Ewitt for the enormous knowledge he has given me during the 2011 Caucasus expedition;

My **family** who supported me in every passage of my experience and **Francesca**, for having joined on the last part of this way.