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ESTIMATION OF WINTER PRECIPITATION IN A HIGH-ALTITUDE CATCHMENT OF THE EASTERN ITALIAN ALPS: VALIDATION BY MEANS OF GLACIER MASS BALANCE OBSERVATIONS

ABSTRACT: CARTURAN L., DALLA FONTANA G. & BORGA M., *Estimation of winter precipitation in a high-altitude catchment of the Eastern Italian Alps: validation by means of glacier mass balance observations*. (IT ISSN 0391-9838, 2012).

This work analyses the estimation of winter season (October 1 to May 31) precipitation over a high altitude catchment (Val di Peio) of the Eastern Italian Alps. The extrapolation of precipitation over ungauged areas is problematic in this basin due to the combined effect of measurement errors and orographic effects. The study is based on the availability of long term series of weather data, snow observations and glacier mass balance measurements. The error in precipitation measurement at the uppermost weather station was assessed by comparison with snow water equivalent data. The error estimates were consistent with the outcomes of a precipitation correction model, with the aerodynamic correction as its main component. This correction procedure was used to compute correction factors for the entire precipitation gauge network. Both corrected and uncorrected precipitation data were used to estimate the spatial distribution of precipitation over the study area by means of a technique that accounts for the precipitation-altitude relationship. Winter balance observations over Careser glacier (the longest mass balance series in Italy), which is within the study area, were used to assess the improvement given by the precipitation correction procedure, showing a reduction of bias from -38% to –2.5% and a reduction of the RMSE from 410 to 171 mm water equivalent. The use of corrected precipitation data led to a 31.7% (182 mm) increase of the winter season basin-averaged precipitation, whereas the winter season average vertical gradient increased from 4.0% km^{-1} to 21.9% km^{-1}. Overall, these results highlight the need for precipitation correction in precipitation analyses over snow-dominated mountain areas. This paper also provides evidence of a considerable interannual variability in the correction factors for snow at the uppermost weather station. Further improvements in precipitation estimations would require an analysis of the dominant processes controlling this variability.

KEY WORDS: Precipitation measurement, Precipitation estimation, Alps, Snowfall correction factor, Precipitation-altitude relationship, Glacier mass balance.

RIASSUNTO: CARTURAN L., DALLA FONTANA G. & BORGA M., *Stima della precipitazione invernale su un bacino di alta quota delle Alpi Orientali Italiane: validazione per mezzo di misure di bilancio di massa glaciale*. (IT ISSN 0391-9838, 2012).

Questo lavoro analizza il problema della corretta quantificazione delle precipitazioni invernali (1 ottobre-31 maggio) sul bacino della Val di Peio, nelle Alpi Orientali Italiane. La situazione è rappresentativa dei bacini alpini, dove l'estrapolazione delle precipitazioni sulle aree in quota prive di pluviometri è problematica, a causa dell'effetto combinato degli errori di misura e dell'orografia. Lo studio si basa sulla disponibilità di lunghe serie di dati meteorologici, nivologici e di bilancio di massa glaciale. L'errore di misura delle precipitazioni presso la stazione meteorologica più elevata è stato quantificato per mezzo di dati di equivalente in acqua della neve fresca. L'errore di misura è risultato confrontabile con le stime ottenute mediante un modello di correzione delle precipitazioni su base aerodinamica. Il modello è stato utilizzato per il calcolo dei fattori di correzione presso tutte le stazioni termo-pluviometriche utilizzate nello studio. Si è proceduto quindi alla spazializzazione dei dati di precipitazione grezza e corretta, mediante una tecnica che tiene in considerazione la relazione tra quota e precipitazione. La serie di bilanci invernali misurata sul ghiacciaio del Careser (la più lunga in Italia), che si trova all'interno dell'area di studio, è stata utilizzata per valutare il miglioramento ottenibile tramite la correzione delle precipitazioni, quantificabile in una riduzione dell'errore medio da -38% a –2.5% e in una riduzione del RMSE da 410 a 171 mm di equivalente in acqua. La correzione delle precipitazioni ha portato ad un aumento del 31.7% (182 mm) della precipitazione media nell'area di studio, mentre il gradiente delle precipitazioni è aumentato da 4.0% km^{-1} a 21.9% km^{-1}, corrispondenti rispettivamente a 23.4 e 174.5 mm km^{-1}. Questi risultati mettono in evidenza l'importanza della correzione delle precipitazioni su aree montane in cui domina l'idrologia nivale. Un altro aspetto degno di nota è la variabilità inter-annuale dei fattori di correzione per la neve presso la stazione meteorologica più elevata. Ulteriori miglioramenti nella stima delle precipitazioni richiedono quindi un'analisi dei processi che controllano questa variabilità temporale.

TERMINI CHIAVE: Misurazione delle precipitazioni, Stima delle precipitazioni, Alpi, Fattore di correzione delle precipitazioni nevose, Relazione quota-precipitazioni, Bilancio di massa glaciale.

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INTRODUCTION

The estimation of precipitation inputs over high-altitude ungauged areas is a key issue for many fields investigating mountainous catchments. Alpine areas play a specific role in the hydrological processes of the planet and the regional hydrology of all continents, and they influence the hydrological processes of much larger downstream basins (Viviroli & *alii*, 2007; Norbiato & *alii*, 2009). Thus, a correct estimation of precipitation is an essential prerequisite for the assessment of runoff timing and amount from high mountain areas.

Distributed models of snow and ice mass balance are useful for: i) understanding the processes involved in glacier hydrology, ii) predicting glacier runoff under possible future climatic scenarios, iii) interconnecting different levels of observations and iv) extrapolating in space and time (Hock & Jansson, 2005, Machguth & *alii*, 2006a). Major improvements in glacier mass balance modelling can be achieved by focusing on accumulation processes (Machguth & *alii*, 2006b; Paul & *alii*, 2008 and 2009). Besides redistribution processes, precipitation extrapolation from gauge data to glacierized areas remains problematic in the absence of direct snow water equivalent observations (e.g., Giada & Zanon, 1985; Klok & Oerlemans, 2002; Ranzi & *alii*, 2010). Permafrost and periglacial processes are also affected by precipitation distribution. In particular, the formation time of a stable snowpack in fall and its depletion in spring and summer, together with its depth, strongly affect the ground thermal regime and the current and past distribution of permafrost (Barsch, 1996; Baroni & *alii*, 2004; Luetschg & *alii*, 2008). Precipitation is also strongly related to hazards (e.g. landslides, debris flows and floods), to ecological processes (dynamics of plants and animal species), and to the touristic appeal of mountain areas (Speranza & *alii*, 1996; Orlove & *alii*, 2008; Surdeanu & *alii*, 2009; Beniston, 2010; Borga & *alii*, 2010).

Given the limitations of weather radar and satellite remote sensing of precipitation in mountainous areas (Borga & *alii*, 2000; Nikolopoulos & *alii*, 2010), data from precipitation gauges remain the main source of information for quantitative precipitation estimation. However, precipitation estimation in mountainous high altitude terrain from gauge data is made difficult by the following: i) the sparseness of the measuring stations and the prevailing distribution of weather stations at the bottoms of the valleys, ii) the increasing measurement error with the increasing fraction of solid precipitation (Schwarb, 2000; Molnar & Trizna, 1989) and iii) the impact of orography on precipitation patterns (Schwarb, 2000).

Systematic measurement errors at precipitation gauges mainly result from the influence of wind. Under high wind speeds, which are common in mountainous settings, precipitation gauges disrupt the boundary-layer atmospheric flow, causing frozen precipitation to preferentially blow over and around, rather than into, the gauge (Goodison & *alii*, 1998). Liquid precipitation is less susceptible to this undercatch problem because it is denser and has a higher falling velocity. Estimates of the snowfall undercatch for some types of gauges are as high as 70% or more (Yang & *alii*, 1998, 2000). Other systematic errors in measuring precipitation include evaporation/sublimation, wetting losses from water sticking inside the gauge, blowing snow, and the tendency of observers to ignore trace events (Sevruk, 1982; Legates, 1987; Hood & *alii*, 1999). All of these systematic biases lead to the underestimation of precipitation, with the exception of blowing snow, which can be recorded as «false» precipitation. Adam & Lettenmaier (2003) provided a useful summary of the relative magnitude of these systematic errors.

Systematic errors in precipitation measurements may strongly affect areal precipitation estimation in high mountainous basins, particularly when the estimation method incorporates precipitation-altitude relationships to describe the effect of the orography, such as the PRISM method proposed by Daly & *alii*, (1994) and the various techniques using elevation as auxiliary data (e.g., Saghafian & Bondarabadi, 2008). It is generally accepted that altitude is the main variable governing the spatial distribution of precipitation in the mountains. The reason, in principle, is the decreasing temperature and increasing condensation with altitude on windward slopes. The precipitation-altitude relationship is the basic assumption on which nearly all regionalization and mapping techniques of precipitation in mountainous areas are based. Because higher altitude gauges are likely to experience higher percentages of snowfall and higher wind speeds, positive vertical gradients of precipitation may be masked by precipitation measurement biases. Therefore, systematic measurement errors influence both the local measurements and the accuracy of the spatial extrapolation at higher altitudes. Thus, a gauge-dependent bias correction of conventional observed precipitation is essential before using these measurements for mean-areal precipitation estimation and water balance assessments.

To assess the national methods of solid precipitation observations, the World Meteorological Organization (WMO) initiated the Solid Precipitation Measurement Intercomparison Project in 1985. This project developed bias correction techniques for a number of precipitation gauges commonly used around the world (Goodison & *alii*, 1998). However, it may be difficult to generalise these bias-correction techniques to precipitation gauges not included in the project. Sevruk (1985) developed a methodology for the correction of the Hellmann gauge measurement errors in Switzerland. The methodology is based on routine observations of wind speed and air temperature, and it takes into account the wind exposure of the gauge sites by means of the average vertical angle of the surrounding obstacles. This correction procedure, developed for gauges located below 2000 m, was applied to correct the raw precipitation data in the Hydrological Atlas of Switzerland (Spreafico & *alii*, 1992).

Glacier mass balance observations have often been used both to surrogate precipitation measurements and to provide a check for precipitation estimates at high altitude (Bucher & *alii*, 2004; Gerbaux & *alii*, 2005; Tangborn, 1999). However, winter mass balance measurements on glaciers are not always easy to interpret in terms of precipitation because of the effect of snow redistribution and the influence of evaporation and sublimation from the surface of the glacier. Thus, measurements should come from areas that are affected little by redistribution processes, and evaporation and sublimation losses must be taken into account in comparisons with precipitation estimates.

This study focuses on winter (October 1 to May 31) precipitation estimation in the Val di Peio catchment (173 km^2, Eastern Italian Alps). The aims of the study were the following: i) to provide an assessment of snowfall measurement errors for a high altitude precipitation gauge located in Val di Peio by comparing snow water equivalent and precipitation measurements; ii) to correct the rain gauge errors of a network of weather stations in the study area; iii) to evaluate the impact of bias correction on areal precipitation estimates; and iv) to use a long time series of winter mass balance observations on Careser glacier to validate the precipitation estimates provided by bias-corrected data. The study was based on the availability of long time series of observations (weather, snow and glacier mass balance), which are unique for the southern European Alps.

STUDY SITE AND DATA

Site description

The study area is the hydrological basin of Val di Peio (173 km^2), located in the Eastern Italian Alps (fig. 1). The basin is drained by the Noce River, a right-hand tributary of the Adige River, which is the main river system in north-eastern Italy. The altitude of the Peio basin averages 2365 m a.s.l and ranges from 952 m a.s.l. at the outlet to 3769 m a.s.l. at Mt. Cevedale. The watershed is bordered to the north by the Ortles-Cevedale massif, with altitudes up to 3700-3800 m a.s.l., and to the south by the Noce River axis and the Adamello-Presanella massif, with peaks around 3600 m a.s.l.. The distribution of the area with altitude (fig. 2) shows that 72% of the basin is located above 2000 m a.s.l.. The basin is located within the transition zone between the main inner Alpine dry zone to the north-east (Val Venosta) and the wetter mountainous ranges of Adamello and Orobie to the south-west. The mean annual precipitation at the Peio station (1565 m a.s.l.) amounts to 920 mm (value uncorrected for precipitation measurement errors). The annual precipitation regime shows a minimum in winter, a principal maximum in autumn and a secondary maximum in spring (fig. 3).

The basin includes a wide glacierized area (12.1 km^2, 7% of the total surface), which significantly affects the annual runoff regime and includes the Careser glacier, where the winter mass balances used in this study have been measured since 1967. The glacier is located in a south-facing cirque, and it covers an area of 2.40 km^2 (surveyed in late summer 2006); it ranges in altitude from 2869 m to 3279 m a.s.l., and its mean altitude is 3057 m a.s.l.. Meltwater from the glacier feeds the Careser dam reservoir (2600 m a.s.l.), which is used for hydropower generation.

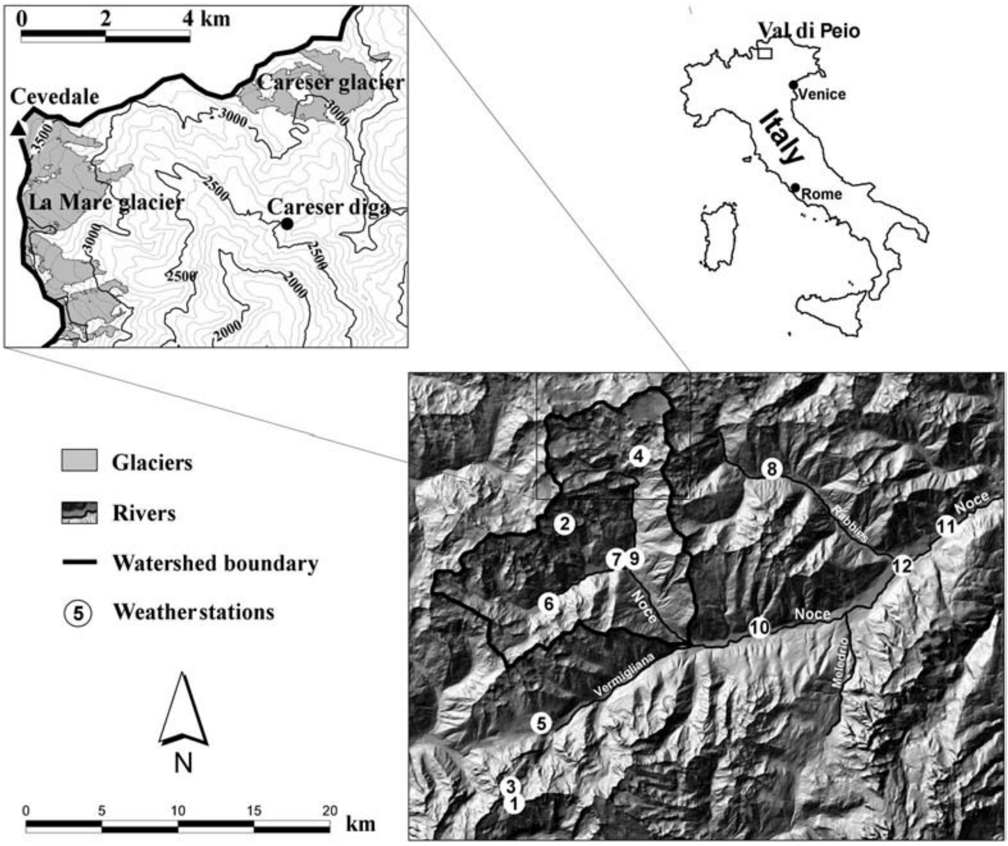

FIG. 1 - Geographical setting of study area. The numbers refer to the weather stations listed in tables 1 and 2.

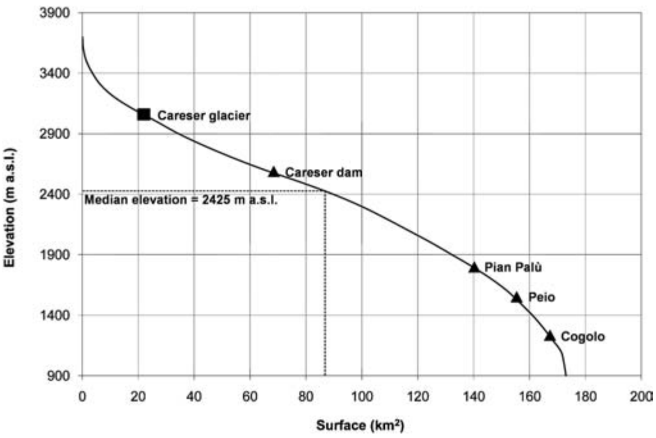

FIG. 2 - Hypsometric curve of the Val di Peio watershed.

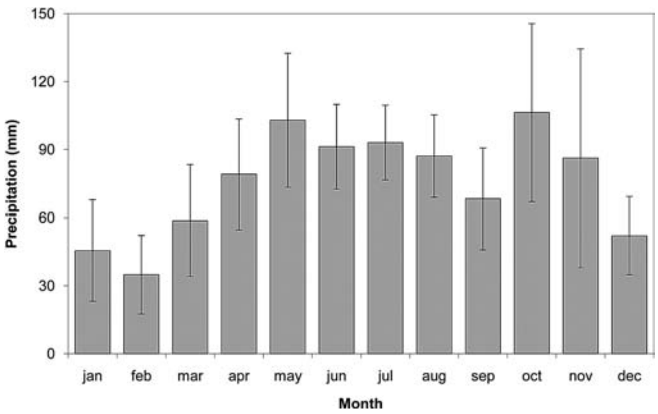

FIG. 3 - Average monthly precipitation at the Peio weather station for the period 1978-2005. The bars indicate ± 1/2 standard deviation.

Data collection

The observations used in this study (table 1) include: i) daily data of wind speed, precipitation and temperature over the period 1978-2005; ii) snow observations over the period 1960 to 2006, including snow water equivalent data collected close to the highest weather station (Careser dam, 2605 m a.s.l.); and iii) winter mass balance data at the Careser glacier for the period 1967 to 2005. The meteorological data were collected at twelve weather stations (fig. 1 and table 1) with altitudes ranging from 735 m (Malè) to 3015 m a.s.l. (Cima Presena). These stations belong to the regional observation network of the Trentino meteorological service. We included weather stations located outside the study basin to improve the estimation of precipitation fields and wind speed. Daily data of temperature and precipitation were available for the period between January 1978 and September 2005, while wind speed data were only available from automatic instruments operated in the period 2001-2009.

Daily measurements of fresh snow depth were regularly carried out over the period 1978-2006 by the personnel

TABLE 1 - Location of the weather stations, meteorological variables used in the analyses (T = air temperature; P = precipitation; W = wind speed; HN = fresh snow; HS = snow depth) and average values in the winter season (raw precipitation data)

Measuring station	Altitude (m a.s.l.)	Location (LAT-LONG, DD)	Variables (sensor height)	Average (Oct-May)
1 - Cima Presena	3015	46.22 - 10.59	W (3 m)	3.6 m s^{-1}
2 - Vioz	2950	46.38 - 10.63	W (5 m)	2.7 m s^{-1}
3 - Capanna Presena	2730	46.22 - 10.58	W (10 m)	2.0 m s^{-1}
4 - Careser dam	2605	46.42 - 10.70	T (2 m)	–3.9 °C
			P (2 m)	545 mm
			W (2 m)	1.9 m s^{-1}
			HN	708 cm
			HS	86 cm
5 - Tonale	1795	46.30 - 10.62	T (2 m)	0.2 °C
			P (2 m)	826 mm
			W (10 m)	2.9 m s^{-1}
6 - Pian Palù	1790	46.34 - 10.62	T (2 m)	–1.4 °C
			P (2 m)	568 mm
7 - Peio	1565	46.36 - 10.68	T (2 m)	3.0 °C
			P (2 m)	553 mm
8 - Rabbi	1350	46.41 - 10.80	T (2 m)	2.8 °C
			P (2 m)	570 mm
			W (5 m)	0.9 m s^{-1}
9 - Cogolo	1200	46.36 - 10.69	T (2 m)	3.9 °C
			P (2 m)	493 mm
10 - Mezzana	935	46.32 - 10.80	T (2 m)	5.4 °C
			P (2 m)	529 mm
			W (10 m)	1.2 m s^{-1}
11 - Caldes	773	46.38 - 10.96	W (10 m)	1.3 m s^{-1}
12 - Malè	735	46.36 - 10.92	T (2 m)	5.6 °C
			P (2 m)	561 mm

of Careser dam (2605 m a.s.l.), using a snowboard. These measurements were used in this work to estimate the precipitation bias at this station, which is the highest site equipped with a rain gauge. To convert the fresh snow depths into heights of water equivalent, we carried out specific field campaigns of several days each during the period 2003-2006. The snow depth was measured by means of a snow probe near to the snowboard, and the snow density was sampled along the vertical profile of the snowpack to calculate its water equivalent.

Mass balance measurements on Careser glacier have been carried out since 1967, and they constitute the longest series for the Italian Alps (Zanon, 1992; Carturan & Seppi, 2007). The «direct glaciological method» was applied (Kaser & *alii*, 2003), which consists of *in situ* measurements of surface level changes at individual points between two dates. The level changes were then multiplied by the near-surface density to calculate the mass gains or losses, and point information was then extended to the entire glacier surface by spatial interpolation. The accumulation season on Careser glacier normally lasts from the beginning of October to the end of May. The winter balance was computed annually at the end of the accumulation season by sampling the water equivalent of snow with graduated probes and density measurements in snow pits. The time series of the Careser glacier mean winter balances is reported in figure 4.

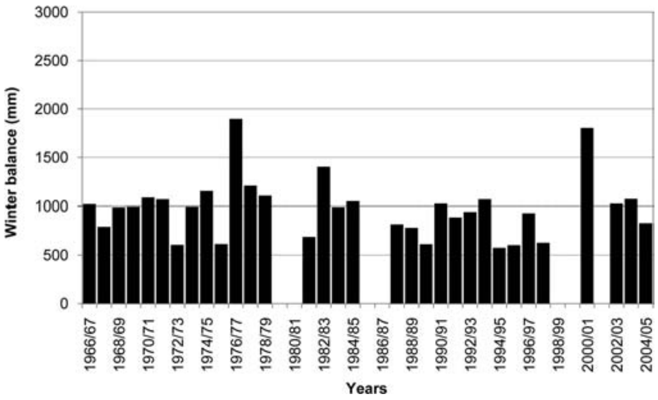

FIG. 4 - Careser glacier mean winter balance time series from 1967 to 2005 (Carturan and Seppi, 2007).

METHODS

Processing of meteorological data

Specific effort was given to collecting metadata for each station, including information on station moves, changes in instrumentation, changes in surrounding environmental characteristics and structures, and observation practices. This effort was essential to checking for errors and providing good quality data.

The data and metadata were inspected to identify potential measurement problems. In the period between 1985 and 1993, the manual precipitation and temperature stations were replaced by new automatic instruments, which introduced changes in the frequency of observations and calculation methods. Two stations, Rabbi and Pian Palù, were relocated. The frequency of observations changed from daily to hourly. Manual precipitation gauges were replaced by heated tipping-bucket gauges (model MTX-PP041) with the same shape and orifice area (1000 cm^2). At Careser dam, before the installation of the automatic gauge, solid precipitation was measured by melting the snow accumulated into the gauge. To do this, the collector was temporarily taken inside the personnel housing. Wind shields were never applied to the precipitation gauges. The mercury minimum/maximum thermometers, previously placed inside a Stevenson's screen, were replaced by digital sensors housed in smaller shields with natural ventilation (model MTX-TU019). We adjusted the temperature time series for inhomogeneities, which were identified by combining the available metadata and statistical analyses (Peterson & *alii*, 1998; Aguilar & *alii*, 2003). According to the indications of homogeneity tests, adjustments were not required for the precipitation time series. Gap filling techniques were applied to the precipitation data based on comparisons with other gap-free stations in the network (Carturan, 2010).

The measurements of snow water equivalent carried out from 2003 to 2006 at Careser dam were used to identify a suitable relationship between daily fresh snow density and air temperature. The experimental data were first compared with the relationships proposed by Pahaut (1975), Hedstrom & Pomeroy (1998), and La Chapelle (1961). The following relationship, which is a modified version of that proposed by La Chapelle (1961), was found to provide the best fit to the data (r = 0.62, RMSE = 27.8 kg m^{-3}):

$$\begin{aligned} \rho(T) &= 50 & T &\leq -15°C \\ \rho(T) &= 6.67T + 150 & T &> -15°C \end{aligned} \tag{1}$$

where T (°C) is the mean daily air temperature, and ρ(kg m^{-3}) is the fresh snow density. This relationship was then used to calculate the water equivalent of the daily fresh snow at Careser dam for the entire study period.

It was necessary to determine the phase of precipitation to apply the appropriate precipitation correction factors for solid or liquid precipitation. Because the correction factors for liquid and solid precipitation differ significantly, the assessment of the phase of precipitation strongly impacts the quality of the precipitation estimates. Routine observations of the prevailing phase of precipitation (rainfall, snowfall, or mixed) were available at Careser dam between 1975 and 1993 for 2231 precipitation days, which allowed a close inspection of this step. Figure 5 shows the relative frequency of the three precipitation types for temperature classes. Rain was almost never recorded with a daily mean temperature below –3.5°C, and snow was almost never observed with temperatures above 7°C. At 0°C, the frequency of solid precipitation matched the frequency of liquid+mixed precipitation, while at +2°C, the frequency of liquid precipitation matched the frequency of solid+mixed precipitation. The solid fraction of daily precipitation was calculated in function of the mean daily temperature (T) by assuming that the «mixed» fraction was composed of snow and rain in equal parts:

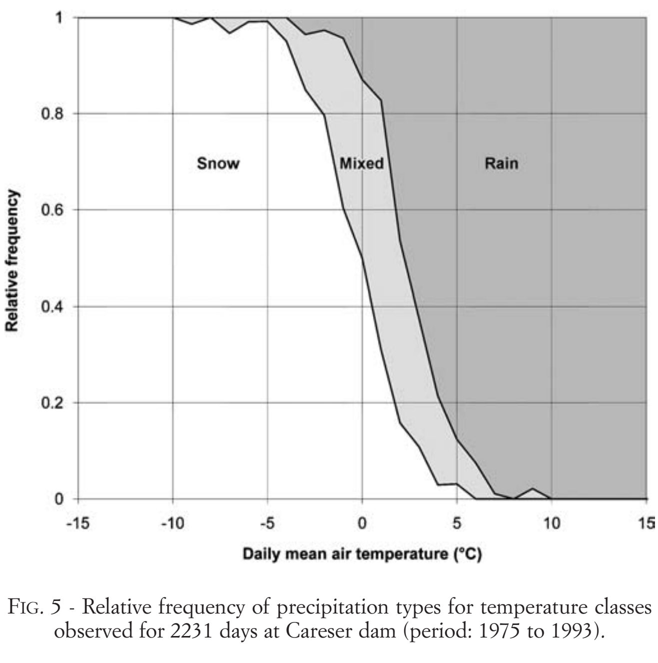

FIG. 5 - Relative frequency of precipitation types for temperature classes observed for 2231 days at Careser dam (period: 1975 to 1993).

$$P_{solid} = P_s + 0.5 \cdot P_m \quad \text{and} \quad P_{liquid} = 1 - P_{solid} \tag{2}$$

where P_{solid} and P_{liquid} are the proportion of solid and liquid precipitation, respectively. P_s and P_m are the fractions of snow and mixed precipitation, calculated by a third order polynomial fitted to the two frequency distributions shown in figure 5:

$$P_s = 1.4 \cdot 10^{-3} \cdot T^3 + 1.5 \cdot 10^{-3} \cdot T^2 - 0.135 \cdot T + 0.466 \tag{3}$$

$$P_m = (2.2 \cdot 10^{-3} \cdot T^3 - 15.3 \cdot 10^{-3} \cdot T^2 - 0.122 \cdot T + 0.852) - P_s \tag{4}$$

This methodology was used over the entire observation network.

Mean monthly wind speed estimates were required for calculating the precipitation bias at the rain gauge locations. An assessment of the average wind speed distribution in the study area was performed with the assumption that the available data were representative of the entire study period, which extends back to 1978. The wind speed was calculated at the level of rain gauge orifices (2 m) by means of the logarithmic wind profile equation, which includes the effects of roughness and nearby obstacles (Sevruk & Zahlavova, 1994):

$$U_{h\alpha} = U_H \left(\log \frac{h}{z_0} \right) \left(\log \frac{H}{z_0} \right)^{-1} \left(1 - 0.024\overline{\alpha}\right) \tag{5}$$

where U_H (m s^{-1}) is the wind speed at the level H (m) of the measuring instrument, h is the level of the gauge orifice above ground (m), z_0 (m) is the roughness length, and α (in degrees) is the average vertical angle of obstacles. The roughness length was estimated according to Sevruk & Zahlavova (1994). Because no observations of snow cover (surface conditions, duration and depth) were available for most weather stations, we assumed fixed z_0 values of 0.03 m for weather stations located on meadow and 0.0125 m for high-altitude weather stations, where snow lasts most of the year. For the same reason, fixed values were assumed for H and h. After direct inspection of the measuring sites, the gauge site exposure class was assessed to be «open» (average $\alpha = 2.5°$) for all weather stations.

Based on the analysis of wind data, two distinct annual wind speed regimes were identified. High altitude stations (above 2000 m a.s.l.) exhibit higher wind speed, higher variability and a winter maximum, while an inverse behaviour was found for the stations near the valley bottoms, which show lower wind speed and a summer maximum (fig. 6). At lower altitudes, the greater roughness (forests and built-up areas) and the topographic barriers reduce the wind speed. The seasonal behaviour is likely due to the location of the measuring sites, the unequal heating of the valley slopes and the coupling between surface winds and the overlying synoptic winds (Barry, 1992; Whiteman & Doran, 1993; Weber & Furger, 2001). The monthly average wind speed calculated for the two groups of weather stations (high altitude and valley bottoms) was extrapolated to the four weather stations without anemometers by considering their altitude.

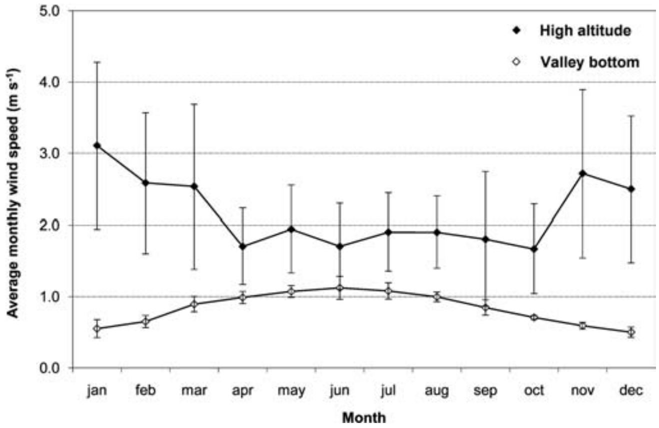

FIG. 6 - Average monthly wind speed regime for high-altitude (above 2000 m a.s.l.) and valley-bottom (below 2000 m a.s.l.) stations. The bars indicate ±1 standard deviation.

Correction of precipitation measurements

The correction of raw precipitation data is a key issue for the estimation of winter precipitation at high altitudes. In our study area, the gauging error was expected to be large at Careser dam due to the high proportion of solid precipitation in the winter season and the high wind speed. In addition, the Careser dam precipitation gauge is the only one located above 2000 m a.s.l., making it crucial for the extrapolation of precipitation data in ungauged high-altitude areas.

We adapted and implemented the correction procedure proposed by Sevruk (1985) (referred to as Sevruk 1985, hereinafter), which was used for the Swiss Hydrological Atlas (Spreafico & *alii*, 1992). The Sevruk 1985 method calculates correction factors for snow and rain and is based on simplified physical concepts. The main variables used in the correction procedure are monthly wind speed, temperature, the fraction of snow in the total precipitation, the number of days with precipitation and the degree of gauge site exposure. This technique was well adapted to the data available in our study site, and it was developed in a Swiss geographic area near Val di Peio. However, this technique's applicability to our rain gauges and at high altitudes (above 2000 m a.s.l.) had to be assessed.

According to Sevruk1985, the corrected amount of precipitation P_k is given by:

$$P_k = k \left(P_g + \Delta P_{wl}\right) \tag{6}$$

where k is the correction factor due to the wind field deformation above the gauge orifice, P_g is the measured amount of precipitation and ΔP_{wl} are the wetting losses. The wetting losses depend on the ratio between the surface of the inner walls of the gauge collector and the orifice area (S_w/S_o). The rain gauges used in the study area are characterized by a ratio of $S_w/S_o = 2.3$; hence, these losses are negligible according to Sevruk & Klemm (1989). Other losses due to evaporation, splashing or drifting snow were not accounted for in the computations. The value of parameter k depends on monthly wind speed, temperature, the fraction of snow in the total precipitation, the number of days with precipitation and the degree of gauge site exposure. Based on these data, it was possible to obtain the monthly values of correction factors for rain and snow, indicated with K_{rain} and K_{snow}, respectively, in the following sections.

For each rain gauge station, we applied the Sevruk1985 procedure in three steps as follows: 1) the calculation of monthly correction factors for snow and rain (K_{snow} and K_{rain}); 2) the computation of raw monthly precipitation divided into solid and liquid fractions based on daily temperature data; 3) the correction of monthly precipitation data.

The suitability of this correction procedure for the gauges used in our study area was assessed by comparing the K_{snow} calculated at Careser dam to a K_{snow_we} (mean winter season precipitation correction factor for snow) derived from comparisons of raw gauge data with snow water equivalent observations at the same site, as follows:

$$K_{snow\_we} = \frac{\sum_{i=1}^{n} we_i}{\sum_{i=1}^{n} P_i} \tag{7}$$

where n is the number of days with solid precipitation in the winter season, we_i is the i-th daily fresh snow water equivalent (calculated from the fresh snow depth by means of the snow density based on Eq. 1), and P_i is the daily solid precipitation measured by the gauge.

Estimation of winter precipitation fields

Monthly and seasonal precipitation fields were estimated for the Val di Peio basin in the period 1978-2005. Various approaches have been reported in the literature to estimate the precipitation distribution from gauge observations in regions with strong orographic control. These approaches include: 1) those that consider orographic and/or atmospheric effects on precipitation occurrence, such as regression (Daly & *alii*, 1994; Michaud & *alii*, 1995; Goovaerts 2000; Drogue & *alii,* 2002); and 2) those that consider both spatial covariance and terrain and/or climatic conditions, such as cokriging precipitation with terrain elevation (Hevesi & *alii*, 1992; Phillips & *alii*, 1992; Goovaerts 2000) or detrended residual kriging (Phillips & *alii*, 1992; Goovaerts 2000; Kyriakidis & *alii*, 2001; Guan & *alii*, 2005; Moral, 2010). Terrain elevation is the most commonly used secondary variable incorporated in the estimation of precipitation based on the physics of orographic effects. However, Goovaerts (2000) also found that the benefits of methods incorporating terrain elevation depend on the correlation coefficient between precipitation and elevation. The author proposed a threshold correlation coefficient of 0.75 for useful precipitation-elevation cokriging. Asli & Marcotte (1995) also reported that the introduction of secondary information in estimation is worthwhile only for correlation coefficients above 0.4, which would restrict the usefulness of the terrain elevation in cases where the correlation coefficient is low. As reported by Carturan (2010), the correlation coefficient between monthly corrected precipitation and elevation is larger than 0.8 in the study area.

Based on this background, we used the detrended residual kriging spatial estimation technique proposed by Gottardi & *alii* (2007) for areas with strong orographic control. This procedure accounts for both the vertical gradient and the horizontal variability of precipitation. The procedure uses, as a measure of the trend, a weighted linear regression of precipitation with altitude. The orography of the study area was described using a Digital Terrain Model with a grid size of 40 m. The weighted linear regression was computed for each grid cell as follows:

$$PE_{ij} = a_{ij} \cdot Z_{ij} + b_{ij} \tag{8}$$

where PE_{ij} is the estimated precipitation at the grid cell (i,j) with altitude Z_{ij}, a_{ij} is the local orographic gradient and b_{ij} is the local intercept value. The weights in the weighted linear regression were computed based on a distance function as follows:

$$W = \lambda \exp\left(-\left(\frac{d_{3D}}{d_0} \right)^{\alpha} \right) \tag{9}$$

where the parameter λ controls the range of the weightings, α controls the shape of the function, d_0 is a limit distance for the selection of gauges, and d_{3D} is the so-called 'crossing distance' from each grid cell to each rain gauge. This «crossing distance» was computed by accounting for both the horizontal Euclidian distance grid-to-pixel and the vertical component of the distance related to the crossing of crests and valleys. A more detailed definition of the «crossing distance» is given by Gottardi & *alii*, (2007). The values of d_0, λ and α (25 km, 5 and 2, respectively) were based on a minimisation of the estimation error in a larger study area. The residuals (i.e., the differences between the trend estimates and the observed values) were then interpolated using ordinary kriging.

RESULTS AND DISCUSSION

Table 2 reports the values of K_{rain} and K_{snow} for the winter months computed by the Sevruk 1985 procedure at the rain gauges located in the study area. Monthly K_{rain} ranged from 1.01 to 1.03 at the valley bottom and from 1.05 to 1.10 at high altitude, while monthly K_{snow} ranged from 1.06 to 1.18 at the valley bottom and from 1.36 to 1.70 at high altitude. The correction factors for snow and rain were nearly identical for low altitude weather stations (below 2000 m a.s.l.) due to the low spatial variability of wind for low altitude sites. The largest values of the correction factors were calculated for Tonale, which is located in a mountain pass, and for Careser dam, which is located at high altitude and in an open po-

TABLE 2 - Minimum, maximum and average monthly values of the correction factors for rain (K_{rain}) or snow (K_{snow}) calculated for the weather stations equipped by precipitation gauges

	Altitude (m a.s.l.)	K_{rain} min	max	avg	K_{snow} min	max	avg
4 - Careser dam	2605	1.05	1.10	1.08	1.36	1.70	1.51
5 - Tonale	1795	1.04	1.04	1.04	1.25	1.29	1.28
6 - Pian Palù	1790	1.01	1.03	1.02	1.06	1.18	1.13
7 - Peio	1565	1.01	1.03	1.02	1.06	1.18	1.13
8 - Rabbi	1350	1.01	1.03	1.02	1.06	1.18	1.13
9 - Cogolo	1200	1.01	1.03	1.02	1.06	1.18	1.13
10 - Mezzana	935	1.01	1.03	1.02	1.06	1.18	1.12
12 - Malè	735	1.01	1.03	1.02	1.06	1.18	1.12

sition. The average value of K_{snow} at Careser dam amounted to 1.51.

The time series of the seasonal K_{snow_we} derived from snow observations at Careser dam is reported in figure 7. The average value of K_{snow_we} over the period 1960 to 2006 (1.57) was nearly identical to the value of K_{snow} obtained by the Sevruk 1985 procedure and was in reasonable accordance with correction factors provided by Goodison & *alii* (1998). The agreement between the experimental values and those given by the Sevruk 1985 procedure, at the highest weather station, increased the confidence of applying the same procedure at the lower stations of the study area. In addition, the correction factors given by this procedure are in good accordance with the calculations of Ranzi & *alii* (1999), which compared gauge measurements with snow water equivalent data in a nearby geographic area.

The change of instrumentation and observation methods at Careser dam in October 1992, as was described in «Methods - Processing of meteorological data», did not lead to significant differences in the average values of K_{snow_we} (1.55 from 1960 to 1992 and 1.60 from 1993 to 2006; fig. 7). There was, however, a notable interannual variability,

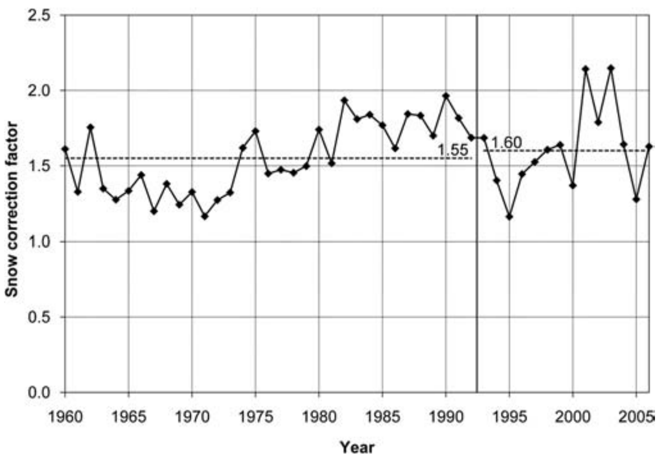

FIG. 7 - Winter season K_{snow_we} computed from 1960 to 2006 at the Careser dam weather station. The vertical line between 1992 and 1993 indicates the time of gauge substitution. The dotted lines indicate the average values of the K_{snow_we} for the two time periods.

with the coefficient of variation (CV hereinafter) of the time series amounting to 0.16. The maximum value of K_{snow_we} (2.15) was calculated for the year 2003, whereas the minimum value (1.16) was calculated for the years 1995 and 1971. The interannual variability increased from 0.148 (before 1993) to 0.175 (after 1993). It is difficult to disentangle instrumental from meteorological causes for this increase. However, episodic malfunctioning, such as the capture of wind blown snow or obstruction during intense snowfall (e.g., the large values of K_{snow_we} reported for the years 2001 and 2003, in figure 7) may have occurred at the automatic tipping bucket rain gauge operated since 1993, as it was checked less frequently than the old manual instrument. On the other hand, we found a direct correlation between K_{snow_we} and the average seasonal intensity of solid precipitation (i.e., the depth of seasonal fresh snow divided by the number of days with solid precipitation). The correlation coefficient was 0.71 before 1993 and 0.84 after 1993. Therefore, a larger variability in snowfall intensity may have caused an increase of interannual K_{snow_we} variability.

The raw precipitation data were corrected by means of K_{snow} and K_{rain} values reported in table 2 with the exception of Careser dam, where a K_{snow_we} of 1.57 was applied to correct the solid precipitation data. The spatial distribution of the average 1979-2005 winter precipitation depth from raw and corrected observations is reported in figures 8a and 8b, respectively. Figure 8c reports the difference between 8b and 8a.

The precipitation correction led to two important effects: i) the average precipitation estimates increased considerably; and ii) the orographic structure arose as a significant control of the precipitation distribution. The average difference in precipitation depth over Val di Peio amounted to 182 mm (+31.7%), which corresponds to a volume of $31.5 \cdot 10^6$ m^3. The uncorrected rain gauge data led to almost no increase of precipitation with altitude (fig. 8a). Consequently, the divergence between estimations from the corrected and raw data increased with elevation, reaching a maximum of 389 mm (66%) in the upper areas.

The quality of estimates at high elevation was assessed by comparing the measured winter balances and seasonal precipitation estimates over Careser glacier provided by raw and processed precipitation data (fig. 9a and 9b). Figure 10 shows the time series of the estimated precipitation depths vs. the Careser glacier winter balances. In these comparisons, we assumed that the mean snow water equivalent measured at the end of the accumulation season on the glacier provides a reasonable estimate of winter precipitation.

The calculations from uncorrected data led to a strong underestimation of the winter precipitation on the glacier and largely smoothed interannual differences, while the estimations by corrected precipitation were in close agreement with the mass balance data. The use of corrected precipitation resulted in a decrease of the mean estimation error from –38% to –2.5% and in a reduction of the Root Mean Squared Error (RMSE) from 410 to 171 mm w.e.. The Nash & Sutcliffe (1970) efficiency index E increased from –1.07 to 0.64. Overall, these results show the out-

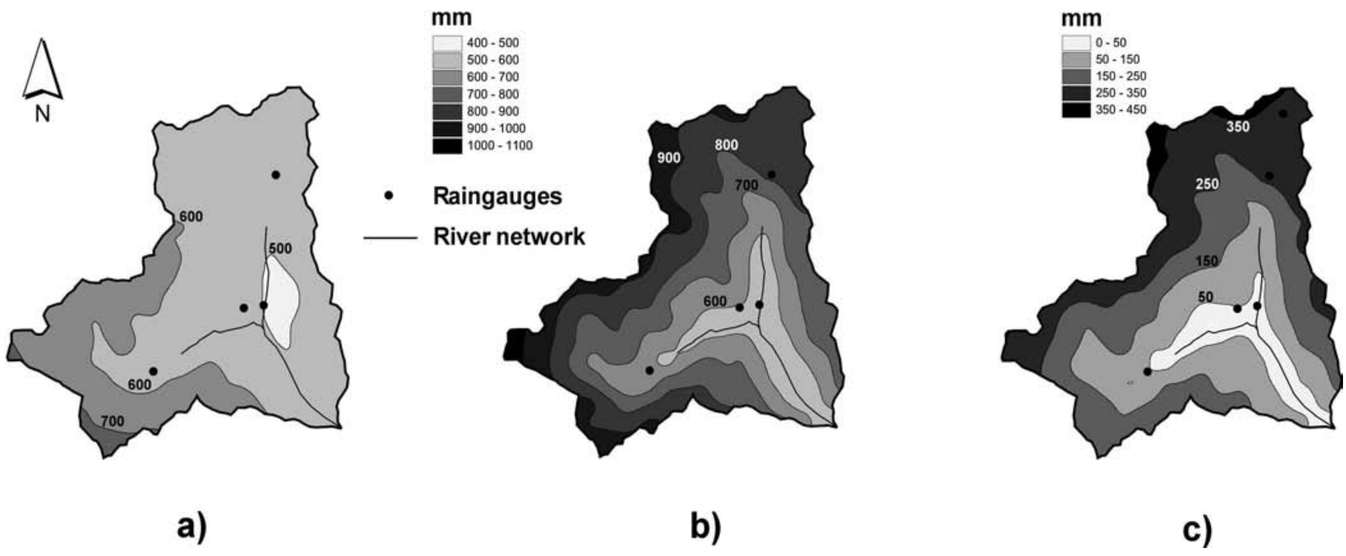

FIG. 8 - Spatial distribution of the mean winter season precipitation obtained by (a) uncorrected and (b) corrected gauge data. The mean values are computed over the period 1979-2005. The map in figure c) shows the differences between figures a) and b).

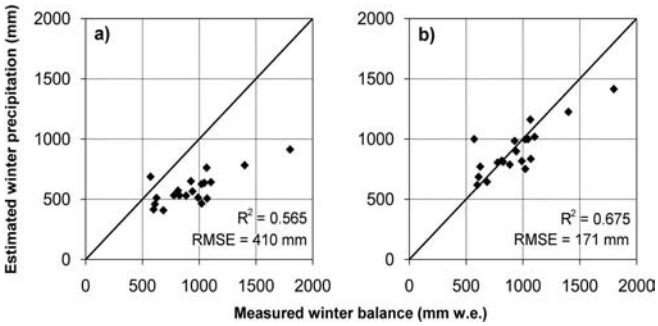

FIG. 9 - Comparison between measured winter balances and estimated winter precipitation on Careser glacier obtained with (a) uncorrected and (b) corrected rain gauge data for the period 1979-2005.

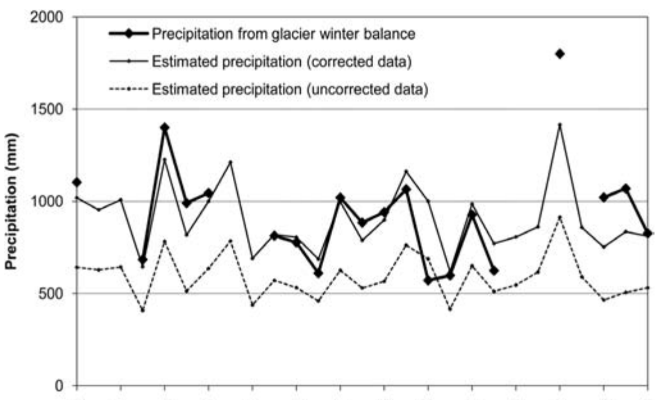

FIG. 10 - Time series of Careser glacier measured winter balance compared with precipitation estimates from corrected and from uncorrected rain gauge data in the 1979 to 2005 period.

standing impact of precipitation correction on high elevation precipitation estimation.

Clearly, there may be differences between the winter mass balance and the actual precipitation depth over the glacier basin, that must be taken into account. These differences may be due to the processes of snow removal and redistribution, such as melt, wind drift and sublimation (Lehning & *alii*, 2008; Déry & *alii*, 2010). Direct measurements of these processes were not carried out during the accumulation season in the frame of the long-term monitoring program of Careser glacier mass balance. On the other hand, an assessment of these processes by means of a mass balance model was beyond the aim of this work. Snowmelt occurred sporadically from October to May on the glacier, and the warm spells were infrequent and brief. Consequently, any percolating melt water likely refroze in the snowpack and formed ice lenses, which were normally sampled during the measurements and included in the water equivalent of snow. Avalanches are of minor importance on Careser glacier. Sublimation losses from snow during the accumulation season are highly variable in space and may locally be important during wind-induced snow transport from suspended snow particles, such as in wind-exposed mountain ridges (Strasser & *alii*, 2008; MacDonald & *alii*, 2010). Nevertheless, direct measurements executed with a lysimeter on Hintereisferner (a glacier in the Ötztal Alps, Austria) indicated a sublimation rate of ~ 100 mm y^{-1} (Kaser, 1983). Therefore, in view of the relatively low importance of wind action on Careser glacier (Carturan & *alii*, 2012), the possible errors due to snow redistribution and sublimation could be neglected because they were indiscernible from the typical errors reported in the literature for snow accumulation measurements on glaciers, which range from 100 to 300 mm w.e. y^{-1} (Cogley & Adams, 1998; Gerbaux & *alii*, 2005; Thibert & *alii*, 2008; Huss & *alii*, 2009).

Table 3 documents the impact of precipitation corrections on the evaluation of the orographic gradient (Eq. 8). The monthly gradients are expressed both in % per unit altitude and in mm per unit altitude. The first expression (values referred to the average altitude of Val di Peio, 2365 m a.s.l.) has the advantage of being independent of the length of the period. The measurement errors greatly affected the estimation of the vertical gradients, and for the months of October and November, even negative values resulted from the raw data. A noticeable increase of the gradients resulted from the correction of precipitation. The winter season average vertical gradient increased from 4.0% km^{-1} to 21.9% km^{-1}. The minimum monthly value (October) rose from –4.3% km^{-1} to 12.7% km^{-1}, and the maximum value rose from 9.7% km^{-1} (May) to 27.9% km^{-1} (April). The winter-season average vertical gradient corresponds to 23.4 mm km^{-1} for uncorrected data, and to 174.5 mm km^{-1} for corrected data. The latest value is in closer agreement with the gradients reported by Sevruk (1997) for the two nearby Swiss regions of Ticino (230 mm km^{-1}) and Engadina (290 mm km^{-1}), and lies in the range of 0 to 330 mm km^{-1} proposed by Ranzi & *alii* (1999) for the Central Italian Alps, even though these gradients concern mean annual precipitation values.

TABLE 3 - Average monthly vertical precipitation gradients in the 1978 to 2005 period (the values expressed in % km^{-1} are referred to the average Val di Peio altitude, 2365 m a.s.l.)

	Vertical precipitation gradient			
	From uncorrected precipitation		From corrected precipitation	
	% km^{-1}	mm km^{-1}	% km^{-1}	mm km^{-1}
October	–4.3	–4.5	12.7	16.7
November	–0.9	–0.7	19.2	22.1
December	0.5	0.3	20.3	14.9
January	3.0	1.4	20.3	14.9
February	6.0	2.3	23.6	12.6
March	9.2	5.8	27.3	24.5
April	8.8	7.8	27.9	34.2
May	9.7	11.1	24.0	34.5
Winter (Oct-May)	4.0	23.4	21.9	174.5

CONCLUSIONS

The correction of gauge data for wind-induced errors strongly impacted the estimation of precipitation in the ungauged, high-altitude areas of the region examined in this study. The correction procedure proposed by Sevruk (1985) and used for the Hydrological Atlas of Switzerland (Spreafico & *alii*, 1992) proved to be applicable to the study area. The reliability of the correction procedure was assessed by comparing snow correction factors with snow water equivalent data at the Careser dam weather station, which is the most affected by errors due to its location, wind exposure and frequency of solid precipitation.

The snow correction factors at Careser dam exhibited a significant interannual variability, and they correlated with the average intensity of snowfalls.

The precipitation correction translated to a strong increase in the areal precipitation depth and in its dependency on elevation, which peaks in the spring months. The difference between the estimates obtained by processed and raw data increased with elevation, and it was most significant over the glacier areas.

The winter balance series of Careser glacier allowed a validation of estimations at high altitude and confirmed the need to correct the raw precipitation data to avoid systematic underestimations of precipitation inputs. The use of fixed values for the correction factors is a simplification, and in extreme years (e.g., 2001), its effect in extrapolations was clearly visible. Nonetheless, reasonably good results were obtained, as can be seen from the agreement between the mass balance observations and precipitation estimations. Further improvements can probably be obtained by means of variable correction factors, but the processes that control this variability are still poorly understood in this area, and they require more detailed investigations, possibly extending to summer months.

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