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New insights in the relation between climate and slope failures at high-elevation sites

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Abstract

Climate change is now unequivocal, however type and extent of terrestrial impacts are still widely debated. Among these, the effects on slope stability are receiving a growing attention in recent years, both as terrestrial indicators of climate change, and for the implications for hazard assessment. High elevation areas are particularly suitable for these studies, because of the presence of the cryosphere, which is particularly sensitive to climate.

In this paper, we analyse 358 slope failures occurred in the Italian Alps in the period 2000-2016, at an elevation above 1500 m a.s.l. We use a statistical-based method to detect climate anomalies associated with the occurrence of slope failures, with the aim to catch an eventual climate signal in the preparation and/or triggering of the considered case studies. We first analyse the probability values assumed by 25 climate variables on occasion of slope failure occurrence. We then perform a dimensionality reduction procedure, and come out with a set of four most significant and representative climate variables, in particular heavy precipitation and short-term high temperature. Our study highlights that slope failures occur in association with one or more climate anomalies in almost 92 % of our case studies. One or more temperature anomalies are detected in association with most case studies, in combination or not with precipitation (47 % and 38 % respectively). Summer events prevail, and an increasing role of positive temperature anomalies from spring to winter, and with elevation and failure size emerges.

While not providing a final evidence of the role of climate warming on slope instability increase at high elevation in recent years, the results of our study strengthen this hypothesis, calling for more extensive and in-depth studies on the subject.

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3 Alto Adige).

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1. Introduction

The effects of climate change on natural systems are the object of worldwide debate, both in science and policy. According to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC 2014), an average global increase of about 0.85 °C in land and ocean surface temperature has been recorded over the period 1880-2012 (IPCC 2014). At the global scale, the first decade of the 21st century has been the warmest one since 1850 and 2016 resulted to be the warmest year on record globally (Northon 2017); further warming (in the range 0.3 - 0.7°C between 2016 and 2030) is expected based on climate projections (Gobiet et al 2014). Changes in extreme weather events, as a rise in high extreme temperatures and a decrease in low ones, have been observed since 1950 almost worldwide (Stocker et al. 2013). Changes in precipitation have also been observed, but confidence is lower, since an evident signal has not been detected worldwide, also because precipitation features at a site are more sensitive than temperature to the orographic and physiographic characteristics of the local territory (Brunetti et al. 2009).

If climate change is unequivocal, the full understanding of its impact on natural environments poses further challenges, due to the inherent complexity of these systems. In this framework, the cryosphere is considered a “*natural thermometer*” (Stocker et al. 2013) of climate change (in particular global warming, Kääb et al. 2007), due to its sensitivity to change in climate variables. Studies on high-elevation/latitude areas show an overall framework of ice degradation at global and regional scales in response to air temperature increase (Zemp et al. 2015; Chadburn et al. 2017). In high-mountain areas, cryosphere degradation, and in particular glacier shrinkage, permafrost thawing and spring snowpack decreasing, have, as direct consequence, the worsening of the mechanical conditions of rocks and soils (Fischer et al. 2006). One of the main consequences is slope destabilization (Gruber and Haeberli 2007; Harris et al. 2009), with shear strength reduction (Davies et al. 2001), opening of deep thaw joints, fractures displacement and change in the stress field (Weber et al. 2017) acting as the triggers in a context of climate warming. Several studies, in fact, confirm an overall growing trend of slope instability worldwide (Huggel et al. 2010).

The European Alpine Region represents a key research hotspot for natural hazards studies, because of its complex climatic, geologic and physiographic setting, and the high touristic value of the entire area (Beniston et al. 2017; Haeberli et al. 2015). Studies on mass-wasting processes have gained an increasing attention in recent years, and a growing number of events is documented in glacial and periglacial areas, in particular since the hot summer of 2003 (Ravanel et al. 2017). The number of events of slope instability is expected to increase in the next future, even if several studies indicate a more complex response of slope stability to climate change (Gariano and Guzzetti 2016; Stoffel et al. 2017).

Among the different factors affecting slope stability, attention is here devoted to precipitation and temperature. Precipitation is known as a main driver of landslides in various geomorphological settings including the European (Italian) Alps (Peruccacci et al. 2017; Palladino et al. 2018). Changes in rainfall duration and intensity, combined with higher temperature, are supposed to enhance mass-wasting processes in high mountains, in particular those driven by water as debris and mud flows (Rebetz et al. 1997; Chiarle et al. 2011; Pavlova et al. 2014). The combination of abundant precipitation (rainfall and snowfall) and higher temperature is supposed to have a role in initiating shallow landslides (Saez et al. 2013) at high altitude as well. The effect of air temperature variations on rock and ice stability is even more complex to understand (Nigrelli et al. 2017). For example, in the Monte Rosa massif Huggel et al. (2010) relate air temperature increase to various rock/ice avalanches and debris flows, occurred in the area in recent years. Recent studies relate the increased trend of rockfall activity affecting high-alpine steep rock walls to anomalies in mean temperatures in the hottest period of the year (Ravanel and Deline 2015), whereas other papers focus on changes in high daily extremes temperature potentially causing rockfalls over the European Alps (Allen and Huggel 2013). Other works highlight the role of summer heatwaves in the increased slope failure activity on permafrost affected rock walls in the European Alps (Ravanel et al. 2017). Thus, the linkage between climate and slope

1 failures at high-elevation is definitely difficult to investigate, and many authors have pointed out the diffuse
2 lack of a shared and standardized strategies to face the problem (GAPHAZ 2017).

3 Assessing how climate variables could effectively influence slope failure initiation and/or preparation
4 mechanisms is crucial in this framework (Huggel et al. 2013). Paranunzio et al. (2015) have proposed a
5 statistical-based approach to detect anomalous values in the climate variables at the date of a slope failure,
6 with application on five slope instability events of different types, occurred in glacial and periglacial areas of
7 the Piedmont Alps (Northwestern Italy). The role of air-temperature anomalies as the main trigger has been
8 clearly spotted. To further investigate the role of air temperature variations at high elevation sites, Paranunzio
9 et al. (2016) analyzed 41 rock falls in the Italian Alps from 1997 to 2013, again demonstrating that the majority
10 of slope-instability events can indeed be associated to air-temperature anomalies. However, the limited size
11 of the considered dataset did not allow to fully disentangle the relations between climate change and slope
12 instability.

13 In the present work, we aim to catch a possible climate signal behind the preparation and/or triggering of
14 slope instability processes in high-mountain areas starting from a robust sample of more than 400 events
15 occurred above 1500 m a.s.l. in the Italian Alps from year 2000 on. By performing this analysis, we aim to
16 address the following research questions:

- 17 i) Can temperature and precipitation be considered as key conditioning and/or triggering factors
18 for slope failures at high-elevation sites? What climate variables are most relevant for slope
19 instability?
- 20 ii) How can the climate signal detected be linked to the process typology and to its spatiotemporal
21 distribution?
- 22 iii) Can climate change be deemed responsible for the observed increasing trend of slope instability
23 at high elevation? What can be expected in the future?

24 The paper is organized as follows. After a brief description of the general setting of the study, we describe
25 how we constructed the inventory and list the climate data used. This is followed by an explanation of the
26 procedure for data preparation and validation; and by a step-by-step description of the methodological
27 approach. Finally, the main results are described and discussed, presenting a critical analysis of the major
28 outcomes, of the unresolved questions and of the possible further steps.

29 2. Study area

30 We focus on the entire Alpine Italian region, stretching 1200 km from E to W and covering about 5200
31 km², i.e. 27.3 % of the European Alps. The study area extends from 6° to 13° E and from 44° to 47° N, from the
32 Mediterranean Sea (Franco-Italian border) stretching eastward to Slovenia (Fig. 1).

33 The Alpine chain developed from the subduction of a Mesozoic ocean and the collision between the
34 Adriatic (Austroalpine-Southalpine) and European (Penninic-Helvetic) continental margins (Dal Piaz et al.,
35 2003). As a result, the Alps may be distinguished into two belts, separated by the Periadriatic (Insubric)
36 lineament, which are characterized by an opposite direction of tectonic transport and have a different size,
37 age, and geological history. The Europe-vergent belt is a thick collisional wedge, dating back to Cretaceous-
38 Neogene and composed of continental and minor oceanic units. The Southern Alps is a minor, non-
39 metamorphic, Neogene belt, displaced to the south. This complex geological history is at the base of the
40 amazing geodiversity of the Alps, which culminates with the Mont Blanc Massif (4810 m a.s.l.), on the Western
41 side. After their formation, the Alps have been sculptured by glaciers, running waters and slope failures, to
42 reach their present configuration. Both the geologic and morphologic setting strongly influence the proneness
43 of slopes to failure. The last important glacier advance (Little Ice Age) ended around 1850: since then,
44 European glaciers suffered a strong area reduction of approximately 50 % (Zemp et al. 2015). Due to the

1 specific topoclimatic and physiographic setting, glacier shrinkage has been particularly marked on the Italian
2 side of the Alps, up to the almost complete disappearance at the lower altitudes and latitudes (Nigrelli et al.
3 2014). Present glaciers (Salvatore et al. 2015; Smiraglia et al. 2015) are mainly located in the Valle d'Aosta
4 region (36 % of the total glacierized Italian area), followed by Trentino Alto Adige (31 %), Lombardia (24 %),
5 Piemonte (8 %) and Veneto (1 %). Permafrost distribution in the Alps is highly complex and affected by a huge
6 spatial variability, mainly due to topographic effects (Harris et al. 2009). Studies carried out on the European
7 Alps have found evidence of permafrost approximately since 2600 m a.s.l., but it can be found at 3500 m a.s.l.
8 in unfavorable conditions, as south-facing rock walls (Cremonese et al. 2011).

9 The complex topographical and geographical context influences the climate of the Greater Alpine Region
10 (HISTALP 2018), which is characterized by a high spatial variability of precipitation and temperature patterns
11 at regional and local scales (Auer et al. 2007). This is even more evident in the Italian alpine region, due to its
12 complex physiography coupled with local atmospheric patterns (Avanzi et al. 2015). Mean annual precipitation
13 ranges from 500 mm (in the Aosta plain and inner Alpine valleys) to 3000 in some prealpine regions (Crespi et
14 al. 2017). Based on areal values maximum (minimum) annual temperature is respectively 5 °C (-3°C) in the
15 Western and 8° C (-1°C) in the Eastern Italian Alpine sector (Esposito et al. 2014).

16 3. Data

17 3.1. Catalogue of slope instability events

18 The catalogue created for this work consists of 401 events of slope instability, which occurred between
19 year 2000 and 2016 on the Italian Alps, at an altitude of more than 1500 m a.s.l. (Fig. 1). We considered all
20 types of landslides, debris/mud-flows, glacial lake outburst floods, and ice-avalanches. The choice of the
21 starting date is related to the greater availability of climate records from weather stations in the last two
22 decades. To build the catalogue, we collected data from multiple sources. First, we relied on databases and
23 technical reports realized by the Italian regional agencies: many data are freely available on the related
24 geoportals (ARPA Piemonte 2018a; RAVdA 2018) other data have been provided upon request. The dataset
25 was then implemented with the information derived from the archives of CNR-IRPI Torino, from scientific
26 papers and local/national newspapers (Luino and Turconi 2017; Paranunzio et al. 2016). Additional information
27 was obtained from fire-fighters reports and online news of Civil Protection of Regione Autonoma di Trento
28 (Protezione Civile – Provincia Autonoma di Trento 2018). More in detail, 53% of the inventoried case studies
29 comes from the Italian regional agencies and was partly collected in the framework of the Italian Landslide
30 Inventory project (IFFI, Trigila et al. 2010), especially for the Veneto and Trentino regions; 27 % of the case
31 studies comes from the archives of CNR-IRPI Torino, while technical documentation and scientific works
32 represent the 13 % of the data sources. The source of information for each case study is shown in Online
33 Resource 1.

34 Two main geographical clusters were identified, corresponding to the Western and Central-Eastern Italian
35 Alps, respectively (Fig. 1). When considering the distribution and density of case studies in our catalogue, one
36 should bear in mind that they strongly depend on the availability of information. This latter, in turn, is mainly
37 linked to the damage / risk associated with slope instability, rather than to its actual space / time distribution.
38 This is particularly true for high mountain areas, often remote and little frequented, where the information
39 relating to slope instability events is often incomplete and fragmented. In these areas, summer events are
40 more documented than events occurring in other seasons, due to the higher frequentation of high-mountain
41 areas in that season. Likewise, information on natural instability events occurring at lower elevations is
42 typically more readily available and more accurate and detailed, because these events more frequently
43 interact with human activities and structures.

44 Online Resource 1 includes the information on the case studies that is relevant for this work (e.g., type of
45 process, date, elevation and season of occurrence, volume and slope aspect). The case studies have been

1 mapped as single points using a Geographical Information System (GIS) and *Google EarthTH* (Fig. 1). As
2 mentioned before, data come from different sources: this could entail a certain degree of inhomogeneity for
3 accuracy and level of detail. This is particularly true for newspapers that, in some cases, report general
4 information and are only seldom precise enough about the location and time of triggering. Whenever
5 available, the exact type of process has been reported: in the absence of precise information, the slope
6 instability process was generically classified as a “landslide”. The accuracy of the spatial localization of case
7 studies varies also according to the type of process. The starting point of landslides (of any type) is usually
8 identified with a good accuracy. This information, instead, is rarely available for debris/mud flows, for which
9 very often only the point of impact on structures and infrastructures is reported: if this point is below 1500 m
10 a.s.l., and no information was available about the starting point, the event was not included in our catalogue.
11 As a result, debris/mud flows are underrepresented in our catalogue. The percentage of the different types of
12 slope instabilities contained in our catalogue is shown in Fig. 1.

13 A digital elevation model with a 20 m resolution (SINANet Ispra 2017) was used for the analysis of the
14 topographical setting of the case studies.

15 **Fig. 1** Map showing 401 slope instability events included in the catalogue (squares) and 131 weather stations used
16 in this work (dots); type and sample size of the instability processes included in the catalogue are represented in the white
17 box; RF: rockfall, BF: blockfall, RA: rock avalanche, DF: debris flow, MF: mud flow, L: landslide, SL: slide, IF: ice fall, IA: ice
18 avalanche, GLOF: glacial lake outburst flood, S: (soil) slip

19 3.2. Climate data

20 In order to reconstruct the climate history of each event of slope instability, we considered climate data
21 from 131 automatic weather stations in Northern Italy (Fig. 1). These stations are managed by the Regional
22 Environmental Protection Agencies (ARPA) in Piemonte (ARPA Piemonte 2018b), Lombardia (ARPA Lombardia
23 2018) and Veneto (ARPAV 2018) regions, by the Centro Funzionale of the Regione Autonoma Valle d’Aosta
24 (Centro Funzionale Valle d’Aosta 2018), the Hydrographic Office of the Provincia Autonoma di Bolzano
25 (Provincia Autonoma di Bolzano - Alto Adige, 2018), and Meteotrentino (Meteotrentino 2018) in the Provincia
26 Autonoma di Trento. We consider: i) mean, maximum and minimum daily air temperature (denoted as T_{mean} ,
27 T_{max} and T_{min} , respectively, or simply T , from now on) and ii) daily cumulated precipitations (rainfall and solid
28 precipitation, denoted as R from now on). Temperature and precipitation values used in this work have been
29 first validated by the regional agencies owning the data. Nevertheless, a further quality check has been done
30 by the authors, in order to find out residual erroneous or anomalous values. Finally, temperature and
31 precipitation data from 130 and 123 weather stations, respectively, were used. Information on the
32 geolocalization, instruments and source of the data is given in the Online Resource 1.

33 3.3. Data preparation

34 We establish a set of criteria to decide if the slope instability event could be included in the final sample,
35 as follows. As anticipated in Section 3.1, we consider events i) initiated above 1500 m a.s.l. and ii) properly
36 localized in space; beyond that, in order to enable a proper climate analysis, iii) the date of occurrence has to
37 be known with a daily accuracy. Details on the time/moment of the day are rarely available (only 12 % of case
38 studies).

39 We attribute a code to each case study, ranging from 1 to 3, describing the level of accuracy of the spatial
40 localization: code 1 refers to events mapped with high accuracy (the detachment point is known); code 2 is
41 mainly attributed to debris/mud flows, for which the exact detachment point is hard to know, but we know
42 the channel where the flow developed; code 3 refers to the lower level of spatial accuracy (information on the
43 elevation and failure zone are available, but not on the exact detachment point).

44 In order to identify the most suitable weather stations in the study area, we base the selection on three
45 criteria, aimed to achieve the best compromise between the spatial distance from the failure zone, the

1 representativeness of the morphological setting and the availability of climate data. More in detail, weather
2 stations have: i) to be as close as possible to the failure zone, both in terms of altitude, and horizontal distance;
3 ii) to be located in a morphological context similar to that of the failure area, and iii) to provide a suitable
4 temporal coverage (including the day of the event).

5 Whenever possible, weather stations located in the same valley where the slope failure occurred and in
6 similar topoclimatic conditions are preferred (Nigrelli et al. 2017). As a second choice, we choose those stations
7 that are located at a similar altitude and as close as possible to the failure area, at a horizontal distance lower
8 than 20 km; otherwise, we select the available stations at lower elevations. The choice of a 20 km buffer-zone
9 is due to the need to rely on data that are representative of the climate conditions of the failure area: this is
10 particularly crucial in complex-orography environments, as the Alpine region (Beniston et al. 2017). Stations
11 with less than 10 years of climate records or with non-continuous data series are discarded. In case of weather
12 stations measuring only one variable (T or R), we base the analysis of the event on two different, but nearby
13 weather stations.

14 Given these strict requirements, in the end 358 events out of the original sample (401) are included in the
15 final subset subject to climate analysis.

16 4. Methodology

17 The first steps of the method are addressed to the identification of climate variables assuming non-
18 standard values at the date when a slope failure occurred (Sections 4.1 and 4.2). For these steps, we mainly
19 rely on the methodology developed by Paranunzio et al. (2015) and modified by Paranunzio et al. (2016). In
20 this way, we end up handling a multidimensional system, which may provide redundant information coming
21 from variables showing similar patterns, making it complicate to find out the effective role of the climate
22 forcing in the initiation of slope failures. We thus make a step forward, by performing a dimensionality
23 reduction of the variables involved, as detailed in Section 4.3. This step is crucial, in order to identify the climate
24 variables that are mostly involved in slope-failure initiation. We complete our study with some additional
25 analyses taking into account the spatiotemporal distribution of the slope failures (Section 4.4), based on the
26 set of climate variables selected in Section 4.3. The main steps of the procedure are illustrated in the flowchart
27 of Fig. 2.

28 **Fig. 2** Flowchart representing the main steps of the method as in Section 4

29 4.1. Detecting the climate anomalies potentially inducing slope failure occurrence

30 The method adopted is a statistical-based approach, aimed to define the climate conditions in the period
31 before a slope failure (the period from one day up to 3-months before the event), compared to the typical
32 climatic conditions for the area of interest. In a first attempt, we perform the method with a bottom-up
33 approach, i.e. without considering any *a priori* information on the event. Thus, by means of a non-parametric
34 analysis based on the use of the empirical distribution function, we scrutinize the available sample data for
35 each climate variable (V) in order to detect possible non-standard values in correspondence to slope failure
36 occurrence. The essential steps are reported below. Further details are available in Paranunzio et al. (2015,
37 2016). Hereinafter, we refer to the date of failure as “date” (e.g., 1 March 2016), whereas the calendar date
38 is referred as “day” (e.g., 1 March).

- 39 i. Selecting the climate variables. We use any easily available climate variable that could act as a
40 trigger/preparatory factor of slope failure in climate-sensitive environments as high-elevation
41 sites. We include temperature T , precipitation R (rainfall and solid precipitation) and temperature
42 variation ΔT (i.e. the difference between the temperature at the day of occurrence of the slope
43 failure and the value recorded in the day before or in an antecedent day).

- 1 ii. Choosing the aggregation scale. V is the time-aggregated variable. For T and R , we consider time-
2 aggregated variables from daily to quarterly scale. For ΔT , it is of interest to consider the
3 temperature excursion between the day of failure and the previous days (1, 3, and 6 days in this
4 work).
- 5 iii. Selecting the weather stations. The choice of the most suitable weather stations for records
6 collection is based on the requirements listed in Section 3.3.
- 7 iv. Selecting the reference sample. We select the reference sample whereon comparing the variable
8 V for the date of failure. The reference sample includes n values ($n \geq 10$) and $V_{(i)}$ is the i th value in
9 the ordered sample, $i = 1 \dots n$. The choice of the reference sample depends on the seasonality of
10 the variable involved. For T , we compare the value recorded at the date of the event with the
11 values in the data series referring to the same period in other years. In the case of ΔT or
12 intermittent processes as R , we extend the sample whereon performing the comparison to
13 include the previous and following 45 days, in order to increase the robustness of the analysis and
14 to obtain larger reference samples.
- 15 v. Non-exceedance probability value computation. We estimate the cumulative probability
16 distribution $P(V)$ in a non-parametric way as $P(V)=i/n$, if $V > V_{(i)}$. We hypothesize that V may be a
17 significant driver/trigger of a slope failure when $P(V) \leq \alpha/2$ (negative anomaly) or $P(V) \geq 1-\alpha/2$
18 (positive anomaly). Here, we set the significance level set α at 0.2, performing a 10 % test on each
19 of the distribution tails. In the end, we consider V as a relevant factor for slope failure occurrence
20 when $P(V) \leq 0.1$ or $P(V) \geq 0.9$.

21 Note that we compute the probability distribution using data as they are recorded at the weather
22 station, without transposing them at the detachment elevation, considering that a simple
23 translation of values to the elevation of the slope failure would not modify the results. This
24 assumption is discussed in detail in Paranunzio et al. (2015; 2016).

25 4.2. Towards the identification of a climate signal in the variables

26 After performing steps (i) to (v) as in Section 4.1 for all the considered variables, we obtain as many as 25
27 probability values per case study. These preliminary outcomes can be summarized in a $M \times L$ matrix, where M
28 is the number of case studies included in the final sample and L is the number of climate variables (25 in this
29 case). Each cell of the matrix reports the non-exceedance probability $P(V)$ associated with the j th variable, $j =$
30 $1 \dots L$, of the k th event, $k = 1 \dots M$. We claim that when $P(V) \geq 0.9$ we are in the presence of a positive anomaly in
31 V , but of course the obtained value could also be the result of a random variation in V , which brought the V
32 value above 0.9 by chance. This will happen on average in 10 % of the cases, which entails that the variable
33 brings a significant information at the regional scale only if the detected positive anomalies for that variable
34 are more than 31, considering the example of Fig. 3, where the total number of considered case studies is 312.
35 Instead of concentrating one's attention on the 0.9 probability, one can perform a graphical verification of the
36 regional-scale significance of the variable V for explaining slope instability occurrence. Consider the ordered
37 sample of M $P(V)$ values, where each value is associated to an event and $M = 312$. $P(V)_{(k)}$ is the k th value in the
38 ordered sample, $k = 1 \dots M$ for the j th variable. We compute the Empirical Cumulative Distribution Function
39 (ECDF) as $q_{(j)}=k/M$, and we plot the q values versus their corresponding $P(V)$ value. We obtain a graph with 312
40 points, one for each event. If the points are close to the bisector, the variable is not significant at the regional
41 scale in explaining slope instability occurrence, because the points are positioned in the graph as if they were
42 sampled randomly.

43 Conversely, curves that deviate significantly from the bisector are an indication of a variable being
44 significant at the regional scale. If the points are positioned below the bisector line, in a considerable number
45 of case studies the statistical analysis of the variable detects a positive anomaly. Similarly, curves above the
46 bisector line refer to a more frequent presence of lower-tail values (negative anomaly). As in the example of
47 Fig. 3, we incur in a 19.5 % of probability of detecting values above the 90th percentile, whereas just 5.4 % of

1 values is located in the lower-tail of the distribution (10th percentile). The former value is almost twice the
 2 expected 10 % significance, indicating the presence of an evident statistical positive climate anomaly at the
 3 daily scale.

4 **Fig. 3** Empirical Cumulative Distribution Function (ECDF), denoted as $q_{(j)}$, based on the probability values $P(V)$ of a j th
 5 variable for the entire sample (312 events in this case), as detailed in Section 4.2. The straight line indicates the bisector,
 6 whereas the curve indicates the ECDF. 19.5 % of values lies above the 90th percentile, whereas only 5.4 % of values are in
 7 the lower-tail (10th percentile).

8 The same approach is applied to all the considered variables. Ideally, we may detect as many as 42 positive
 9 and negative temperature anomalies, from the daily to the quarterly scale, for $Tmean$, $Tmax$ and $Tmin$, and 4
 10 positive anomalies, from the daily to the quarterly scale, in precipitation values R . In this framework, a
 11 dimensionality reduction is of help in promoting the most important variables, by improving the
 12 interpretability of the final outcomes.

13 4.3. Reducing the dimensionality

14 A dimensionality reduction is a procedure that allows one to reduce the initial number of considered
 15 variables (L) by obtaining a set of most important variables (S). We first perform a procedure based on
 16 correlation thresholds in order to detect variables that are highly correlated with others. The main steps are
 17 illustrated hereinafter. In the following points, we present some examples to make the methodology more
 18 clear.

- 19 i. Computing the pairwise correlation coefficient. We calculate the Pearson correlation coefficient
 20 ρ , a dimensionless index measuring the linear correlation between couples of variables in columns
 21 i and j , as:

$$22 \quad \rho_{ij} = \frac{\sigma_{ij}}{\sigma_i \sigma_j} \quad (1)$$

23 where σ_{ij} is the covariance and $\sigma_i \sigma_j$ is the product of the standard deviations. The coefficient ρ
 24 ranges between -1 and 1, with 1 indicating perfect correlation, and -1 perfect anticorrelation.
 25 Based on the number of investigated variables (25 in this case), the output table is a 25x25
 26 correlation matrix, i.e. the matrix with the correlation coefficients for all pairs of data columns.

- 27 ii. Detecting highly correlated variables. We define a minimum correlation threshold $|\rho|=0.5$, as
 28 suitable to identify pairs of variables with a high correlation. If the absolute value of the
 29 correlation coefficient is larger than 0.5, we mark the couple of variables as potentially redundant,
 30 because they bring a similar information into the system.

- 31 iii. The pruning process starts from the couple of variables with the highest correlation; the variable
 32 is eliminated which contributes less to explain slope instability occurrence. As an example, $Tmax_1$
 33 and $Tmean_1$ show a high correlation ($\rho=0.87$); we eliminate $Tmax_1$, since $Tmean_1$ recognizes 24.9
 34 % of potential climatic anomalies, compared to 18% for $Tmax_1$. In cases when the potential
 35 anomalies recognized by the two variables are similar, we eliminate the variable with lowest
 36 number of available data. As an example, although $Tmin_{30}$ provides almost the same number of
 37 positive anomalies compared to $Tmean_{30}$ (17.5 % and 17 % respectively), we prefer the latter
 38 because $Tmean$ is measured at 312 weather stations, versus 275 for $Tmin$. We proceed with the
 39 pruning until all variables in the final subset have a correlation coefficient lower than 0.5 (in
 40 absolute value). We finally come out with a pool of the most significant variables, achieving a
 41 lower-dimensional representation of a dataset, capable of preserving as much as possible the
 42 initial information.

4.4. Climate anomalies versus spatiotemporal factors

Finally, we perform a multivariate analysis in order to highlight complementary factors that, in combination with climate anomalies, could help in the interpretation of processes leading to slope failure.

For this study, we considered: i) season of occurrence, ii) scar elevation, iii) slope aspect and iv) detached volume. Events are divided in four seasonal classes based on the meteorological seasons. Four homogenous classes of elevation have been defined, ranging from the minimum to the maximum height of occurrence of the case studies. Aspect is expressed in degrees (0-359°) and divided in four classes: north (315°-45°), east (45°-135°), south (135°-225°) and west (225°-315°) facing slopes. Case studies are finally grouped in two classes of volume, i.e. small events ($< 10^3 \text{ m}^3$), and large events ($\geq 10^3 \text{ m}^3$).

5. Results

5.1. Statistical analysis of the climate variables

The results of the analyses as in Section 4.1 are fully reported in the Online Resource 1 as a $M \times L$ matrix, where M is the number of case studies (358 in this case) included in the final sample and L is the number of climate variables (25 in this case). Thus, each cell of the matrix includes the non-exceedance probability $P(V)$ associated with the j th variable, $j = 1 \dots L$, of the k th event, $k = 1 \dots M$. Positive anomalies are in red, whereas negative ones are in blue. Positive anomalies refer to the upper-tail of the distribution ($P(V) \geq 0.9$), whereas negative ones refer to the lower tail ($P(V) \leq 0.1$). Information on the selected weather stations used for the analyses are also reported.

Fig. 4 synthesizes the outcomes. Fig. 4a displays the number of stations showing a positive/negative anomaly in the considered variables, in association with slope failure occurrences. There, we report the results of the data analysis of one station for each investigated event, being the total sample size equal to 358. Note that most stations provide T_{mean} and precipitation R data (312 out of 358 case studies), whereas T_{min} and T_{max} data are available for fewer stations (275 out of 358 case studies). We considered only the upper-tail of the distribution for precipitation, since low precipitation values are not a trigger of slope instability. Bold numbers immediately above the stacked bars indicate the number of stations for which the variable was available and has been analyzed, whereas upper/lower regular font numbers indicate the number of negative/positive anomalies detected for the considered variable. As an example, the variable T_{mean}_1 is recorded by 312 weather stations, thus this variable was analyzed for 312 case studies (one event = one weather station), and a positive/negative anomaly was detected for 66 and 19 weather stations, respectively. In total, almost 27 % of case studies show a statistical anomaly in T_{mean}_1 .

In general, as one can see from Fig. 4a, almost 30 % of the weather stations used for this analysis show a positive/negative anomaly in most of the variables. Positive temperature anomalies mainly refer to T , in particular to T_{mean} , T_{max} and T_{min} , at daily and weekly scales. Ranging from T_{mean}_1 to T_{min}_{90} in Fig.4a, at least 18 % of the analyzed weather stations provide a positive temperature anomaly (from 18.5 % for T_{mean}_{30} to 26.2 % for T_{min}_7). Negative anomalies are mainly referred to ΔT data: some variables present about 13 % of stations with a negative temperature-variation anomaly (ΔT_{mean}_1 , ΔT_{min}_1 and ΔT_{max}_3). More in detail, negative anomalies for these variables refer to a significant drop of temperature between the day of the failure and the previous one, three or six days. More than 20 % of the analyzed stations show at least one precipitation anomaly (from 21 % for R_1 to 29.4 % for R_7).

Fig. 4b reports the number of climate anomalies, associated with the occurrence of slope instabilities, for the considered case studies. In total, at least one anomaly is detected in 329 out of 358 events (91.9 %). 47 % of the analyzed weather stations shows from one to four anomalies in the long-term series, and 34.5 % from five to nine anomalies. Only 8.1 % of the events (29) does not show any anomaly in the climate variables, in association to slope instability occurrence. Based on the results shown in Online Resource 1, in 47 % of the

1 case studies (168) we detect one or more temperature anomalies, in just 7 % (25 case studies) only a
2 precipitation anomaly, and in almost 38 % a combination of the two (136 case studies).

3 **Fig. 4 a)** Number of climate anomalies per variable and percentage out of the total number of events (358). Lighter
4 colors refer to positive anomalies (high extremes, heavy precipitation), darker color to negative anomalies (low
5 temperatures). Numbers above the stacked bar, from the bottom to the top: number of available weather stations per
6 variable, number of positive anomalies per variable, number of negative anomalies per variable. **b)** Number of events (in
7 bold) showing from 0 (no anomaly) to 19 climate anomalies and percentage out of the total number of events (358)

8 5.2. Identifying the key variables

9 As detailed in Section 4.2, we aim to find out the evidence of a climate signature in the occurrence of a
10 slope instability event. To this aim, we plot the empirical distribution functions of the values obtained by each
11 variable in all case studies, for each of the 25 investigated variables, to have a visual validation of the
12 significance of the considered variable in explaining slope instability at a regional scale.

13 Fig. 5 shows the results for the entire sample of 25 variables. The percentages of values allocated in the
14 upper/lower tails of the sampling distributions are indicated in the graphs (upper right and lower left sectors,
15 respectively). As can be seen, focusing on the 90th percentile, the most significant variables are T_{mean} , T_{max}
16 and T_{min} and R , at all temporal scales (day, week, month, 3-months), together with ΔT_{min} (6 days). Confirming
17 the results of Fig. 4, a major presence of negative anomalies is detected for ΔT . In this last case, the non-
18 exceedance probability exceeds the expected significance in ΔT_{mean} (1, 3 days), ΔT_{max} (1 day) and ΔT_{min} (3,
19 days). Percentages above 10 % are not of interest, since they are below the significance level.

20 **Fig. 5** Empirical Cumulative Distribution Function (ECDF), $q_{(j)}$, based on the probability values $P(V)$ of the j th variable
21 ($j=1\dots L$, $L=25$) as detailed in Section 4.2. The straight lines indicate the bisectors, whereas the curves indicate the ECDF.
22 Numbers in the lower left and upper right sectors refer to percentages of data below the 10th and above the 90th
23 percentile, respectively.

24 A large number of variables is found to be statistically significant and, thus, potentially worth to be
25 included in further analysis. On the one hand, this confirms the presence of evident climate signals associated
26 with the occurrence of a slope failure in more than one variable; on the other hand, this entails a redundancy
27 of information. For this reason, we tried to synthesize the information coming from variables showing similar
28 patterns by performing the dimensionality reduction described in Section 4.3.

29 First, we quantify the correlation existing among all considered climate variables, by running a pairwise
30 Pearson correlation analysis as described in Section 4.3 (Fig. 6). Darker colors indicate an increasing positive
31 and negative correlation between the probability values associated to each pair of variables. Coefficients
32 above the threshold $|\rho|=0.5$ (indicating strong correlation) are in bold. As can be seen, T_{mean} correlates
33 strongly at all temporal scales with T_{max} and T_{min} . A strong positive correlation across different temporal
34 scales is detected in T_{mean} (daily/weekly and weekly/monthly scales), as well as in T_{min} . Strong correlations
35 at different scales and between different variables are also detected (as for example $T_{mean_{30}}/T_{max_{90}}$ and
36 $T_{mean_{30}}/T_{min_{90}}$). Similarly to T_{mean} , ΔT_{mean} correlates strongly with ΔT_{max} and ΔT_{min} at the same temporal
37 scale (1, 3 and 6 days), and across different temporal scales (e.g., $\Delta T_{mean_3}/\Delta T_{mean_6}$, $\Delta T_{min_1}/\Delta T_{min_3}$ and
38 $\Delta T_{min_3}/\Delta T_{min_6}$). Also precipitation values strongly correlate at the daily/weekly and monthly/quarterly scales.
39 As can be seen in Fig. 5 and 6, ΔT shows values that are often below the expected significance: for this reason,
40 ΔT is discarded from the following analyses.

41 Starting from the couple of variables with the highest correlation (Fig. 6), we perform the pruning process
42 as in Section 4.3. We first eliminate those variables contributing less to the explanation of slope failure
43 occurrence. The pruning process proceed until all variables in the final subset have an absolute value of the
44 correlation coefficient lower than 0.5. In case of similar potential anomalies recognized by the two variables,
45 we select the variable with the highest availability of data. Considering the greater availability of weather
46 stations recording T_{mean} compared to T_{max} and T_{min} , we select the former variable as being the most

1 representative, along with R . For these variables the non-exceedance probability is greater than the expected
 2 significance only in the upper-tail (Fig. 5), thus we focus solely on positive anomalies. The choice of the
 3 variables to be selected is thus a matter of achieving a suitable number of representative variables and the
 4 need to minimize redundancy. In the end, we decrease the dimensionality of the problem to four variables,
 5 two for the short-term range and two for the longer one: i) $Tmean_1$, ii) $Tmean_{30}$, iii) R_1 , and iv) R_{30} .

6 **Fig. 6** Pairwise Pearson correlation coefficients among the 25 climate variables used for this work. Darker colors indicate
 7 stronger positive (numbers in bold) and negative (numbers in bold italic) correlation, respectively, whereas lighter colors
 8 indicate weaker correlation.

9 The final results related to the four selected variables are illustrated in the form of contingency table as
 10 in Table 1. Each cell represents the frequency distribution of one or more variables at a time. Values on the
 11 diagonal refer the percentage of events whereon only the variable in the i th row ($i=1..4$) resulted to be
 12 significant, whereas each off-diagonal element is the percentage of events whereon the variable in the i th row
 13 is significant in concomitance with another variable in the j th column ($j=1..4$), $j \neq i$. The last column indicates
 14 the percentage, out of the total sample size, for which the variable in the i th row is detected as statistically
 15 significant, alone or in association with the other variables in the j th columns. In other words, this percentage
 16 indicates the percentage of case studies where the variable is a potential driver of slope instability.

17 In order to guarantee a statistical homogeneity of the sample, we now consider only the case studies for
 18 which both $Tmean$ and R data are available i.e., 317 out of the total number of case studies (358). As an
 19 example in Table 1, $Tmean_1$ is significant in 10.7 % of the analyzed 317 events, 1.6 % of events show a double
 20 anomaly in both $Tmean_1$ and R_1 , while 18.5 % of events in total are related to a significant anomaly in the
 21 $Tmean_1$. As can be seen from Table 1, results are well above the expected significance for all the variables.
 22 Precipitation at the long-term shows the most evident signal, which is associated with 21.3 % of case studies,
 23 followed by a short-term high temperature (18.5 %), heavy short-term precipitation (17.6 %) and, finally, long-
 24 term positive temperature anomaly (15.7 %).

25 **Table 1** Contingency table among the selected climate variables and related statistics out of 317 case studies

Selected climate variables					
	$Tmean_1$	$Tmean_{30}$	R_1	R_{30}	Total
$Tmean_1$	10.7 %	4.4 %	1.6 %	1.9 %	18.5 %
$Tmean_{30}$	4.4 %	9.1 %	1.6 %	0.6 %	15.7 %
R_1	1.6 %	1.6 %	6.9 %	7.5 %	17.6 %
R_{30}	2.2 %	0.6 %	7.5 %	11.0 %	21.3 %

26

27 5.3. Spatiotemporal distribution vs key variables' climate anomalies

28 Results of the statistical analysis are shown in the following Figs. 7-8. Based on the results of the
 29 dimensionality-reduction performed in Section 5.2, we select two climate variables that are representative of
 30 short-term temperature and precipitation anomalies (T_1 and R_1) and two for the longer-term (T_{30} and R_{30}). A
 31 slope failure event could be related to more than one climate variable at a time. In other words, a combination
 32 of climate anomalies could be detected for a specific case study e.g., T_1 and R_{30} or T_1 and T_{30} . Thus, to better
 33 interpret the climatic framework leading to slope failure, a clusterization of the selected variables in eight
 34 groups is performed, as follows.

- 35 • T_1 (short-term positive temperature anomaly);
- 36 • T_{30} (long-term positive temperature anomaly);
- 37 • T_1 and T_{30} (wide-spread positive temperature anomaly);
- 38 • R_1 (short-term precipitation anomaly);
- 39 • R_{30} (long-term precipitation anomaly);

- 1 • R_1 and R_{30} (wide-spread precipitation anomaly);
- 2 • T and R (both temperature and precipitation anomaly of any type);
- 3 • Other or no anomaly (no anomaly in the four selected variables).

4 Note that 180 events out of 317 are related to almost one anomaly in the four selected variables. The
 5 remaining 137 are linked to statistical anomalies in the other 21 discarded variables or unrelated to climate
 6 forcing (Other or no anomaly group). In this Section, we present the main results, further analyses are included
 7 in the Online Resource 2, and briefly illustrated at the end of this paragraph.

8 Results shown in Fig. 7 highlight that summer events definitely prevail (58.7 % out of 317 events), whereas
 9 winter events are the smallest group (3.2 % out of 317 events). Proportions are almost the same if compared
 10 to the initial sample of 358 events, with 20 %, 57.5 %, 19 % and 2.8 % of case studies occurring in spring,
 11 summer, autumn and winter, respectively. Most of the events occurred in spring (Fig. 7a) are associated with
 12 an anomaly in precipitation values (41 %), mainly in combination with prolonged precipitations (R_{30}), while T
 13 anomalies are detected only for 15 % of the case studies. Summer events (Fig. 7b) are almost equally
 14 distributed between T (25 %) and R (24 %) anomalies. In autumn, temperature plays a major role, in
 15 combination or not with precipitation (Fig. 7c), and this is even more clear for winter events, even if in this
 16 latter season the sample of case studies is very limited (Fig. 7d). In general, positive temperature anomalies
 17 are relevant in spring and summer months. The case studies not associated with anomalies in the four variables
 18 selected are distributed homogeneously among seasons.

19 **Fig. 7** Climate anomalies based on the selected climate variables across season of occurrence. Short-term
 20 temperature anomaly: T_{mean1} , long-term temperature anomaly: T_{mean30} , wide-spread temperature anomaly: T_{mean1}
 21 and T_{mean30} , short-term precipitation anomaly: R_1 , long-term precipitation anomaly: R_{30} , wide-spread precipitation
 22 anomaly: R_1 and R_{30} , both temperature and precipitation anomaly: R and T , no detected anomaly in the four selected
 23 variables: Other or no anomaly

24 In Fig. 8 we analyze climate anomalies' distribution across four elevation ranges, obtained from a
 25 homogeneous subdivision based on the minimum and maximum height of occurrence of the case studies. It
 26 has to be pointed out that most debris/mud flows are located in the lowest elevation range, since, in general,
 27 the exact initiation point is hardly documented and only information on the deposition area, where damage
 28 usually occurs, is available. As can be seen, precipitation anomalies prevail at lower elevations, whereas the
 29 presence of positive temperature anomalies is more and more evident at higher elevations. Precipitation
 30 anomalies of different type (short/long term, widespread) are detected in the lowest range (Fig. 8a), whereas
 31 the long-term one is predominant in the mid-range (Fig. 8b). Case studies showing no anomaly in the four
 32 selected variables are mainly located in the mid-range (Fig. 3c).

33 **Fig. 8** Climate anomalies based on the selected climate variables across elevation of occurrence. Short-term
 34 temperature anomaly: T_{mean1} , long-term temperature anomaly: T_{mean30} , wide-spread temperature anomaly: T_{mean1}
 35 and T_{mean30} , short-term precipitation anomaly: R_1 , long-term precipitation anomaly: R_{30} , wide-spread precipitation
 36 anomaly: R_1 and R_{30} , both temperature and precipitation anomaly: R and T , no detected anomaly in the four selected
 37 variables: Other or no anomaly

38 With regard to the type of slope instability, we grouped the case studies in two main clusters, the first
 39 including rock/blockfalls, rock avalanches, landslides, icefall/avalanches, soli slips and slides (from now on,
 40 "landslides") and the second including debris/mud flows and Glacial Lake Outburst Floods (from now on,
 41 "debris/mud flows"). Landslides are mainly associated to positive temperature anomalies (25 %, if considering
 42 short/long term and wide-spread anomalies) and to long-term precipitation anomalies (12 %), whereas short-
 43 term precipitation anomalies slightly prevail in association with the occurrence of debris/mud flows. Overall,
 44 in the case of debris/mud flows, a major combined contribute of precipitation and temperature is detected
 45 (12 % of events) with respect to landslides (5 %). Most part of case studies showing no anomaly in the four
 46 selected variables are landslides (48 %). Graphs are available in the Online Resource 2.

1 Case studies have been grouped in four classes according to slope aspect at the initiation point (Online
2 Resource 3): no strong indication was detected of a preferential distribution of the events among the different
3 climate anomalies in relation to slope aspect, with the exception of a 35 % of N-facing slope instabilities
4 occurred in association with a temperature anomaly (mainly long-term), in combination or not with
5 precipitation. The case studies not associated to anomalies in the four variables selected are distributed
6 homogeneously among slope aspects.

7 With regard to the magnitude of the events (Online Resource 4), we have data only for about 40 % of the
8 case studies, i.e. 128 events out of 317: for this reason, and for the uncertainty that is sometimes associated
9 with volume assessment, only two classes of volume have been defined, i.e. small ($< 1000 \text{ m}^3$) and large (\geq
10 1000 m^3) slope instability events. Small-volume events are more numerous and almost equally distributed
11 among temperature and precipitation anomalies. Large-volume events are more often related to extreme
12 temperatures (33 % of events, by summing short, long and wide-spread anomalies). As can be seen in the
13 Online Resource 4, almost half of the case studies in both classes of volume are not associated to anomalies
14 in the selected variables.

15

16 6. Discussion

17 The main results of the study are hereinafter discussed by re-connecting to the main research questions
18 raised in the introduction.

19 i) Can temperature and precipitation be considered as key conditioning and/or triggering factors for slope
20 failures at high-elevation sites? Which climate variables are the most relevant for slope instability?

21 According to our analyses, almost 92 % of the case studies is associated with one or more
22 anomalies in the climate variables. Obviously, this does not necessarily imply a direct and univocal
23 cause-effect relation between anomalous values of the climate variables and initiation of slope
24 instability. However, this result is a clear evidence that climate variables (both T and P , at various time-
25 scales) are key factors for slope instability. At the same time, our analysis sheds light on the high
26 heterogeneity of the type of anomaly/ies detected on occasion of the occurrence of a slope instability
27 event, confirming what is known about the complexity of the diverse mechanisms leading to slope
28 failure. In such a framework, the number of case studies collected in this dataset provides the ground
29 for robust and significant conclusions about the nexus between climate variables and slope failure
30 occurrence. This result is an improvement with respect to Paranunzio et al. 2016, where the number
31 of considered events was 41 compared to 401 in the present paper. Paranunzio et al. (2016) were able
32 to associate a climate anomaly to 85% of the case studies, which is consistent with the result obtained
33 here (92%), also in view of the fact that in the present paper we have enlarged the set of considered
34 climatic variables.

35 Through the dimensionality reduction, we detect four key climate variables: T_{mean_1} , $T_{mean_{30}}$, R_1
36 and R_{30} . Significant positive climate anomalies prevail as the drivers of slope instabilities, i.e. heavy
37 precipitation and high temperatures. Precipitation at the long-term shows the most evident signal.
38 Heavy, prolonged precipitations act in depth, altering soil moisture and slope hydrological conditions
39 (Ma et al. 2014). Rainfall is a recognized trigger of landslides (e.g., Jakob and Lambert, 2009), and the
40 role of heavy prolonged rainfall has been widely discussed in the past (e.g., Luino 2005). In the recent
41 years, the role of temperature as a potential trigger of slope instability has been investigated (e.g.
42 Allen and Huggel 2013) and statistically verified by Paranunzio et al. (2016). The results of this work,
43 based on a larger sample, confirm this evidence. Our results show that most of the events is associated
44 with a temperature anomaly, in combination or not with precipitation non-standard values (47 % and
45 38 %, respectively). These results strongly support the hypothesis, already put forward by many

1 authors (e.g., Geertsema et al., 2006), that global warming can be deemed responsible for an increase
2 of slope instability, in particular in high elevation/latitude areas, where the cryosphere plays a crucial
3 role in conditioning slope dynamics.

4 The dimensionality reduction, on the one hand, entails a cost in terms of ability to associate a
5 specific case study to a climate anomaly, but, on the other hand it has many advantages. First of all,
6 this approach allows one to identify climate anomalies also when few climate variables are available.
7 This is the case of most of the weather stations located in high-mountain areas: often these weather
8 stations register only few variables (mostly *Tmean*, as evidenced in the previous sections) and we can
9 thus rely on a limited set of climate records. Moreover, this dimensionality reduction procedure
10 facilitates the visualization, classification and interpretation of the results.

11 ii) How can the detected climate signal be linked to the process typology and to its spatiotemporal
12 distribution?

13 As already mentioned, in our sample “landslides” (including all different types, 239 case studies)
14 sharply prevail on “debris/mud flows” (78 case studies). This asymmetric distribution is mainly due to
15 the fact that most information about debris/mud flows is related to the transition/accumulation area
16 (where damage is produced), rather than to the starting zone. This means that, if the
17 damage/accumulation/transition area was at an elevation lower than 1500 m, the event was
18 discarded. Keeping this bias in mind, we notice that debris/mud flows are better explained by the four
19 key climate variables than landslides, with a major role played by precipitations, at all temporal scales.
20 This picture, on one side, confirms some already well-known facts. Debris/mud flow initiation is mainly
21 driven by precipitation (e.g., Jakob et al. 2012), even if their occurrence is the result of a combination
22 of not only water (mainly from rainfall, but occasionally from snow/ice melt or GLOF), but also debris
23 availability (from landslides, scarp and/or channel bed erosion; at high-elevation, (post)glacial deposits
24 may represent an important sediment source areas, Turconi et al. 2010). Landslides, instead, are the
25 result of more complex processes, which depend on landslide type. On the other side, our results
26 suggest that, at high elevations, temperature plays a crucial role not only in landslide initiation, as it
27 has been widely recognized, but also in debris/mud flow initiation: temperature’s contribution to the
28 initiation of this latter type of process is something that is less understood and debated (Stoffel et al.,
29 2011). On this regard, it is interesting to notice that, among temperature anomalies, long-term ones
30 prevail on short-term ones for debris/mud flows, while the opposite is true for landslides: according
31 to this, we may speculate that high temperatures are mainly a preparatory factor for debris/mud flows
32 (e.g. through snow/ground-ice melt which saturate the debris, Wieckzoreck and Glade 2005,
33 Geertsema et al. 2014) and a triggering factor for landslides (e.g. by snowmelt, Cardinali et al. 2000).

34 Half of the case studies are concentrated in the lowest elevation range, and the number of case
35 studies decreases rapidly with elevation. Low-elevation areas are in fact wider and more frequented,
36 and the probability that slope instabilities are reported is therefore greater. Despite the relative lack
37 of data at the highest elevations, an increasing role of positive temperature anomalies with elevation,
38 on extended timescales, clearly emerges: in particular, at the highest elevations (> 2890 m a.s.l.)
39 precipitation is not a significant forcing for slope instability. This is a clear indication of the crucial role
40 played by cryosphere dynamics in the development of slope instability in high mountains (Deline et al.
41 2015). Long-term temperature anomalies may be responsible for permafrost degradation/thawing in
42 depth (Gruber and Haeberli, 2007): interestingly, this is the most significant climate variable at the
43 highest elevations (> 3500 m a.s.l. approximately). Short-term temperature anomalies affect near-
44 surface dynamics: in permafrost environments, temperature variations and short-term extremely
45 warm conditions could affect rock stability within hours through rapid thawing processes (Hasler et al.
46 2012). It is also interesting to notice that only a small part (39%) of case studies in the elevation range

1 2890-3585 are associated with anomalies in the four key climate variables. An explanation for this may
2 be in the fact that at this elevation range, in recent times, the most important changes in permafrost
3 and glacier extent occurred: additional processes, developing at time scales larger (pluriannual) than
4 those investigated here, may be responsible for slope instability occurrence, e.g. slope debulking
5 as a consequence of glacier retreat (Geertsema and Chiarle 2013).

6 As for the seasonal distribution, summer events sharply prevail, whereas spring and autumn ones
7 are almost balanced; winter events are the clear minority. The concentration of case studies in
8 summer is partly due to some inhomogeneity in data reporting, considering that frequentation of high
9 mountain areas is the highest in this season. Besides this, we can observe that most of the spring
10 events occur in the presence of some extraordinary precipitation (41 %), whereas this type of anomaly
11 decreases gradually in summer and autumn, to almost disappear in winter. In this last case, we have
12 to consider that winter precipitation recorded by high-elevation weather stations is only partly
13 reliable, because of undercatch bias when precipitation is in solid form (Buisan et al. 2016). It is
14 interesting to notice how the relative importance of precipitation among seasons only in part reflects
15 the pattern of climate variables during the year. In particular, precipitation is as important as
16 temperature during summer, when we might expect a predominant role of temperature. On the
17 opposite, temperature is more relevant than precipitation in autumn, a season generally associated
18 to heavy precipitations. In this regard, it has been observed that permafrost active layer reaches
19 maximum depths in late autumn, when the ground surface is already in freezing conditions (Magnin
20 et al. 2015): this situation may lead to water pressure build-up in the slope, up to its failure. The
21 association of winter events with short-term temperature anomalies is the most difficult to explain,
22 even if the number of case studies for this season is so little, that any outcome has to be considered
23 very carefully. For these cases, we may speculate that, on steep slopes, where only little snow can
24 accumulate, short-term warm conditions may cause snowmelt, able to trigger the slope instability.
25 Since winter events are in general large events, we should consider a water input from snowmelt only
26 as the trigger of unstable conditions, generated perhaps by water pressure build-up, as discussed for
27 autumn events.

28 The distribution of case studies and of the related anomalies in relation to slope aspect does not
29 reveal any relevant pattern. Temperature and precipitation anomalies at various temporal scale are
30 quite homogeneously distributed on the different slope aspects. Only the north-facing slopes show a
31 slightly higher sensitivity to temperature anomalies, in particular to long-term ones (15%). This is in
32 agreement with the findings of some studies that identify north slopes as the most sensitive to
33 temperature increase, because of the thinner permafrost active layer. For east-facing slopes, a relative
34 importance of short-term temperature (15%) and long-term precipitation (16%) anomalies is
35 highlighted: the significant association of events with short-term temperature anomalies may be
36 related to the higher solar radiation received by these slopes. For these types of analyses it would be
37 important to consider among climate variables also solar radiation, for which however, at the moment
38 only few data are available for the Italian Alps.

39 The analysis of anomalies' distribution in relation to the size of slope instability highlights how the
40 four key climate variables are less able to catch a climate signal in association with large events, than
41 with small events. This quite predictable outcome can be explained by the higher complexity of
42 processes and mechanisms involved in the occurrence of large-scale slope instabilities (Crozier, 2010).
43 What is interesting, however, is that, quite surprisingly, very few large-volume case studies are
44 associated with precipitation, while temperature, and in particular short-term anomalies (18%),
45 appears to be a significant climatic driver. We may conclude that, for large-volume events, our
46 approach is able to catch the climate anomaly eventually associated to the triggering of slope
47 instability but cannot shed light on the complex set of processes involved in its setup. A different

1 approach, considering more extended (annual/pluriannual) time scales and predisposing factors, such
2 as the lithological and structural setting (Fischer et al. 2013), not directly related to climate, would be
3 necessary, but this is out of the scope of this work.

4 iii) Can climate change be deemed responsible for the observed increasing trend of slope instability at
5 high elevation?

6 The answer to this question is very complex. In order to respond unambiguously, we should have,
7 as is the case of climate data, long-term datasets, allowing the identification of trends in the
8 occurrence of slope instability. Unfortunately, even today, the reporting of these events strongly
9 depends on the associated damage/risk, so that the available data series are incomplete and
10 inhomogeneous, and thus unsuitable to provide trends to be compared with climate variations. Even
11 if some authors attempted to fill this gap using different approaches/techniques, results are
12 nevertheless partial and/or of local value. In addition, as already mentioned, slope instability is the
13 result of a complex set of processes, that respond with different velocity and amplitude to climate
14 change.

15 In this complex framework, the results of this study, while not being able to unambiguously
16 prove/disprove the role of climate change on slope instability increase at high elevation, strengthen
17 this hypothesis. Three out of the four key variables detected can be attributed to pattern of climate
18 change and global warming scenarios (i.e., T_1 , T_{30} and R_1). The robust sample investigated in this work
19 and the high number of case studies occurred in the presence of positive temperature anomaly
20 support the hypothesis of a climate signal in the initiation of mass-wasting processes at high-elevation
21 sites, as suggested by previous studies (e.g. Allen and Huggel 2013). Based on the results of this work,
22 precipitation at the longer-term scale (R_{30}) is the main climate forcing related to slope failure
23 occurrence. According to the scenarios of climate change, reduced total amount of precipitation in
24 the Southern European Alps are expected (Brunetti et al. 2009; Gobiet et al 2014) and, thus, also a
25 reduction of slope instabilities induced by prolonged abundant precipitation is hypothesized (e.g.
26 Dehn et al., 2000). Conversely, the effect of intense short-term precipitation (as R_1) on the initiation
27 of slope instability events could be more and more evident in the next future (Gariano and Guzzetti
28 2016).

29 As cryosphere degradation proceeds up to its complete disappearance, we might anticipate a
30 decreasing impact of global warming on slope stability at high elevation/altitude. However, taking into
31 account recent studies highlighting the role of diurnal thermal stressing in unstable rock masses also
32 in non-cryospheric areas (Collins and Stock 2016), where warming-cooling cycles can gradually affect
33 rock mechanics, leading to slope failure, we may conclude that T , and in particular global warming will
34 continue to impact on slope instability also in a scenario of a vanishing cryosphere.

35 If the role of extraordinary warmth in destabilizing rock mass has been widely analysed by recent
36 works (e.g. Gruber and Haeberli 2007; Collins and Stock 2016), little attention is paid to negative
37 anomalies as potential drivers of slope instability. Our analysis pointed out that a significant number
38 of events were associated with negative values of ΔT , i.e. with sudden temperature drops in the day(s)
39 preceding the failure. Build-up of water pressure (e.g. freezing of water springs) or rock damage due
40 to freezing-thaw cycles are among the different mechanisms that, in association with temperature
41 drops, can lead to slope failure, depending, among others, on the type of instability process, season
42 of occurrence, lithological and geomorphological features (Fischer et al. 2012). Climate change may
43 influence also these processes, through an increase of temperature variability (Schar et al. 2004), and
44 thus of the probability of sudden temperature drops/raises.

1 In the end, we recall some important points and constraints that have to be kept in mind, when analyzing
2 the results of this work. First of all, we base our analysis on a relatively short-period (from 2000 on): this is due
3 to the need of disposing of sufficient information on temporal and spatial localization of the events, and of
4 reliable and consistent data from the weather stations.

5 We are aware that, in mountain regions, many factors affect the measure of the climate variables at a
6 site, complicating the climatological framework whereon we operate, as the scarce coverage of long-term
7 weather stations at high-elevation in Italy. In these remote areas, automatic weather stations have been
8 installed recently and, in general, only cover the last 15-20 years (Pepin et al. 2015). This entails relying
9 sometimes on measuring stations far from the study area and, thus, not fully representative of the climate
10 conditions of the detachment area. This is particularly evident for precipitation, which is affected by a larger
11 spatial variation in high-mountain areas with respect to temperature (Isotta et al. 2014). To limit these
12 problems, as illustrated in Section 3.3, we first fix a series of requirements when selecting weather stations, in
13 terms of data availability, length of the historical data series and distance from the failure area. Moreover, the
14 method as is allows one to detect the climate anomaly directly at the station, thus overcoming the problem
15 related to the elevation of the instrument.

16 The data heterogeneity, due to the fact that in Italy there are several meteorological data source, and the
17 different length of the historical data series could introduce bias into the records (Merlone et al. 2015).
18 However, this does not affect the estimation of the probability values, since we compare the value of the date
19 of failure occurrence to climate data recorded at the same reference instrument and not among different
20 weather stations. The availability of new products based on a merging and/or combination of gridded data
21 and in situ climate records could be a way to partially overcome the shortcoming related to lack and
22 inconsistency of climate data, but the relatively low spatiotemporal resolution of remotely sensed records with
23 respect to our scale of analysis is another major limiting factor in this context (Mountain Research Initiative
24 EDW Working Group 2015).

25 6. Conclusions

26 In this work, we collected an inventory of 401 slope instability events and finally analyze 358 case studies,
27 documented from 2000 to 2016 in the Italian Alps, at elevations above 1500 m, with the aim to assess the role
28 of climate forcing, and an eventual signal related to climate change. This dataset is something unique in the
29 world for mountain regions, considered not only the number of case studies, but also the quality of their
30 spatiotemporal localization. First of all, we tested on this robust and diverse dataset the statistical approach
31 proposed by Paranunzio et al. (2015, 2016), which allows one to define in a standardized way the climate signal
32 behind a slope failure event. In order to make a step forward towards a quantitative attribution of the effect
33 of global warming in the observed increasing mass-wasting activity at high elevation, in this work we
34 implemented a procedure for the identification of the essential climate variables associated with slope failure
35 occurrence. Although some critical points still remain, as outlined in the previous paragraph, some important
36 conclusions can be drawn from this work, as listed hereinafter.

- 37 i) More than 90 % of the 358 investigated events occur in the presence of one or more of the 25
38 climate anomalies considered. For this high-elevation dataset, temperature was confirmed as a
39 fundamental climate variable: in 47 % of cases, we detect a temperature anomaly, in 38 % a
40 combination of temperature and precipitation anomalies, and only in 7 % of cases solely a
41 precipitation anomaly.
- 42 ii) The dimensionality reduction from 25 to 4 key climate variables reduced the sample size to 317,
43 but allowed us to decrease the noise created by redundant information and to catch the most
44 evident climate signals behind slope instability.
- 45 iii) The key climate variables that resulted to have positive anomalies in association with 57% of the
46 317 case studies are: $Tmean_1$ and $Tmean_{30}$, R_1 and R_{30} . Precipitation at the long-term shows the

1 most evident signal (21.3 % of case studies associated to anomalous values of this variable),
2 followed by short-term temperature anomalies (18.6 %).

- 3 iv) Considering the four above-mentioned key climate variables, an evident signal related to the
4 season and elevation of occurrence and to the type of process and size of the event, emerges.
5 More specifically: the role of precipitation decreases (and that of temperature increases) from
6 spring (41%) to winter (0%), and from the low (42%) to the high (5%) elevations. Debris/mud flow
7 occurrence is well related to precipitation anomalies (47%) compared to landslides (27%), but,
8 surprisingly, the same percentage (25%) of debris/mud flows and landslides occur in association
9 to positive temperature anomalies. Finally, small volume events are better explained by the 4 key
10 selected climate variables than large events: however, these latter appear to be much more
11 sensitive to temperature anomalies (33%) than to precipitation (4%).

12 The high occurrence of positive temperature anomalies in the lead-up of a failure, associated or not with
13 heavy precipitation, supports the hypothesis of a role of climate warming in the occurrence of mass-wasting
14 processes at high-elevation sites in recent years. This evidence is also confirmed by the different distribution
15 of temperature and precipitation anomalies across season and elevation of occurrence. According to past
16 climate trends and future projections, we can expect that, for the Italian Alps, slope instability driven by
17 positive temperature anomalies will become more and more important, while processes related to long-term
18 precipitations will lose relevance.

19 In conclusion, the statistical approach proposed here represents a standardized method, which can be
20 applied to different contexts, implemented with additional variables (e.g. solar radiance) and used to compare
21 climate change impact on different natural processes. In the field of geohazards, the interpretation of the
22 mechanisms leading to slope failure in the light of the main climate anomalies detected represents a
23 challenging avenue for future research and an essential step for the knowledge and management of global
24 warming impacts.

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14 Figure Captions

15 **Fig. 1** Map showing 401 slope instability events included in the catalogue (squares) and 131 weather
 16 stations used in this work (dots); type and sample size of the instability processes included in the catalogue
 17 are represented in the white box; RF: rockfall, BF: blockfall, RA: rock avalanche, DF: debris flow, MF: mud flow,
 18 L: landslide, SL: slide, IF; ice fall, IA: ice avalanche, GLOF: glacial lake outburst flood, S: (soil) slip

19 **Fig. 2** Flowchart representing the main steps of the method as in Section 4

20 **Fig. 3** Empirical Cumulative Distribution Function (ECDF), denoted as $q(j)$, based on the probability values
 21 $P(V)$ of a j th variable for the entire sample (312 events in this case), as detailed in Section 4.2. The straight line
 22 indicates the bisector, whereas the curve indicates the ECDF. 19.5 % of values lies above the 90th percentile,
 23 whereas only 5.4 % of values are in the lower-tail (10th percentile).

24 **Fig. 4 a)** Number of climate anomalies per variable and percentage out of the total number of events (358).
 25 Lighter colors refer to positive anomalies (high extremes, heavy precipitation), darker color to negative
 26 anomalies (low temperatures). Numbers above the stacked bar, from the bottom to the top: number of
 27 available weather stations per variable, number of positive anomalies per variable, number of negative
 28 anomalies per variable. **b)** Number of events (in bold) showing from 0 (no anomaly) to 19 climate anomalies
 29 and percentage out of the total number of events (358)

30 **Fig. 5** Empirical Cumulative Distribution Function (ECDF), $q(j)$, based on the probability values $P(V)$ of the
 31 j th variable ($j=1\dots L$, $L=25$) as detailed in Section 4.2. The straight lines indicate the bisectors, whereas the curves
 32 indicate the ECDF. Numbers in the lower left and upper right sectors refer to percentages of data below the
 33 10th and above the 90th percentile, respectively

34 **Fig. 6** Pairwise Pearson correlation coefficients among the 25 climate variables used for this work. Darker
 35 colors indicate stronger positive (numbers in bold) and negative (numbers in bold italic) correlation,
 36 respectively, whereas lighter colors indicate weaker correlation

37 **Fig. 7** Climate anomalies based on the selected climate variables across season of occurrence. Short-term
 38 temperature anomaly: T_{mean_1} , long-term temperature anomaly: $T_{mean_{30}}$, wide-spread temperature
 39 anomaly: T_{mean_1} and $T_{mean_{30}}$, short-term precipitation anomaly: R_1 , long-term precipitation anomaly: R_{30} ,
 40 wide-spread precipitation anomaly: R_1 and R_{30} , both temperature and precipitation anomaly: R and T , no
 41 detected anomaly in the four selected variables: Other or no anomaly

1 **Fig. 8** Climate anomalies based on the selected climate variables across elevation of occurrence. Short-
2 term temperature anomaly: $Tmean_1$, long-term temperature anomaly: $Tmean_{30}$, wide-spread temperature
3 anomaly: $Tmean_1$ and $Tmean_{30}$, short-term precipitation anomaly: R_1 , long-term precipitation anomaly: R_{30} ,
4 wide-spread precipitation anomaly: R_1 and R_{30} , both temperature and precipitation anomaly: R and T , no
5 detected anomaly in the four selected variables: Other or no anomaly

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