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# GEOMORPHIC VARIATIONS OF DEBRIS FLOWS AND RECENT CLIMATIC CHANGE IN THE FRENCH ALPS

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**Abstract.** Much work has been done to show that there is a relationship between the triggering of debris flows and the recorded increase in temperatures or in the number of intense rainy events over the last few decades. The question addressed in this paper is that of the impact of these climate changes on the dynamics of debris flows since the 1950s. 319 debris flows in the Dévoluy and Ecrins massifs located in the French Alps, the triggering of which is independent of the current glacial retreat, have been analysed. In the Dévoluy a reduction in the number of debris flows was observed in the periods 1950–1975 and 1975–2000. In the Massif des Ecrins, we have observed a shift of the triggering debris flow zone toward higher elevations and a lack of significant variation in the number of debris flows. But in the Massif des Ecrins this global result masks two different trends depending on elevation. At low altitude (<2200 m) the number of debris flows and the frequency of debris flows less than 400 m in length have decreased significantly since the 1980s whereas no significant variation was observed at high altitude (>2200 m). At the same time, we have observed a significant increase in the annual and seasonal temperatures for these 20 last years combined with a significant reduction in the number of freezing days. A significant increase in summer rains higher than 30 mm/d has also been observed. In Dévoluy and at low altitude in the Massif des Ecrins, these variations can be explained by the decrease in the number of freezing days related to the increase in the temperatures, which implies a slower reconstitution of the volume of debris stored between two events. But at high altitude it is currently difficult to establish the link between the climatic change and the dynamics of the debris flows because very little is known about the two variables controlling the triggering of the debris flows, i.e., on the one hand intense precipitations and on the other hand the volume of rock debris.

## 1. Introduction

Hill slope debris flows can be defined as a rapid mass movement and are common in the Alpine environment. The geometrical and sedimentological characteristics of these fast flowing mixtures of sediment and water have been observed in a wide range of mountain environments (Van Steijn et al., 1988; Kotarba, 1989; André, 1990; Haeberli et al., 1990; Nieuwenhuijzen and Van Steijn, 1990; Luckman, 1992; Van Steijn, 1991, 1996).

At the same time laboratory measurements have enabled a better understanding of how this process operates and have provided various models (Johnson and Ro-



dine, 1984; Coussot, 1994; Bertran and Texier, 1994; Inverson, 1997; Major, 1997). Specifically, studies of the initiation of debris flows have shown the importance of stored volumes of debris and of climatic conditions (Caine, 1980; Blijenberg, 1998). Generally, an intense rainfall event triggers this process, although debris flows have also been triggered by the sudden release of water stored under glaciers or by the breaching of a morainic dam (Evans and Clague, 1994; Antoine and Fabre, 2001) or by melting snow (Wenshou and Cunhai, 1992).

Over the last few decades an increase in temperatures and changes in the amount and frequency of rainfall have been observed (Katz and Brown, 1992; IPCC, 1996; Bryant, 1997), and depend on the altitude (Beniston et al., 1997; Diaz and Bradley, 1997). As a consequence, many researchers have considered the impact of climatic change on periglacial processes (Nyberg and Lindh, 1990; Beniston, 1994; Evans and Clague, 1994; Haeberli and Beniston, 1998) especially on the triggering of debris flows. Some studies on the relation between climate and debris flows have therefore been carried out in northern Europe (Innes, 1983, 1985a,b; Rapp, 1995; Sandersen, 1997; Nyberg and Rapp, 1998) and in the Alps (Haeberli et al., 1990; Zimmerman and Haeberli, 1992; Rebetz et al., 1997). These studies show a relationship between the triggering of debris flows and an increase in the number of intense rainfall events and/or glacier retreat and permafrost degradation due to an increase in temperatures since the middle of the 19th century.

However, research into the impact of this climatic change on debris flow frequency-magnitude or on occurrence remains insufficient. In the French Alps, for example, the only research done on this subject was in a valley located in the southern part of the French Alps (Van Steijn, 1996; Blijenberg, 1998). This work showed that activity has been high since 1980 (especially in 1987) but results do not necessarily point to a recent increase in debris flow activity. In the Dolomites, Strunk (1992) says 'there is no evidence of any trend towards an increase or a decrease in the frequency since the end of the last century'.

In the Swiss Alps, studies concerning a large number of events over a wide area focused on the year 1987 because of its exceptional character (Haeberli et al., 1990; Zimmerman and Haeberli, 1992). Rebetz et al. (1997) have observed that the frequency of large debris flows has increased since the late 1980s but unfortunately the low number of events (9 cases), limited geographically to the Ritigraben area, and the link between the triggering of these debris flows and the melting of permafrost, means that these observations cannot be generalised. Consequently, additional observations made from a wide sample of debris flows in other sectors of the Alps are needed to confirm or refute these preliminary predictions especially in locations not recently deglaciated.

The aim of this paper is to illustrate geomorphic variations of debris flows since the 1950s with respect to climatic change in two areas of the French Alps. This study was conducted on a large sample of debris flow deposits located in two areas, the south of Dévoluy (plateau de Bure and its neighbouring valleys) and the northern part of the Massif des Ecrins.

## 2. Area Studied

Debris flows were selected in two areas. The first area includes the plateau de Bure and its neighbouring valleys (between  $44^{\circ}36' N$  and  $44^{\circ}49' N$ ), and is about  $26 \text{ km}^2$  in area. It is located on the south edge of the Massif du Dévoluy, which is the highest massif of the French Pre-Alps (Pic de Bure, 2709 m asl, Obiou, 2790 m asl) (Figure 1).

The sides of the plateau de Bure and the neighbouring valleys are made up of Senonian limestone which forms rock faces over 200 m high dominating active talus slopes (Pech, 1993, 1996) on which debris flows occur (Figure 2). These slope deposits are located between 1900 and 2400 m asl. Meteorological data acquired over 6 years on the plateau (2700 m asl) show that the  $0^{\circ}C$  isotherm is around 2500 m asl (Pech, 1996). At this elevation precipitation is around 2000 mm/y of which 70% occurs as snow essentially between November and May. In contrast to other studies in the Alps (Haeberli et al., 1990; Zimmerman and Haeberli, 1992; Rebetz et al., 1997), the originality of this site is that it has not been glaciated since the last glaciation. Thus glacial recession since the end of the Little Ice Age has not affected debris flow activity. However, we cannot exclude the influence of a possible melting of the permafrost.

The second area of  $28 \text{ km}^2$  corresponds to the northern part of the Massif des Ecrins ( $45^{\circ}00' N$ ,  $6^{\circ}30' E$ ). In this area (Figure 3), 11 valleys not covered by glaciers since the Little Ice Age (according to old documents and geomorphic observations) have been studied. These valleys consist of broad steep granite walls 200–300 m above talus slopes (Figure 4), which have already been studied (Francou, 1988; Jomelli, 1997, 1999; Jomelli and Francou, 2000; Jomelli, 2002). These deposits are situated between 1600 m asl and 2400 m asl which is close to the  $0^{\circ}C$  mean annual isotherm.

In both areas, the debris flows studied are triggered at the base of a cliff which has been eroded by a gully, and flow on slope deposits at the base of the cliffs. Consequently the flow is only partially chenalized except where debris flow uses an old channel formed on the slope by an old debris flow.

## 3. Data Sources

### 3.1. DEBRIS FLOW DOCUMENTATION

In the Dévoluy area, debris flows were observed by combining two approaches. Aerial photographs from 1948, 1952, 1956, 1971, 1978, 1980, 1982, 1989, 1993, 1997, 1999 and 2000 were analysed. The scale of these photographs was between 1:20,000 and 1:30,000. Field observations were made every year from 1995 onwards. These two approaches were used to date debris flows. However, old documents (*Office National des Forêts, Restauration des Terrains en Montagnes, Archives départementales*) and dendrochronology (Shroder, 1978) were

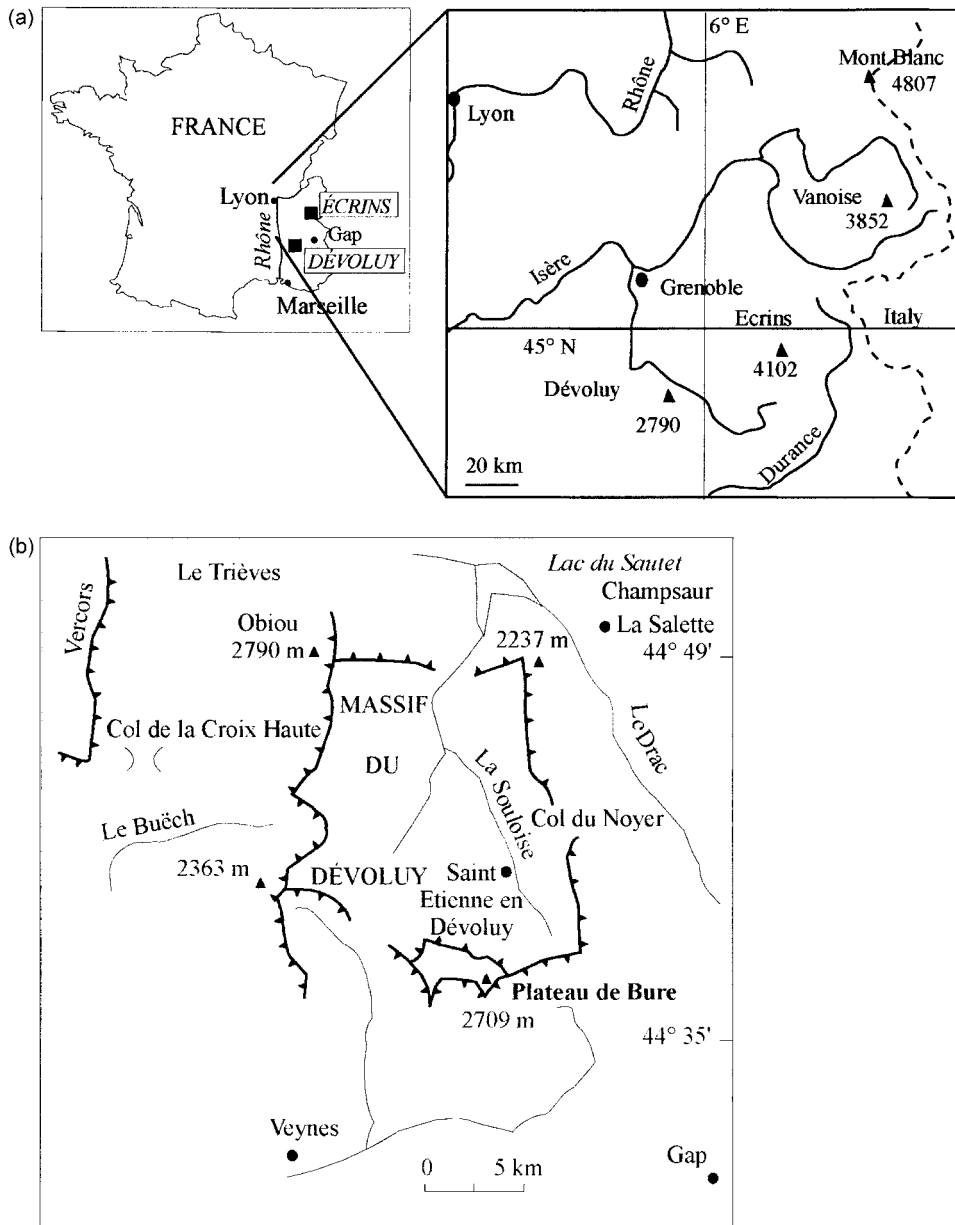


Figure 1. (a) General location map with the Massif du Dévoluy and Massif des Ecrins; (b) Detail of the Massif du Dévoluy.



Figure 2. Debris flows in the southern part of the Dévoluy (Combe Ratin, Pic de Bure).

also used to date the debris flows triggered between two aerial photographs. In dendrochronology, we worked from scars and wood reaction of *Larix decidua*. Samples were processed in accordance with the classical methods of dendrogeomorphology. Debris flows predating 1951 and those that were impossible to date (13, mainly before 1975) were not taken into account in the analysis.

For the Massif des Ecrins, the same methods were used. Aerial photographs from 1952, 1960, 1967, 1971, 1974, 1981, 1986, 1989, 1994, 1996, 1999 and 2000 were analysed. Seventeen debris flows (11 before 1975) which it was impossible to date to the year and those predating 1953 were rejected. For each dated debris flow, its length was measured in the field and the minimum altitude of the starting zone was estimated from photo interpretation and field observations (accuracy  $\pm 20$  m) (Tsao, 2000; Le Parcq, 2001; Chochillon, 2002). This minimum altitude of the starting zone corresponds in the couloir to the point from which the slope exceeds  $41^\circ$ . This value corresponds to the angle of repose of granite rock debris.

If the couloir had a regular slope but a gradient higher than  $41^\circ$  or if, on the contrary, it presented vertical sections intersected with subhorizontal sections, the limit which was selected corresponded to the altitude of the apex of the deposit.

### 3.2. METEOROLOGICAL DATA

The criteria when choosing the meteorological stations were three in number: they had to be the closest to the study site, the highest in altitude and have been observed

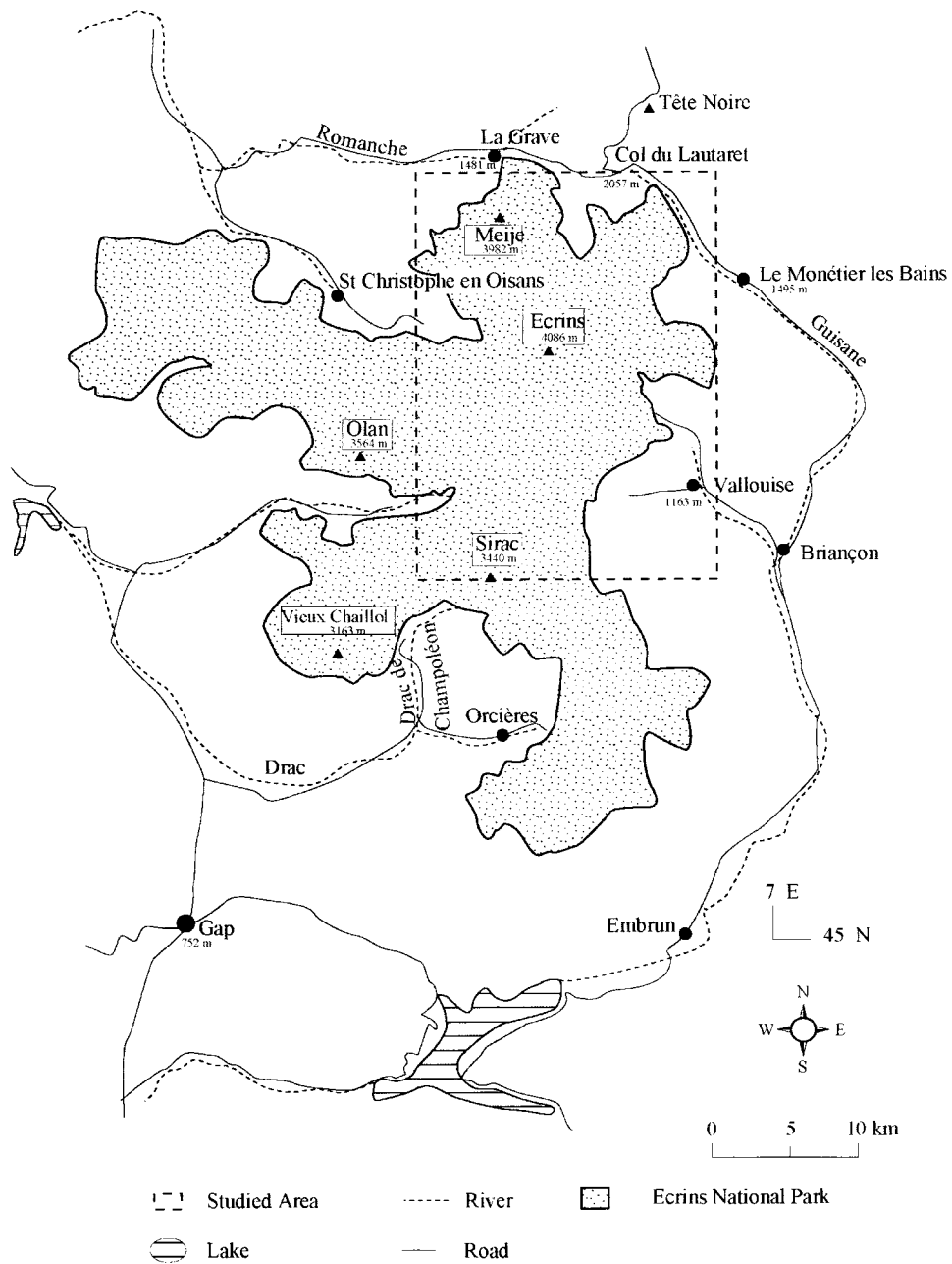


Figure 3. Location map of the Ecrins sector.



Figure 4. Northern face of the Fontenil valley on which several debris flows can be observed.

for a period of at least thirty years. In the Dévoluy area, we worked with La Salette (1770 m asl) (Figure 1) station, which provided temperature and precipitation data going back to 1965. For the Massif des Ecrins, we worked with Le Monétier-les-Bains (1490 m asl) and Saint Christophe en Oisans (1570 m asl) stations, whose data go back to 1961 and 1964 respectively.

We chose to use data collected from stations covering very large areas, in order to try to document the relationship between climatic change and debris flow variations. This has been done to improve the reliability of the results because there is not necessarily a relationship between local variation of the events studied and more global climatic trends.

### 3.3. WHY A REGIONAL APPROACH?

In this analysis we have chosen to work from a sample distributed over a larger spatial area, though the link between climatic change and variation of the dynamics is indirect, since weather measurements cannot be taken on the same site. However, given the number of observations taken into account and the spatial scale considered, the results are more representative.



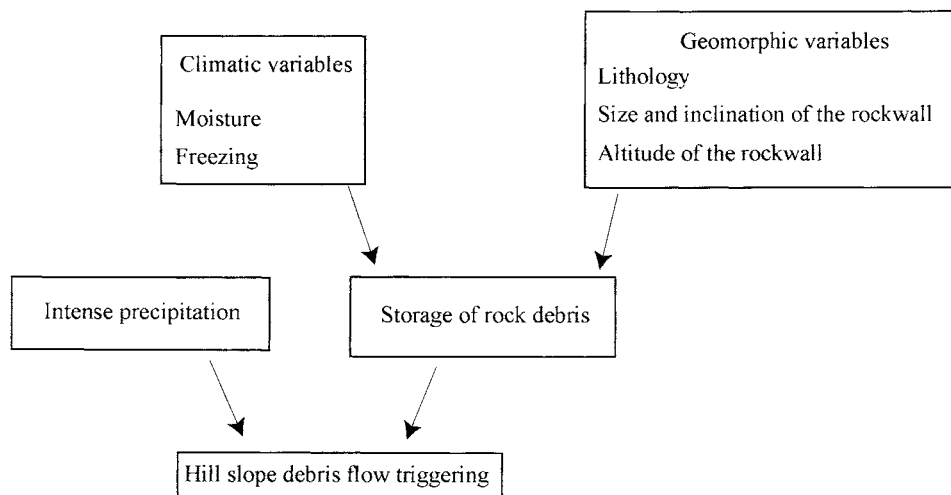


Figure 5. Factors controlling the triggering of hill slope debris flows in the French Alps.

#### 4. Triggering Factors in Debris Flow Occurrence

It is well known that two conditions are needed for debris flow to be triggered (Caine, 1980; Johnson and Rodine, 1984; Van Steijn, 1996; Iverson, 1997): long duration or high intensity of heavy rainfalls and a large volume of debris (Figure 5). The relationship between debris flow triggering and the duration – intensity of rainfalls has often been studied (Zimmermann, 1990; Zimmermann and Haerberli, 1992; Blijenberg, 1998). Caine (1980) and other authors have shown that the higher the intensity the shorter the duration. Even if the importance of stored volumes of debris has been well known for a long time (Statham, 1976), few authors have studied the relationship between stored volumes of debris and debris-flow triggering (Pech and Jomelli, 2001). These volumes of debris are due either to morainic accumulations on a slope (Haerberli et al., 1990; Rickenman and Zimmermann, 1993) or to debris accumulated in a couloir or at the apex of a slope deposit (Pech and Jomelli, 2001) by frost weathering.

The importance of spatial and temporal overlap between heavy rainfall and a stored volume of rock debris has already been underlined, its absence being a possible explanation for the absence of relationship between heavy rainfall and the triggering of debris flows (Pech and Jomelli, 2001). In other words, if at year  $t$  a flow is triggered, the probability of a flow being triggered at year  $t + 1$  will be low even if heavy rainfalls are observed because, if the stored volume is completely mobilised when the first flow is triggered, several years will *a priori* be necessary for it to be reconstituted.

After a flow has been triggered, the time needed for the store of rock debris to be reconstituted, when it is not related to glacial dynamics, depends on climatic and geomorphological parameters. The climatic parameters are linked to the tem-

peratures and precipitations. Generally the rate of this type of weathering depends on the frequency of freeze – thaw alternations inside the rock-wall, their duration and their intensity (Coutard and Francou, 1989). However, fragmentation of the rock by freezing occurs only when the rock has a certain water content (Matsuoka et al., 1997). This moisture can be related to rain or the melting of the snow cover. However, the time necessary for the reconstitution of the stock also depends on lithology, contributive surface and the altitude of the rock-wall (since temperatures vary according to altitude).

We have shown (Pech and Jomelli, 2001) that a debris flow needs both heavy rainfalls and the accumulation of debris. When a debris flow occurs, no other can be triggered immediately afterwards, even if high intensity rainfall events occur, because the debris was evacuated with the first event. The reconstitution of the accumulation is a long process which depends on climatic and geomorphic parameters, especially frost activity in the cliff (Coutard and Francou, 1989; Pech and Jomelli, 2001). The rate of this type of weathering depends on rock strength and local climatic conditions such as freeze – thaw alternation intensity.

In order to better understand the relationship between these parameters and the occurrence of debris flows, we recently (Jomelli et al., 2003) used a statistical model with data collected from the Massif des Ecrins. To build this model we needed a logistic regression (Aldrich and Nelson, 1984). The dependent variable  $Y_i$  is ordered and has values from 1 to  $k$ . The model based on cumulative probabilities is:

$$\text{Logit}(p_i) = f(\text{Pr}(Y_i \leq i|x)) = \alpha_i + \beta'x + e \quad (1)$$

$f(x)$  is the logistic distribution function, where  $i$  varies from 1 to  $k$ , the intercept  $\alpha_i$  varies from  $\alpha_1$  to  $\alpha_k - 1$ ,  $\beta'$  is the slope coefficient and  $e$  the error. The logistic distribution constrains the estimated probabilities to lie between 0 and 1. The cumulated probability  $p_i$  of the occurrence  $i$  is calculated from the equation:

$$p_i = \frac{e^{\text{Logit}(p_i)}}{1 + e^{\text{Logit}(p_i)}} \cdot \quad (2)$$

This model shows a clear statistical relationship between the dynamics of debris flows and climate. Between 1961 and 2000, the probability of the occurrence was due to the cumulated number of freezing days since the last event (nfd) and the number of rainfall events greater than 30 mm/day (nrd) between June 15th and October 15th, with an error ( $e$ ).

The model is:

$$\text{Logit}(p_1) = -4.7236 + (0.388 * \text{nfd}) + (0.5717 * \text{nrd}) + e \cdot \quad (3)$$

## 5. Results

### 5.1. CLIMATIC VARIABILITY

First of all, we studied the evolution of the annual and seasonal mean precipitations for the periods 1960–1980 and 1980–2000. No significant variation appeared from the annual data taken since 1980, for all weather stations studied.

On a seasonal time scale, we can see at Monétier station (Figure 6) a small increase in the amounts of rainfall in spring, summer and autumn since 1961. Winter is the only season which shows a small decrease in precipitation for this period. However, these variations between 1960–1980 and the period after 1980 are not significant (Mann–Whitney test), whatever the meteorological station.

We also studied the highest daily total rainfall during the warmest period of the year. Figure 7 shows the evolution between 1964 and 2000 of the ten highest daily totals occurring each year between June 15 and October 15 at St Christophe. There would seem to be a 20 mm/d threshold. Above 20 mm/d no significant difference was observed over the period from 1965 to 2000 ( $z$ -test). However, above 30 mm the number of rain-days was observed to have increased significantly ( $z$ -test) since the 1980s for the thresholds of 30 and 40 mm (Table I). At La Salette 13 events between 30 and 40 mm/d occurred between 1970 and 1985, as opposed to 29 after 1985. For the more intense events, the difference was less marked and not significant, with 9 events as opposed to 11 between 40 and 50 mm and 9 events as opposed to 10 above 50 mm/d. These observations are in concordance with the results of meteorological analyses which concern wider areas in the Alps, such as the northern Italian Alps (Brunetti et al., 2001) or other sectors in the Alps (Rebetez et al., 1997; Frei and Schär, 2001).

Our analysis of the mean annual temperatures, which have increased since the 1980s, confirms other observations which concern the whole of the European Alps using homogenised instrumental time series (Böhm et al., 2001). Two stations (with the exception of Le Monétier station) clearly show such an increase (Table II). With ten-year mean temperatures, there is a decrease of 0.3 °C for the period 1960–1970, followed by an increase of 0.4 °C between 1970 and 2000.

From the seasonal data we can observe that in spring the temperatures have increased significantly ( $t$ -test) since the beginning of the 1980s, (with the exception of Le Monétier station). In summer, the mean value of this increase is around 0.9 °C. The increase in temperatures in autumn (0.6 °C) is only significant ( $t$ -test) in Saint Christophe and La Salette stations. Finally, for winter period we observed a general significant increase (0.7 °C) concerning the periods 1960–1980 and 1980–2000 for all the stations.

When we focus on the evolution of the number of freezing days since the 1960s, a significant decrease ( $t$ -test) in the number of freezing days (around 14%) is observed at La Salette station since the beginning of the 1980s (Figure 8) due to the increase in temperatures. Such a trend is also observed at Le Monétier (12%) and at Saint Christophe (14%).

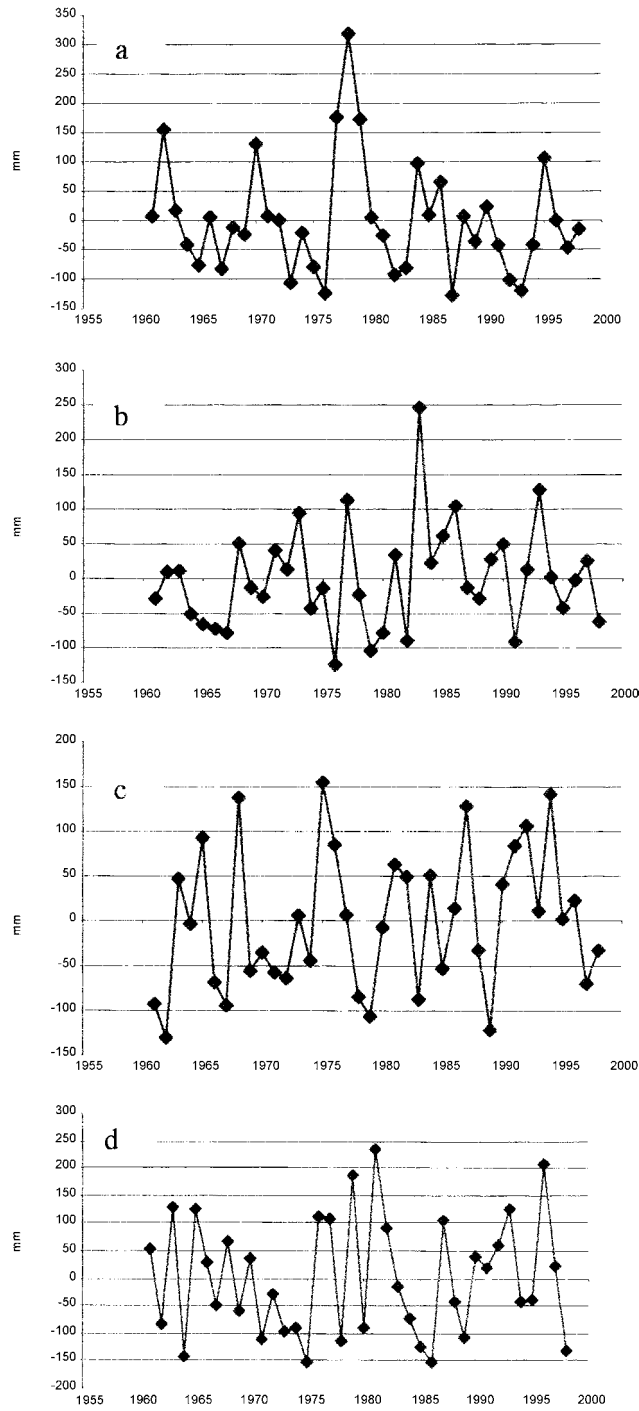


Figure 6. Deviations from the mean of precipitations at Monétier station between 1961 and 2000. (a) Winter (d, j, f); (b) Spring (m, a, m); (c) Summer, (j, j, a); (d) Autumn (s, o, n).

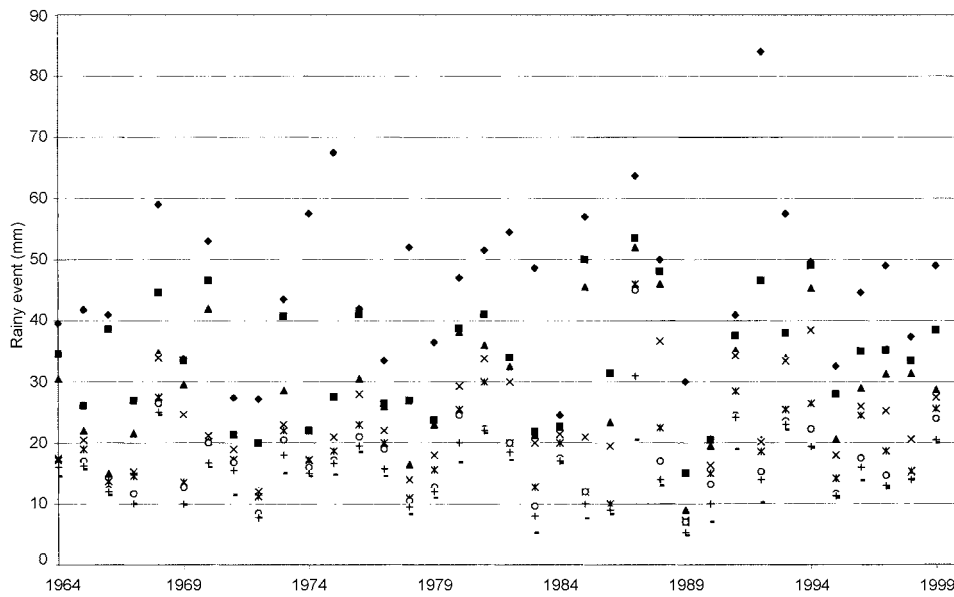


Figure 7. The 10 most important rainy events per day occurring each year between the 15th June and the 15th October since 1964 at St Christophe station.

Table I  
z-Test values (bold = significant)

		1965–2000
St Christophe	>30 mm	<b>0.006</b>
	30–39 mm	<b>0.019</b>
	40–49 mm	<b>0.049</b>
	>50 mm	0.167
Monétier	>30 mm	<b>0.02</b>
	30–39 mm	<b>0.03</b>
	40–49 mm	0.67
	>50 mm	0.8
La Salette	>30 mm	<b>0.001</b>
	30–39 mm	<b>0.005</b>
	40–49 mm	0.722
	>50 mm	0.835

Table II  
*t*-Test values (bold = significant)

	1970–1980	1980–1990	1965–2000
St Christophe	0.061	0.052	<b>0.001</b>
Monétier	0.368	0.522	0.163
La Salette	0.301	0.662	<b>0.001</b>

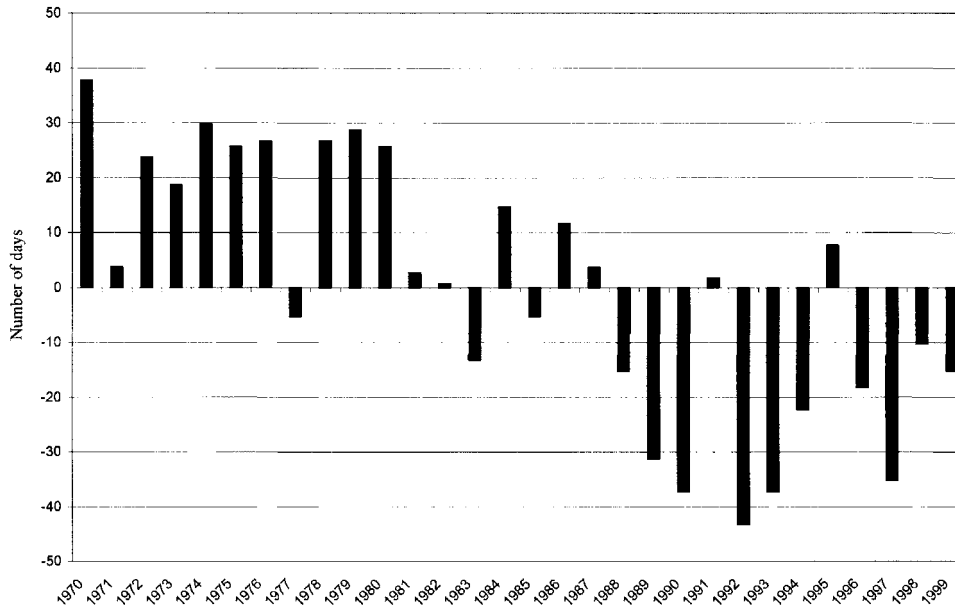


Figure 8. Deviations from the mean annual number of freezing days at La Salette station between 1970 and 2000.

To summarise, it is clear that climatic changes are attested by both a significant increase in the annual mean temperatures and a decrease in the number of freezing days. The analysis of the rainfall shows a significant increase in daily rainfall greater than 30 mm. Consequently, if we consider the relationship between heavy rainfalls and the triggering of debris flows, the logical evolution would be an increase in the number of debris flows since the 1980s.

## 5.2. VARIATIONS IN THE DYNAMICS OF DEBRIS FLOWS

The analysis of variations in debris flow occurrence in the Dévoluy area is based on 128 events which have occurred since 1948. A significant decrease (0.048 Mann–Whitney test, 0.05 level) in the number of debris flows has been observed since the mid-seventies (Figure 9). 84 debris flows were triggered between 1950 and 1975

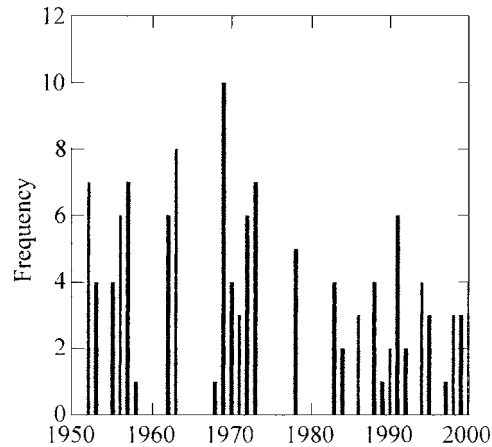


Figure 9. Frequency of debris flows in the Dévoluy sector ( $N = 121$ ).

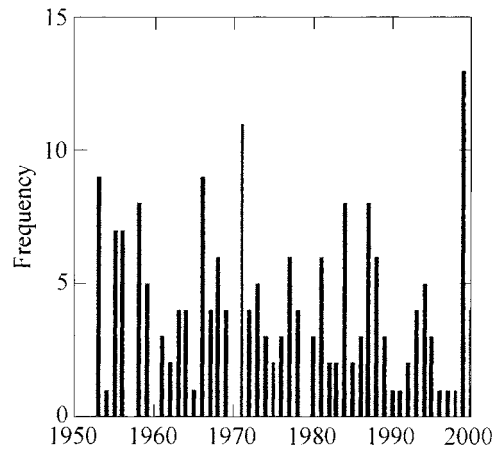


Figure 10. Frequency of debris flows in the northern part of the Massif des Ecrins ( $N = 219$ ).

and only 48 since 1975. If this analysis is conducted over ten-year periods, the total number of debris flows was not statistically different (0.05 level) between 1950 and 2000.

For the Massif des Ecrins, data analysis was based on 191 debris flows which occurred between 1953 and 2000. If we look at two comparable periods, 102 debris flows were triggered between 1953 and 1976 and only 89 since 1976. However, as shown in Figure 10, this evolution is not statistically significant.

Considering that a relationship exists between the length of debris flows and their frequency, the data have been classified here by their length. The analysis of the frequency distribution (Figure 11) shows that the modal value of the debris flows is between 200 m and 299 m, and that length values under 599 m account for 74% of the debris flows.

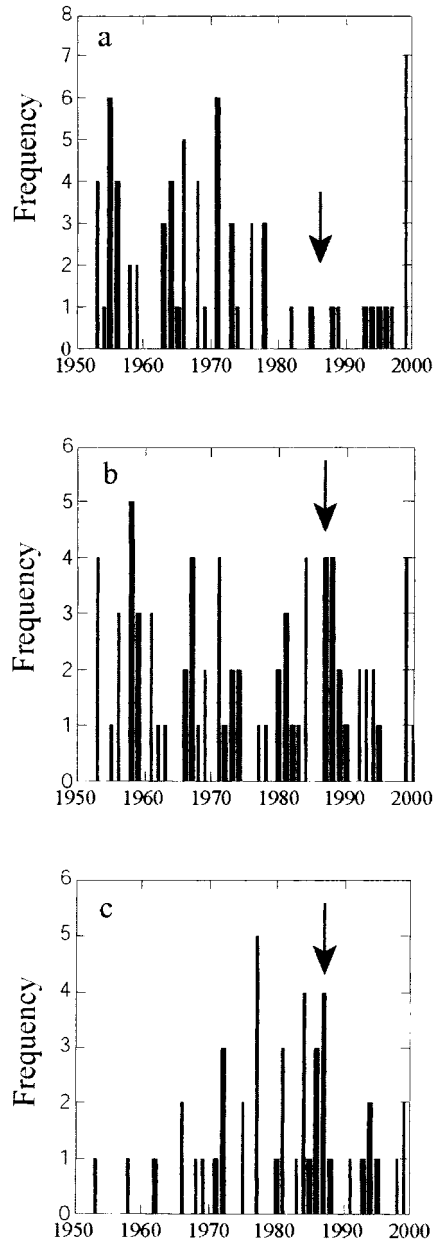


Figure 11. Frequency of debris flows in the Massif des Ecrins according to the track length of deposits. (a) <400 m long ( $n = 70$ ); (b) Between 400 and 699 m long ( $n = 75$ ); (c) >700 m ( $n = 46$ ). Arrow = 1987.



Table III

Return period for the different debris-flow class lengths between 1952 and 2000 in the Massif des Ecrins

	<100 m (yr)	100– 199 m (yr)	200– 299 m (yr)	300– 399 m (yr)	400– 499 m (yr)	500– 599 m (yr)	600– 699 m (yr)	700– 799 m (yr)	>800 m (yr)
1953–1975	5	3.5	4	4.4	6	4.8	9.5	18	–
1976–2000	4.7	2	2	4.5	5	5	10	17	–
1953–2000	4.8	2.7	3	4.5	5.5	4.9	9.7	17.5	>52

A significant decrease (0.01, Mann–Whitney test) in the number of debris flows can be observed between 1953–1975 and 1976–2000 for the debris flows less than 400 m in length (Figure 11). On the contrary, there are no significant variations in the number of the debris flows longer than 400 m between these two periods (Figures 11b,c).

When taking the data over ten-year periods, the variations are hardly noticeable. There was a high activity during the 1970s but the number of debris flows is not significantly different from the other decades. Moreover, 1987 is not characterised by a great number of debris flows, unlike observations elsewhere in the Alps (Zimmermann and Haerberli, 1992; Rickenmann and Zimmermann, 1992; Van Steijn, 1996, 1999) even though the number of rainfall days greater than 30 mm was the highest over the 1960–2000 period (7 events).

In all the valleys studied, the mean return period, defined as the time between two events whether or not they occurred in the same system for each valley separately, was around 19.6 years. It has decreased for debris flows with lengths between 100 and 300 m for the periods 1953–1975 and 1976–2000 (Table III). For the other lengths, we have not observed clear significant variation between the two periods. As shown in Table III, this return period increases with length of debris flow: 4 years for debris flows less than 100 m in length and more than 50 years for debris flows longer than 800 m. Moreover, we can observe that this return period is correlated with the mean elevation of debris flow triggering zones (Figure 12).

For the whole period, the minimum mean elevation of triggering zones is around 2100 m, between a maximum at 2650 m and a minimum at 1650 m. However, if we compare the distribution of the elevation of triggering zones during the two periods 1953–1975 and 1976–2000 (Figure 13), it seems that a shift has occurred towards higher elevation zones. The triggering zones lower than 2200 m represent less than 40% between 1976 and 2000 as opposed to more than 65% for the period 1953–1975.

Finally we studied variations in minimum altitude of the debris flow triggering zones between 1952 and 2000 for all sites of the Massif des Ecrins (Figure 14).

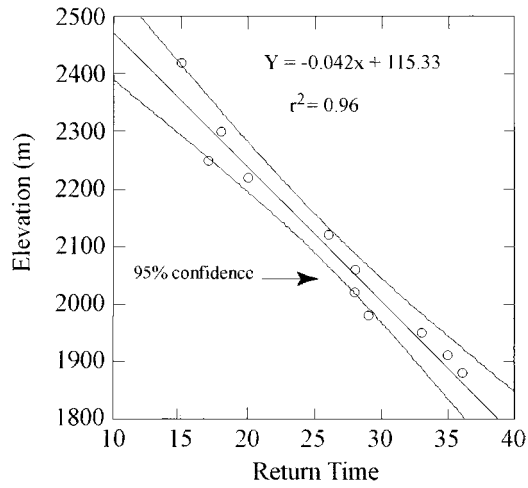


Figure 12. Mean altitude of the triggering zone of debris flows smaller than 400 m versus return time over the period 1952–2000 for each of the 11 valleys in the Massif des Ecrins.

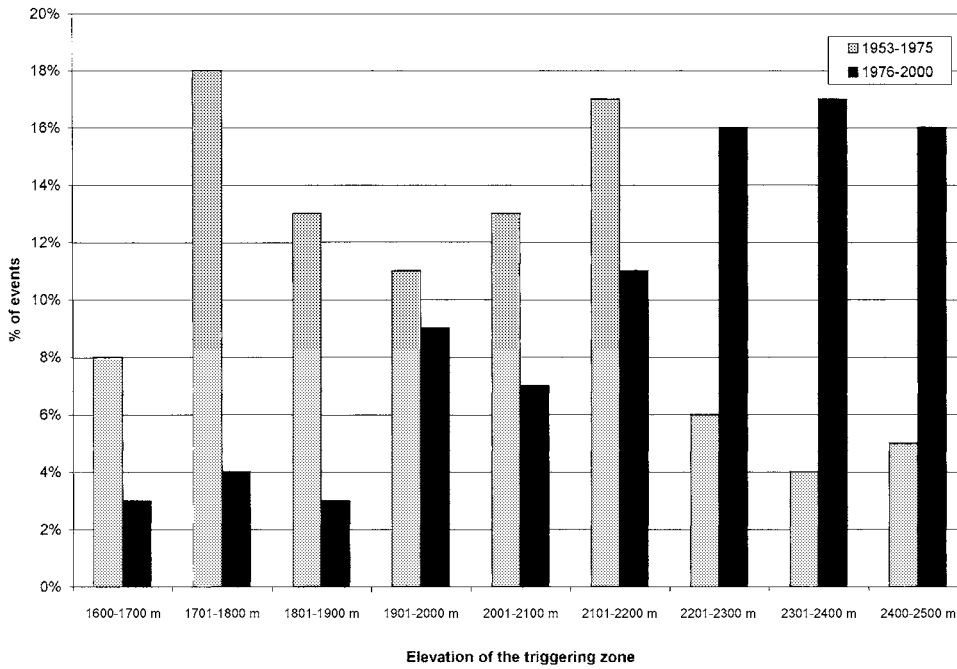


Figure 13. Number of debris flows smaller than 400 m between 1952 and 2000 according to the elevation of the triggering zone.

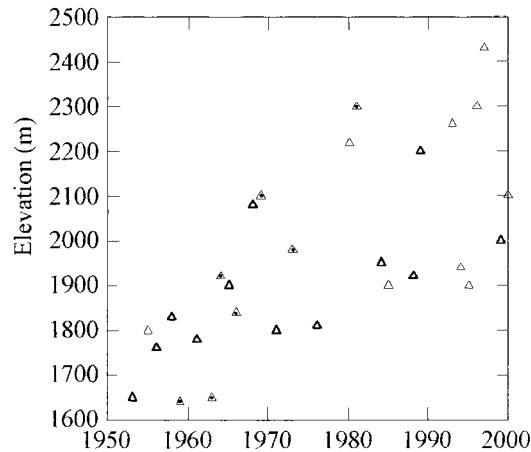


Figure 14. Lowest triggering zone altitude of debris flows between 1952 and 1999 in the Massif des Ecrins. Bold triangle > 6 measurements; Triangle with a point = between 3 and 5 measurements; Triangle < 3 measurements.

According to our observations, minimum altitudes increased by more than 100 m during this period. Triggering zones lower than 1800 m asl have been inactive since 1976.

Finally, in the Dévoluy area a reduction in the number of debris flows has been observed since the 1980s. In the Massif des Ecrins, we have shown an upward shift in the triggering debris flow zone and a lack of significant variation in the total number of debris flows. At the same time, the frequency of debris flows under 400 m has decreased significantly.

## 6. Discussion

In various places in the Alps, recent studies have shown a relationship between the triggering of debris flows and the increase in the number of intense rainy events and/or in temperatures. This rise in temperatures has caused a glacial recession and the melting of the permafrost over the last few decades (Haeberli et al., 1990; Zimmerman and Haeberli, 1992; Rebetz et al., 1997). In the Dévoluy, the Massif des Ecrins and in other sites of the French Alps (Hang, 1994; Thevenon, 1999; Bocquet, 2001), it seems that signs of the recent climatic change can also be observed. This climatic change can be seen in an increase in temperatures and an increase in the number of daily rainfalls above 30 mm occurring in the summer.

Our analysis, based on 319 debris flows for which the triggering is not linked to glacial retreat, is characterised by variations in debris flow dynamics since the 1980s. Some are apparently in contradiction with the nature of the climatic changes observed. Why has the increase in rainfall above 30 mm not led to an increase in the number of debris flows? Here, we propose different explanations.

### 6.1. HOW GOOD ARE THE DATA CONCERNING DEBRIS FLOWS?

First of all, does the number of debris flow deposits correspond to the actual number of debris flows triggered? It is indeed possible that certain low amplitude triggered flows were not listed, even though debris flows under 100 m in length were taken into account in the analysis protocol. As for the debris flows which follow the exact path of the channel created by a previous debris flow but are not as long, and whose frequency is sufficiently high not to be identified by the analysis of the aerial photographs or field observations, their number is difficult to evaluate but is probably low. Consequently, without excluding these two possible skews, their influence on the results remains weak and in no way explains the trend observed previously.

### 6.2. THE REPRESENTATIVITY OF THE RAINFALL DATA

It is well known that the intensity and duration of the rainfall play an important role in the triggering of debris flows (Caine, 1980; Van Steijn, 1996; Blijenberg, 1998) and that therefore the representativity of the rainfall data can be discussed. Several doubts arise concerning the extrapolation of meteorological data collected at low elevation. For example, Sevruck (1997) showed in Switzerland a clear spatial heterogeneity of the precipitations. Such a situation seems to be particularly true for our field area because heavy rainfalls, which occur in summer, are mainly due to local storms. In order to evaluate the impact of the elevation on the variations in amounts of rainfall, we compared daily data from Le Monétier with nine years of recorded data at the Tête Noire station, which is in the same valley but at a higher altitude (2030 m asl; Figure 3). The result of the comparison (*z*-test) between data does not show significant differences in the number of days for which the rainfall is more than 30 mm between June 15th and October 15th during the 1976–1985 period. The number is strictly the same, with the exception of 1984, for which there is one event with rainfalls greater than 30 mm at the bottom of the valley and 4 at the top. It is well known that rainfall is greater at higher altitudes than at lower ones (Gerrard, 1990; Barry, 1992). In the stations studied, 13 daily events between 30 and 40 mm occurred in lower areas and 14 events at higher elevation; these patterns were 5 and 10 respectively, for rainfall between 40 and 50 mm, and 5 and 7 respectively, for rainfall above 50 mm/d. Consequently this undervaluation does not account for the absence of increase in the number of debris flows.

Moreover, it is necessary to recall the differences between the data observed on the flat surface collected by classical instruments and those which were recorded in a couloir or at the base of a wall acting as an impluvium. Francou (1988) has shown that one debris flow had been triggered on August 9th 1983, in Combe de Laurichard, but the rainfalls had been 12.8 mm at Le Monétier, less than 10 mm at Saint Christophe and 17 mm (in 3 hours) at Tête Noire. However because of the size of the impluvium, the author considered the event to be equivalent to a 50 mm runoff when deliberately taking a high infiltration rate (50%). In addition,

some researchers have highlighted the role of intensities lasting only a few minutes (Blijenberg, 1998) in the triggering of debris flows, but unfortunately this type of observation does not exist near the sites selected in this study. Consequently this undervaluation does not account for the absence of increase in the number of debris flows either.

### 6.3. THE POSSIBLE ROLE OF ANOTHER PARAMETER

The results show an apparent contradiction between the decrease in number and frequency of debris flows, and the increase in the number of days during which rainfall is more than 30 mm. However, it has already been shown that the amount of rainfall may be a necessary (Caine, 1980; Johnson and Rodine, 1984; Innes, 1985b) but not sufficient factor to trigger debris flows.

The volume of stored debris at the foot of the cliff is the second factor involved in the triggering of debris flows which must be taken into account in this analysis. A decrease in this volume or in the time needed for its reconstitution may be responsible for the variation. If the cliff supplies less debris, even if heavy rainfall events remain high in number, there is no evidence that a debris flow occurs. However, for the same lithology, the volume or the rate of reconstitution of the stock depends on the frequency, intensity and duration of freezing and consequently on the temperatures.

The question is whether the increase in temperatures reduces the volume and/or rate of reconstitution of the stock. At low altitude, the stations selected showed a significant reduction in the frequency of freezing days (*t*-test). This observed decrease in the number of freezing days may explain the decrease in debris flow frequency in the Dévoluy Massif (assuming that changes in air temperatures are high enough to provoke a change inside the rock faces which it is currently not possible to check).

In the Massif des Ecrins, we separated the debris flows whose triggering zone is below 2200 m from those with a higher elevation (this 2200 m value corresponds to the mean of all watersheds) because they did not have the same dynamics. At low elevation, we have observed a significant decrease (0.01, Mann–Whitney test) in debris flow occurrence (Figure 15a) and also in the frequency of small debris flows since the 1980s. Consequently, this decrease in occurrence and frequency is due to the decrease in the number of freezing days at low elevation. This fact may also explain the general increase in elevation of the triggering zone.

On the contrary, the number and frequency of debris flows have not changed at high altitude (>2200 m asl) (Figure 15b). Are these results in contradiction with climatic variations observed before?

We assume that it is not the case because it is still difficult to extrapolate this reduction in the number of freezing days at high altitude. Indeed, although the increase in temperatures at high altitude has clearly been shown over the last few decades by different work done on homogenized series (Beniston et al., 1997; Diaz

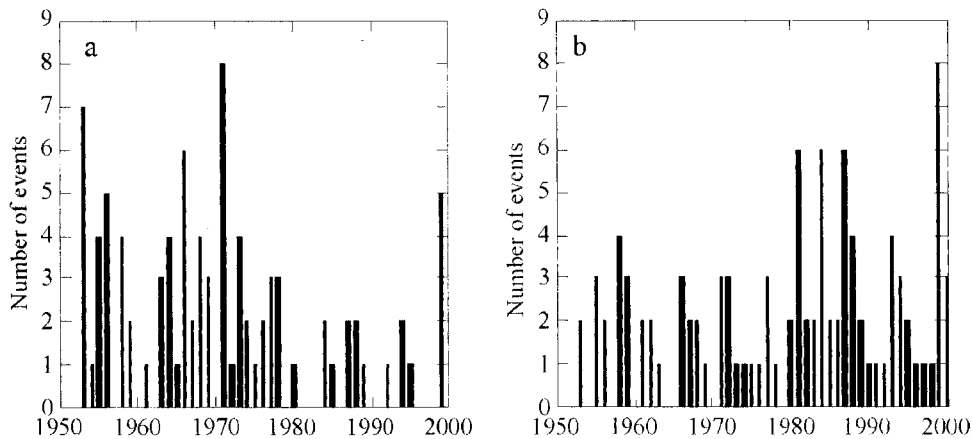


Figure 15. Frequency of debris flows in the Massif des Ecrins according to the elevation of the triggering zone. (a) Triggering under 2200 m ( $n = 90$ ); (b) Triggering above 2200 m ( $n = 101$ ).

and Bradley, 1997; Böhm et al., 2001), a detailed analysis of the amplitude of the variations in temperatures with respect to altitude is complex. For example, Beniston et al. (1997) have shown that temperature anomalies show a significant linear relationship with altitude in the Swiss Alps. Such a trend might prove that the frequency of freeze/thaw alternations and thus frost-weathering is increasing. However, Diaz and Bradley (1997) have also shown that it is clear that maximum temperature trends are lower above around 2000 m asl, in comparison with those at lower elevations.

Therefore, if we assume that the increase in temperatures is to be found at high altitude, the upward shift of the isotherm  $0^{\circ}\text{C}$  may be at the origin of increasing weathering. The temperature increase could provoke a degradation in the permafrost in rock walls as has been observed elsewhere in the Alps (Haeberli et al., 1993; Haeberli and Beniston, 1998; Wegmann et al., 1998; Imhof et al., 2000) and this trend may increase frost weathering. Moreover, the temperature increase may reduce snow cover duration and expose the high slopes to greater temperature variations.

Consequently, without local measurements it is very difficult to estimate the impact of the increase in temperatures on the frequency, intensity and duration of the freezing of the walls at high altitude and thus on the volume or rate of stored debris.

## 7. Conclusion

The aim of this study was to analyse the response of debris flow dynamics to climatic change since the 1950s in large sample areas. Two sectors of the French Alps were selected, the Plateau de Bure and neighbouring valleys ( $20 \text{ km}^2$ ) and

the eastern part of the Massif des Ecrins (20 km<sup>2</sup>). In the two sectors 319 debris flows have occurred since the beginning of the 1950s, the triggering of which is independent of the current glacial retreat. A climatic analysis was carried out on 3 stations located in the vicinity. This analysis has shown a significant increase in the annual and seasonal temperatures for the 20 last years combined with a significant reduction in the number of freezing days. A significant increase in summer rains above 30 mm/d was also observed.

However, in the Dévoluy area a reduction in the number of debris flows has been observed since the 1980s. In the Massif des Ecrins, we have shown an upward shift in the triggering debris flow zone and a lack of significant variation in the number of debris flows. But this global result masks two different trends, depending on elevation.

At low altitude (<2200 m) the general number of debris flows and the frequency of debris flows that are less than 400 m length have decreased significantly whereas no significant variation has been observed at high altitude (>2200 m).

In the Dévoluy area and at low altitude in the Massif des Ecrins these variations can be explained by a decrease in the number of freezing days related to the increase in temperatures, which implies a slower reconstitution of the debris stored between two events. But at high altitude it is currently difficult to establish the link between the climatic change and the dynamics of the debris flows because very little is known about the two variables controlling the triggering of the debris flows, i.e., intense rainfalls and the stored volume of debris.

Until now most research has been devoted to the relationship between rainfall and triggering of debris flows. In the future a major issue will be to link this research on rainfall with a quantification of the volume of debris with respect to climatic parameters. It is necessary to acquire a better understanding of the mechanisms linking climate and the triggering of debris flows in view of future climatic forecasts (GICC, 2001) and of the risks which the debris flows represent for the populations living in the valleys.

Finally this study underlines the heterogeneous character of the response of the debris flows to the climate in the periglacial zone. The low elevation zone is explicitly sensitive to climatic variations because of its marginal position which corresponds to the altitudinal limit of distribution of the process in this area. Consequently, in the study of the relationship between climate and periglacial processes it now seems necessary to take these different sensitivities into account.

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