Two Decades of Responses (1986–2006) to Climate by the Laurichard Rock Glacier, French Alps

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ABSTRACT
The Laurichard active rock glacier is the permafrost-related landform with the longest record of monitoring in France, including an annual geodetic survey, repeated geoelectrical campaigns from 1979 onwards and continuous recording of ground temperature since 2003. These data were used to examine changes in creep rates and internal structure from 1986 to 2006. The control that climatic variables exert on rock glacier kinematics was investigated over three time scales. Between the 1980s and the early 2000s, the main observed changes were a general increase in surface velocity and a decrease in internal resistivity. At a multi-year scale, the high correlation between surface movement and snow thickness in the preceding December appears to confirm the importance of snow cover conditions in early winter through their influence on the ground thermal regime. A comparison of surface velocities, regional climatic datasets and ground sub-surface temperatures over six years suggests a strong relation between rock glacier deformation and ground temperature, as well as a role for liquid water due to melt of thick snow cover. Finally, unusual surface lowering that accompanied peak velocities in 2004 may be due to a general thaw of the top of the permafrost, probably caused both by two successive snowy winters and by high energy inputs during the warm summer of 2003.

INTRODUCTION
Rock glaciers are creeping mixtures of ice and debris that exist mainly in areas of discontinuous permafrost in fairly dry mountains (Haeberli, 1985; Barsch, 1996). The high thermal inertia of permafrost and the slow motion of rock glaciers (typically decimetres per year) lead the latter to react to decadal to centennial climatic trends (Haeberli et al., 2006). However, recent observations also suggest that the sensitivity of rock glaciers to climate may increase when ground temperatures approach 0°C (Kääb et al., 2007), and this is partially confirmed by studies of inter-annual fluctuations in the movements of surveyed rock glaciers.
glaciers (Roer et al., 2005; Hausmann et al., 2006; Delaloye et al., 2008; Ikeda et al., 2008).

The study of rock glaciers, and especially of the relations between their thermal and mechanical dynamics and climate, requires annual monitoring over the long term. This includes ground temperature measurements in boreholes, or at shallower depths (Delaloye, 2004; Lambiel, 2006; Bodin, 2007) which may indicate how the thermal state of permafrost is reacting to climatic warming (Harris et al., 2003). The deformation of ice-rich permafrost on mountain slopes, which is thought to be partially controlled by the thermal state of permafrost (Arenson et al., 2002; Ladanyi, 2006; Kääb et al., 2007), can be examined using remote sensing (for a synthesis, see Kääb, 2005) or in situ geodetic methods (e.g. Lambiel and Delaloye, 2004).

In the French Alps, data collected over two decades on and around the Laurichard rock glacier (RGL1) are invaluable for examining this Alpine landform, which is located close to the lower limit of discontinuous permafrost (Bodin, 2007). This paper uses these data to address the following questions: (1) How has surface velocity of the RGL1 changed over the last two decades? (2) Have perceptible changes occurred in its internal structure over the same period? (3) What are the relationships between surface movement, sub-surface temperatures and climatic parameters over the last few years?

STUDY SITE

The Combe de Laurichard catchment (45.01°N, 6.37°E) is a well-investigated area of mountain permafrost (Francou, 1981, 1988; Bodin, 2007) located in the southern French Alps (Figure 1). Particular attention has been paid to the main active rock glacier, RGL1 (Francou and Reynaud, 1992; Bodin, 2005; Thibert, 2005; Bodin et al., 2008).

The Combe de Laurichard is characterised by high rock faces extending from 2700 m to 3000 m asl, composed of densely fractured granite that generates large amounts of debris. Between 2400 and 2700 m asl (bottom of the Combe), the north-facing slopes are affected by permafrost, as shown by abundant landforms produced by the long-term creep of ice-rich debris (Figure 2a). The RGL1 extends from the rooting zone (2650 m asl) at the contact with the rock face to 2450 m asl at its front (Figure 2b), and is advancing onto Holocene glacial and periglacial deposits. It is 490-m long, between 80 and 200-m wide, and has an apparent thickness (based on the vertical height of the sides and front) of 20–30 m. It displays morphological features typical of an active landform: longitudinal ridges in the central steep part and a succession of transverse ridges and furrows in the compressive part of the tongue.

Datasets from the four closest meteorological stations (Briançon, Monêtier-les-Bains, Saint-Christophe-en-Oisans, La Grave, located, respectively, at 1324, 1459, 1570 and 1780 m asl, and at 20, 18, 18 and 8 km from the study site) for the 1960–91 period were used to analyse the main characteristics of the regional climate. The mean extrapolated elevations of the 0°C and -2°C isotherms are, respectively, at 2560 and 2910 m asl. Mean winter snow accumulation (October to May) recorded from 1960–2006 on the Sarennes Glacier (45.11°N, 6.12°E, 2900 m asl), located 20-km west of the Combe de Laurichard, is 1.7-m water equivalent (Thibert et al., 2008). Measurements of snowpack thickness in the Combe de Laurichard between 1979 and 1986 (Francou, 1988) show that the snow cover generally remained thin (<0.5 m) until December, after which it could remain thin or reach 1 m. A second phase of thickening (to about 2 m)
Figure 2. (a) Perspective view (July 2003, Orthophoto Institut Géographique National) of the northern slopes of the Combe de Laurichard; (b) morphology of the Laurichard rock glacier and location (on 1-m DEM, Lambert two metric coordinate system, Bodin et al., 2008) of the geodetic survey points, geoelectrical measurement lines and ground surface temperature loggers.
generally took place from March to May, before rapid melt from May to June.

The meteorological station data show that air temperatures in the region increased by 1.3°C over the 20th century, following the general trend in the Alps (Casty et al., 2005). A major break (Figure 3a) occurred after 1984, when the mean annual temperature increased by 1 to 2°C, with the spring and summer seasons experiencing the strongest increase. An increase in the ablation flux at the Sarennes glacier (Thibert et al., 2008), starting in the mid-1980s is highly correlated with the regional increase in temperature and was estimated to be 11–20 W/m² (Vincent et al., 2004).

Total precipitation recorded at the meteorological stations increased by 24–48 mm between 1960 and 2006, but no clear trend was present. Winter accumulation on the Sarennes glacier, however, increased suddenly in 1977 (Figure 3c). December snow depths measured at the closest high elevation site (La Toura, 44.98°N, 6.17°E, 2590 m asl) also indicate an increase over the period 1983–2000, and snow in this month can significantly influence the ground thermal regime.

Figure 3  Air temperature trends in the study area between 1960 and 2006: (a) 12-month and 36-month running means at the Briançon and Saint Christophe stations; (b) seasonal means at the Monêtier station (Météo France data); (c) winter snow accumulation recorded at the Sarennes glacier (Thibert et al., 2008).
METHODS

A geodetic survey was initiated in 1979 by Francou (1981) to regularly monitor surface velocities of the RGL1 along longitudinal and transverse transects. Only two of the six original lines of blocks (longitudinal line L of 13 blocks, along the main flowing axis, and transverse line A of 11 blocks, at the root) are still being surveyed with an accuracy estimated at ±0.025 m using a total station (Sokkia Set C) from control points on the bedrock (Thibert, 2005). The survey was carried out every two to three years from 1979 to 1999, and subsequently every year. Four marked blocks were added at the upper end part of the longitudinal profile (Figure 4, points L14 to L17) in 1999. Annual velocity (downslope, vertical and horizontal), actual vertical change after removing total displacement of the block downslope and variation in inter-block distance can be calculated from these data.

A total of 12 vertical electrical soundings (VES) and two electrical resistivity tomography (ERT) profiles have been undertaken on the RGL1 to investigate its internal structure: two VES in 1986 (Francou and Reynaud, 1992), five VES in 1998 (V. Jomelli and D. Fabre, unpublished work), four VES and two ERT in 2004, and one VES in 2006. All of the VES were carried out using the Schlumberger configuration and the results were interpreted with two or three-layered resistivity models, whereas a pole-dipole configuration proved to be the best for ERT.

Seven temperature loggers (UTL1, © Geotest, Zollikofen, CH), installed just below the blocky surface to protect them from direct solar radiation, have been recording bi-hourly since October 2003. Ground surface temperatures recorded from 2000–04 in a rock glacier in Valais (140-km north in the Swiss Alps, 2500 m asl, see Delaloye et al., 2008) were used to help assess evolution of the thermal regime in the Laurichard area. The use of the Valais data, to assess the ground thermal state before 2003, is justified by concurrence of the trends at the two sites from 2003 to 2006. However, local variation in snow thickness may also strongly influence ground temperature, as occurred in winter 2004–05 when very thin snow cover in Laurichard led to greater ground cooling than in Valais.

RESULTS

Temporal Change in Surface Movement

Mean annual velocities measured along line L range from 0.39 to 1.44 m/yr (Figure 4), and are typical for surface movements induced by deep-seated creep of ice-rich debris (Barsch, 1996; Haeberli et al., 2006). Surface velocities are influenced by topography of the
bed with a maximum just upslope of the maximum gradient, as is typically observed for glaciers (Paterson, 1994). Where the slope is less steep, longitudinal transmission of strain also affects surface dynamics (Kääb, 2005), leading to extensional and compressive flow zones that correspond, respectively, to surface lowering or rising (Figure 5). In addition, it has been shown (Bodin et al., 2008) that movements from 1978 to 2005 have not only been concentrated along the main flow axis of the RGL1 but also on its right side. Therefore, the relatively low vertical rising and even local lowering, evidenced by geodetic measurements on the tongue (Figures 4 and 5, points L8 to L13), may relate to this lateral loss of mass.

The mean velocity of the RGL1, based on points L1 to L3, L5 to L8 and L10 to L12, increased from 0.48 m/yr ($\sigma = 0.12$) in 1986 to a maximum of 1.2 m/yr ($\sigma = 0.32$) in 2004 (Figure 6). Four main phases are recognised: (1) low velocities of about 0.5 m/yr from 1986–91; (2) a marked acceleration up to 1.0 m/yr from 1991–97; (3) fluctuating high velocities up to 1.2 m/yr from 1997–2004; and (4) a deceleration to 0.7 m/yr from 2004–06. These phases appear synchronous with regional air temperature variations (Figure 6). Although geodetic measurements were not carried out annually prior to 1999, it appears that interannual variability observed after this year is significantly lower than the multi-year change beforehand. Hence, the increase in rock glacier surface velocities between 1986 and 1999 may be linked to increasing air temperatures, the greatest velocities of 1999–2004 are synchronous with the highest air temperatures, and the decline in velocity after 2004 is associated with falling air temperatures. Similar observations have been made at rock glaciers elsewhere in the European Alps (Roer et al., 2005; Kääb et al., 2007; Delaloye et al., 2008).

Changes in Internal Structure

The electrical tomographies from 2004 provide the best available estimate of current thickness and ice content of permafrost at the root of the RGL1 (where the ice-rich mixture is generated), as well as at its tongue (Figure 7). The vertical structure of the permafrost was confirmed at one location by direct field observations.

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Figure 5  Annual vertical changes (grey line: 2002–03; black line: 2003–04) and annual inter-block distance changes (in grey; positive values = extension, negative values = compression) of the Laurichard rock glacier surface in 2003–04 along line L of the geodetic survey. The sectors where both compression and lowering of the surface occurred in 2003–04 are underlined.
within a 5–7 m-deep cleft which appeared between the rock glacier root and the rock face in 2003. Though varying spatially in detail, the overall vertical structure comprises, from top to bottom:

i) An ice-free upper active layer with apparent resistivities of $10^4$–$10^5$ $\Omega \cdot$m that thickens downslope from 1–2 to 5–6 m. The surface is generally extremely coarse (decimetric to metric blocks) with large voids, except on some ridges where fine materials may emerge. Sandy material washed down to the base (i.e. on top of the icy layer) is also generally present.

ii) An icy layer with resistivities of $10^4$–$10^6$ $\Omega \cdot$m with variable ice/debris ratios, the thickness of which ranges from 15–30 m. At the rooting zone, this layer probably comprises a core of massive ice (resistivity $>2 \cdot 10^6$ $\Omega \cdot$m), 15–20-m thick or more, with very little debris, the sharp upper boundary of which is clearly delineated by ERT (Figure 7). At the tongue, both VES (Figure 8) and ERT (Figure 7) suggest that less ice is present (resistivities of $2 \cdot 10^6$–$1 \cdot 10^6$ $\Omega \cdot$m) and that it is laterally discontinuous. Zones of very low resistivity may indicate the presence of liquid water.

iii) An ice-free layer possibly composed of loose debris (thickness unknown) or bedrock, the electrical properties of which are difficult to determine.

There was a general decrease in the maximum apparent resistivity ($\rho_{\text{max}}$) at the VES sounding sites located at the root and on the tongue (accuracy of relocation of the 1986 sites is judged to be about ±5 m for the lower one and ±10 m for the upper one) from 1986 to 2004–06 (Figure 9). At the root (for an inter-electrode spacing $AB/2 = 50$ m), $\rho_{\text{max}}$ decreased by 35–50 per cent from $8 \cdot 10^5$ $\Omega \cdot$m in 1986 to $3-4 \cdot 10^5$ $\Omega \cdot$m 18 years later. At the tongue, $\rho_{\text{max}}$ decreased by 20–50 per cent from $5 \cdot 10^4$ $\Omega \cdot$m in 1986 ($AB/2 = 30$ m) to $1.5-4 \cdot 10^4$ $\Omega \cdot$m in 2004 ($AB/2 = 20$ m).

The relatively large variations in resistivity over the period 1998–2006 suggest that it may be difficult to distinguish resistivity changes due to long-term modification of the permafrost (e.g. a thickening of the active layer, an increase in ground temperature, a decrease in ice content) from those relating to seasonal changes in ground properties (e.g. differing water contents during the sounding campaigns in the active layer and/or within the icy layer), and/or localised changes in the internal structure related to permafrost deformation. Despite these limitations, the diachronic VES measurements can be interpreted using inversion models to depict hypothetical vertical changes in structures composed of homogeneous horizontal layers (Figure 8). Taking into account the infinity of solutions and the well-established electrical properties of similar ground materials (Fabre and Evin, 1990; Hauck, 2001; Delaloye, 2004), changes at the root can be interpreted as being due to thickening of the active layer from 1–2 m to 2–3 m with no noticeable modification of the icy layer. In contrast, those at the tongue could represent a thickening of the active layer and a lower ice content, and/or a thinning of the icy layer. These results are less reliable, however, due to strong lateral variations of the internal structure which are not ideal for layered modelling (cf. Figure 7).

**Relations between Velocities, Ground Surface Temperatures and Climate**

The temperature at the front of the RGL1 is thought to be close to 0°C based on a MAAT of 0.6°C (1975–2006)
Figure 7: Inversion models of the electrical resistivity tomography profiles at the root (top) and tongue (bottom) of the Laurichard rock glacier in summer 2004.
extrapolated for the tongue and potential global solar radiation in summer around 200 W/m². These topoclimatic conditions also suggest that the seasonal supply of liquid water from rainfall or snowmelt inside the rock glacier may influence its deformation. The possible relationships between climate and rock glacier kinematics were examined at three different time spans: 1986–2006, 2000–06 and 2003–04.

The influence of air temperature on rock glacier deformation was first examined by correlating surface velocity with Monétier monthly air temperature during the months preceding geodetic measurements. The highest Bravais-Pearson coefficient of correlation was observed for the preceding December when a relation significant at \( p = 0.02 \) and \( R^2 = 0.69 \) existed for blocks L6 to L12. However, this statistical pattern is strongly influenced by extreme events (especially two very cold winters).

There was no clear relationship between weather station precipitation or annual snow accumulation on the Sarennes glacier and rock glacier velocity, but there was a significant correlation between snow depth measured in December at La Toura station (2590 m asl, 16 km to the W) and velocity (Figure 10; \( R^2 = 0.71 \)) prior to 2000 when the record ended. It appears that rock glacier creep was enhanced by a thicker December snow cover, probably because of the importance of early winter snow conditions on internal temperatures and hence rock glacier movements (e.g. Haeberli et al., 2006).

In addition, movements of the rock glacier from 2000–06 can be correlated in detail with precipitation parameters and ground surface temperatures.
These interpretations both support and are supported by the longer term relations outlined above for 1986–2000:

- Precipitation in October, November and March 2000–01 exceeded monthly normals (1960–91) by 125–180 mm, and snow accumulation on the Sarennes glacier was 630 mm w.e. above the 1984–2006 mean during winter 2000–01. The early, thick snow cover and its melt in spring and early summer would have, respectively, limited winter cooling of the ground and provided large amounts of liquid water into the rock glacier body, both of which may have caused the high measured rock glacier velocities.
- In contrast, movement slowed in 2001–02 when insulating snow was sparse: precipitation was less than normal and total winter accumulation of 1040 mm w.e. on the Sarennes glacier was 750 mm w.e. below the 1984–2006 mean.
- The 2004 peak in velocity may have been caused by two successive winters with a thick, early snow cover and the very warm summer of 2003 in between (see below).
- Slowing again occurred from 2004–06 and is linked to the lack of snow and cold air temperatures of 2005 and 2006 which led to a decrease in ground surface temperatures on the RGL1 (Figure 11).

Finally, 2004 was not only characterised by a peak in velocity, but also by unusual behaviour of the RGL1 surface. In other years, vertical changes show good agreement with changes in the inter-block distance: compression at the tongue is accompanied by rising up of the surface while extension at the root leads to surface lowering. However, in 2004, lowering occurred over almost the entire surface (mean of 100 mm), even in the compressive parts of the rock glacier (see Figure 5). This may have been a response to: (1) a topographically controlled change in flow direction of the tongue towards the orographic right, as mentioned above (Bodin et al., 2008); or (2) a climatically controlled melt of ice in the body of the rock glacier, especially at the permafrost table (Lambiel and Delaloye, 2004). In the absence of adequate geodetic data to fully assess the direction of flow, only the second hypothesis is discussed further.

Surface mass-balance measurements on the Sarennes glacier show that ice ablation was 1.90 m w.e. higher in summer 2003 than the 1954–2002 average (Thibert et al., 2008), corresponding to an increase of about 42–65 W/m² of surface melt flux (Vincent et al., 2007). The potential melt of 0.1 m of ice within the RGL1 would have required only a surplus of atmospheric power flux of 2.3 W/m² over a period of five snow-free months with an additional 0.03 W/m² if it is assumed that the ice temperature was -2°C rather than 0°C. If this hypothesis is correct, lowering must have taken place after the summer 2003 survey (carried out on 27 August 2003), which had not shown any significant change (Figure 11). This indicates that lags of several weeks or months are involved, implying that the influence of snow cover during 2003–04 might also have added to that of the warm air temperature of summer 2003.

DISCUSSION

The long-term datasets analysed in this study suggest a synchronous increase in surface velocity and DC resistivity between the mid-1980s and the mid-2000s. Both parameters may relate to an increase in permafrost temperatures, which would be coherent with the increase from around –2.5°C to –1.5°C at the depth of 11.6 m in the Murtel borehole over the same period (Harris et al., 2009). However, the Murtel rock glacier is estimated to be between 400 and 600 m above the regional 0°C isotherm (Hanson and Hoelzle, 2004), whereas the RGL1 is located very close to the same isotherm, raising the crucial question of potential degradation of its frozen core. As demonstrated by Kääb et al. (2007), Ladanyi (2006) and Arenson
(2002), at an annual time scale the deformation rate of ice-rich permafrost is probably mainly related to its thermal state, which itself represents a complex transient response to changes in climatic parameters (mainly air temperature) and thickness and history of snow cover (first appearance during early winter, development of a thick mantle, melting duration). Where permafrost is close to (or at) the point of thaw, the role of liquid water may also become important (Hausmann et al., 2006; Ikeda et al., 2008), by either changing the thermal properties of the ground (role of latent heat), by reducing friction between particles, or allowing/increasing basal sliding on a water film.

It appears possible that temperatures in the RGL1 are not in equilibrium with the current climate conditions, and this may also be the case for other rock glaciers in the southern Alps. In the Argentera Massif, Italy (around 44°N) for example, geophysical investigations and ground thermal monitoring indicate that permafrost at elevations of 2450–2550 m asl is probably close to thawing (Ribolini and Fabre, 2006). Furthermore, Krysiecki et al. (2008) recently reported destabilisation of the Béard rock glacier in the southern French Alps, similar to the cases described by Roer et al. (2008) but which finally ended in collapse of the rock glacier during summer 2006. According to Roer et al. (2008), this likely results from higher internal deformation and changes in shearing and basal sliding that relate to modifications of the rheological properties of warming ice.

Although the RGL1 is probably not in a suitable topographic context for major destabilisation, its location at a relatively low elevation in a Mediterranean-influenced climate likely makes it representative of...
changes being experienced by other rock glaciers in
the southern Alps. Therefore, besides its continued
long-term monitoring, future efforts should be focused
on: (1) characterisation of its internal structure in
terms of ice and water content and thermal state;
(2) monitoring of relevant parameters controlling the
ground thermal state, including snowpack develop-
ment and melt, water seepage in the ground and
thermal effect of the coarse debris layer; and (3) survey
of its deformation, possibly combining high temporal
resolution with seasonal geodetic measurements and
high spatial resolution with a laser scanning campaign
every three to five years.

CONCLUSIONS

The two main datasets of surface velocity measure-
ments and geoelectrical soundings enabled us to
characterise the internal structure and movement of
the RGL1 and to examine relationships between its
dynamics and climate. First, it appears that surface
velocities increased during the 1990s and then
decreased after 2004, as has been noted for other
rock glaciers in the Alps. Second, between 1986 and
2006, a decrease in resistivity and changes in the
resistivity curves suggest that changes in the internal
structure of the RGL1 took place. However, a potential
thickening of the active layer and reduction in ice
content cannot be fully confirmed in the absence of
other geophysical and/or mechanical investigations.
Both the increase in surface velocity and the decrease
in resistivity can, nevertheless, be linked to a
hypothesised increase in permafrost temperature.

The influence of snowpack development in early
winter on surface velocity was apparent for the 1986–
2006 period. In the last six years of the record, inter-
annual variability of the latter tracked ground surface
temperature, which itself is related to air temperature
and snow conditions. A large amount of meltwater
due to a thick snow cover may also explain the peaks
in velocity in 2001 and 2004. However, 2003–04 was
unusual not only because of the peak in velocity, but
also because it was accompanied by a generalised
lowering of the rock glacier surface attributed to high
surface energy inputs causing the melt of 10 cm of
ground ice.

Further geophysical investigations and direct
observations of the RGL1 are required, particularly
to ascertain if the permafrost is at the point of thaw.
Such long-term monitoring is needed to understand
the way that mountain permafrost in the southern Alps
is responding to atmospheric forcing. It is especially
important to examine further the short- to mid-term
sensitivity of perennially frozen, but warming, debris
deposits, as these may become hazardous if destabi-
lised on steep terrain.

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REFERENCES

Arenson L. 2002. Unstable alpine permafrost: a poten-
tially important natural hazard - variations of geotechnical
behaviour with time and temperature. PhD thesis,
ETH, Zürich. 270 pp.
deformation measurements and internal structure of
some rock glaciers in Switzerland. Permafrost and
Barsch D. 1996. Rockglaciers: Indicators for the Present
and Former Geocology in High Mountain Environ-
ments, Springer Series in Physical Environment 16.
Bodin X. 2005. Laurichard rock glacier thermal state in
2003–2004 : analysis of ground surface temperature
data, Combeynot Massif, French Alps. Shifting Lands,
Etienne S (ed.). Seteun: Clermont-Ferrand; 57–558.
Bodin X. 2007. Géodynamique du pergélisol de
montagne: fonctionnement, distribution et évolution récente.
L’exemple du massif du Combeynot (Hautes
272 pp.
Bodin X, Schoeneich P, Jaillet S. 2008. High resolution
DEM extraction from Terrestrial LIDAR topometry
and surface kinematics of the Alpine permafrost: the
Laurichard rockglacier case study (French Southern
Alps). 9th International Conference on Permafrost,
DOI: 10.1002/ppp
Fabre D, Evin M. 1990. Prospection électrique des milieux à très forte résistivité: le cas du pergélisol alpin. 6ème Congrés International de AIGI, Rotterdam.