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Measuring and interpreting heat fluxes from shallow volcanic bodies using vertical temperature profiles: a preliminary test

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Abstract To test the potential of heat flux prospecting in active volcanic areas using shallow temperature data taken along vertical profiles, we carried out two thermal profile surveys, one not far from Yasur cone on Tanna Island, and another inside the caldera of Ambrym (New Hebrides arc, southwestern Pacific). The basic steady heat flux of internal volcanic origin was determined, taking into account both conductive and convective heat transfers. At both locations there exists, over small distances, significant differences in the heat flux. These differences correspond to shallow sources of heat. The use of a network of vertical profiles allowed: (a) heat flux mapping; (b) location of shallow volcanic heat sources; and (c) observation of the detailed structure of the heat release at quiescent but active volcanoes.

Key words Shallow temperature profiles · Surface heat flux

Introduction

In areas where fluids can circulate through pores and fractures (convective heat transfer), significant temperature anomalies are observed at the surface of the ground or at shallow depth. In active volcanoes the presence of a magma body near the surface may generate important heat fluxes (from 10 Wm^{-2} to 10 kWm^{-2}) by conduction, single-phase convection or two phase convection (Hardee 1982). Such areas where significant transfer occurs can be mapped by infrared remote sensing (Bonneville and Gouze 1992; Gaonac'h et al. 1994; Mongillo 1994) or by indirect effects such as the melting of snow cover (Sekioka and Yuhara 1974). The extent and magnitude of these temperature anomalies can be monitored to evaluate variations in the shallow hydrothermal–magmatic system and thus eruption potential. Surface temperature, gas fluxes, and temperature gradient in boreholes have also been used to characterise shallow magmatic bodies (Harris and Stevenson 1997).

For other areas (e.g. where magmatic sources are deeper), the low thermal conductivity and diffusivity of rocks drastically limit the usefulness of shallow temperature monitoring; the signal induced at the surface by any thermal event at a given depth is small, damped, and delayed. However, it is possible to determine the heat flux from temperature measurements along shallow vertical profiles. In previous papers (Tabbagh and Trézéguet 1987; Tabbagh and Lardy 1993), attention was drawn to the possibility of separating the "stationary" or " basic" component of the heat flux of internal origin from the unsteady component reflecting "external" climatic changes (solar heating, wind, etc.). Mapping of this basic component would constitute a particular method of exploration, a thermal prospecting technique. This can be used in combination with other types of geophysical prospecting to investigate the internal structure of a volcano.

Heat-flow mapping has been applied successfully at the bottom of volcanic lakes (Whiteford and Graham 1994). When the water is sufficiently deep, the climate-originated external components of the heat flux are reduced so that the differences they induce between measurements at different points are negligible. In such conditions measurements can be performed with sediment-penetrating probes similar to those used for oceanic studies. On land, the filtering out of these external variations are based on two combined concepts (Tabbagh and Trézéguet 1987), time integration and spatial differences. If the sensors are several decimetres deep, the dominant term in the time-temperature re-

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cord is the so-called annual wave, the variation through the year. When the record lasts one or several annual cycles, this term can be rejected by integration. However, external changes contain lower-frequency terms that are not rejected by integration over one or several years. On one hand, these slower terms are of relatively small amplitude, within the –0.3 to 0.3 Wm^{-2} interval (Tabbagh and Trézéguet 1987), and in active geothermal areas the amplitude of upward heat flux may be significantly greater. On the other hand, they have the same value at different points of a surveyed area, so that flux differences would be totally free of their influence.

This paper presents the results of a preliminary experiment to validate heat flux determination from vertical temperature profiles as a prospecting method in active areas. It follows our previous work on Etna, Matthews and Hunter volcanoes (Tabbagh and Trézéguet 1987; Tabbagh and Lardy 1993) and is based on two sets of data acquired on two volcanoes in Vanuatu. It also allows us to evaluate the role of a possible convective downward steady term (infiltration of rain water) on the total flux determination. The experiments were carried out on Yasur and Ambrym volcanoes.

Fig. 1 Photograph from the northwest of the installation of the measurement station (A) on Yasur Volcano. Distance to cone is appoximately 2 km

Fig. 2 Photograph from the south of the measurement station on Ambrym Volcano, showing the transmitting antenna and several meteorological sensors

The volcanoes

Both volcanoes belong to the quaternary volcanic chain of the New Hebrides arc in the southwest Pacific. This arc corresponds to an intra-oceanic subduction zone. Tanna Island has a basaltic–andesite volcanic base. Located in its southeastern part, Yasur Volcano is an active strombolian cone 360 m high and less than 2 km in diameter at its base; it is several centuries or even 1000 years old (Carney and MacFarlane 1976). Ambrym is the most extensive active volcano of the arc after Ambae (Aoba) and is a 35×50-km-wide basaltic volcano, rising 1800 m above the surrounding sea floor (1200 m above sea level). The top of the cone is crowned by a roughly circular, 12-km-diameter caldera. The age of this caldera is less than 2000 years (Robin et al. 1993).

Sites of measurement and data

The measurement site selected on Tanna Island is 2 km north of the crater of Yasur Volcano, inside the ash plain just beside Siwi lake (Fig. 1) at Nayanamakel (called Naya hereafter). The site comprises three vertical profiles: the station (A) where the electronics and transmitting system are located, and two points (B) and (C) linked to the station by cables and located 100 m away in orthogonal directions. On Ambrym Volcano, the measurement site was installed inside the caldera (Fig. 2), also with three profiles; AB and AC are orthogonal, 80 m (AB) and 25 m (AC) apart.

To place the sensors at each profile, a hole with a depth and diameter of approximately 1 m was dug and copper thermistors were inserted approximately 0.1 m horizontally into the wall of the hole at three depths, 0.30, 0.60 and 0.90 m. The holes were then backfilled. The depths were chosen to reduce the effect of the rapid daily external variations that dominate temperature in the upper 20 cm; consequently, aliasing was avoided that would have been introduced by the low sampling rate imposed by the transmission. At each site the data were collected by the station, transmitted to NOAA satellites (using the Argos system) up to eight times a day (if two satellites were operating), collated at the Argos receiving station in Toulouse, France, and processed by the "Centre de Téléobservation Informatisée des Vol-

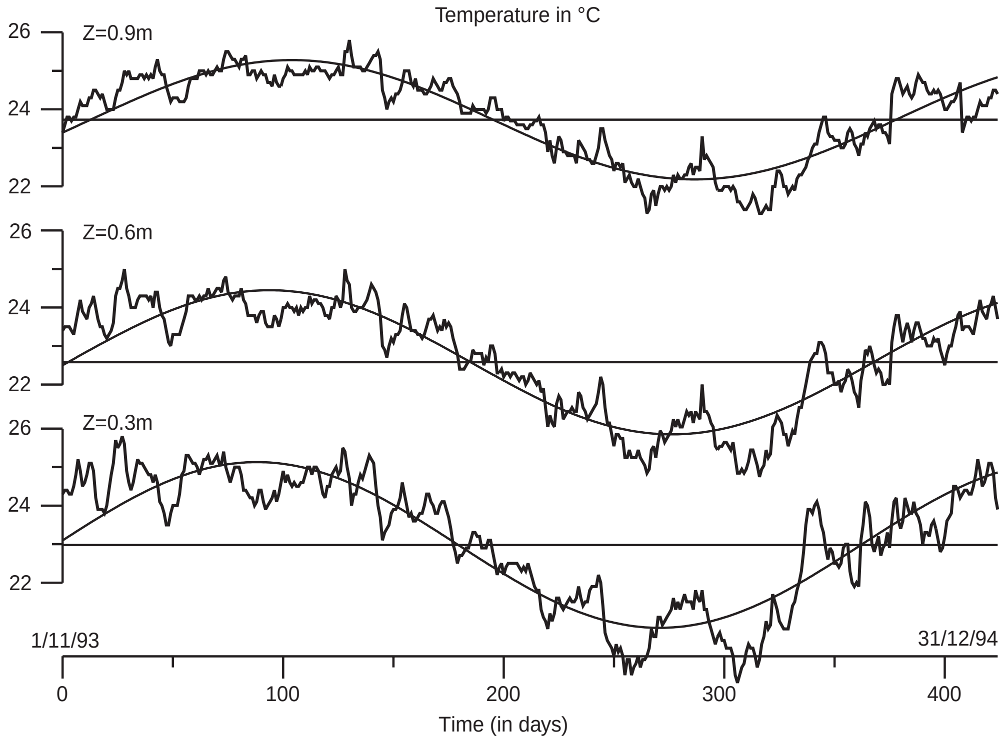

Fig. 3 Daily temperature records (in °C) at various depths at Ambrym profile (A) for a period of 425 days starting on 31 October 1993. The sum of an annual sine variation and of a steady term is fitted to the data

cans" (CTIV; "Centre de Recherches Volcanologiques", CNRS, Garchy, France). The sensitivity of the measurements (least significant bit) corresponds to 0.01 K. The Naya station has been in operation since the end of 1992, except for the 0.30-m sensor at point (B). The Ambrym station has been in place since October 1993; an example of the temperature records is seen in Fig. 3.

Thermal properties of the ground

To determine the heat flux, the ground thermal properties must be known. They were measured using active methods (Tabbagh 1985) when placing the thermistors. Measurements were performed by one of us (M.L.) by making three measurements at each depth of each profile. The results are summarised in Table 1. The variability of soil constitution at the different stations likely explains the dispersion of the data; thus, it is reasonable to group all these data and use global means. By eliminating the two outliers from the thermal diffusivity values, we obtained a mean of $0.362 \pm 0.054\ 10^{-6}\ m^2/s$ for the remaining values. By eliminating the outlier in thermal conductivity values, we obtained a mean of 0.51 ± 0.16 W/(mK).

These experimental data can be checked by two independent methods: firstly, by considering the empiri-

Table 1 Results of measurements (by active methods) of the thermal diffusivity and conductivity at the different places on Yasur and Ambrym. Each result is the median of three measurements

Location	Depth (cm)	Thermal diffusivity (10^{-6} m^2/s)	Thermal conductivity ($Wm^{-1}K^{-1}$)
Naya, station (A)	30	0.29	0.45
	60	0.30	0.70
	90	0.30	0.45
East (B)	30	0.31	0.30
	60	0.35	0.51
	90	0.58	0.37
North (C)	30	0.36	0.42
	60	0.39	0.52
	90	0.33	0.38
Yasur, top of the volcano	30	0.43	0.63
	60	0.46	0.83
	90	0.65	0.43
Ambrym (medians of the three stations)	30	0.40	0.80
	60	0.36	0.42
	90	0.43	1.06

cal laws governing the thermal properties of mixtures of solids, water and gas, and secondly, by calculating the diffusivity from the phase shift and the amplitude damping of the annual variation in temperature for the different profiles.

The empirical laws are the following:

1. Volumetric heat capacity, Cv, is calculated using a simple weighted average mixing law:

$$Cv = (Cw)(Xw) + (Cs)(Xs) \tag{1}$$

where Cw and Cs are the volumetric heat capacities of the water and solid fractions, respectively. We used $Cw = 4.18 \times 10^6$ J/(m^{3}K) and $Cs = 1.8 \times 10^6$ J/(m^{3}K) (Al Naksabandi and Kohnke 1965); Xw and Xs are the respective volumetric contents.

Thermal conductivity, k, is calculated using a geometric average law:

$$k = ks^{Xs} kw^{Xw} ka^{Xa} \text{ (Woodside and Messmer 1961)}, \tag{2}$$

where ks, kw and ka are the conductivities of the solid, water, and air fractions, respectively [we used ks = 2.9 W/(mK), kw = 0.6 W/(mK) and ka = 0.024 W/(mK)]. From these conductivity and diffusivity data, the volume fractions are found to be Xs = 0.60, Xw = 0.08 and Xa = 0.32, resulting in 40% porosity. These fractions are in good agreement with what is expected for a well-drained superficial glassy ash in volcanic areas (Tabbagh and Trézéguet 1987; Tabbagh and Lardy 1993).

Our calculations of diffusivity from the phase shift or from the amplitude damping with depth were performed by fitting the recorded data at each depth to the sum of a steady temperature and of a sinusoidal annual temperature variation. At Naya, the thermal diffusivity values derived from this calculation are too high, exceeding 1×10^{-6} m^2/s for the determination from the amplitude damping. This was due to the existence of an additional convective heat transfer term (as explained later) that is not considered in this approach, which as a result is not a relevant check on the thermal properties measured by us at Naya.

Transient heat flux

The transient flux corresponds to the non-sine variable flux. The interest in studying it is twofold: (a) to observe the possible occurrence of transient heat discharge originating from changes in the hydromagmatic system; and (b) to identify the main heat transfer mechanisms. Following Tabbagh and Trézéguet (1987), two methods of calculation are used: (a) the finite element, by considering each sensor location as a node in the vertical direction and by using an iterative procedure to follow the time variations; and (b) an analytical method in which the flux variations at the top and bottom of each profile are approximated by a series of step functions, the amplitudes of which are determined from temperature variations. Both methods allow the recovery of the heat flux from the surface and at the deepest sensor. Knowledge of the flux at the two ends of the profile means that external and internal transient fluxes can be separated. The transient flux at depths z = 0 and z = 90 cm for profiles (A) and (C) on Ambrym, calculated using a conduction transfer model, are presented in Fig. 4 (for a period of 425 days). The flux at 90 cm is of small amplitude; the transient flux is therefore of external (e.g. of climatic origin), and the conduction is a convenient mechanism to fit the experimental data. (The flux is proportional to the temperature gradient but of opposite sign.) The situation is identical at Naya. Here, as at Ambrym, variations in the transient heat flux are currently dominated by external (climatic) rather than internal (hydrothermal–magmatic system) factors.

Influence of a convective heat flux

From transient variations in heat flux, we can establish that the main heat transfer mechanism is conduction (if not, the preceding calculation would diverge), but a correct fit between a conduction model and the experimental data does not prove a total absence of convective transfer. A small convection term would correspond to a limited modification in the calculated heat flux value, and we need criteria to establish the presence or absence of a convective term. By studying the annual variation, however, we can define such a criterion and propose a value for an average water flow rate (infiltration) for the year (if infiltration happened).

For a homogeneous soil, the heat-balance equation corresponds to:

$$k\frac{\partial^2\theta}{\partial z^2} - qC_w\frac{\partial\theta}{\partial z} - C_v\frac{\partial\theta}{\partial t} = 0, \tag{3}$$

where t is time, z is vertical distance, θ is temperature variation, q is volumetric fluid-flow rate, C_w is volumetric heat capacity of the fluid, C_v is volumetric heat capacity of bulk soil, and k is its thermal conductivity. For the annual temperature variation, the solution of this equation is:

$$\theta = \theta_0 e^{\gamma z} e^{i\omega t}, \tag{4}$$

with $\omega = \frac{2\pi}{T}$, T = 365.25 days, $\gamma = \frac{C_w q - \sqrt{q^2 C_w^2 + 4i\omega C_v k}}{2k}$, $i^2 = -1$, θ_0 being the temperature amplitude at the ground surface.

Using Eq. (4), we can compute synthetic data and calculate the diffusivity from both the amplitude damping and the phase shift of the annual variation under the assumption of conductive transfer. Two different diffusivity values are obtained. The one deduced from the phase shift is practically unchanged, but the one deduced from the amplitude damping is significantly increased if q > 0 (downward fluid flow), or decreased if q < 0 (upward fluid flow). If we consider, for example,

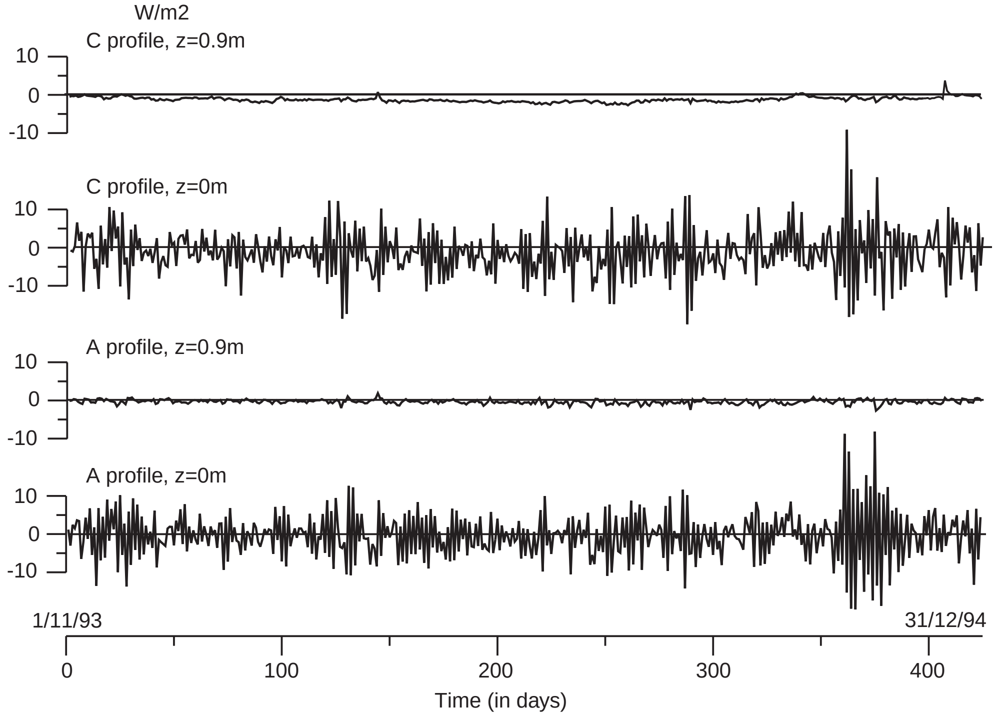

Fig. 4 Transient heat flux (in Wm^{-2}) at $z=0$ and $z=0.90$ m for profiles (A) and (C) on Ambrym. The reduction of the flux amplitude between 0 and 0.9 m depths implies that the flux diffuses from 0 to 0.9 m and is of external origin

$k=0.51$ W/(mK), $C_v=1.41\times10^{-6}$ J/(m^3K) and $q=1.6\times10^{-8}$ m^3/(sm^2) (corresponding to an annual average infiltration of 500 mm), we obtain 0.466×10^{-6} m^2/s from the amplitude damping and 0.363×10^{-6} m^2/s from the phase shift. In the presence of downward fluid flow, the amplitude damping is reduced, indicating the presence and orientation of convective transfer. In this example the Peclet number, $P_e=\frac{qC_wL}{k}$, which is the ratio of the order of magnitude of the convective heat flux to the order of magnitude of a conductive heat flux for a given range of distance, L, is equal to 0.13, if we consider a distance L of 1 m. The convective transfer is thus small but cannot be neglected.

For the steady case, the heat balance equation is written:

$$k\frac{\partial^2\theta}{\partial z^2}-qC_w\frac{\partial\theta}{\partial z}=0, \tag{6}$$

to which a solution is obtained from:

$$=\theta_n+\frac{\varphi_n}{qC_w}\left(1-e^{-\frac{qC_w}{k}(z_n-z)}\right), \tag{7}$$

with θ_n and φ_n being the temperature and the heat flux at the deepest point, z_n, of the vertical profile. The heat flux expression is:

$$\varphi=-k\frac{\partial\theta}{\partial z}+qC_w\theta. \tag{8}$$

In homogeneous ground, the temperature gradient varies with depth,

$$\frac{\partial\theta}{\partial z}=-\frac{\varphi_n}{k}e^{-\frac{qC_w}{k}(z_n-z)}, \tag{9}$$

and for downward fluid flow it is lower than in the conductive case $\left(\frac{\partial\theta}{\partial z}=-\frac{\varphi_n}{k}\right)$. Knowing the steady temperature at several points along the vertical profile, it is possible to calculate q and then φ_n.

In the present experimental results, we observe no increase in the thermal diffusivity deduced from the damping of the annual variation at the three stations on Ambrym. On Naya, however, the apparent diffusivity values at the station (A) profile, obtained by damping of the amplitude of the annual variation, 1.77×10^{-6} m^2/s, corresponds to downward liquid water flow of 8.6×10^{-8} m^3/(m^2s); at (B) the apparent diffusivity of 0.63×10^{-6} m^2/s corresponds to a liquid water flow of 3.4×10^{-8} m^3/(m^2s) and at (C) the apparent diffusivity of 0.99×10^{-6} m^2/s corresponds to 5.6×10^{-8} m^3/(m^2s). At the three points, the Peclet number approaches one, and convective transfer is important.

Steady heat flux

Previous calculations allow determination of the steady temperatures at each depth of a profile. They correspond to the nonvariable terms, but they may depend on both conductive and convective transfer. At Naya, where we have shown that the convective heat flux must be considered, the determination of the steady heat flux is achieved using Eq. (8). For $k=0.51$ W/(mK) and the different q values, we obtain for the flux, φ_n, at the deepest sensor for a period of 760 days between 1993 and 1994 (the minus sign means upward flux):

Station (A) (x = 0 m,y = 0 m) $\varphi_n = -0.70$ W/m^2

Station (C) (x = 0 m,Y = 100 m) $\varphi_n = -1.55$ W/m^2;

At point (B) the flux determination was not possible, because of the absence of measurements at 0.3 m and of an unknown offset of the sensor at 0.6 m. The heat flux values that would have been obtained by neglecting the convective transfer, –0.60 W/m^2 at (A) and –1.44 W/m^2 at (C), would correspond to smaller values. At Ambrym we have shown it reasonable to consider conductive heat flux only. By applying Fourier's law, we obtain, for a 400-day period (November 1993 to January 1995),

Station (A) (x = 0 m,y = 0 m) $\varphi_n = -0.64$ W/m^2

Station (B) (x = 80 m,y = 25 m) $\varphi_n = -0.92$ W/m^2

Station (C) (x = 0 m,Y = 25 m) $\varphi_n = -1.61$ W/m^2.

It is difficult to estimate precisely the error on these heat flux values. The principal source of errors, however, is the thermal conductivity value for which the standard deviation is approximately 30%; an equivalent relative standard deviation can be adopted for the flux values.

These heat flux values clearly indicate an abnormal upward heat flux, of greater amplitude than would be expected for external climate-induced slow variations, normally within (-0.3 ± 0.3 Wm^{-2}; Tabbagh and Trézéguet 1987). The magnitude of these climate-induced components can be compared to those obtained from the temperature data recorded at 50 cm and 1 m depth at the Tanna meteorological station, 20 km northwest of Yasur Volcano. We adopt a value of 0.3 W/(mK) for soil conductivity, deduced from a diffusivity value of 0.2×10^{-6} m^2/s and from the same volumetric heat capacity as on Yasur Volcano (because the soil is similarly composed of recent andesitic ash with high porosity). The calculation of the diffusivity indicated that, at this location, the convection transfer is negligible. We obtain for 1993 a value of +0.21 W/m^2, and for 1994, –0.25 W/m^2.

By comparing the results of the three profiles at each site, we observe a significant lateral change of the upward heat flux over a 100-m distance at Naya, and over 25 and 80 m at Ambrym. This implies the existence of heat sources at shallow depth below or not far from the measurement points. Inside the caldera of Ambrym, these sources are not surprising, because the presence of a nearby young lava flow is evidence for a shallow cooling magma body. But heat sources are unexpected at Naya, on a flat plain 2 km away from Yasur crater; here a hydrothermal–magmatic system is suggested.

Three points are insufficient for the determination of the location and strength of the heat sources, and drastic speculations would be necessary to propose a model. Moreover, the general inverse problem of steady conduction has an infinite number of solutions. Nevertheless, it can be informative to determine for both sites the power of an equivalent point source generating the observed values. At Naya, the ratio of the strength of the source to the square of the distance to the points, Q/R^2, is approximately 12 W/m^2, which corresponds, for instance, to 120 kW at a 100-m-distant source. At Ambrym, the ratio, Q/R^2, is approximately 33 W/m^2. In both cases the sources are at small distances (100 m) and not far from the surface, which suggests a complex structure of heat release in the study areas. These sources are not very powerful; a 100-kW magnitude would correspond, for instance, to the power generated by a 16-m cube of magma whose temperature decreases 1 °C per day.

Conclusion

At the two sites studied we obtained results showing the presence of a complex system of heat release, with important variations over several tens of metres. We illustrate the possible influence of convective transfer corresponding to infiltration related to heavy tropical rainfalls, which also exhibit a high spatial variability. This can be determined by the depth variation of the annual wave and removed to allow the calculation of the steady heat flux from internal (hydrothermal–magmatic) sources. We suggest that heat flux measurements using shallow vertical temperature profiles be adopted as an operational procedure for mapping the flux on land. If made over a sufficiently wide area and with a sufficient number of profiles, it will be much cheaper than using drill holes.

In the experiment described herein, the data were transferred using telemetry (the Argos system). To lighten the procedure, it would also be possible to record and store the data in situ and to recover them at the end of a given period (1 or 2 years). This would reduce the cost of installation but eliminate real-time monitoring. A network of stations would be a very efficient tool for constraining the detailed internal structure of volcanoes in combination with other types of surface geophysical measurements.

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