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Geophysical studies of the recent 15-year eruptive cycle at Poás Volcano, Costa Rica

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Abstract

The recent eruptive cycle at Poás Volcano was notable for the dramatic disappearance and subsequent reappearance of the summit crater lake. This cycle consisted of discrete phases of activity associated with a range of geophysical and geochemical signatures that illustrate the relative value of the various techniques for identifying precursory phenomena. Intrusive episodes in 1980 and 1986–1989 at Poás were preceded by A-type seismicity. Magma rose close to the surface on both occasions but the focus shifted from the dome (1980), when the lake remained stable, to the crater lake (1986–1989). The 1986–1989 event, which culminated in the complete loss of the crater lake and explosive eruptions, was characterised by concurrent increases in micro-gravity (on the southern crater floor), B-type seismicity and lake temperature and by changes in lake geochemistry. The calculated mass of magma intruded in this period is far too small to account for the observed increase in surface heat flux and subsequent loss of the lake; we suggest that a series of magma-filled dendritic conduits intruded beneath the lake facilitated enhanced heat and gas flux from a deeper magma feeder body. A model is envisaged where brittle fracture of the magma carapace at about 500 m depth allows magma to rise up through the conduit system beneath the crater and to fall again or solidify in situ when pressure drops. Whilst active, this process transfers heat and gas upwards driven by the convection of buoyant, volatile-rich magma displacing colder, relatively volatile-poor magma. As magma pressure from below decreases, the link between the deeper magma feeder and the upper conduit system is broken and the hydrothermal system resumes its role of cooling the magma feeder. The role of the lake as a physical and chemical buffer to the volcanic system was clearly demonstrated when its disappearance in 1989 was accompanied by enhanced eruptive activity and gas emissions with considerable local environmental impact. The lake therefore acts as both a moderator and index of volcanic processes at Poás.

Keywords: crater lake; geophysics; Poás volcano

1. Introduction

Poás Volcano is a large basaltic-andesite stratovolcano with a hot, acidic crater lake and shallow hydrothermal system confined within the active crater. It has been persistently active throughout historical times, undergoing cycles of phreatic and phreato-magmatic activity focussed on the crater lake. Volcanoes with hot crater lakes are not very common (only about 12% of Holocene volcanoes; Rowe et al., 1992a), and in the classification of Pasternack and Varekamp (1997), Poás Lake is an example of a 'high-activity' volcanic lake that became a 'peak-activity' lake during the period 1984–1990.

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The steady-state energy loss from volcanoes such as Poás (10^8–10^{10} W), totalled over tens or hundreds of years (neglecting more energetic short-term events), is equivalent to the energy typically released (10^{18}–10^{19} J) by a single globally-significant explosive eruption (Brown et al., 1991).

The crater lake at Poás has been relatively small (approximately 10^6 m^3 in volume) since 1965 and in the non-erupting state has been typically well above ambient temperatures (i.e. 40–60°C). The lake is sensitive to variations in volcanic activity, meteoric fluxes and outflows and where these are not balanced, the lake either becomes deep and cool or becomes hot and disappears (Brown et al. 1989; Pasternack and Varekamp, 1997). Consequently, cycles of eruptive activity in the past have been associated with the disappearance and re-establishment of the lake.

Apart from some geyser-like activity during the late 1970s, Poás was in a stable state after the eruption of 1953–1954 when a new pit crater and lake formed together with a pyroclastic cone. Increased surface activity at Poás in 1980 was preceded by a regional tectonic earthquake. Since 1980 the lake has changed from a deep tranquil aqueous body, to a rapidly convecting lake with geysers, then a sulphur-rich boiling mud pool and finally backs to a deep tranquil lake.

The accessibility of Poás Volcano, its location close to the capital city (San Jose) and its designation as a National Park make it a popular tourist destination and therefore increase the risk posed by increased eruptive activity. In the absence of the lake the effects on local vegetation, environment, health and economy are serious. Acid aerosols and rain, damage rainforest and commercial plantations (Rowe et al., 1995) and have increased the incidence of skin and breathing disorders in populations downwind of the plume (Barquero and Fernandez, 1990; Smithsonian Institution, 1990a, 1991).

Prior to 1980 there were very few geophysical measurements made on Poás however in the past 15 years, the period spanning the latest eruption cycle, the volcano has been a site of intense geophysical research, mainly centred on the summit and crater lake area. Consequently, there is an outstandingly comprehensive set of data from different types of measurements from this period. Although each type of measurement alone provides useful information on the state of the volcano, by combining the different data sets we are able to develop a detailed model of the subsurface processes leading up to and following the most recent eruptive phase.

Here we present new micro-gravity, ground deformation, and seismic data and review and synthesise other summaries of geophysical and geochemical data collected prior to 1990 (Brown et al., 1991; Rowe et al., 1992a,b) to build an integrated model of the processes responsible for the cycle of activity observed at Poás over the last 15 years. We show the unique role played by the lake and illustrate its critical importance to the nature and progress of eruptive activity at Poás.

2. Geological background

Poás Volcano is located within the Cordillera Central, the southernmost volcanic front associated with the subduction of the Cocos plate beneath the Caribbean plate. It is a composite basaltic-andesite cone rising from a base at an elevation of about 1000 to about 2700 m (a.s.l.), and like its near neighbour Barba, stands on upper Tertiary pyroclastic avalanche deposits, which in turn lie unconformably on the ‘intra-canyon lavas’ of the Aguacate formation (Weyl, 1975).

Aerial photography, geological mapping (Prosser and Carr, 1987), and detailed gravity surveying (Thorpe et al., 1981; Rymer and Brown, 1986; Brown et al., 1987) have been used to define five centres of activity in the summit region of Poás. An outer caldera structure of some 9–16 km in diameter has been largely infilled by subsequent lavas and tephras. The Poás lapilli tephra found extensively on the flanks of Poás is thought to be related to this feature, and has been dated at <40 000 years (Prosser, 1985). The other centres lie within this oldest and largest caldera. A second caldera of about 3 km diameter formed in the northern part of the older caldera at about 2400 m (a.s.l.), and the three summit craters lie within this second caldera. The oldest summit crater is von Franzius, a vegetated and eroded cone that lies to the north of the presently active crater (Fig. 1). Botos, a younger cone that formed to the south is a well preserved feature with a freshwater lake about 15 m deep in the centre; lavas from this

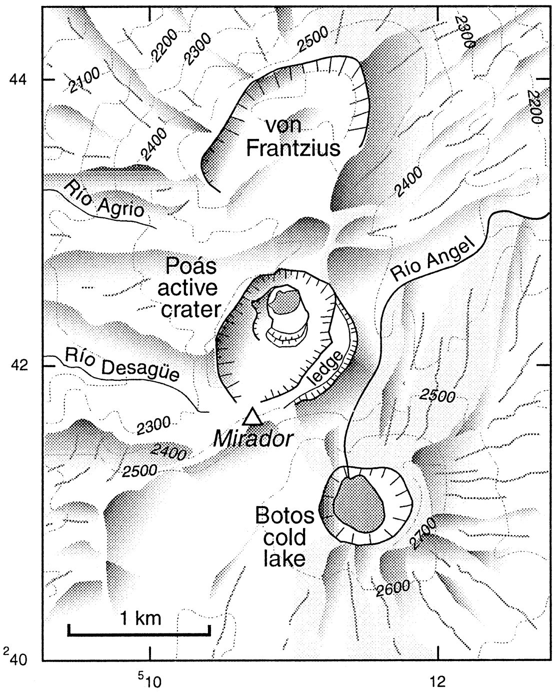


Fig. 1. Map of the summit area of Poás Volcano.

crater outcrop in the present crater walls and have been dated at 7540 ± 100 years (Prosser, 1985).

Several faults converge in the summit region and at least 400 m of vertical downthrow occurred to the west of the summit during the active lifetime of the Botos cone producing a prominent scarp and valley (Rio Desague, Fig. 1). This feature gives Poás an asymmetry with steep cliffs to the west compared with relatively gentle slopes to the north, south and east. A distinctive tephra layer visible on the southwest rim of the presently active crater is thought to be associated with collapse of the active crater, some time after cessation of activity from Botos (Prosser, 1985). There have been two distinct episodes of collapse in the presently active crater; a ledge remaining on the eastern side represents an earlier level of the crater bottom (Fig. 1). There have been several episodes historically of tephra, phreatic, strombolian, vulcanian and lava eruptions within the active crater, demonstrating the predominantly explosive nature of the volcano.

The hydrodynamics of the Poás system are complex, comprising the fresh water lake (Botos)

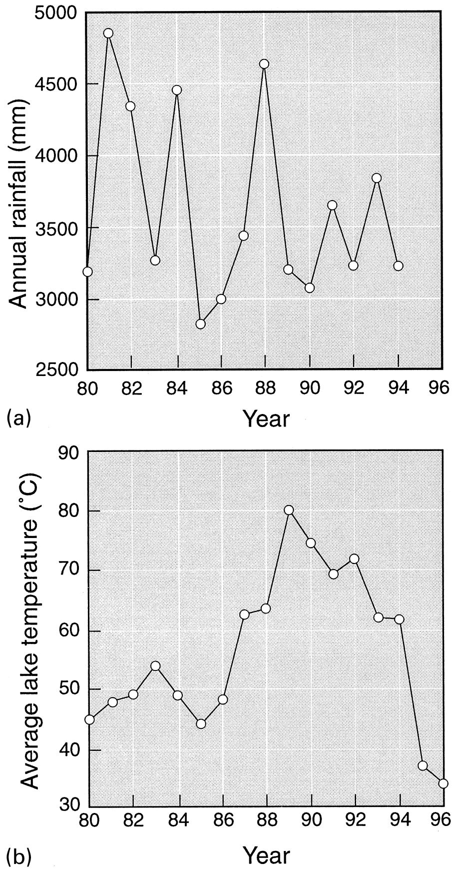
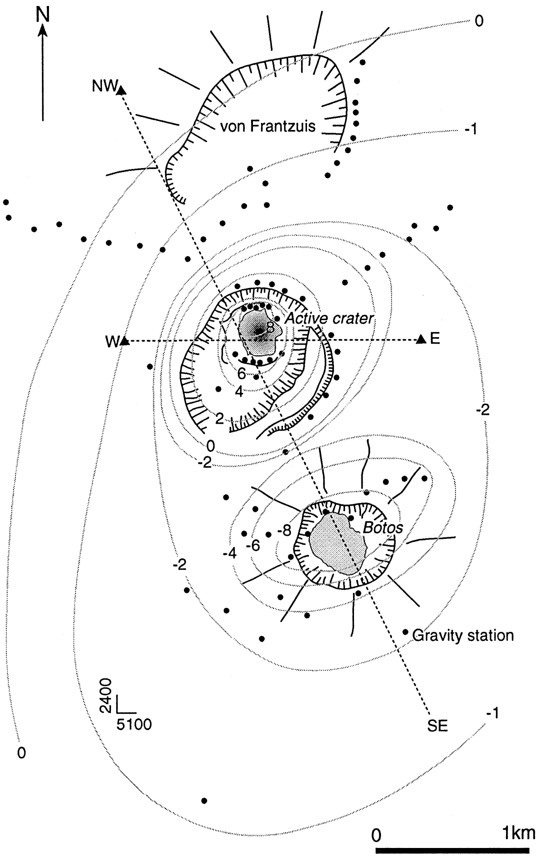


Fig. 2. (a) Annual rainfall at the summit of Poás Volcano (1985–1995). (b) Annual average temperature of Poás crater lake (1980–1996).

about 200 m above an acidic brine lake in the active crater. The fresh water lake drains into the Rio Angel and ground water seepage from Botos probably recharges the acidic crater lake (Sanford et al., 1995). There is evidence that the hydrothermal system at Poás has existed at least for much of the lifetime of the presently active crater; there are areas of hydrothermally altered material inside the crater, but it is interesting in that it appears to be tightly confined within the active crater. Outside the active crater there are no fumaroles, nor is there evidence that there ever have been. Rainfall at Poás is very heavy (Fig. 2a), particularly on the north side, averaging about 3.5 m per year at the summit (on average, about 60 kg s^{-1} of rain water enters the lake—Brantley et al., 1987). A hot spring and a set of springs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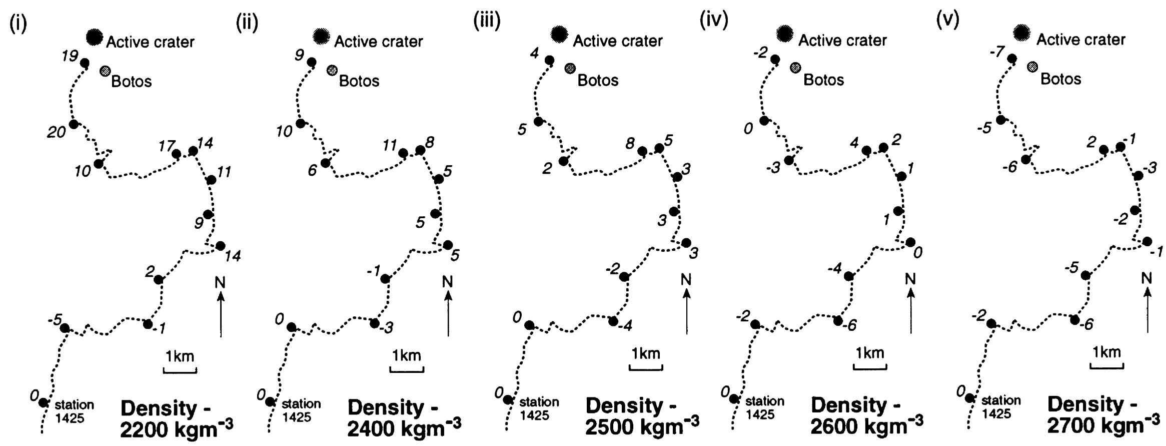
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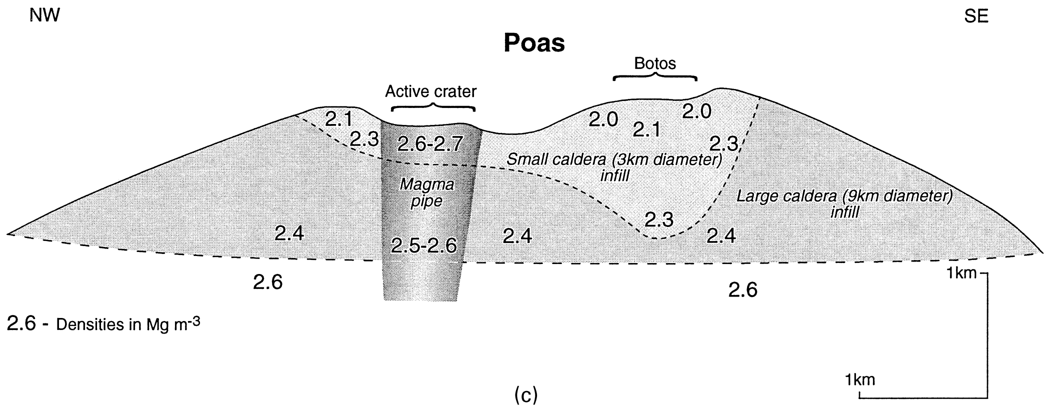

Fig. 3. (*continued*)

only slightly above ambient groundwater temperature feed the Rio Agrio (Fig. 1); chemical and isotopic data indicate that these springs represent diluted summit lakes brines (Rowe et al., 1995).

3. Eruptive activity at Poás

Historical activity at Poás is summarised by Krushensky and Escalante (1967), Raccichini and Bennett (1977), Vargas (1979) and Boza and Mendoza (1981). The largest historical eruption occurred on 25 January 1910, when a large mushroom-shaped cloud of ash and steam rose 4 km (intermittently to ca. 8 km) above the volcano summit and rained ca. 800 000 m^3 of ash over the summit and nearby villages; after this eruption, the crater lake was recorded as being about 150 m across.

Intermittent small-scale activity occurred between 1910 and 1952, when the eruptive characteristics of the volcano began to change dramatically with the onset of a phreatomagmatic phase. Geyser-like eruptions preceded the eruption of a large mushroom-shaped cloud in September 1952, which rose to a height of 6 km. The lake disappeared and was replaced by a small pyroclastic cone in January 1953. This cone, usually referred to as the dome, probably grew taller than its present 27 m and about a third of it collapsed on its northern side into a new pit crater formed during this eruption. It is probable that a lake grew intermittently in this pit crater, and by 1961 geyser-like eruptions resumed. The crater lake within this pit attained its maximum size of about 3×10^4 m^2 surface area and 40–50 m depth by 1967. A hot, acid brine crater lake in the pit within the active crater was fairly stable from 1965 to 1987 although there was occasional geyser activity and a large explosive eruption of sulphur-encrusted blocks in 1978 (Bennett and Raccichini, 1978; Bennett, 1979; Francis et al., 1980).

During the early 1980s the main site of activity moved to the dome, where fumarole temperatures rose from 92°C in December 1980 to 960°C during March–November 1981 (Casertano et al., 1987). Fumarole temperatures gradually fell until by 1986 we recorded maximum temperatures of 450°C. Fumarole temperatures remained fairly high until 1989 when the fumaroles all but disappeared on the dome; they reappeared in 1997–1998 close to the lake associated with large areas of bright yellow

Fig. 3. Summit gravity map showing negative Bouguer anomaly associated with Botos crater, positive anomaly at active crater and larger wavelength negative anomaly. Data are reduced using a density of 2400 kg m^{-3} (from Brown et al., 1987). (b) Nettleton (1939) profiles for the southern flank of Poás illustrating the importance of the choice of density for data reduction. About 4 km down the southern flank from the active crater there is point at which the best-fit Bouguer density changes from 2400 to 2600 kg m^{-3} (from Brown et al., 1987). (c) Cartoon summarizing the density structure of the summit region of Poás Volcano deduced from static gravity studies. Density values are in kg $m^{-3} \times 10^{-3}$ (from Brown et al., 1987).

sulphur deposits. Lake temperature fluctuated about the $50 \pm 20°C$ level until March 1986, when the temperature began to increase steadily, reaching almost boiling point by March 1989 (Brown et al., 1991). The level of the lake was also reasonably constant, with seasonal variations of a few meters until about 1986 when the level of the lake began to fall. It has been suggested that this was at least in part due to unusually low rainfall (Rowe et al. 1992a; Stevenson, 1992), although this was followed by a period of anomalously high rainfall (Fig. 2a) and this did not reverse the trend. As soon as the lake temperature (Fig. 2b) exceeded 65°C, conditions were again favourable for geysering to begin (Dowden et al., 1991); this occurred in mid 1987. The continued reduction in lake level and the increase in lake temperature culminated in an ash and dry steam eruption (450°C) in April 1989 (Smithsonian Institution, 1989) from the base of the dried crater lake. The eruption continued for 2–3 weeks reaching a maximum height of 1.5–2 km.

Activity after the 1989 ash eruption was strongly seasonal; during the rainy season (May–December), the crater lake began to reform, reaching a few meters depth. The lake took on a much yellower appearance than the deeper lake that existed before, due to the increased SO_2/H_2S gas ratio injected from below producing a concentration of ferric chloride complexes (Rowe et al., 1992b; Rowe, 1994) and sulphur particles. The next dry season saw a gradual reduction in lake level and the formation of discrete boiling or near boiling mud pools and mud and sulphur volcanoes a few meters tall. An eruption of gas, ash and blocks with a plume to 300 m occurred in April 1990 (Smithsonian Institution, 1990b). Vigorous fumarole activity, gas emission and phreatic eruptions from mud and sulphur pools was maintained and the level of the lake was seasonal; no further significant eruptions occurred until 1994. In April 1994, the lake sediments were erupted over a large area. Activity persisted until blocks were ejected (July–August) within the crater and a series of ash falls covered an area of 56 km^2 (Smithsonian Institution, 1994).

The crater lake began to re-establish in October 1994 apparently marking the end of this active cycle; the lake was approaching its pre-eruption level by the end of 1995 and by February 1998 had exceeded it. Its characteristic greenish hue (reflecting the dominance of ferrous iron (Fe^{2+}) over ferric iron (Fe^{3+}) as the gases injected into the lake bottom are richer in H_2S than SO_2; Rowe, 1994) has returned. Continued gas injection at the base of the lake maintains its temperature well above ambient (about 30–40°C). Fumarolic activity on the dome increased in 1997–1998 as the lake level rose.

The prevailing wind direction is from the east and noxious fumes are transported downwind to farms and villages on the northern, southern and western flanks. Before 1989, when the lake disappeared, volcanic gases were discharged into it and became hydrated and diluted. Once the lake no longer provided this buffer (Rymer and Brown, 1989), gases and particulates vented directly into the atmosphere. Acid aerosols are a particular problem, because they are small enough to be inhaled. Acid rain has been a significant problem in the region for at least the last decade and there is clearly a need to be able to predict the level of eruptive activity and pollution in the future for the purposes of urban planning and hazard mitigation in Costa Rica.

4. Static gravity

Static gravity surveys have been carried out in the summit area and in the region surrounding Poás (Thorpe et al., 1981; Locke et al., 1985; Rymer and Brown, 1986; Brown et al., 1987). Surveys reveal a positive Bouguer anomaly of ca. 10 mGal and 1 km wavelength centred on the active crater and a similar wavelength negative anomaly of ca. −6 mGal associated with Botos cone, both superimposed on a broader (3–4 km) negative anomaly of ca. −2 mGal over the entire summit area (Fig. 3a).

Modelling of these gravity data indicates that there are distinct density structures within the Poás edifice (Fig. 3b). The lower part of the edifice is characterised by the highest average density of 2600 kg m^{-3} interpreted in terms of lava flows, avalanche deposits and mud flows. The summit area has an average density of 2400 kg m^{-3} and comprises poorly consolidated pyroclastic material. The radius of this lower density summit region is about 9 km and is thought to represent an old caldera structure (Rymer and Brown, 1986; Brown et al., 1987). Within this summit region, a still lower density zone south of the active crater, centred on Botos cone, ranges from 2000–2300 kg m^{-3} and represents low-density crater-infill beneath this prehistoric centre. The large positive anomaly centred on the active crater is interpreted to be a relatively dense magma pipe (2500–2700 kg m^{-3}) intruding the lower density debris of the summit area.

5. Seismicity

Seismic data have been recorded with nearly continuous coverage at the summit of Poás since 1976, mainly by a single station located near the Visitor Centre (1 km from the southern crater rim) though more recently supplemented with temporary portable networks installed in the summit region. Seismicity at Poás falls into the usual categories for volcanoes: A-type (high frequency >3 Hz), B-type (low frequency, typically <2 Hz) and tremor (e.g. Minakami, 1969). There is a complex relationship between seismicity and activity at Poás, because of the interaction of a shallow magma body with a hydrothermal system which includes the crater lake.

Annual variations of seismicity since 1980 (Fig. 4) illustrate the long-term overall trends rather than the more 'instantaneous' (e.g. monthly or weekly) changes which can sometimes be correlated with specific eruptive events. These patterns of annual seismic activity clearly correlate with the overall eruptive cycle at Poás during the past 15 years.

A-type events have been erratic in their frequency of occurrence, with notable peaks in 1980 and 1990 typically resulting from short periods of swarm activity (Fig. 4a). The 1980 and 1990 events closely follow and were perhaps triggered by regional earthquakes (Rowe et al., 1992a). These events may therefore represent externally triggered stress release within the volcanic edifice. An increase in B-type events (Fig. 4b) following the 1980 A-type swarm and the subsequent significant increase in tremor (Fig. 4c) and fumarole temperatures on the dome (some 5 months later) led Casertano et al. (1987) to suggest that the swarm signified a 'hydrofracturing' of the magma carapace, with the time delay caused by the buffering effect of the overlying hydrothermal system. A-type events in 1986 do not correlate with

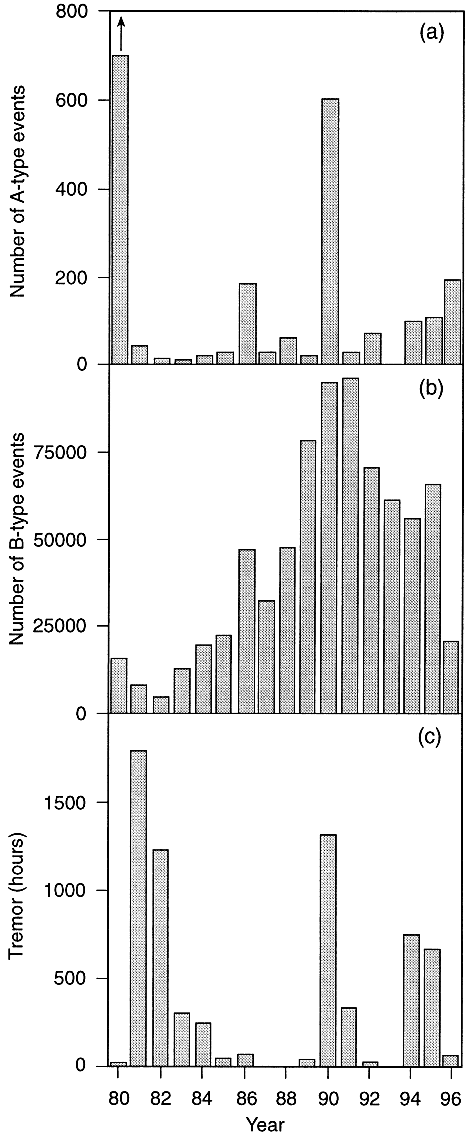

Fig. 4. Seismicity recorded at Poás Volcano (1980–1996). (a) Annual number of A-type earthquakes (>3 Hz) (1980 data unknown but >800). (b) Annual number of B-type earthquakes (<2 Hz). (c) Hours of tremor recorded annually (including data from Fernandez, 1990).

regional seismicity and probably were caused by a similar hydrofracturing of the magma carapace; these events occurred during a period of subdued surface activity and mark the onset of notable changes in several other physical phenomena (e.g. temperature, micro-gravity, lake level—see later). The events in 1990 are thought to reflect tectonic readjustment of the volcanic edifice after a moderate regional earthquake (Rojas et al., 1990). A-type activity has steadily increased over the last few years to a relatively high level in 1996 but this activity does not correlate with increases in other geophysical indicators.

The most common seismicity at Poás is B-type. Annual changes are more gradual than for A-type events (Fig. 4b). In late 1985 there was a sharp increase in the relatively steady background level of 50–100 events per day and the annual rate of activity rose to a peak in 1990–1991. Event foci determined during a 6-week period in 1989 show that low frequency events are generated within a few hundred metres of the surface and are concentrated below the crater floor (Fernandez, 1990). B-type seismicity gradually returned to background levels by 1996 as the crater lake filled to its highest level in almost 10 years.

Source mechanisms for low frequency events are much disputed especially regarding the role of fluids versus that of gases, though it is generally accepted that changes in the rate of B-type activity reflect changes in the heat and volatile flux from the underlying magma (e.g. Ferrucci, 1995). At Poás, both the interaction of fluid/volatile flow within fracture conduits and bubble formation/collapse in the hydrothermal system have been postulated as sources (Fernandez, 1990). Although there are short-term correlations between increased B-type activity and increased water saturated conditions (Rowe et al., 1992a), the longer-term trend of increasing B-type activity and its correlation with the disappearance of the crater lake suggests multiple sources for such seismicity operating at different depths.

Tremor activity has been very variable at Poás over the last 15 years, having declined from a maximum in 1981 to low levels at the start of the 1990s. Analysis of peaks in the FFT spectrum from 1981 data were interpreted in terms of eigen-frequencies of the ‘organ pipe’ response of a 1.4-k-long vertical magma conduit (McNutt, 1986). A tremor peak in 1990 followed the 1989 eruption and was associated with increased B-type seismicity. The nature of tremor during this period was relatively weak and continuous compared with the activity in the 1980s and may reflect enhanced degassing through the crater floor.

6. Power output

There have been several attempts to quantify the power output of Poás volcano (Brown et al., 1989, 1991; Rowe et al., 1992a). The lake may be used as a calorimeter; estimates of the heat lost by the lake may be derived from measurements of water temperature and surface area, wind speed and ambient air temperature. However, there are other heat sinks within the summit hydrothermal system such as fumaroles and geysers that are more difficult to quantify (Brown et al., 1989; Dowden et al., 1991). Rainwater adds mass to and removes energy from the system. Volcanic gases entering the lake from below add mass and energy to the system (Brantley et al., 1987). Surface losses can be estimated, but seepage from the lake bottom and sides and from the underlying hydrothermal system can only be estimated by studying the fumaroles and the Rio Agrio (Rowe et al., 1992a). Only rainwater can actually be measured although the catchment area for calculation of the input to the lake is a controversial assumption (Stevenson, 1992; Sanford et al., 1995). Temperature and chemical compositions of the Rio Agrio can also be measured and these provide an estimate of the seepage from the hydrothermal system (Rowe et al., 1995). In the steady state, of course, mass and energy inputs are balanced by outputs. Throughout much of the 1980s when the crater lake was stable, the power output of the system was estimated to be about 200 MW on average (Brown et al., 1989, 1991). However, for short periods after 1978 and for an extended period in 1981–1983 when the SO_2 flux was several hundred tonnes per day and dome fumerole temperatures were $>800°C$, the power output exceeded 500 MW (Rowe et al., 1992a). This increased to 300–400 MW in 1987–1989 and peaked at 1000 MW during geysering episodes (Brown et al., 1991). After this time, the model breaks down because the lake almost disappeared and no longer acted as a calorimeter. At that point, volcanic gases were venting directly to the atmosphere and a new phase of activity had begun.

7. Ground deformation

A triangulateration line running radially (south) from the crater was measured on three occasions, in 1983, 1988 and 1989. Between 1983 and 1988 deflation was observed up to 5 cm (8.3 μrad) 3–9 km from the crater. Between 1988 and 1989 general inflation (up to 5 cm, 4.8 μrad) occurred along the whole line. Levelling lines running from the Visitor Centre to the Mirador and from Poste to Cobro (Fig. 5a) were measured between 1991 and 1994 but again no systematic trend is found; the total movement over this period is <15 μrad (Fig. 5b). The distance to reflector prisms placed in and around the active crater (Fig. 5a) has been measured using an edm theodolite. The changes are small but there is evidence for a contraction of 10 ppm between 1991 and 1994 (Fig. 5c). Dry tilt measurements (summit stations located on Fig. 5a) have revealed no systematic trend and no significant changes (<5 μrad).

Height control for the micro-gravity stations has been provided by edm theodolite and (since 1994) dual frequency GPS. Apart from deflation on the dome (probably due to collapse) of up to 34 cm between 1987 and 1988 (Brown et al., 1991) vertical deformation has been less than ± 10 cm for all stations (equivalent to an uncertainty in micro-gravity value of ± 20 μGal).

In general, deformation at Poás over the last 10 years has been small-scale and seems to be related to local variations in the hydrothermal system; a general tendency for deflation as the hydrothermal system lowered was observed, but there have been no major ground movements over this period.

8. Micro-gravity changes

Systematic repeat measurements of micro-gravity at stations in and around the active crater at Poás began in 1985. Observations are made in the same sequence on each occasion, starting at points well away from the volcano (approximately 30 km south of the summit), travelling up the southern flank to the reference station (1445a; the penultimate station located on Fig. 3b) 2 km south of the crater rim and then around the crater rim and down to the crater floor. This procedure is typically repeated five times resulting in a standard deviation on gravity differences at each station relative to the value at station 1445a of <20 μGal. Data were collected on at least one occasion and sometimes three in every year between 1985 and 1996 except 1995 (Fig. 6). There appears to be no significant seasonal effect in the data, but there are long term changes, the largest of which are observed at crater bottom stations.

Simultaneous and persistent gravity increases of up to 360 μGal occurred between 1985 and 1989 at stations on the dome and on the southern crater floor (D1, E1, E3, E5, E6, G1; located in Fig. 6 inset). After the 1989 eruption, gravity stopped increasing and by 1992 it was beginning to decrease again reaching 1985 values by 1998 at E6, E1 and D1 but at E3 the recovery seems to have been more rapid. In contrast, the northern crater floor stations (D2, D3) showed consistent gravity decreases in the period 1985–1991, followed by more recent gravity increases. In 1998, station D2a was below the lake surface.

The largest elevation change detected at any station over this period was -34 cm at G1 which occurred between 1987 and 1988 (Brown et al., 1991) and was probably a local effect of dome re-adjustment. Elevation changes were an order of magnitude less than this for all other stations at all times and within observation error (± 5 cm) for the techniques used (EDM theodolite 1985–1993; differential GPS 1994–1998). The 34 cm fall at G1 can account for only 90 μGal of the observed 360 μGal increase, and so ground deformation is clearly not the main source of gravity changes at Poás.

Between 1985 and 1989, while the level of the crater lake was falling and thermal activity was migrating from the dome to the lake, culminating in explosive eruptions from the dried lake bed, gravity increased dramatically in the southern part of the crater floor. Over the same period, there is relatively little variation in gravity on the crater rim, but some decrease in the northern part of the crater bottom. The dramatic fall in crater lake level between 1985 and 1989 would presumably be accompanied by a fall in the surrounding water table. The gravity signature in the northern part of the crater bottom (D2, D3) is consistent with this as a fall in the water table of 30 m (lake level actually fell by 50 m) could account for the observed 250 μGal decrease (assuming 20% porosity). Gravity at these stations has increased since 1992–1923 suggesting that the water table was rising at this time and that this heralded the recovery of the hydrothermal system although it preceded the re-appearance of the lake by 2–3 years. By 1998 when the lake level was at its highest since gravity measurements began and station D2a was submerged, gravity at D3 was almost as high as in 1986. If the gravity changes at this station are mainly due to the level of the hydrothermal system, it is clear that it has been modified and disrupted by the eruption cycle.

The fall in lake level and water table must also have

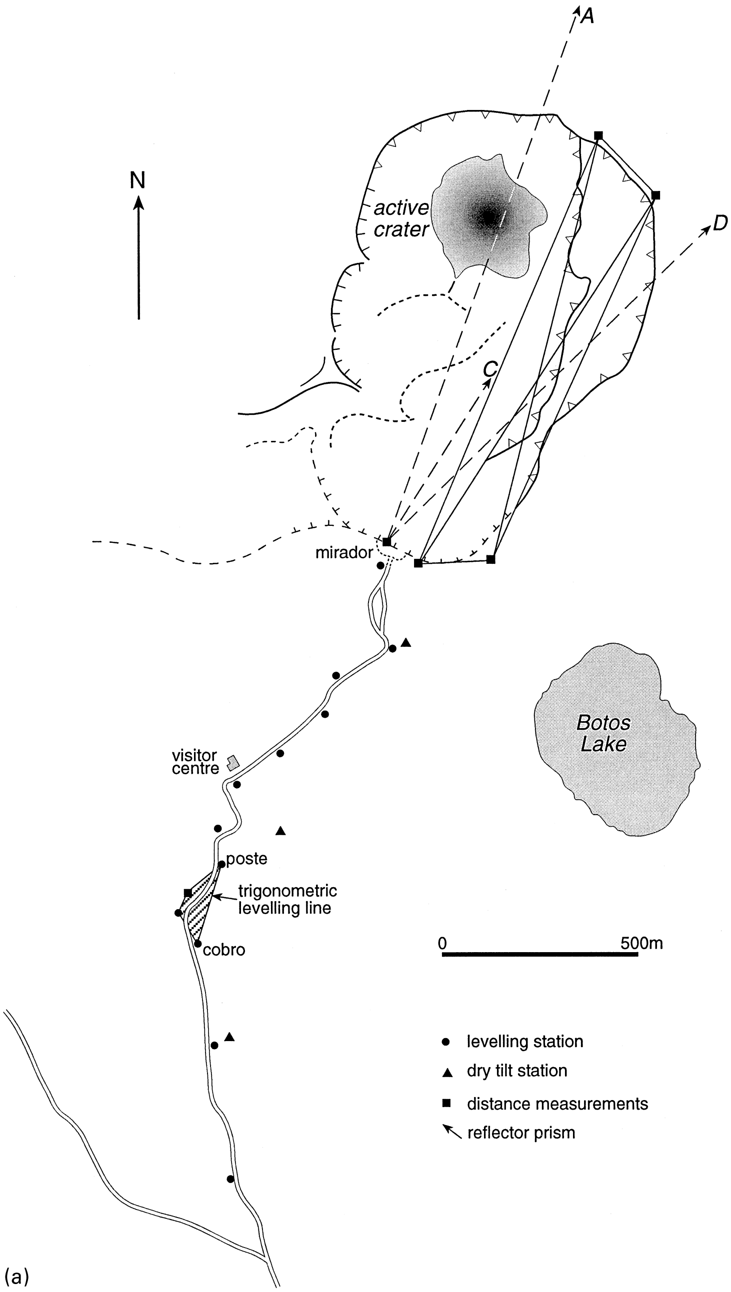

Fig. 5. (a) Map showing location of stations for levelling, dry tilt and distance measurements. (b) Total tilt observed along levelling lines. (c) Changes in line length between summit crater prism reflectors.

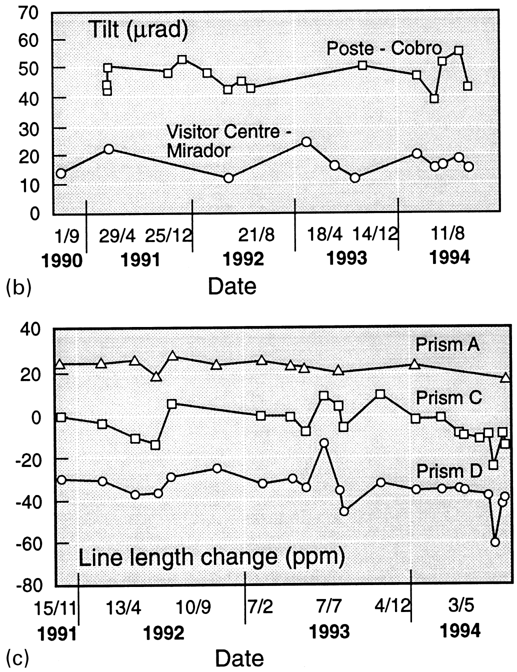

Fig. 5. (*continued*)

affected stations near the dome and the gravity increases observed at these stations is therefore likely to have been subdued. Since the gravity changes in both the dome and south crater regions are very similar in the pre-eruption period and share similarities in the post-eruption period, they are likely to have a common cause. These variations cannot be explained in terms of ground deformation and therefore they must reflect sub-surface mass changes. A simple point-source approximation indicates that a minimum mass increase of about 10^8 kg is required to account for the observed gravity increases, other geometries require a greater mass change. For example, if the change is assumed to occur within a vertical cylinder (radius 100 m extending from the surface to 400 m depth), then a mass change of 10^9 kg is required. A rise in water table is an unlikely explanation for the new mass in view of the fall in lake level and the increased surface activity over this period, but a rise in magma level or the deposition of sulphur and other minerals are possible mechanisms. These gravity increases are widespread in the southern part of the crater from E3 to D1 but the magnitude of the changes is greatest at the stations on the dome. Deposition of sulphur or other minerals is unlikely to be so widespread and given the focus of activity and hence the highest heat flow at the lake and dome, it would seem unlikely that deposition would be concentrated in these areas.

The post-eruption gravity decreases are a further indication that mineral deposition is an unlikely mechanism since a sudden remobilisation of the previously deposited material would be required to explain the gravity observations. Rather, we suggest that the pre-eruption gravity increases result from an increase in magma pressure causing the magma level to rise in the upper part of the conduit system. After the 1989 eruption the pressure reduced and the magma level dropped, draining completely by 1991 from peripheral conduits in the west (E3) and later (1993) from the east (D1). However, the magma remains elevated centrally below the dome (E5, E6, G1) compared with its pre-eruption level.

Although the gravity increase data indicate that there was an intrusive event, the heat energy available in the calculated mass (10^{14} J assuming a 500 K temperature decrease and complete crystallisation of 10^8 kg of magma) is insufficient to account for the observed power output increase at the summit. Indeed, it is not large enough (1 MW over 3 years) to produce even the background energy flux observed at Poás (typically 200 MW, Rowe et al., 1992a).

Gravity decreases from 1994–1998 at southern crater bottom and dome stations accompany the re-establishment of the lake, although they are of the opposite sign to that expected due to a water table rise. New and vigorous fumaroles have appeared on the northern face of the dome and in the inner walls of the main crater, suggesting that a tightly confined aqueous system (the lake is immediately north of the dome) is exerting enough hydrostatic pressure on the underlying magmatic gas that some of the fumaroles have migrated south of the lake. A gravity decrease would be expected if water saturated rocks became dehydrated and acted as gas conduits rather than a fluid store. Thus, the gravity increase followed by a decrease observed at these stations might simply reflect aqueous fluid movements. However, we consider it likely that there was a magmatic contribution to the gravity increases.

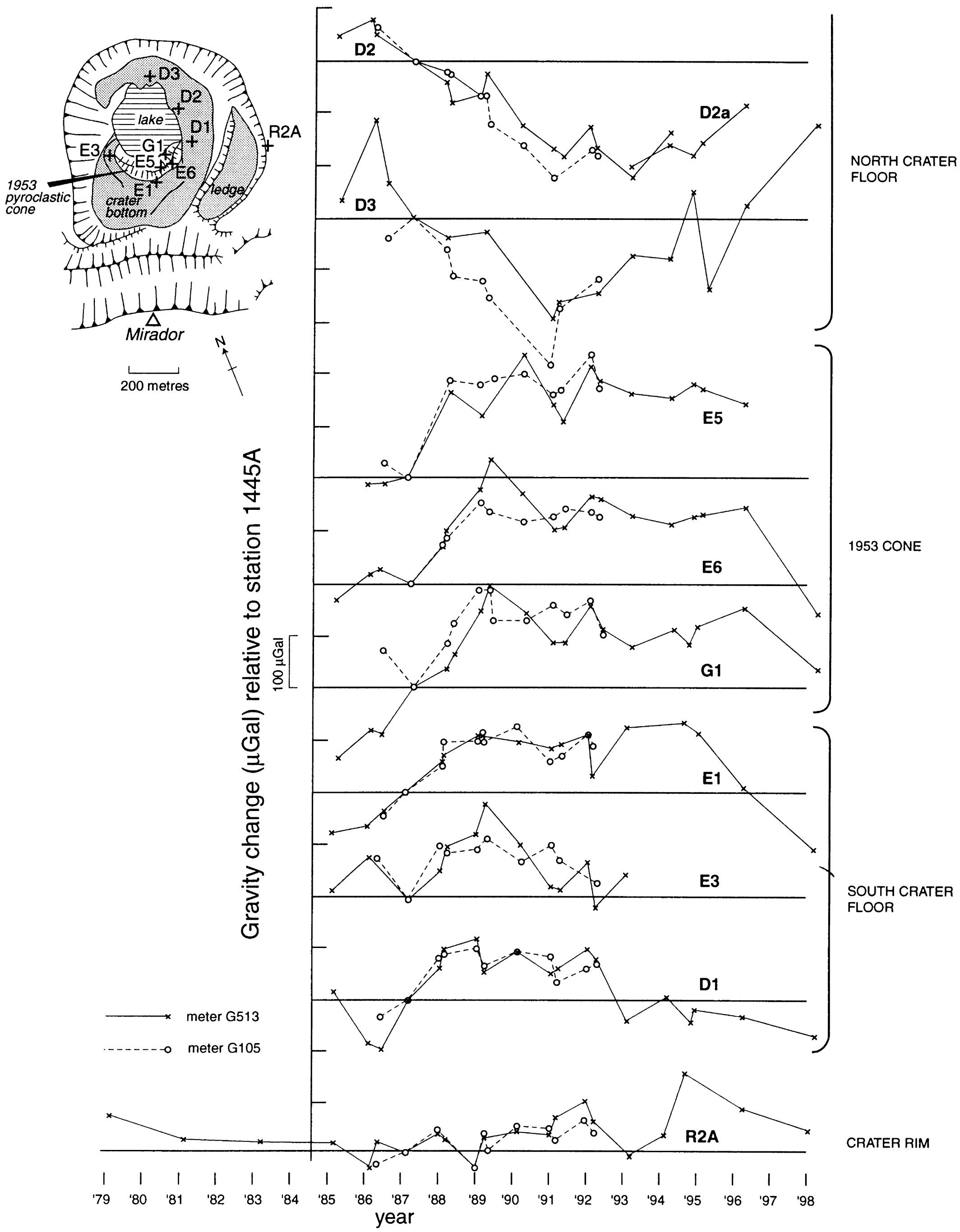

Fig. 6. Microgravity variations within the crater and on the crater rim of Poás Volcano. Locations of the gravity stations are shown; station D2a is located adjacent to station D2 (which was destroyed).

9. Magnetic signature

A total-field magnetic survey of the summit area in early 1979 (Locke et al., 1985) showed a local negative anomaly of −400 nT (approx. 300 m wide) surrounded to the south and east by a more extensive positive anomaly of 200 nT (approx. 750 m wide), all superimposed on a broad-scale negative anomaly of at least −600 nT (Fig. 7). These anomalies were initially interpreted in terms of a cylindrical, relatively magnetic body beneath the crater surrounded by less magnetic material (Locke et al., 1985). However, the local negative anomaly centred on the lake may be more important, reflecting either the presence of material above its Curie temperature beneath the crater or the presence of altered but not necessarily hot material in which magnetite has been destroyed by the circulation of hydrothermal brines. Hot material beneath the crater could result from either a small intrusion of fresh magma (which might have been responsible for the phreatic activity coincident with the survey) emplaced to less than 100 m below the lake or, from heating by magma intruded at greater depth. We calculate that the dimensions of the hot zone, consistent with the local negative magnetic anomaly, are a few tens of metres across by ca. 500 m thick; such a body lies easily within the boundaries of the magma conduit system modelled by Rymer and Brown (1986) and Brown et al. (1987).

A detailed study of the magnetic total-field over the crater floor between March 1980 and October 1981 was made by Casertano et al. (1987). These two surveys, comprising 33 stations, of which six were within 230 m of the fumaroles on the dome, bracketed a seismic event and a dramatic rise in fumarole temperatures. These data yielded an anomaly pattern similar to that described above, but of particular interest are the changes in the magnetic field between the two sets of measurements. A maximum total magnetic field decrease of almost 200 nT occurred between the two sets of observations (Fig. 8). This was interpreted as a decrease in the magnetization of the dome as the magmatic gases venting through it raised its interior temperature above the Curie point. The mid-1980 A-type seismic swarm apparently triggered the rise in fumarole temperatures, which by October 1981 had significantly heated the interior of the dome.

We propose therefore a re-interpretation of the Locke et al. (1985) data. It seems most likely that the region of depressed magnetic signature in our

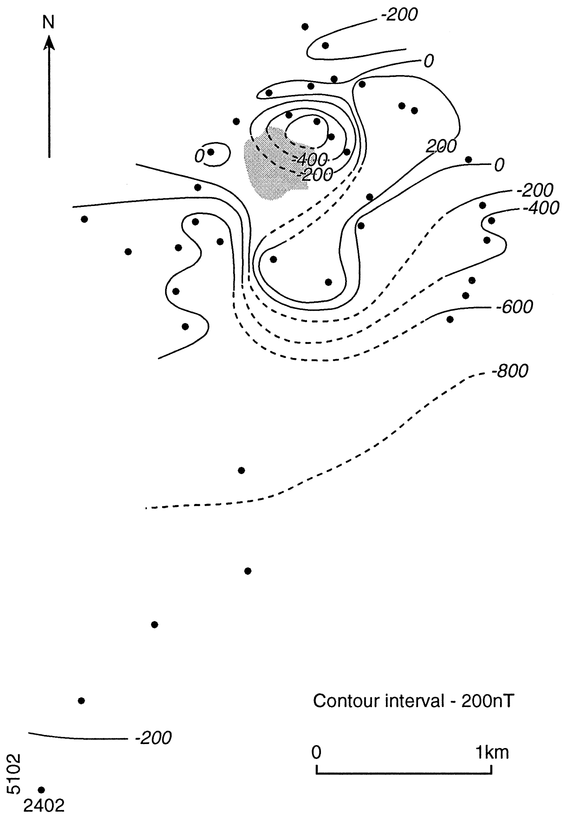

Fig. 7. Total field magnetic anomaly map of the Poás summit area. Stippled region represents the crater lake (from Locke et al., 1985).

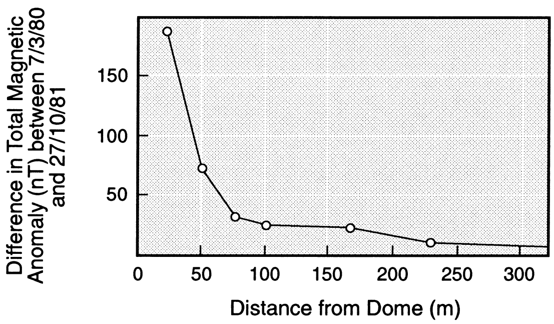

Fig. 8. Total-field magnetic anomaly changes recorded between 7/3/80 and 27/10/81 (20 months) at a series of stations radially distant from the 1953 dome (from Casertano et al., 1987).

model reflects some combination of enhanced alteration and heating from a magma source at greater depth. Intrusion of hot material to within 100 m of the surface would produce elevated fumarole temperatures; dome fumarole temperatures at this time were near boiling point, not several hundred degrees Celcius as expected if magma were close to the surface (Stevenson, 1993).

Thus these combined data sets indicate that magma was located at least 500 m below the surface in 1979, but that it rose after the A-type seismicity in 1980 to shallow depth below the dome by October 1981. Dome fumarole temperatures of 900°C observed after this event imply a depth to magma of just a few metres (Stevenson, 1992, 1993).

10. Geochemical variations within the lake

Volcanic gases venting beneath the lake form complex compounds with the aqueous mixture and variations in some chemical species through time have been correlated with activity. The crater lake at the summit of Ruapehu Volcano, New Zealand, for example, has been closely monitored over a period of several years and variations in the Mg and Cl concentration have been linked to eruptions (Hurst et al., 1991; Christenson and Wood, 1993). On Poás, chemical variations within the lake itself and in the Rio Agrio, which is connected underground to the hydrothermal system, have been investigated (Rowe et al., 1995). Classical geochemical parameters such as the Mg/Cl and SO_4/Cl ratios have been found to give little indication of impending eruptive activity at Poás (Rowe et al., 1992b). Changes in the SO_2/H_2S content of sub-aqueous fumaroles, detectable indirectly through polythionate concentrations however, correlate strongly with seismicity and thus with volcanic processes and eruptive activity. Total polythionate concentrations increased dramatically in late 1985 (Fig. 9) and the number of B-type low frequency earthquakes increased at the same time (Rowe et al., 1992b). The peak in polythionate concentration in mid-July 1986 correlates with increased A-type seismic activity (Figs. 4a and 9). Rowe et al. (1992b) considered that the seismic activity resulted from hydrofracturing of the carapace above the magma which led to an increase in gas flux.

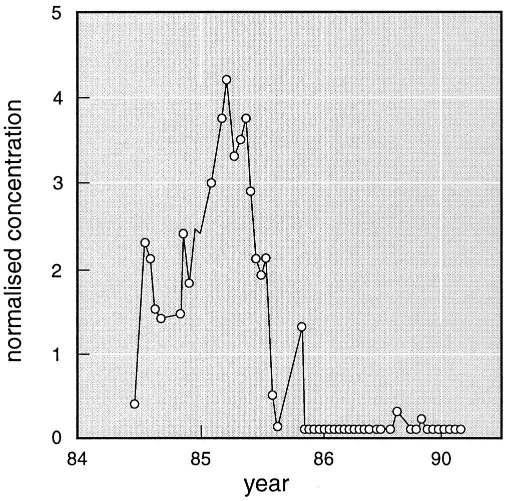

Fig. 9. Variation (1985–1990) in polythionate concentration normalised to fluoride in Poás crater lake (after Rowe et al., 1992b).

11. Summary and discussion

A compilation of key data sets covering the period 1980–1996 illustrates a striking coherence in the behaviour of the various observed physical phenomena (Fig. 10). Elevated levels of A-type seismicity in 1980 coincide with the beginning of a trend of elevated lake temperatures that peaked in 1983 but lake depth was not affected. A significant increase in A-type activity in 1986 marked the beginning of a period of annually increasing B-type activity. Lake temperature also increased steadily over this period and lake level decreased, slowly at first, then rapidly until it disappeared at the onset of explosive activity in April 1989. Concurrently, micro-gravity in the southern part of the crater floor (typified by data from station G1) showed consistent increases from 1986 to 1989 and on the northern crater floor (stations D2 and D2a) gravity decreased along with lake level. A further increase in A-type seismicity in 1990 and a maximum in B-type activity in 1990–1991 was associated with minimal gravity change on the southern crater floor. The lake covered the pit crater floor only seasonally at this time although lake temperatures were beginning to fall. Gravity in the northern part of the crater increased consistently from 1993 in response to a rising water-table while in the southern part of the crater disparate rates of gravity decrease

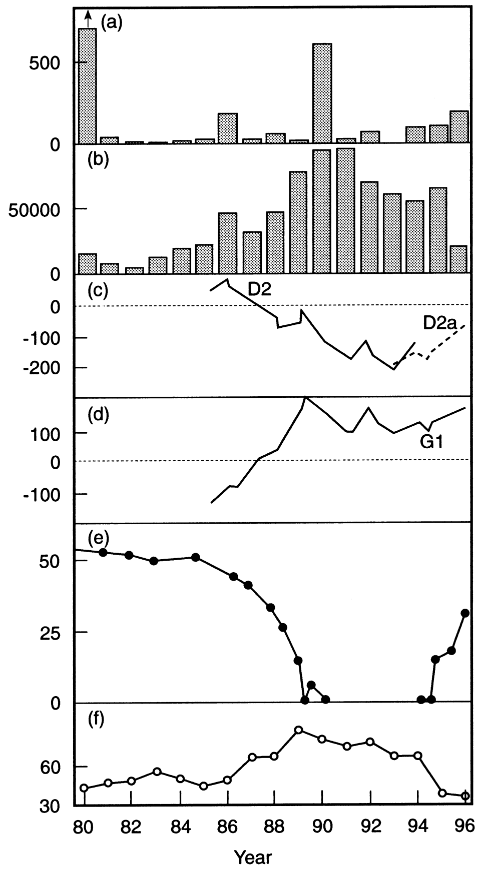

Fig. 10. Summary of observations 1980–1996 showing: (a) annual number of A-type earthquakes; (b) annual number of B-type earthquakes; (c) gravity variations (μGal) at station D2/D2a; (d) gravity variations (μGal) at station G1; (e) lake depth (m); and (f) mean annual lake temperature (°C).

occurred reflecting non-uniform changes of magma level in the conduit system or removal of water from this region. Lake level rose from late 1994 and the temperature began to fall quickly. B-type seismicity also began to decrease over this period, but A-type seismicity continued to increase gradually. The approximate correlation between the number of B-type events and lake level suggests that these events may be associated with the interaction between fluids and vapour below the lake bottom rather than with magmatic intrusion.

A high-frequency seismic event in 1980, attributed to hydrofracturing of the carapace (Casertano et al., 1987) was followed by the development of incandescent fissures at temperatures of up to 1000°C on the dome. Decreases in the local magnetic field in the vicinity of the dome (Casertano et al., 1987) after this event are consistent with heating of material above its Curie temperature at shallow depth. Since the high temperature of the fissures is indicative of magma at depths of only a few metres (Stevenson, 1992), we suggest that there was an intrusive event in 1980 beneath the dome. This model is consistent with the available geophysical data and the transfer of the main focus of degassing from beneath the lake to the dome area at this time. The intrusion cooled by at least 700°C over the following 5 years (Brown et al., 1991) and activity once again returned to 'background' degassing through the hydrothermal system beneath the lake.

The micro-gravity increases from 1986 to 1989 indicate that magma was also intruded at this time. Polythionate ion concentrations within the lake peaked in 1986 with an increase in A-type seismicity (Rowe et al., 1992b) which was attributed to a hydrofracturing event. We suggest that this hydrofracturing event in 1986 allowed the rise of magma. The calculated mass of the intrusion is several orders of magnitude too small to account for the observed increase in heat flux at Poás, indeed it would not even provide enough energy to keep the lake above ambient temperature. Magma intrusion, therefore, was not the cause of the disruption of the hydrothermal system and the disappearance of the lake but an expression of increased heat flux from the deeper feeder system.

It is not possible to quantify the relative sizes of the intrusive events in 1980 and 1986–1989, although more A-type seismicity and a greater power output were associated with the former, however the effects at the surface of these two events were very different. The location of the intrusion seems to be critically important; an intrusion beneath the dome allows excess heat energy and some gases to be dissipated away from the lake with only a minor increase in lake temperature, but a sub-lake intrusion has a more dramatic effect on the hydrothermal system. The location of the 1986–1989 intrusion meant that the lake evaporated away and the shallow hydrothermal system was severely disrupted. High temperature and pressure volcanic gases subsequently vented directly to the atmosphere and the lake was not able to re-establish until the heat and gas flux reduced. The longevity and magnitude of the surface manifestations

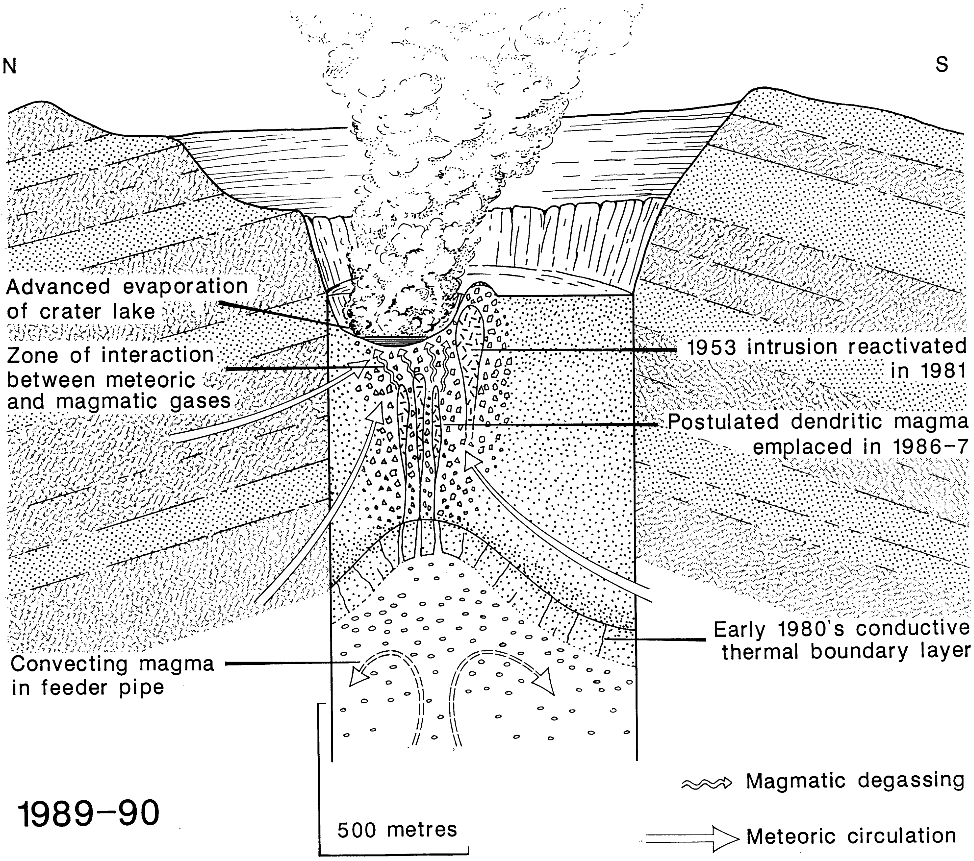

Fig. 11. Schematic model of the upper plumbing system at Poás. An intrusion to shallow depth beneath the dome in 1980 was rapidly separated from the underlying magma feeder. A series of dendritic bodies intruded between 1986–1989 remained connected to the magma feeder until at least 1992 facilitating heat and gas transport beneath the crater floor. A thermal boundary layer separates the magma feeder from the hydrothermal system which is recharged by meteoric water.

of the 1986–1989 event may mean that the transfer of heat and gas flux from the parent magma body was greater than during the earlier event.

A schematic model for the upper part of the Poás edifice (Fig. 11) illustrates the relationship between the shallow conduit system and the deeper magma feeder. The depth to this magma body is around 500 m (Rymer and Brown, 1987); this is the source of heat and gas for the summit region. Even during quiescent periods there is a significant heat and gas influx to the crater lake. During these periods, the hydrothermal system probably percolates downwards cracking the boundary layer between convecting magma and overlying rubble; heat and gases are transported upwards by the hydrothermal system (Rowe et al., 1992b). Intrusive events beneath the dome (1980) and the lake (1986–1989) brought magma closer to the surface. The latter intrusion, in the form of several dendritic pathways remained active for about 3 years and the magma remained hydraulically linked to the underlying body allowing convection to occur. This increased the efficiency of heat and gas transport to the surface as buoyant, volatile-rich magma rose displacing cooler, volatile-depleted magma. Scaled laboratory experiments show that an individual 10 m diameter conduit of the type envisaged at Poás (Fig. 11) can maintain convective processes over a height of several hundred metres (Stevenson, 1992). In the waning stages of the eruptive cycle, more magma in these small intrusions moved down than was replaced by upwelling new material and the surface activity subsided. As the lake reappeared, the hydrothermal system evolved, possibly migrating at least partially towards the northern end of the crater.

This paper has illustrated the benefit of integrated geophysical and geochemical studies for developing an understanding of the dynamic processes at active volcanoes. Not all techniques have been equally effective in identifying the mechanisms involved at the various stages of the eruptive cycle. A-type seismicity for example, is commonly used as an early-warning indicator of renewed activity at volcanoes but has been the least reliable precursory phenomenon at Poás. A-type events are usually associated with fracture mechanisms related to increased magma or volatile pressure, and this is the case in 1980 and 1986. However, the 1990 seismic event was not associated with magmatic intrusion and there is no obvious correlation between this recent increase and other geophysical indicators. We suggest that these A-type events are associated with continued downwards fracturing of the carapace over the partially molten magma by cooling due to the hydrothermal system (Fig. 11). Such continued fracturing is necessary to maintain the background ca. 200 MW power output at Poás. From the variable number of A-type events observed over the last 15 years, hydrofracturing is clearly not a steady process.

The minimal deformation observed is consistent with a degassing hydrothermal system without magma intrusion. This technique, in contrast to results obtained at numerous other volcanoes (e.g. Murray et al., 1995) is apparently not useful as a primary monitoring tool at Poás.

Poás crater lake quickly re-established once the heat flux reduced after the last phreatic eruption in August 1994. The volcano now appears to be in a state of dynamic equilibrium in which the lake is able to condense incoming volcanic gases thus buffering variations in heat and mass flux and allowing the large body of andesitic magma beneath the active crater to degas with minimal effect on the local environment (see however, Brantley et al. (1987) and Rowe et al. (1995)). Thus the crater lake has acted as both a moderator and index of volcanic processes at Poás and has now moved from Pasternack and Varekamp's 1997 designation of 'peak activity' to 'medium activity'. The main hazards at Poás now are therefore atmospheric pollution and the seepage of toxic fluids (Pasternack and Varekamp, 1997); phreatic eruptions and lahars are unlikely to occur until the power output increases again following a hydrofracturing event and magma intrusion.

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