



SECTION 1 (and 3)

SUBMARINE RIFT ERUPTIONS
AND OCEANIC
ISLAND ACTIVITY

Convenor :

Prof. H.-U. SCHMINCKE
University of Bochum
Germany (Fed. Rep.)

MOTION OF THE PLATES AS FUNCTION OF THE LOW-Q ZONES BENEATH SPREADING CENTERS

by

J. A. CANAS

Department of Earth & Atmos. Sci.,
Saint Louis University,
St. Louis, MO 63156

ABSTRACT

Fundamental-mode Rayleigh waves generated by several earthquakes situated along great-circle paths between pairs of seismograph stations on the American and European coastal regions and on Atlantic islands have been analyzed to obtain anelastic attenuation coefficients. Inversion of the attenuation data yield a model for the Mid-Atlantic Ridge.

The mean results obtained in this study are : (1) A strong low-Q zone is situated in the upper mantle beneath the bottom of the lithosphere (2) the bottom of the Asthenosphere is situated at a depth of around 250 km.

Comparison of the above results with earlier studies in the Pacific Ocean suggest that the velocity of plate motion is slower when attenuation coefficients and internal friction values are smaller and vice versa.

(This paper was not read at the Symposium)

SCALES OF RECENT SUBMARINE VOLCANISM

by

G. P. AVDEIKO

Institute of Volcanology,
Petropavlovsk-Kamchatsky, USSR

ABSTRACT

Submarine volcanic activity is confined to (1) rift zones of mid-ocean ridges (MOR) and transform faults, (2) hot spots (HS) and belts of hot spots («hot lines»), (3) systems of island arcs (IA) and (4) marginal seas (MS). Volcanic activity may occur on the oceanic flanks of deep-sea trenches and marginal swells between trenches and the ocean floor.

Rift zones of MOR exhibit the deep-sea fissure eruptions of low-potassium oceanic tholeiites with some variations of mineralogical and chemical compositions. They produce pillow and ropy flows, volcanic cones generally being not formed. As a result of these eruptions, more than 1×10^{10} t of lava per year pour out on the ocean floor.

HS volcanoes pass through the submarine and insular stages of their development. During the submarine stage outpourings of tholeiites occur which differ from oceanic tholeiites in high content of TiO_2 , K_2O , P_2O_5 and some other characteristics.

They form shield volcanoes rising from depths of 4 to 6 km. Rocks of alkalic and nepheline series constituting commonly not more than 1-2 % pour out at the final stages. However, for the HS belts alkalic basalts are more characteristic. HS volcanoes supply to the surface about 1×10^9 t of volcanic material per year, more than 90 % appearing as lava.

IA submarine volcanoes are close in composition of erupted products to terrestrial volcanoes. They are composed of rocks of calc-alkalic series from basalts to dacites. Lava outpourings are characteristic of submarine volcanoes whose tops are located deeper than the first hundreds of meters. Explosive eruptions are indicative of shallow-sea volcanoes whose tops appear periodically above sea level. According to approximate estimates, the «output» of IA submarine volcanoes is $1-1.5 \times 10^9$ t/yr. Terrestrial volcanoes, according to calculations made by E. K. Markhinin, supply about 3×10^9 t of volcanic material per year.

Fissure outpourings of basalts differing in composition from other types of submarine basalts are characteristic of MS. The total amount of MS volcanites does not exceed $0.5 - 1 \times 10^9$ t of lava per year.

The total «output» of all the Recent submarine volcanoes is more than 1.3×10^{10} t of volcanic material per year, that is four times greater the «output» of terrestrial volcanoes of IA and inner parts of continents.

(This paper was not read at the Symposium)

MAGNETIC PROPERTIES OF SOME THOLEIITIC BASALTS DREDGED FROM THE GULF OF CALIFORNIA, MEXICO

by

J. URRUTIA-FUCUGAUCHI

School of Physics,
The University, Newcastle upon Tyne
NE1 7RU Great Britain

ABSTRACT

This study is based on dredged samples collected from the mouth of the Gulf of California, a young opening oceanic basin. The samples are distributed between the East Pacific Rise crest and the Baja California continental slope. Ages are assigned on their position relative to magnetic anomalies and bathymetric profiles. Results from (1) non-destructive measurements, e.g. NRM intensity and direction, initial susceptibility, magnetic anisotropy and viscosity effects ; (2) semi-destructive measurements, e.g. ARM acquisition and AF demagnetization of NRM and ARM ; and (3) destructive measurements, e.g. viscous-partial TRM acquisition, are presented. Geochemical results of major and trace elements, including REE, are also presented. The samples show increasing alteration effects and

enrichment of light-REE with distance away from the rise crest. Among the results we may mention: (1) the intensity pattern does not conform that usually expected for profiles normal to spreading centers; (2) low-temperature weathering is of considerable importance, and its effects correlate well with time; (3) viscous effects at high temperatures (up to 200-300° C) are important in controlling the intensity pattern; and (4) magnetic anisotropy may constitute an alternative feature for determining orientation of dredged basalts.

(This paper was not read at the Symposium)

OBSERVED GROUND DEFORMATION DURING THE KRAFLA ERUPTION OF MARCH 16, 1980

by
EYSTEINN TRYGGVASON
Nordic Volcanological Institute
University of Iceland

ABSTRACT

The Krafla volcano has been monitored with continuously recording tiltmeters and seismometers and frequent geodetic measurements since the beginning of the present episode of activity in 1975. The inflation-deflation sequence of the volcano showed striking regularity in 1977 and 1978 but it became more erratic in late 1979 with slow inflation interrupted by small deflations.

The eruption of March 16, 1980 was preceded by rapid deflation which started at 15:15 (GMT) and intense volcanic tremor started simultaneously. The subsidence became very rapid at about 16:00, about three times more rapid than had been seen in any previous subsidence event. The eruption was first seen at about 16:20, but it may have started 10 to 20 minutes earlier. It lasted until about 22^h that same night. The deflation of the volcano ceased at about 03^h next morning,

March 17, and a new inflation started within a few hours. Tilt observations indicate that roughly $30 \times 10^6 \text{ m}^3$ of magma left the Krafla magma chamber, but only some 10 % of this came to the surface as very fluid basaltic lava. New fissures extended through the Krafla volcano over a distance of about 21 km and the widening of the fissure zone was about 1.5 m.

INTRODUCTION

A period of volcanic and tectonic activity started in 1975 in North Iceland, and the center of activity is located in the Krafla Central Volcano and the associated fissure swarm (Björnsson et al. 1977). This activity has been characterized by alternating inflations and deflations of the volcano, and episodic widenings of the fissure swarm (Fig. 1). The inflation of the volcano progresses at a rate of 0.5 to 1.0 cm per day for one to seven months between deflations, which last from

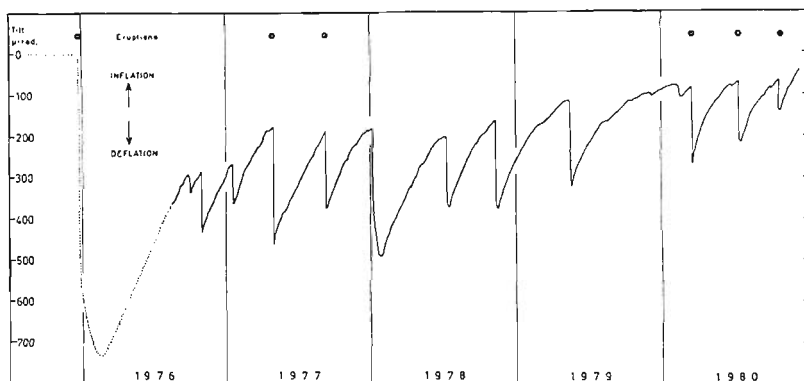


Figure 1 — North component of tilt at the Krafla power station. Before August 20, 1976 (dotted curve), no tiltmeter was operated, but several levelings give the approximate tilt values. Filled circles give the times of volcanic eruptions.

a few hours to 20 days. The rate of deflation varies greatly from one deflation event to another (Björnsson et al. 1979, Tryggvason 1980). Widening of the fissure swarm associated with opening of new and old fissures, vertical fault displacements, earthquake swarms in the fissure swarm, and sometimes outpouring of basaltic lava, occurs only during deflations of the Krafla volcano. Only a fraction of the fissure swarm widens during each deflation event, often about 20 km section, but the part of the fissure swarm which has widened during the sequence of events since 1975 extends from Axarfjörður in the north to the southern part of the Mývatn area in the south, or over an 80 km distance. The southernmost part of the fissure swarm has not been affected by the present activity as yet (Björnsson et al. 1979).

After the deflation event of May 13 to 18, 1979 (Tryggvason 1980) inflation proceeded at the usual rate, and in mid October 1979 the Krafla volcano had reached the same stage of inflation as immediately before the May deflation. The inflation rate in October and November 1979 was rather slow or only 1.0 to 1.5 mm per day, and in early December 1979 a small deflation of about 4.5 cm occurred and another small deflation of 10 to 12 cm occurred in early February 1980. The rate of inflation remained rather slow between these mini-deflations and also after the February 1980 deflation, and in mid March the ground surface in the central part of the Krafla volcano was about 12 cm higher than immediately before the May 1979 deflation.

On March 16, at about 15^h 15^m, rapid deflation of the Krafla volcano started. The maximum rate of deflation of about 50 cm per hour occurred at 16^h 30^m. This was faster deflation rate than in any previous deflation event. The deflation stopped at about 03^h on March 17. The total deflation amounted to about 61 cm according to tiltmeters in the Krafla power house, using the average relation between tilt stage and land elevation. As fissures opened near the power station during this deflation, the relation between tilt and land elevation may have deviated

somewhat from the average relation, making both total deflation and maximum rate of deflation somewhat in doubt.

At about 16^h 20^m, one hour after the deflation started, a volcanic fissure opened about one km north of the center of the Krafla volcano and basaltic lava poured out. This rift propagated northwards during the next 30 minutes and the volcanic fissure reached about 4 km in length. The eruption lasted only some 6 hours during which time an estimated amount of 3 million cubic meters of lava was formed.

Inflation of the Krafla volcano started in the morning of March 17, and the rate of inflation was very high during the following days, or about 2 cm per day.

The subsidence event of March 16-17, 1980 differs from all previous subsidence events in the Krafla volcano as the maximum rate of subsidence was about three times higher than had been observed before. The highest rate of deflation previously observed was on September 8, 1977, about 16.5 cm per hour and about 15.5 cm per hour on April 27, 1977. In all other subsidence events the maximum rate of deflation was observed as 5 cm per hour or less (Tryggvason 1980).

Two subsidence events have occurred since March 1980 until the writing of this paper in December 1980. Both events were accompanied by outpouring of basaltic lava in much greater quantity than in previous eruptions during the present sequence of events in the Krafla area.

OBSERVED TILT

There were 4 recording tiltmeters in the Krafla area during the deflation event of March 16-17, 1980, in addition to one water tube tiltmeter where daily readings were made. Furthermore, 11 «dry tilt» stations were occupied about three months before the event and about two months after the event. These

stations can rarely be occupied during winters because of snow cover. All these stations, with possibly one exception, showed clear indication of tilt during the deflation event.

At five of the «dry tilt» stations, which are located within 7 km from the center of deflation, good correlation is found between their tilt, and that of the water tube tiltmeter in the Krafla power house. This correlation is used to estimate the tilt stage immediately before and after the event. At the remaining six «dry tilt» stations no or very poor correlation is found with the water tube tiltmeter, and the observed tilt between the last observation before the event and the first observation after the event is taken as the tilt during the event. Errors in the event tilt due to this assumption are considerable or roughly 10 μ -rad as judged from other observations at the same stations.

Fig. 2 shows the location and the observed tilt at all the tilt stations in the Krafla area during the deflation event of March 16-17, 1980. Tilt at stations less than 5 km away from the center of subsidence is primarily caused by the nearly circular subsidence bowl, while stations farther away are affected by deformation along the fissure swarm to the south of the center of subsidence.

GROUND SUBSIDENCE AND VOLUME OF THE SUBSIDENCE BOWL

Two methods are available to estimate the ground subsidence from observed tilt at the tilt stations. One method uses the relation between observed north component of tilt at the Krafla power station and land elevation determined by precise leveling, and the relation between tilt at different tilt stations. The other method uses a deformation model and seeks the best fit between observed tilt and theoretical tilt.

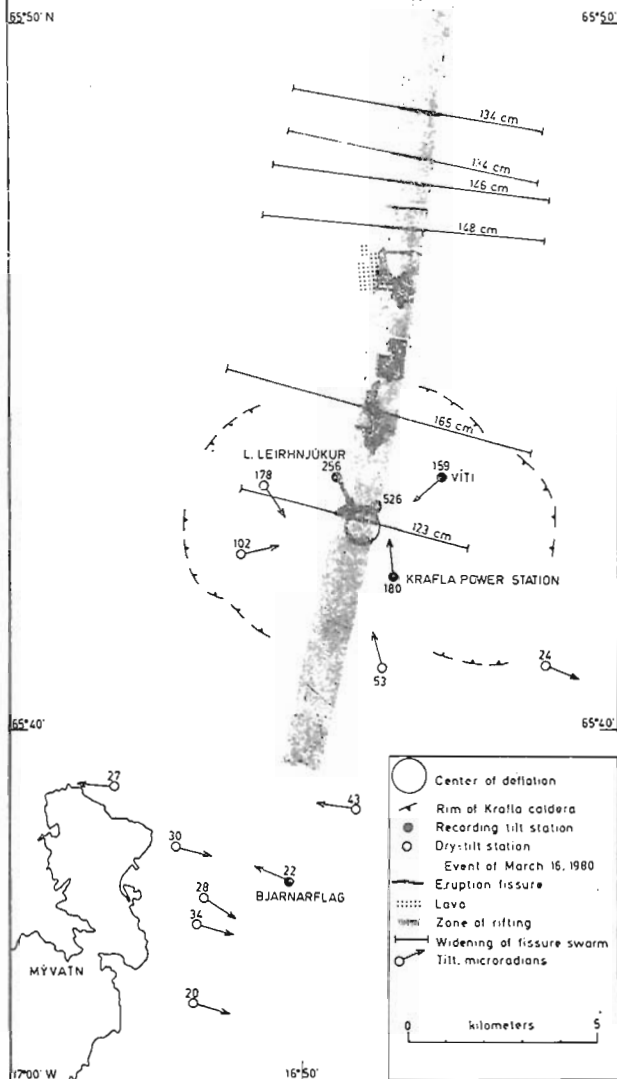


Figure 2 — Map of the Krafla area showing tilt stations with observed tilt during the subsidence event of March 16-17, 1980. Bars show displacements of one side of the rift zone relative to the other during this event. The center of deflation is shown with an open circle and the size of this circle indicates the uncertainty of location of this center. All geodimeter lines which crossed the rift zone (stippled) or extended into it in E-W direction increased in length by more than 20 cm during the event.

The four tilt stations, Krafla power station, Hlidardalur (53 on Fig. 2), Bjarghóll (102 on Fig. 2), and Hvannstód (178 on Fig. 2), which always show considerable tilt during inflation and deflation of Krafla, and are located at considerable distance from fissures formed during the March 16-17, 1980 deflation event, are used to estimate the maximum subsidence by the first method. The north component of tilt at the Krafla power station has been correlated with precise leveling, and one microradion of tilt corresponds to 3.4 mm vertical movement in the center of inflation/deflation (Björnsson et al. 1979). The tilt of $180 \mu\text{-rad}$ thus corresponds to 61 cm subsidence. The tilt at the other stations indicate maximum subsidence of 43 to 62 cm, and the average value of maximum subsidence thus calculated is 55 cm, with a standard error of about 5 cm.

A model of spherical body of increasing or decreasing pressure inside a homogeneous elastic halfspace (Mogi 1958) has been compared with the tilt observations and a reasonable conformance is found. The best correlation between observed tilt and the model is obtained if the depth to the center of the spherical body is 2.9 ± 0.1 km, the volume of the subsidence bowl 35 ± 2 million m^3 , and the maximum subsidence 65 ± 5 cm.

The discrepancy between the empirical value for maximum subsidence of 55 ± 5 cm, and the theoretical value of 65 ± 5 cm is not significant and the theoretical model seems not to deviate greatly from the actual condition. However, one obvious error in the model is the assumption of homogeneous elastic halfspace, while the geology of the Krafla area is rather chaotic, and increasing temperature and pressure with depth causes vertical gradient in the elastic properties. It is also very unlikely that the inflating and deflating body is spherical, and its diameter is probably large compared with its depth, but the theory accounts for a small body relative to its depth below the surface of the halfspace.

The conclusion of this discussion is that the subsidence bowl formed during the March 16-17 deflation of the Krafla volcano had a maximum depth of 60 ± 10 cm, and its volume

was probably between 30 and 40 million m^3 . The shape of the bowl can be roughly approximated by a theoretical model of decreasing pressure in a spherical body centered at 2.9 km dept. This is taken as indication of removal of material (magma) at approximately this depth. The volume of material removed is probably somewhat greater than the volume of the subsidence bowl as the decreased stress causes increase in volume. However, the stress change is not known, and the volume of material removed from below the area of subsidence can best be estimated as 30 to 40 million m^3 . Tilt in the Mývatn area, 7 to 13 km south of the center of deflation (Fig. 1) is mostly towards the Krafla fissure swarm, 20 to 40 microradians. This shows that the fissure swarm subsided to the south of the region where surface rifting was observed. One station, off the north coast of the lake Mývatn, shows tilt away from the fissure swarm. This may indicate uplift of the flanks of the fissure swarm in the region where surface rifting occurred, similar as observed during the subsidence events of April 1977 and January 1978 (Björnsson et al. 1979, Sigurdsson 1980).

HORIZONTAL DEFORMATION

Distance measurements in the Krafla area have been made many times since early 1977, and an effort has been made to cover the area with measurements during every inflation period. These measurements have shown that inside the Krafla caldera the distances between bench marks vary regularly with the stage of inflation, and each distance varies linearly with the observed tilt at the Krafla power station. This linear relation is broken only during deflation events with rifting across the caldera. This allows us to estimate the length of each measured line immediately before and after the March

16 to 17, 1980 deflation event, and to estimate further how much of the observed length difference is due to the deflation and how much is due to permanent deformation in form of rifting.

A total of 79 lines were measured with a geodimeter at times before and after the March event which allowed estimate of the length change during the event. Of these, 41 had increased in length by more than 20 cm. All these 41 lines cross the rift zone shown on Fig. 2, or extend into this zone. No line which lies wholly outside the rift zone had increased in length by more than 10 cm.

Where two or more lines crossed the rift zone at different angles, the total displacement of one side of the zone relative to the other can be determined. Six such determinations of displacements across the rift zone are shown on Fig. 2. One line, about one km south of the southernmost such determination, gave increase in length of 142 cm, but the exact direction of displacement was not obtained.

The direction of displacement is found to be nearly perpendicular to the rift zone. In the northern part of the area, this direction is 99 ± 3 degrees and in the southern part is about 104 degrees. Thus a slight change in the direction of the rift zone is reflected in the direction of displacements.

The north and south ends of the zone where rifting occurred during the March 1980 event were not determined with the geodimeter measurements, but open fissures were observed in the snow as far north and south as shown on Fig. 2.

Fig. 2 shows that the observed widening of the rift zone during the March 16-17 deflation event ranged between 123 and 165 cm. The average widening of the rift zone over the 13 km of length, where it was measured is about 140 cm. The length of the zone where surface rifting was seen is about 21 km, and it may be assumed that the widening was less towards the end of the rifted zone, than in the central part. If it is assumed that the 8 km length of the fissured zone, which lays outside the region of distance measurements was widened 70 cm on the average, the increase in area of the rift zone was 23800 m².

The increase in area of the Krafla fissure swarm during all rifting events since December 1975, is probably more than 300000 m^2 (Björnsson et al. 1979).

The volume of the subsidence bowl formed during the March 16-17 deflation event is estimated as $30\text{-}40 \times 10^6 \text{ m}^3$, and less than 10 per cent of that volume was erupted. Therefore, it is estimated that some 30 million cubic meters of magma were deposited in a dike of horizontal cross section similar to the areal increase of the rift zone of 23800 m^2 . This gives the average vertical extension of the dike about 1.3 km.

CONCLUSION

The subsidence event in the Krafla area, North Iceland, on March 16-17, 1980, is one of 13 similar events since December 1975 (Fig. 1) (Björnsson et al. 1979, Einarsson & Brandsdóttir 1980). Common features of all these subsidence events are

1. Rapid subsidence of the Krafla area, with center of subsidence about 1.5 km north or northwest of the Krafla geothermal power station.
2. Rifting and widening of a section of the Krafla fissure swarm.
3. Swarm of earthquakes with epicenters in the area of rifting.
4. Continuous tremor on local seismometers.

In six of the subsidence events, basaltic lava has reached the surface, but in only small amounts as compared with the volume of the subsidence bowl, until in July and October 1980, when the volume of lava was similar to the volume of the subsidence bowl.

SYMPOSIUM ON THE ACTIVITY OF OCEANIC VOLCANOES

It is assumed that the subsidence in the Krafla area is caused by removal of magma from beneath that area, and depositing it in fissure or fissures in the elastic crust below the zone which developed surface rifting (Björnsson et al. 1977). The areal increase of this zone is considered to reflect the horizontal cross sectional area of the dike or dikes formed and the average height of the dike is calculated as 1.3 km by equating the volume of the subsidence bowl and the volume of the dike. The depth to the top of the dike is zero along the eruption fissure, but elsewhere the depth of the dike can be roughly estimated as equal the width of the fissured zone (Tryggvason 1980), ou about one wm. Thus the bottom of the dike should be at somewhat less than 3 km depth.

ACKNOWLEDGEMENTS

The National Energy Authority operates the water tube tiltmeter at the Krafla power station, which is the most important single measurements made to monitor the tectonic events in the Krafla area.

REFERENCES

- BJÖRNSSON, A., SAEMUNDSSON, K., EINARSSON, P., TRYGGVASON, E., and GRÖNVOLD, K., 1977 : Current rifting episode in north Iceland, *Nature*, 266, 318-323.
- BJÖRNSSON, A., JOHNSEN, G., SIGURDSSON, S., THORBERGSSON, G., and TRYGGVASON, E., 1979 : Rifting of the plate boundary in north Iceland, 1975-1978, *J. Geophys. Res.*, 84, 3029-3038.
- EINARSSON, P., and BRANDSDÓTTIR, B., 1980 : Seismological evidence for lateral magma intrusion during the July 1978 deflation of the Krafla volcano in NE-Iceland, *J. Geophys.*, 47, 160-165.
- MOGI, K., 1958 : Relations between the eruptions of various volcanoes and the deformations of the ground surfaces around them, *Bulletin of the Earthquake Research Institute (Tokyo)*, 36, 99-134.
- SIGURDSSON, O., 1980 : Surface deformation of the Krafla fissure swarm in two rifting events, *J. Geophys.*, 47, 154-159.
- TRYGGVASON, E., 1980 : Subsidence events in the Krafla area, north Iceland, 1975-1979. *J. Geophys.*, 47, 141-153.

SEISMIC ACTIVITY AND RIFTING IN THE KRAFLA FAULT SWARM IN NE-ICELAND

by

BRYNDÍS BRANDSDÓTTIR

and

PÁLL EINARSSON

Science Institute, University of Iceland,
Dunhaga 3, Reykjavík

ABSTRACT

The Krafla volcano in the rift zone of NE-Iceland has been going through a series of inflation-deflation cycles since 1975. Magma accumulates beneath the volcano during slow inflation periods and is injected laterally into the Krafla fault swarm during deflation events. Each deflation event has a characteristic pattern of seismic activity. It typically begins with continuous volcanic tremor and the tremor amplitude is dependent on the rate of deflation. Earthquake activity increases shortly after the deflation starts and the epicentral area is soon extended from the caldera region, along the fault swarm to the north, the south or both. The propagation speed of the seismic activity is highest in the beginning, but decreases with decreasing deflation rate and increasing length of the epicen-

tral zone. Typical speed is 0.5 m/s, but may reach values as high as 1.2 m/s.

Although the hypocentral zones of the different deflation events often overlap, the largest earthquakes are located within a well defined, but each time different section of the fault swarm. The earthquakes occur in the uppermost 10 km of the crust, but the depth range is different for different deflation events.

The earthquake activity culminates after the maximum of deflation rate and tremor amplitude is reached, with earthquakes typically reaching magnitude 4. Extensive fault movements and fissuring usually occur in the area of maximum earthquake activity.

The seismological data strongly support a model where the present events are assumed to be the result of interaction between magma pressure under the Krafla volcano and rifting of the plate boundary. The rifting is triggered by increasing magma pressure in the reservoir and a fluid filled extensional crack propagates laterally along the Krafla fault swarm. The driving force of this process is the tectonic stress at the plate boundary, but the mode of strain release is modified by the presence of fluid.

WHY DO LONG RIFT ZONES DEVELOP BETTER IN HAWAIIAN VOLCANOES — A POSSIBLE ROLE OF THICK OCEANIC SEDIMENTS

by

KAZUAKI NAKAMURA

Earthquake Research Institute, University of Tokyo
Bunkyo-ku, Tokyo 113. Japan

ABSTRACT

Rift zones are one of the characteristic features of Hawaiian volcanoes. They are long narrow zones of flank fissure eruptions but are distinct from ordinary flank eruption sites on polygenetic volcanoes in that eruptions, and therefore dike intrusions, occur repeatedly at the same general place for a long time and thus cause a considerable distance of horizontal spreading. This spreading should somehow be accommodated and the direction of the minimum compressive stress should remain the same after accommodation in order for a new dike to intrude in the same orientation.

The Krafla spreading events in Iceland between North American and European plates revealed that the process of the lithospheric spreading is similar to that observed for Hawaiian volcanic activities, including rift zone eruptions.

Accreting plate boundaries may be understood as consisting of chain of linear rift zones and their source polygenetic centers where the magma supplied from the asthenosphere is temporarily stored.

Horizontal spreading caused by repeated dike intrusions has been accommodated in the case of the accreting plate boundaries by the lateral separation of lithosphere over asthenosphere. In the case of Hawaii sliding of the volcanic edifice over deep sea sediments may be the analogous mechanism such as appears to have occurred during the 1975 Kalapana earthquake which was anticipated by SWANSON et al. (1976) as one of the repeated events as the east rift zone has continuously dilated. Lack of rift zones in otherwise similar Galapagos shields which sit over the young ocean floor with higher relief is consistent with this view.

HAWAIIAN RIFT ZONES

Hawaiian rift zones are the channelways to bring the lava to the flanks of shield volcanoes. According to MACDONALD (1972), typically, there are three rift zones radiating from the summit with angles of about 120° between them. Usually, however, two of the rift zones are much more prominent than the third (Fig. 1). The major rift zones range from one to four kilometers in width and contain hundreds of fissures that served as eruptive vents. These are marked at the surface by open cracks, spatter ramparts, rows of spatter cones, collapse craters, and shields of Icelandic type (i.e. monogenetic). At depth, where they have been exposed by cutting of canyons into the older shields, the rift zones are seen to consist of innumerable parallel thin dikes.

Some rift zones extend more than a hundred kilometers from the summit. FURUMOTO (1978) estimated the depth of

the parallel dikes of the rift zone as deep as 5 to 7 kilometers, who constructed a structural model of Kilauea volcano on the basis of seismic velocity structure (HILL, 1969), Bouguer gravity anomaly (KINOSHITA et al., 1963) and some other geophysical observations (FURUMOTO and KOVACH, 1979; BROYLE et al., 1979).

Rift zones are one of the major characteristics of Hawaiian shields. WILLIAMS and MCBIRNEY (1979) classified polygenetic shield volcanoes into Hawaiian and Galapagos types. Rift zones are not observed in smaller Galapagos shields, otherwise similar oceanic, basaltic shields volcanoes, nor in stratovolcanoes (composite volcanoes), the commonest type of polygenetic volcanoes.

This short paper is an abbreviated and modified version of my previous paper in Japanese (NAKAMURA, 1980).

PROBLEMS

The aggregate thickness of the parallel dikes composing a rift zone frequently attain the width of the order of a kilometer or more (MACDONALD, 1956). The room occupied by the width of the dikes should somehow have been accommodated without much mechanical difficulty. What is the mechanism responsible for such an accommodation under an intraplate situation?

The same problem may be addressed in terms of the stress field near the rift zone. Dikes tend to develop in the plane perpendicular to σ_3 ($\sigma_3 < \sigma_2 < \sigma_1$), or parallel to σ_1 and σ_2 . On the other hand, the magnitude of stress in the direction of the previous σ_3 should increase due to the shortening caused by dike intrusions. Then there should be a certain mechanism to maintain the same direction of σ_3 after repeated

dike intrusions, so that a new dike can intrude in the same orientation. What is such a mechanism for maintaining the same direction of σ_3 ?

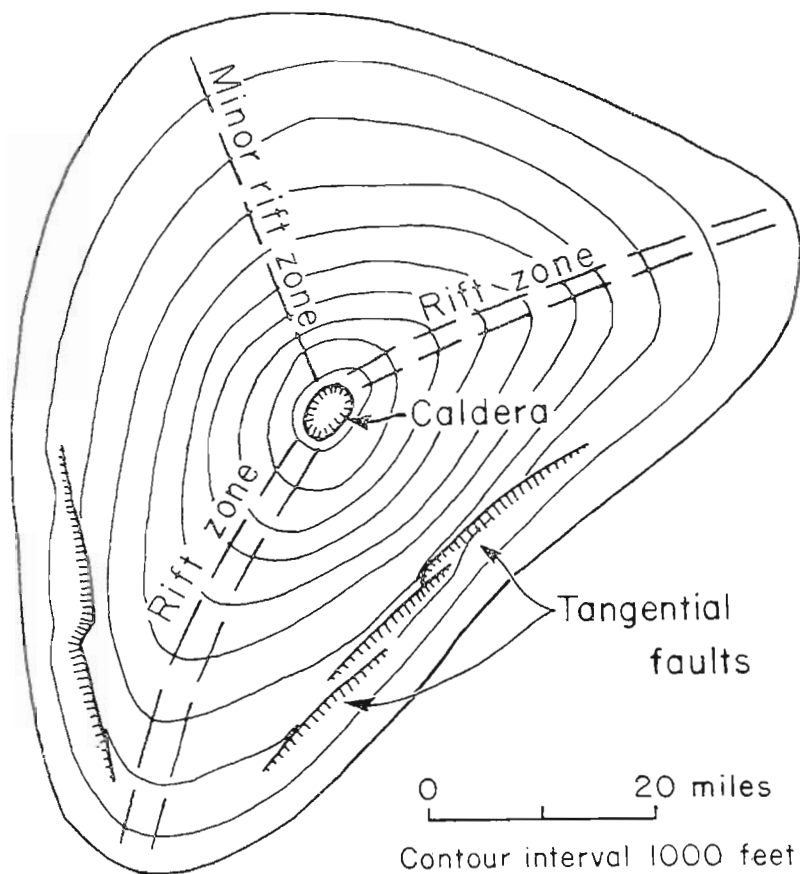


Fig. 1 — Plan of a typical Hawaiian shield volcano showing the radiating rift zones, caldera and tangential faults. MACDONALD, 1956.

PREVIOUS WORKS

MACDONALD (1956) attributed the radial pattern of the rift zones to a magmatic force thrusting up the central portion of the volcanic edifices. He explained the tangential faults (Fig. 1) by the same mechanism. Later studies revealed, however, that the volcanic edifices have subsided (HILL, 1969; ZABLOCKI et al., 1974) and is subsiding (MOORE, 1970), when the rift zone eruptions are vigorously going on. It may be concluded that the radial pattern certainly indicates the point source mechanism which works principally as horizontally expanding force (ODÉ, 1957; NAKAMURA, 1977) rather than an upthrusting one. The horizontally expanding source may be able to explain the flank (fissure) eruptions but it does not explain the reason why the extensive diking repeated many times in a few particular directions is possible.

FISKE and JACKSON (1972) demonstrated the role of gravitational force in determining the orientation of the rift zones. They classified Hawaiian shields into two categories, the one grew over the relatively flat ocean floor (isolated type) and the other grew on the flanks of the adjacent shield (buttressed type) (Fig. 2). In the case of the buttressed shields, they argue that radial dikes propagate in the direction perpendicular to the maximum inclination of the underlying slopes. In the case of isolated shields, FISKE and JACKSON (1972) maintain that the strike of repeated dike intrusions follow the crest of the initially formed ridges.

Because FISKE and JACKSON (1972) are concerned mainly with the plan view configuration of the rift zones, they did not explain how the aggregate thickness of the dikes is accommodated and the stress field remains the same in orientation of the principal axes after repeated intrusions.

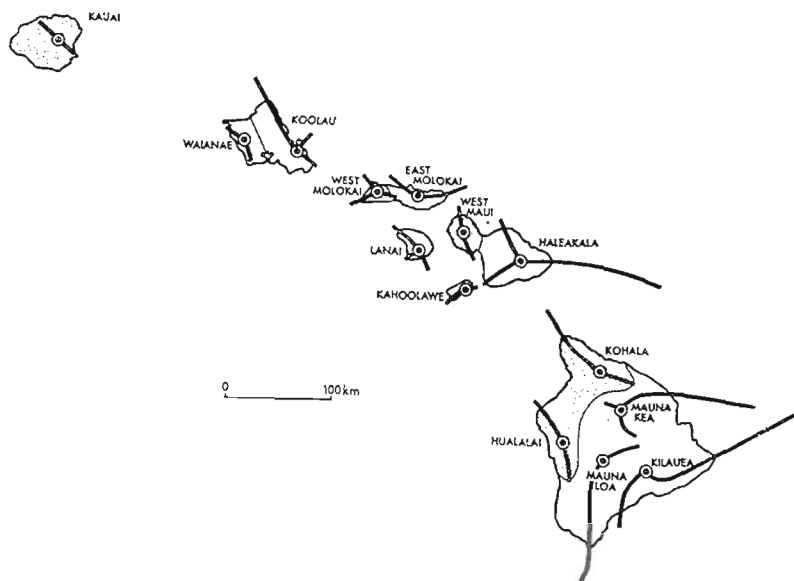


Fig. 2 — Map showing the southeastern part of the Hawaiian Archipelago. The 6 volcanoes shown in the stippled pattern grew as isolated edifices. The 7 unstippled Volcanoes grew later and were buttressed by the earlier formed edifices. FISKE and JACKSON, 1972.

WYSS (1980) regarded the plan view pattern of Hawaiian rift zones as essentially controlled by a point source of force, i.e. a hot spot, and thus radiating from individual volcanoes in the same three directions with an angle of about 120° . He also argues against the interpretation by FISKE and JACKSON (1972) citing the north-south direction of the elongation of Loihi seamount located 30 km south of the island of Hawaii. Loihi should elongated, according to WYSS (1980) in a northeast-southwest direction parallel to the regional contour lines, if FISKE and JACKSON's interpretation holds. However, the length of the long axis of Loihi seamount (ca 20 km) appears to be too short to justify the argument of WYSS.

It may be rather difficult to view the rose diagram of Hawaiian rift zones (Fig. 3) as radiating in three principal directions. Moreover, his argument fails to explain the lack of similar rift zones in other volcanoes of the hot spot origin, like Galapagos.

Briefly, it seems to me that the problem raised in this paper has not been considered seriously.

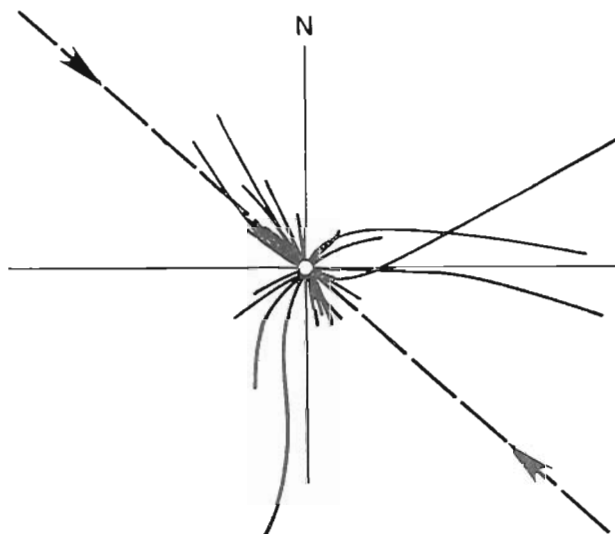


Fig. 3 — Rose diagram showing 33 Hawaiian rift zones radiating from a single imaginary center. From FISKE and JACKSON, 1972.

The dashed line is the strike of the Hawaiian island chain as drawn by WYSS, 1980.

IMPLICATIONS OF RECENT KRAFLA EVENT IN ICELAND

Currently occurring event at Krafla in Iceland showed us detailed processes of spreading and accretion of the lithospheric plate (BJÖRNSSON et al., 1979). The process is essentially the same with the rift zone activity of Hawaiian volcanoes,

that is, the magma supplied nearly constantly (TRYGGVASON, 1980) from the deep source is temporarily stored in a shallow magma reservoir beneath the center of the Krafla caldera (EINARSSON, 1978), one of the polygenetic volcanic centers in the neovolcanic zone in Iceland, the accreting plate boundary (SAEMUNDSSON, 1978). The storage causes inflation of the caldera region (TRYGGVASON, 1980). Intermittent deflation events are associated with extensive lateral diking from the reservoir into the axial zone of the plate boundary (BRANDSDOTTIR and EINARSSON, 1979; EINARSSON and BRANDSDOTTIR, 1980; TRYGGVASON, 1980; LARSEN et al., 1979), resulting in the spreading and accretion of the European and North American plates.

The Krafla event implies that the axis of the accreting plate boundary, in general, consists of a series of polygenetic centers with shallow magma reservoir and linearly radiating rift zones (SAEMUNDSSON, 1978; BJÖRNSSON et al., 1979). Here the rooms for dike intrusions have been accommodated by the separation of the lithosphere «sliding» over the asthenosphere. Therefore, if there is an analogous mechanism in Hawaii, it will be an answer to our problem.

THE 1975 KALAPANA EARTHQUAKE

The 1975 Kalapana earthquake (M_s : 7.1) occurred at the southeast coast of the island of Hawaii and was studied by ANDO (1979) and FURUMOTO and KOVACH (1979). The fault parameters given by these authors are generally similar. In Fig. 4, the two possible fault planes by ANDO (1979) are given. According to these authors, a significant portion of the Kilauea's south flank slid south-southeastward by several meters over a nearly horizontal fault plane, with the depth 7 to 10 km. This depth coincide with the upper surface of the oceanic

crust underlying the volcanic edifice (HILL, 1969). Therefore, the sliding which caused the Kalapana earthquake may have occurred in the oceanic sediment layer beneath the volcanic edifice (FURUMOTO and KOVACH, 1979; ANDO, 1979).

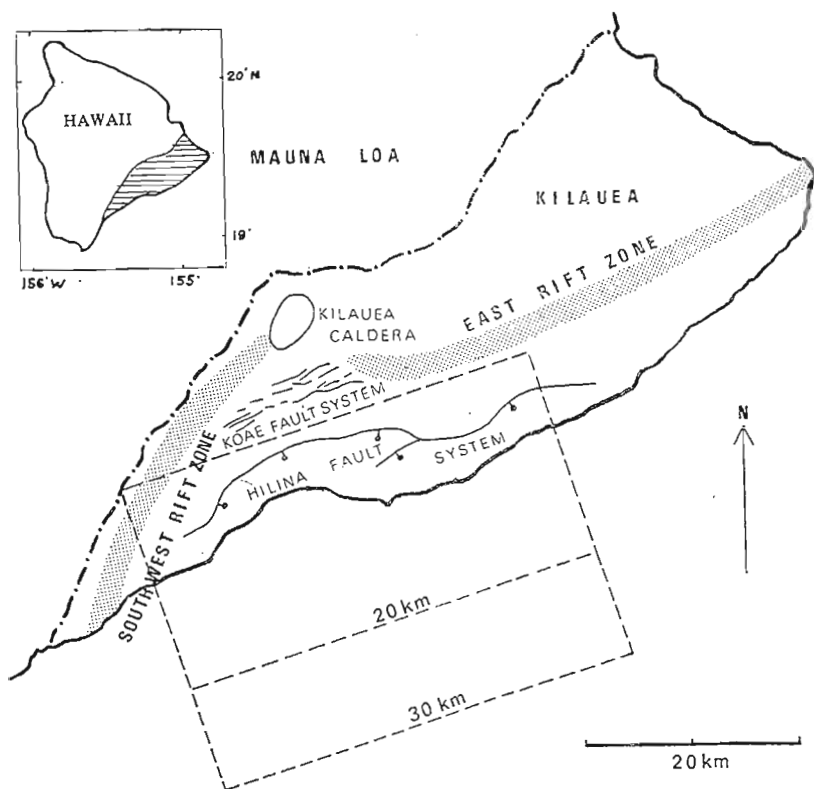


Fig. 4—Simplified map showing the location of east and southwest rift zones, Koae and Hilina fault systems of Kilauea volcano and the nearly horizontal two possible 1975 fault planes (squares). After ANDO (1979).

Moreover, the 1975 Kalapana earthquake had been anticipated by SWANSON et al. (1976) as one of the repeated steps to accommodate the south-southeast contraction of the Hilina

fault zone, which, in turn, had been associated with the south-southeast dilation of the east rift zone and the Koa'e fault system.

After the Kalapana earthquake, the activity of Kilauea volcano appears to be characterized by more frequent intrusive events without surface eruptions, similar to the state of the volcano in the earlier half of this century (SWANSON et al., 1976) which followed the previous Kalapana type earthquake in 1868 (ANDO, 1979). These observations also support the interpretation by SWANSON et al. (1976) and ANDO (1979) that the 1975 Kalapana earthquake was one step of repeated process to accommodate the long continued rift zone dilation.

Thus, the case which has enabled to accommodate the room for repeated dike intrusion into rift zones may be found in the existence of thick enough oceanic sediment beneath the volcanic edifice (NAKAMURA, 1980). The sudden loading of volcanic material over the sediment may produce anomalously high pore pressure which reduces the frictional resistance for the overlying body to slide easily.

DISCUSSION

The thickness of the oceanic sediments beneath the Hawaiian volcanoes may be estimated from the age of the crust at the time of the initiation of the volcanism at the site and from the average rate of accumulation of sediments. Age of the crust in the above sense may be 80 Ma or more in Hawaii (PITMAN et al., 1974). Average rate of accumulation over the crust of present Hawaiian region is estimated at several to 10 mm/ka (LISITZIN, 1972). Then the thickness of sediment layers may be calculated as several hundred meters. Observed thicknesses of sediments are 580 m at DSDP Site 313, 1600 km west of Hawaii (THE SHIPBOARD SCIENTIFIC PARTY, 1975) and 280 m at Site 163, 1200 km southeast of Hawaii (VAN ANDEL et al., 1976).

In the equatorial Galapagos region, accumulation rate attains 35-50 m/Ma (LONSDALE and KLITGORD, 1978). Age of the oceanic crust underlying Galapagos volcanoes is estimated to range one to eight Ma (HEY, 1977). Age of the volcanoes is a few Ma or younger (COX and DALRYMPLE, 1966). Therefore, the period of deep sea sedimentation before the initiation of the volcanic activity may be a few to 5 Ma at most. Thus, the sediments may be well below 250 m in thickness in Galapagos region, less than one half of that of Hawaii.

In addition to the thinner sediment layer in Galapagos, basement topography appears to be unfavorable in Galapagos to form an extensive sliding plane in the sediments. When observed near the triple point between Pacific, Cocos and Nazca plates, the portion of the plates that was formed at the Nazca-Cocos (Galapagos) spreading center is much richer in topographic relief than the Pacific plate (HEY et al., 1977). Northwest of the Galapagos islands, the ocean floor is cut by closely spaced faults with the scarp 10 to 280 m in height averaging 50 m (KLITGORD and MUDIE, 1974).

The comparison of sediment thickness and basement topography also support the present hypothesis that the existence of thick enough sediment layer plays an important role in the formation of long rift zones of Hawaiian volcanoes. If this hypothesis holds in general, it may be expected that rift zones are expected to develop better in Oceanic volcanoes formed on older crust than those formed on younger crust near the spreading centers

CONCLUSION

Hawaiian rift zones may have developed under the gravitationally controlled stress field as FISK and JACKSON indicated. What have kept the gravitational stress field function for a long time during the repeated dike intrusions into the rift zones, may well be the existence of thick enough oceanic sediment beneath the volcanic edifice. Rapid accumulation to form volcanic edifices may have formed anomalously high pore pressure, and thus reducing the frictional resistance for the edifice to slide. Smoother basement topography beneath the sediments may also be a factor to form an extensive sliding plane in the sediments.

ACKNOWLEDGEMENTS

Critical comments to NAKAMURA (1980) by Masanori Sakuyama and Shun'ichiro Karato were helpful and are deeply acknowledged. Interest of Donald A. Swanson and Hans-Ulrich Schmincke to the present study was encouraging. The work was mostly carried out during my 1980 stay in the University of Lille 1 in France (host: Jacques Charvet) and in the Ruhr University of Bochum in Germany (host: Hans-Ulrich Schmincke) supported by the funds of the French Ministry of the Universities and the German Research Association.

REFERENCES

- ANDO, M., 1979 : The Hawaii earthquake of November 29, 1975 : low dip angle faulting due to forceful injection of magma. *J. Geophys. Res.*, 84, p. 7616-7626.
- BJÖRNSSON, A., JOHNSEN, G., SIGURDSSON, S., THORVERGSSON, G., TRYGGVASON, E., 1979 : Rifting of the plate boundary in north Iceland. *J. Geophys. Res.*, 84, p. 3029-3038.
- BRANDSDOTTIR, B. and EINARSSON, P., 1979 : Seismic activity associated with the September 1977 deflation of the Krafla central volcano in NE Iceland. *J. Volcanol. Geotherm. Res.*, 6, p. 197-212.
- BROYLES, M. L., SUYENAGA, W. and FURUMOTO, A. S., 1979 : Structure of the lower east rift zone of Kilauea volcano, Hawaii, from seismic and gravity data. *J. Volcanol. Geotherm. Res.*, 5, p. 317-336.
- COX, A. and DALRYMPLE, G. B., 1966 : Paleomagnetism and potassium-argon ages of some volcanic rocks from the Galapagos Islands. *Nature*, 209, p. 776-777.
- EINARSSON, P., 1978 : S-wave shadows in the Krafla caldera in NE-Iceland, evidence for a magma chamber in the crust. *Bull. Volcanol.*, 41, p. 1-9.
- and BRANDSDOTTIR, B., 1980 : Seismological evidence for lateral magma intrusion during the July 1978 deflation of the Krafla volcano in NE-Iceland. *Jour. Geophys.*, 47, p. 160-165.
- FISKE, R. S. and JACKSON, E. D., 1972 : Orientation and growth of Hawaiian volcanic rifts: the effect of regional structure and gravitational stresses. *Proc. R. Soc. Lond., Ser. A*, 329, p. 299-326.
- FURUMOTO, A. S., 1978 : Nature of the magma conduit under the east rift zone of Kilauea Volcano, Hawaii. *Bull. Volcanol.*, 41, p. 435-453.
- FURUMOTO, A. S. and KOVACH, R. L., 1979 : The Kalapana earthquake of November 29, 1975 : an intraplate earthquake and its relation to geothermal processes. *Phys. Earth Planet. Int.*, 18, p. 197-208.

- HEY, R., 1977 : Tectonic evolution of the Cocos-Nazca spreading center. Geol. Soc. Amer. Bull., 88, p. 1404-1420.
- , JOHNSON, G. L. and LOWRIE, A., 1977 : Recent plate motions in the Galapagos area. Geol. Soc. Amer. Bull., 88, p. 1385-1403.
- HILL, D. P., 1969 : Crustal structure of the island of Hawaii from seismic refraction measurements. Bull. Seismol. Soc. Amer., 59, p. 101-130.
- KINOSHITA, W. T., KRIVOY, H. L., MABEY, D. R. and MACDONALD, G. R., 1963 : Gravity survey of the island of Hawaii. U.S. Geol. Surv. Prof. Pap., 475-C, p. C114-C116.
- KLITGORD, K. D. and MUDIE, J. D., 1974 : The Galapagos spreading center : A near-bottom geophysical survey. Roy. Astro. Soc., Geophys. Jour., 38, p. 563-586.
- LARSEN, G., GRÖNVOLD, K. and THORARINSSON, S., 1979 : Volcanic eruption through a geothermal borehole at Namafiall, Iceland. Nature, 278, p. 707-710.
- LISITZIN, A. P., 1972 : Sedimentation in the world ocean. Soc. Econ. Paleont. Mineral., Spec. Publ., 17, 218 pp., Tulsa.
- LONSDALE, P. and KLITGORD, K. D., 1978 : Structure and tectonic history of the eastern Panama Basin. Geol. Soc. Amer. Bull., 89, p. 981-999.
- MACDONALD, G. A., 1956 : The structure of Hawaiian volcanoes. Verh. Koninkl. Nederl. Geol. Mijnb. Gen., 16, p. 274-295.
- , 1972 : Volcanoes, 510 pp., Prentice-Hall, New Jersey.
- MOORE, J. G., 1970 : Relationship between subsidence and volcanic load, Hawaii. Bull. Volcanol., 34, p. 562-576.
- NAKAMURA, K., 1977 : Volcanoes as possible indicators of tectonic stress orientation — principle and proposal. J. Volcanol. Geotherm. Res., 2, p. 1-16.
- , 1980 : Why do long rift zones develop in Hawaiian volcanoes — a possible role of thick oceanic sediments. Bull. Volcanol. Soc. Japan., 25, p. 255-269.

SYMPOSIUM ON THE ACTIVITY OF OCEANIC VOLCANOES

- ODÉ, H., 1957 : Mechanical analysis of the dike pattern of the Spanish Peaks area, Colorado. Bull. Geol. Soc. Amer., 68, p. 567-576.
- PITMAN, W. C., III, LARSON, R. L. and HERRON, E. M., 1974 : Isochron map and age map of ocean basins. Geol. Soc. Amer., Boulder.
- SAEMUNDSSON, K., 1978 : Frissure swarms and central volcanoes of the neovolcanic zones of Iceland. In Crustal evolution in north-western Britain and adjacent regions, p. 415-432.
- SWANSON, D. A., DUFFIELD, W. A. and FISKE, R. S., 1976 : Displacement of the South flank of Kilauea volcano : The result of forceful intrusion of magma into rift zones. U.S. Geol. Surv. Prof. Pap., 963, 39 pp.
- THE SHIPBOARD SCIENTIFIC PARTY, 1975 : Site 313 : Mid Pacific mountains. Initial Rept. of the DSDP, 32, p. 313-390.
- TRYGGVASON, E., 1980 : Subsidence events in the Krafla area, north Iceland, 1975-1979. J. Geophys., 47, p. 141-153.
- VAN ANDEL, T. H., HEATH, G. R. and MOORE, T. C. JR., 1976 : Cenozoic history of the central equatorial Pacific : A synthesis baser on Deep Sea Drilling Project Data. A.G.U. Geophys. Monogr., 19, p. 281-295.
- WILLIAMS, H. and MCBIRNEY, A. R., 1979 : Volcanology. Freeman and Cooper, San Francisco, 397 pp.
- WYSS, M., 1980 : Hawaiian rifts and recent Icelandic volcanism : expression of plume generated radial stress fields. J. Geophys., 47, p. 19-22.
- ZABLOCKI, C. L., TILLING, R. I., PETERSON, D. W., KELLER, G. W. and MURRAY, J. C., 1974 : A deep research drill hole at the summit of active volcano, Kilauea, Hawaii, Geophys. Res. Lett., 1, p. 323-326.

SUBMARINE PYROCLASTIC ROCKS OF THE LA PALMA «OPHIOLITE» COMPLEX

by
S. STAUDIGEL
and
H.-U. SCHMINCKE
Institut für Mineralogie
Ruhr-Universität Bochum

ABSTRACT

The basement of La Palma (Canary Islands) consists of plutonic and hypabyssal intrusives and a submarine series, 2.5 km thick. Pyroclastic rocks increase in thickness upwards in the section and dominate at the top where they are several 100 meters thick. The clastic rocks in the pillow-dominated section are well bedded to cross-bedded hyaloclastites forming layers generally < 2 m thick, and thicker beds (up to 5 m) of coarse breccias which range from incipiently fractured pillows, that came to rest nearly in situ, to coarse, well sorted breccias deposited some distance from their source. Clastic rocks in the upper section are massive to poorly bedded, dominantly lapilli- to sand-sized whit clasts, being generally highly vesicular and larger fragments being reddish oxidized displaying shapes and vesicularity intermediate between pillows and sub-

aerial scoria. We think these clastics formed close to and, perhaps, just above sea level when the seamount precursor to La Palma island was emerging. Low overburden permitted explosive disruption of magma resulting in much higher portion of clastic basalt. Instability of thick clastic masses resulted in periodic slumping and downslope mass transport, possibly feeding widespread submarine pyroclastic debris flows, such as those described by Schmincke & v. Rad (1979) from south of Gran Canaria. The presence of a thick clastic section near the top may also have influenced the pattern of hydrothermal circulation systems: greenschist facies minerals such as epidote have not been found above the lower few meters of the thick hyaloclastite in the upper section, possibly because convecting fluids were dammed at the base of the hyaloclastites that had become impermeable due to wholesale low temperature alteration to smectite and other phases.

A comparison will be made between the La Palma complex and constructional processes in oceanic crust found at Mid Oceanic Ridges.

(This paper was not read at the Symposium)

LE CAPELINHOS (FAIAL, AÇORES)
VINGT ANS APRES SON ERUPTION :
LE MODELE ERUPTIF « SURTSEYEN »
ET LES ANNEAUX
DE TUF HYALOCLASTIQUES *

par

GUY CAMUS

Département de Géologie et Minéralogie
Université de Clermont-Ferrand II
et L.A. 10. C.N.R.S.

I/ LE CAPELINHOS

En août 1978, vingt ans après son éruption le Capelinhos était amputé par l'érosion marine et éolienne de plus de la moitié de sa surface initiale. Les coupes observables dans les falaises permettent de corréler les phases d'activité observées lors de l'éruption¹ et les faciès des produits ainsi que leur disposition.

27/9-5/11/57: Activité sous-marine à gerbes cypressoïdes et déferlantes basales, centrée L'île I édiflée pendant cette période s'est effondrée progressivement à partir du 24/10

* CAMUS G., BOIVIN P., GOËR de HERVE A. de, GOURGAUD A., KIEFFER G., MERGOIL J., VINCENT P. M. : Note acceptée pour publication au Bull. Volc. Int., Vol. 44, 1.

¹ MACHADO, F. et al. (1959 et 1962) : Mem. Serv. Géol. Portugal, n° 4 et 9.

1000 m à l'Ouest de l'île de Faial.

6/11/57 - 25/10/58 : Activité centrée 500 m à l'Est de l'île I.

6/11/57 - 12/5/58 : Activité comparable à la précédente à gerbes cypressoides et déferlantes basales jusqu'au 17/12, puis entrecoupée de quelques épisodes à laves incandescentes à partir du 18/12.

Rythmicité de l'éruption.

Présence d'eau en abondance au niveau de l'évent.

Effondrements intracrateriques.

Déferlantes basales peu chargées.

13/5-25/10/58 : Après la fermeture du cratère de l'anneau de tufs ; activité à laves incandescentes.

Fontaines de lave.

Lacs de lave temporaires.

Coulé du 23/8.

et a disparu le 1/11. Un îlot éphémère est apparu le 5/11.

L'île II édifiée pendant cette période s'est rapidement reliée à l'île de Faial. C'est elle qui subsiste.

Edification d'un anneau de tufs de 1 km de diamètre, 150 m de haut, avec un cratère de 450 m de diamètre, en communication avec la mer.

Stratification régulière.

Pulvérisation des produits hyaloclastiques, lapilli vitreux, bombes en chou-fleurs.

Absence locale de strates à pendage centroclinal.

Epannage de produits fins sur l'île de Faial.

Edification d'un cône de scories, coulées.

Spatter-cone à cratère central, cendres squelettiques allochtones accompagnées de quelques bombes.

Accumulations de lave massive dans l'atrio.

Visible au Sud-Est du volcan.

La base de l'anneau de tufs hyaloclastiques est palagonitisée. Or l'éruption s'est produite il y a seulement 20 ans. La palagonitisation a donc été rapide : elle pourrait être

syn-éruptive. A l'appui de cette hypothèse existent plusieurs arguments :

- vers le centre de l'appareil des panneaux de tufs de plusieurs mètres, voire plusieurs dizaines de mètres, décalés par failles, sont basculés. Si les matériaux n'avaient pas été consolidés par palagonitisation et zéolitisation, ils se seraient disloqués pendant le mouvement.
- les tufs hyaloclastiques sont des matériaux poreux qui permettent l'arrivée d'eau dans la cheminée même si le cratère est fermé (observation en cours d'éruption) : l'imperméabilisation des produits par palagonitisation et zéolitisation explique mieux la disparition de l'activité phréatomagmatique.

II/ LA COSTA DA NAU

Il s'agit de l'ancienne côte ouest de l'île de Faial, antérieure à l'éruption de 1957-1958. Elle montre la superposition de deux appareils de type Capelinhos.

L'appareil inférieur coupé presque diamétralement montre un anneau de tufs hyaloclastiques haut de 90 m à pendages périclinaux et centroclinaux. Des écroulements intracratériques en cours d'éruption sont incontestables. La phase à laves incandescentes a engendré un cône scoriacé qui remplit et débordé partiellement le cratère de l'anneau de tufs.

L'appareil supérieur, sans doute centré au niveau des anciens « îlots de Capelinhos », est coupé tangentiellement, il montre la même superposition tufs hyaloclastiques-scories.

CONCLUSIONS

Les anneaux de tufs hyaloclastiques caractérisent un « volcanisme d'émersion ». Ils abondent le long des côtes des Açores (Faial, Terceira, São Miguel, etc.). Parfois la phase terminale à laves incandescentes peut manquer. Les brèches basanitiques du Velay et du Cantal (France) sont des édifices comparables à des degrés d'érosion plus avancés, résultant d'un volcanisme sous-lacustre.

Il est nécessaire de bien différencier les éruption phréato-magmatiques sensu stricto qui engendrent les maars, des éruptions sub-aquatiques « surtseyennes » qui engendrent les anneaux de tufs.

La différence entre les deux types d'appareils a été soulignée par V. Lorenz².

<i>Maar</i>	<i>Tuff-ring</i>
Croissant pyroclastique surbaissé.	Anneau complet et proportionnellement plus élevé.
Pendages centroclinaux exceptionnels.	Pendages centroclinaux fréquents.
Vaste cratère.	Cratère réduit.
Abondance (50-70%) dans les projections d'éléments repris au substratum.	Moins de 10 % d'éléments énallogènes dans les produits.
Diatrème sous-jacent large et profondément enraciné.	Diatrème sous-jacent de taille réduite.

Tous ces caractères découlent d'un fait essentiel : la subsidence est discrète dans le diatrème sous-jacent à un anneau

² LORENZ, V. (1973) : Bull. Volc., Vol. 37, 2, p. 183-204.

de tufs alors qu'elle est très importante dans le diatrème sous-jacent à un maar.

Cette différence est due à la différence de quantité d'eau disponible dans un cas ou dans l'autre :

- dans le cas d'une éruption surtseyenne le rapport eau/magma est élevé et la vapeur faiblement surchauffée se condense rapidement. Les mélanges fluidifiés — moteur de l'éruption — ont un rendement énergétique médiocre.
- dans le cas d'une éruption phréatomagmatique continentale le rapport eau/magma est bas et la vapeur est d'autant plus surchauffée que le foyer des explosions s'approfondit rapidement avec la subsidence générale. Les mélanges fluidifiés vapeur d'eau — particules solides ont un rendement énergétique élevé, favorisant de ce fait l'éjection et la dispersion des éléments lithiques, et en corollaire la subsidence générale.

THE HISTORIC ERUPTIONS OF LA PALMA ISLAND (CANARIES)

by

A. HERNANDEZ-PACHECO

and

M. C. VALLS

Dpt. Petrologia. Facultad Ciencias Geológicas
Universidad Complutense de Madrid

ABSTRACT

From the first contacts of the European navigators who first visited the Canary Islands, around the middle of the XIV century, 18 eruptions have taken place of which seven occurred on the island of La Palma.

A historic-bibliographic study as well as a detailed field work have enabled us to accurately determine the places, dates, duration and volcanological patterns as well as the petrology and geochemistry of this historic eruptive cycle.

In every instance, alkaline basaltic lavas were emitted: true basalts, basanitoids and/or ankaramites. A differentiation process always took place in the magma chamber and a sequence from amphibole to olivine-bearing lavas was erupted. These variations of the chemistry and mineralogy of the lavas were related to the different stages of the eruption and the

height over sea level, where the corresponding eruptive vents opened.

The two main structural trends of the historic subhistoric volcanism of La Palma are N 5° W and N 35° W. Both trends correspond to the predominant ones of the dike swarms of the Basal Complex of the island. The secondary trends, N 80° E and N 15° E, likewise coincide with the corresponding main directions of the dike swarms of the Basal complexes of La Gomera and Fuerteventura.

The duration of these historic eruptions was between 1 to 3 months and the area covered with lava and pyroclasts was 37 km², 5 % of the total surface of the island.

THE HISTORIC ERUPTIONS OF THE CANARIAN ARCHIPELAGO

From the first contacts of the European navigators who visited the Canary Islands around the middle of the XIVth century, 18 eruptions have taken place (Table 1).

Of the 18 eruptions, 7 occurred on the island of La Palma. A new historic-bibliographic study as well as a detailed field-work have enabled us to accurately determine the places, dates and duration of these eruptions as well as their volcanological, petrological and geochemical features. The detailed study of all these aspects will be published separately; here we will give only a brief summary.

Nearly all historic eruptions on the Canary Islands, were relatively moderate, restricted to local areas of limited extension and their duration ranged only from a few days to 3 months. Notwithstanding, some of the eruptions were of a bigger magnitude, specially the Timanfaya eruption of Lanzarote island in the XVIII century, that lasted six years practically without interruption, covered with lava and tephra

a third of the island and obliged most of its inhabitants to emigrate. In other instances, as in the eruption of Montaña Negra, Tenerife, in 1706, the volcanic outburst, that lasted 9 days only, completely destroyed the town of Garachico.

THE HISTORIC ERUPTIONS OF LA PALMA

The seven historic eruptions of La Palma (Table 2) have taken place in the southern half of the island, in the mountain spine known as Cumbre Vieja, from its beginning in the El Paso valley till the southern point of the island, in Fuenca-liente, near the sea (Fig. 1).

All the eruptions began after a more or less prolonged period of earth tremors, whose magnitude never exceeded No. 6 of the Richter scale. These tremors were always restricted to some zones of the island, increased their intensity and frequency on the days and hours preceeding the eruptions, restricting to the area where the eruption took place thereafter. The eruption s.s. starts with the opening of little fissures of the ground following directions prefixed by the main structural patterns of the island. From the first moments, these fissuration is accompanied by the emission of gases and little lava fountains from several points along the whole extension of the main fissure that can attain several kilometers in length. Within a short time during the first hours of the event, these multiple incipient volcanic vents remain restricted to a few ones, more and more active, where the construction of heaps of tephra grow little by little and coalesce to the typical volcanic cones with their corresponding craters.

When the fissure opens in a terrain with a considerable slope and in its direction, high pressure lava fountains, pyroclastic materials and gases are emitted from the higher volcanic vents, while from the lower vents only more or less degasified lava pours out with a much lower explosivity.



Fig. 1 — Location of the historic eruptions of La Palma. 1. — Tacande (1470-92). 2. — Tahuya (1585). 3. — Tegalate (1646). 4. — San Antonio (1677). 5. — El Charco (1712). 6. — San Juan (1949). 7. — Teneguía (1971).

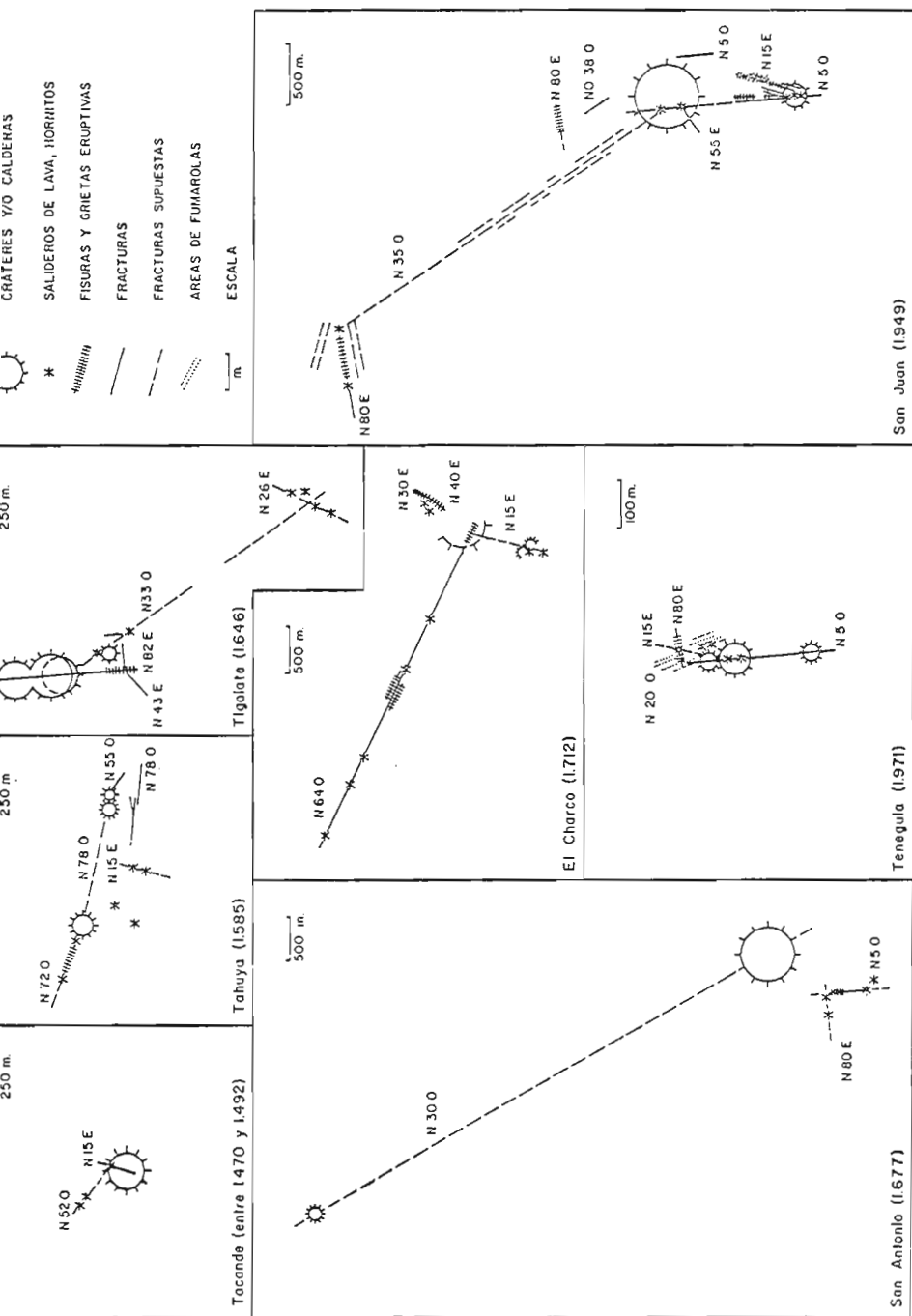


Fig. 2 — Structural patterns of the historic eruptions. Legend from top to bottom : craters and/or Calderas, Lava vents, hornitos, Eruptive fissures and fractures, Supposed fractures, Fumaroles, Scale.

This pattern is more noticeable when the difference in height of the extreme volcanic vents is greater, and the duration of the eruption is longer. In the most typical instances the higher vents grow to volcanic cones of a couple of hundreds of meters, while in the lower ones there only remain some eruptive fissures outpouring lava. This pattern is very clearly exposed in the eruptions of the volcanoes Tigalate, El Charco or San Juan, where the difference in height of the different vents is very noticeable.

During the course of the eruption it is also frequent that secondary cracks and fissures develop near the main volcanic vents, following also the main structural trends of the island. (Fig. 2 and 3).

The eruption continues with changes in the activity of the several vents, until suddenly and without a marked and gradual diminution of the explosivity and lava outpour, the activity practically stops. Subsequently, all the eruptive area suffers a period of slow degasification, becoming weaker and weaker, that lasts for 2 or 3 years. After this period, all volcanic manifestations cease, starting a new eruption, after an irregular period of time that normally lasts for several years, in other part of the same island or in other island of the Archipelago. (Table 1).

STRUCTURAL PATTERNS OF THE VOLCANIC ERUPTIONS

All these historic eruptions and in general all the recent eruptions of the island are conditioned in its geometric patterns by the directions that rule the whole of the Archipelago and La Palma in particular.

The main structural trends that rule the several eruptive areas (Fig. 2 and 3) are N 5° W and N 35° W. Both are the

SYMPOSIUM ON THE ACTIVITY OF OCEANIC VOLCANOES

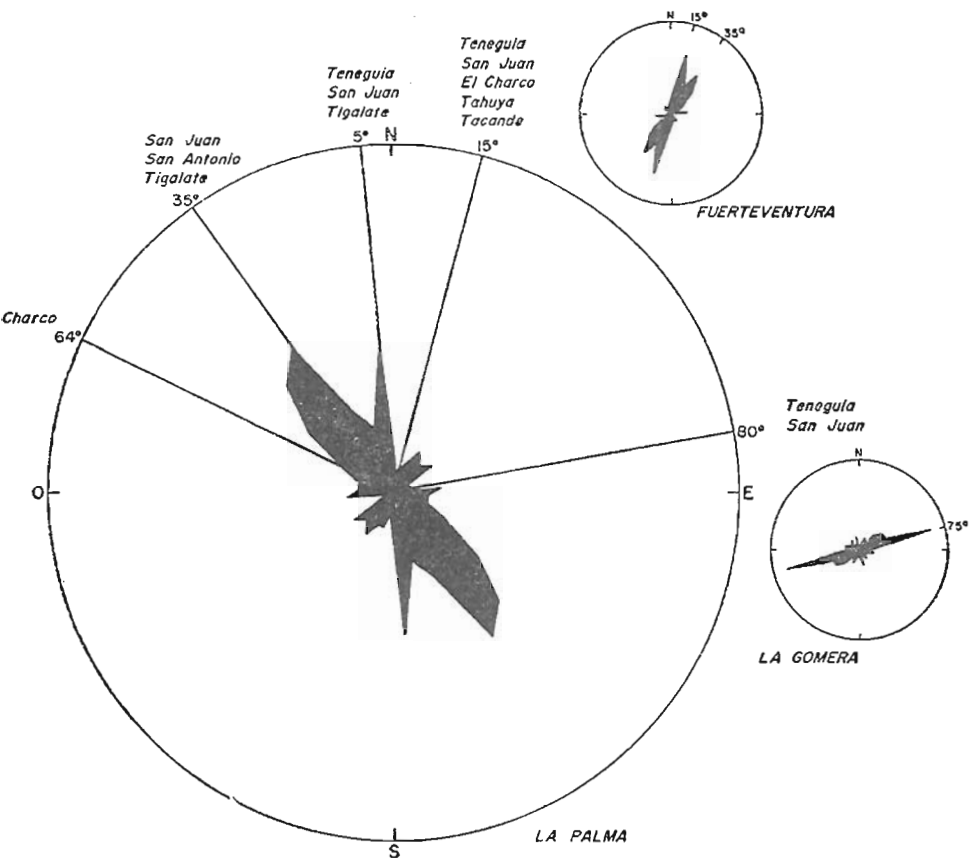


Fig. 3 — Main directions of historic eruptions of La Palma related to the dike-swarms of the Basal Complex of the Island. The orientation diagrams corresponding to the Basal Complexes of Fuerteventura and La Gomera Islands are also drawn. (Data of Basal Complex of La Palma from NUEZ, J. de la (pers. communication) and data of the basal complexes of Fuerteventura and La Gomera from HERNANDEZ-PACHECO, A., 1979).

predominant ones in the eruptions of Tigalate, San Antonio, San Juan and Teneguía volcanoes. Both coincide with the maxima for the trends of the dike swarms of the Basal Complex of the island that appears in the Caldera de Taburiente.

Other trend, N 15° E, of secondary magnitude, appears in the eruptions of Tacande, Tahuya, El Charco and also San Juan and Teneguía volcanoes. It coincides with a maximum of the dike swarms of the Basal Complex of Fuerteventura. Lastly, the approximate direction N 75-80° E, fundamental in San Juan eruptions but also existing in San Antonio, Tigalate and Teneguía volcanoes, coincides with the corresponding maximum of the dike swarms of the Basal Complex of La Gomera. Special consideration deserves the fracture originated by the fissural eruption of El Charco. Its direction follows a N 64° W trend that could be likened to the direction of the Atlantic fracture system that exists to the W of the island. It marks the alignment of the volcanic edifices of Gran Canaria, Tenerife and La Palma.

The fact that these directions are the same that control the volcanic phenomena is logical, but the important point is that they do not follow a simple, unique pattern NE-SW or NNE-SSW as it was previously said. The above mentioned directions predominate in the whole extent of the Archipelago and exist at least from the Lower Miocene, age of the magmatic episodes of the Basal Complexes until the present time. It can be said that, for the whole Archipelago in general and for La Palma in particular, the recent vulcanism is ruled by complex structural lineaments, of regional extent, that come from about 30 M.a. ago and are periodically reactivated in connection with volcanic eruptions dependent from one or other of these structural trends.

PETROLOGY AND GEOCHEMISTRY
OF HISTORIC LAVAS

The materials emitted by the historic volcanoes of La Palma belong to the group of alkaline basalts with porphyritic texture with pyroxene and/or olivine and/or plagioclase and/or amphibole phenocrysts in a dark cryptocrystalline to microcrystalline matrix.

Following the criteria of IBARROLA, E. (1970) and according to the abundance of CIPW normative feldspar (Feld.) and nepheline (Ne), only a few percentage of the samples are basalts ($> 45\%$ Feld, $< 10\%$ Ne). Most of them are basanitoids ($< 45\%$ Feld, $> 10\%$ Ne) and ankaramites ($< 45\%$ Feld, $< 10\%$ Ne), as it can be seen in Table 2.

Generally each volcano has two or more types of basalts depending on their existing minerals as phenocrysts. At Teneigua volcano, the two types of basalts, pyroxene amphibole basalts and pyroxene olivine basalts, correspond to two eruptive phases, delimited in time.

In the other volcanoes it is not always possible to establish this correspondence but in general terms, it can be said that the crystallization has taken place at three stages. In a first stage of magma consolidation, amphibole (kaersutite type), greenish pyroxene (ferrisaltite type), acidic plagioclase and haüyne were formed. These minerals were mainly preserved in the scorias and pyroclastic materials, while in compact lavas they appear corroded and with resorption phenomena. Haüyne is only found as inclusions in the very scarce and corroded plagioclase crystals. The main crystallization stage follows and olivine, augite and plagioclase are then formed. Lastly, there is a final consolidation stage, only observed in rather few samples and formed by interstitial analcime, alkaline feldspar and biotite or by glass.

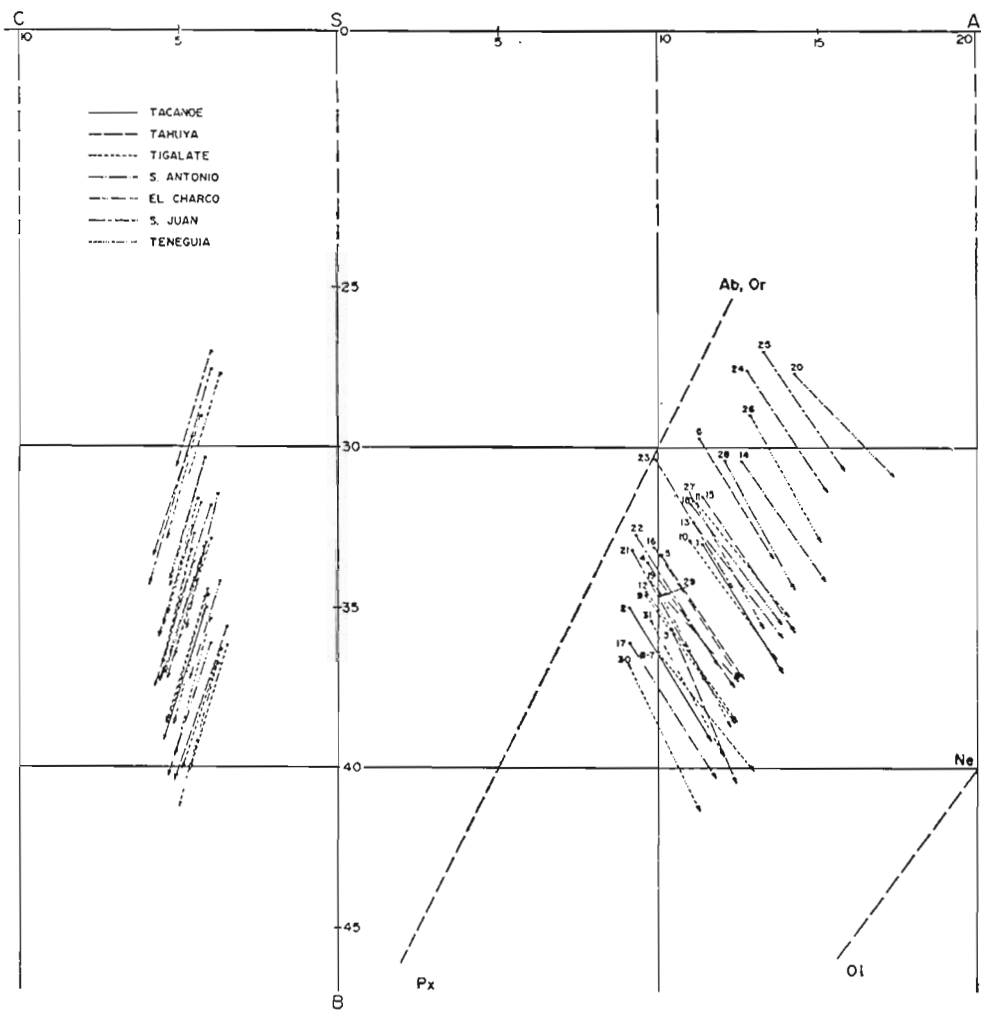


Fig. 4 — Zavaritzkii diagram of the historic lavas of La Palma.

TABLE 1

HISTORICAL ERUPTIONS OF THE CANARIAN ARCHIPELAGO

Year	Start-up	End	Island	Site and name	Main References	Observations
1341	—	—	Tenerife	Not located	RECCO's report, according to BRAVO, T., pers. com.	It is doubtful.
1393-1394	—	—	Tenerife	Not located	Report of Biscayan seamen. BRAVO, T., pers. com.	It is doubtful.
1430	—	—	Tenerife	Taoro eruption (Orotava valley) : Mña de las Arenas or Horca Mña de los Frailes, Mña de Gañanías.	Aboriginal guanche tradition. BRAVO, T., pers. com.	
between 1470 & 1492	—	—	La Palma	Tacande Volcano (Montaña Quemada)	Aboriginal guanche tradition SANTIAGO, M. (1960) HERNANDEZ-PACHECO, A.	Corroborated by C ¹⁴ dating 1530 ± 60 years
1492	Aug. 24 volcanic activity existed.	—	Tenerife	SW slopes of Pico Viejo : Montaña Reventada ?, near Montaña Bilma : Montañetas Negras ?	« Historias del Almirante » by Fernando Colon and Summary of Fray Bartolomé de las Casas according to BRAVO, T., pers. com.	The volcano has not been localized with absolute certainty.
1585	May 20	July ?	La Palma	Tahuya Eruption (Roques de Jedey)	Historical researches, by SANTIAGO, M. (1960) Geolog. researches by BRAVO, T. pers. com. & HERNANDEZ-PACHECO, A.	The Roques de Jedey made extrusion during this eruption
1646	Oct. 2	Dec. 18 or 21 st.	La Palma	Tigalate or Martin Volcano	SANTIAGO, M. (1960) and later geological researches by HERNANDEZ-PACHECO, A.	Some volcanic vents opened at the Eastern coast between the Puertito and La Baja del Agua.
1677-1678	Nov. 17	Jan. 21	La Palma	San Antonio or Montaña de las Cabras volcano, near Fuencaliente. La Caldereta Volcano ?	SANTIAGO, M. (1960) and later geological researches by HERNANDEZ-PACHECO, A.	
1704-1705	Dec. 31 Jan. 5 Febr. 2	Jan. ? Jan. 13 Febr. 26	Tenerife	Siete Fuentes or Llano de los Infantes volcano Fasnia, Almarchiga or Dos Roques volcano Montaña Arenas or Güimar volcano	FRITSCH, K. v. & REISS, W. (1868) BRAVO, T., pers. com.	The distance between the furthest vents is about 12 Km.
1706	May 5	May 14	Tenerife	Montaña Negra or Garachico volcano	FRITSCH, K. v. & REISS, W. (1868). BRAVO, T., pers. com.	Garachico village was destroyed.
1712	Oct. 9	Dec. 2	La Palma	El Charco or Montaña Lajiones eruption	SANTIAGO, M. (1960). AFONSO, A. (1974) and later geolog. researches by HERNANDEZ-PACHECO, A.	
1730-1736	Sept. 1	April 16	Lanzarote	Timanfaya eruption	HERNANDEZ-PACHECO, E. (1910, 1960)	Large number of vents opened. One third of the surface of the island was covered with lavas & pyroclasts.
1793	March ?	July ?	El Hierro	At El Golfo, near the N. coast	DARIAS & PADRON, D. V. (1929) BRAVO, T., pers. com.	The eruption was probably submarine. Many earthquakes affected the island.
1798	June 9	Sept. 8	Tenerife	Chahorra or Narices del Teide volcano	FRITSCH, K. v. & REISS, W. (1868). BRAVO, T., pers. com.	
1824	July 31 Sept. 29 Oct. 16	Oct. 16 Oct. 4 Oct. 24	Lanzarote	Tao or Clerigo Duarte volcano Volcán Nuevo del Fuego Tinguaton volcano	HERNANDEZ-PACHECO, E. (1910, 1960)	During the last stages of the eruption, sea water was emitted by Tinguaton volcano.
1909	Nov. 18	Nov. 27	Tenerife	Chinyero volcano	FERNANDEZ NAVARRO, L. (1911)	
1949	June 24 July 8 July 12	Aug. 9 July 26 July 31	La Palma	San Juan or Duraznero volcano Llano del Banco or Las Manchas volcano Hoyo Negro volcano	ROMERO ORTIZ, J. (1951) MARTEL, M. (1960) and geolog. researches of HERNANDEZ-PACHECO, A.	Also named « Eruption del Nambroque ».
1971	Oct. 26	Nov. 18	La Palma	Teneguía volcano	Teneguía Volume. Est. geol. 1974 unpublished Journal of the eruption by HERNANDEZ-PACHECO, A.	Several vents named Teneguía I to Teneguía VI.

TABLE 3

CHEMICAL ANALYSIS OF HISTORIC LAVAS OF LA PALMA ISLAND

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
SiO ₂	43.49	44.54	43.69	43.69	44.76	45.67	42.99	43.49	43.94	43.95	44.39	44.60	43.09	43.90	45.04	45.47
Al ₂ O ₃	14.83	13.42	14.19	13.65	14.19	15.48	13.39	12.98	13.81	14.58	14.96	13.81	15.03	15.73	14.58	14.06
Fe ₂ O ₃	4.00	3.65	4.10	6.72	5.04	6.11	4.62	4.10	4.46	4.58	5.85	3.85	4.47	4.79	5.43	4.06
FeO	8.33	9.04	8.42	5.74	7.51	6.25	8.15	8.18	7.61	8.03	6.66	8.17	7.81	6.99	4.81	7.56
MnO	0.21	0.21	0.08	0.22	0.22	0.22	0.20	0.22	0.21	0.22	0.23	0.21	0.22	0.33	0.23	0.21
MgO	8.06	8.83	10.68	7.92	8.08	6.67	8.27	8.50	8.43	7.48	6.87	8.42	6.95	6.77	8.02	8.28
CaO	10.24	11.27	9.25	11.36	10.91	9.79	12.76	12.65	11.84	10.91	11.19	11.84	11.22	10.07	10.72	11.19
Na ₂ O	3.94	3.21	3.70	3.41	3.56	3.91	3.50	3.54	3.40	3.91	3.94	3.50	3.81	4.42	4.26	3.50
K ₂ O	1.97	1.64	1.89	1.61	1.74	1.95	1.76	1.66	1.61	1.83	1.84	1.58	1.84	2.02	1.58	1.73
TiO ₂	3.51	3.66	3.57	3.70	3.55	3.55	3.60	3.64	3.66	3.75	3.76	3.68	3.72	3.68	3.70	3.57
P ₂ O ₅	0.87	0.95	0.85	0.81	0.83	0.82	0.95	1.19	1.26	1.01	1.11	1.02	1.06	1.03	0.96	0.91
H ₂ O	0.27	0.27	0.37	0.19	0.16	0.33	0.29	0.05	0.32	0.19	0.30	0.21	0.14	0.19	0.29	0.24
Total	99.72	100.69	100.25	99.58	100.55	100.75	100.48	100.20	100.17	100.44	100.90	100.89	99.36	99.92	99.62	100.78

	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31
SiO ₂	43.84	44.99	45.11	45.40	45.55	46.12	46.79	46.79	47.02	44.80	44.30	44.49	43.20	43.50	43.27
Al ₂ O ₃	13.29	14.56	13.42	16.12	14.19	14.24	14.96	15.73	16.12	16.06	15.09	15.44	13.99	12.99	13.68
Fe ₂ O ₃	3.86	3.56	4.04	4.39	3.38	2.69	2.74	2.82	3.95	5.10	4.31	4.38	4.73	3.36	3.92
FeO	8.14	8.23	7.89	6.55	9.15	9.08	8.73	8.01	7.17	7.08	8.55	7.96	8.85	9.76	9.39
MnO	0.22	0.24	0.21	0.23	0.21	0.20	0.21	0.22	0.21	0.21	0.22	0.22	0.22	0.22	0.22
MgO	9.39	7.19	8.69	5.26	8.44	8.22	7.46	6.18	5.91	6.68	7.63	7.18	8.70	10.19	9.22
CaO	12.03	10.97	11.19	10.07	10.55	11.08	10.24	9.23	9.04	8.77	9.52	9.19	10.07	10.37	10.28
Na ₂ O	3.23	3.91	3.50	4.91	3.46	3.50	3.56	4.53	4.70	4.58	3.99	4.33	3.71	3.42	3.60
K ₂ O	1.61	1.91	1.71	2.38	1.30	1.35	1.58	2.02	2.15	1.90	1.68	1.80	1.50	1.37	1.47
TiO ₂	3.57	3.70	3.79	3.36	3.34	3.36	3.50	3.47	3.49	3.30	3.88	3.61	3.90	3.54	3.71
P ₂ O ₅	0.91	0.94	0.93	1.07	0.73	0.71	0.78	0.85	0.88	0.98	0.89	0.98	0.89	0.84	0.88
H ₂ O	0.25	0.22	0.23	0.35	0.18	0.29	0.18	0.10	0.14	0.13	0.04	0.25	0.17	0.16	0.18
Total	100.34	100.42	100.71	100.09	100.48	100.84	100.75	99.95	100.75	99.59	100.10	99.83	99.93	99.74	99.82

1. — Oliv. px. basalt. Tacande

2. — Oliv. px. basalt. Tacande

3. — Px-oliv. basalt. Tahuya

4. — Px-oliv. basalt. Tahuya

5. — Px-oliv. basalt. Tahuya

6. — Px-amphib. basalt. Tahuya

7. — Px-oliv. basalt. Tigalate

8. — Px-oliv. basalt. Tigalate

9. — Px-amphib. basalt. Tigalate

10. — Px-amphib. basalt. Tigalate

11. — Px-amphib. basalt. Tigalate

12. — Px-oliv. basalt. Tigalate

13. — Px-oliv. amphib. basalt. San Antonio

14. — Px-oliv. amphib. basalt. San Antonio

15. — Px-oliv. amphib. basalt. San Antonio

16. — Px-oliv. amphib. basalt. San Antonio
17. — Px-oliv. basalt. El Charco.

18. — Px-oliv. basalt. El Charco

19. — Px-oliv. basalt. El Charco

20. — Px-amphib. basalt. El Charco

21. — Px-oliv.-plag. basalt. San Juan

22. — Px-oliv.-plag. basalt. San Juan

23. — Px-oliv.-plag. basalt. San Juan

24. — Px-plag. basalt. San Juan

25. — Px-oliv.-plag. basalt. San Juan

26. — Px-amphib. basalt. Teneguía

27. — Px-amphib. basalt. Teneguía

28. — Mean of 15 analysis of px-amphib. basalts. Teneguía

29. — Px-oliv. basalt. Teneguía

30. — Px-oliv. basalt. Teneguía

31. — Mean of 13 analysis of Px-oliv. basalts. Teneguía

Analysts : No. 1 to 25 : M. C. VALLS and A. HERNANDEZ-PACHECO.
Nos. 26 to 31 : E. Ibarrola (IBARROLA, E., 1974).

TABLE 2

Historic eruptions of La Palma	— Type of eruption — Number of eruptive vents — Maximum distance between them	— Days of eruption — Surface covered with lava & tephra (Km2)	— Morphology of the lavas — Type of material	— Proportion of ankaramites/basan- itoids/basalts, according to CIPW normative minerals (number of analysis)
TACANDE (between 1470 & 1492)	— Eruptive vent — 3 — —	— ? — 5.6	Aa, blocky lava Oliv. — px, basalt.	1/ 1 /— (2)
TAHUYA (1585)	— Eruptive fracture — 9 — 1100 m	— 50 at least — 4.8	Aa Px-Oliv. basalt Px-Amphib. basalt	2/ 1 /1 (4)
TIGALATE or MAR- TIN (1646)	— Eruptive fracture — 7 (9 ?) — 4900 m	— 79 or 82 — 7.5	Aa Px-Oliv. basalt Px-Amphib. basalt	1/ 5 /— (6)
SAN ANTONIO (1677)	— Eruptive fracture — 5 — 3700 m	— 66 — 6.5	Aa, dalles Px-Oliv-Amphib. basalt	1/ 3 /— (4)
EL CHARCO (1712)	— Eruptive fracture — 9 (12 ?) — 2600 m	— 55 — 4.9	Aa, ropy Px-Amphib. basalt Px-Oliv. basalt	1/ 3 /— (4)
SAN JUAN (1949)	— Eruptive fracture — 5 — 3450 m	— 47 — 4.5	Aa, pahoe-hoe, ropy Px-Oliv-plag. basalt Px-Oliv. basalt Px-plag. basal	2/—/3 (5)
TENEGUIA (1971)	— Eruptive fracture — 6 — 730 m	— 24 — 3.1	Aa, Blocky lava Px-Amphib. basalt Px-Oliv. basalt	2/ 3 /1 (6)

Comagmatic or xenolithic inclusions are common in the historic volcanoes. The comagmatic inclusions are pyroxenic and amphibolic cumulates and the xenolithic ones are not only peridotites and gabbros pulled out from great depths but also syenites, phonolites and basalts corresponding to more superficial substratum of the island. Some of these inclusions show partial fusion phenomena and metasomatic transformations such as alkanization and haüynization.

The geochemical characteristics can be appreciated in Table 3 and Figure 4. The alkaline and subsaturated character is strong. All the samples consolidated from a basic melt scarcely differentiated. However, a differentiation process always happened in the magmatic chamber and the emission of a sequence from amphibolic to olivine types can be related to the different stages of the eruption and the height over the sea level where the corresponding eruptive vents opened.

REFERENCES

- AFONSO, A. (1974) : *Geological sketch and historic volcanoes in La Palma, Canary Islands*. Estudios Geol. Volume Teneguía, p. 7-13.
- DARIAS y PADRON, D. V. (1929) : *Noticias Generales históricas sobre la isla de El Hierro, una de las Canarias*. Imprenta Cubelo, San Cristóbal de La Laguna, p. 181-184.
- FERNANDEZ NAVARRO, L. (1911) : *Erupción volcánica del Chinyero (Tenerife) en noviembre de 1909*. Anales Junta Ampliación de Estudios e Investigaciones Científicas, V. Memoria 1ª, 98 pp.
- FRITSCH, K. V. and REISS, W. (1868) : *Geologische Beschreibung der Insel Tenerife, ein Beitrag zur Kenntniss Vulkanischer Gebirge*. Winterthur 1868, 494 pp.
- HERNANDEZ-PACHECO, A. (1979) : *Lineaciones estructurales y vulcanismo en el Archipiélago Canario*. Comunicaciones III Asamblea Nacional de Geodesia y Geofísica. Madrid, 1979.
- HERNANDEZ-PACHECO, E. (1910) : *Estudio geológico de Lanzarote y de las isletas Canarias*. Mem. Real Soc. Hist. Nat. T. VI, 235 pp.
- HERNANDEZ-PACHECO, E. (1960) . *En relación con las grandes erupciones volcánicas del siglo XVIII y 1824 en Lanzarote*. El Museo Canario, Las Palmas, p. 239-254.
- IBARROLA, E. (1970) : *Variabilidad de los magmas basálticos en las Canarias Orientales y Centrales*. Estudios Geol. 26, p. 337-399.
- IBARROLA, E. (1974) : *Temporal modification of the basaltic materials from 1971 eruption of the Teneguía volcano (La Palma, Canary Islands)*. Estudios Geol. Volume Teneguía, p. 49-58.
- MARTEL SAN GIL, M. (1960) : *El volcán de San Juan, también llamado de «Las Manchas» y del «Nambroque», La Palma (Canarias)*. Taller de Artes Gráficas TPA, Madrid, 239 pp.
- ROMERO ORTIZ, J. (1951) : *La erupción del Nambroque en la isla de La Palma*. Boletín del Inst. Geol. y Min. España, 63. Madrid, 165 pp.
- SANTIAGO, M. (1960) : *Los volcanes de La Palma (Islas Canarias)*. El Museo Canario, 75-76. Las Palmas de Gran Canaria, p. 281-346.

INTERPRETATION OF GROUND DEFORMATION IN THE AZORES

by

F. MACHADO *

Azores University
9900 Horta, Azores

ABSTRACT

Geodetic measurements in Fayal, Pico and San Jorge showed some horizontal expansion of these islands. The total expansion during the interval 1936-1979 amounts to several metres, being considerably greater than the average sea floor spreading for that area of the Atlantic Ocean. This suggests that the spreading could be a discontinuous phenomenon, depending on short lived epochs of quick strain release. A vertical pulsation on the central part of Pico volcano was also detected; the pulsation has amplitude of about one metre and period close to one year. It seems due to periodic changes of pressure inside the magma chamber.

* Present address : Dept. of Geosciences, University of Aveiro, 3800 Aveiro, Portugal.

INTRODUCTION

Plate displacements at the triple junction of the Azores are probably responsible for the tectonic activity of the islands, as well as for the associated seismic swarms.

A tectonic model of the Azores (Fig. 1a) has been proposed previously (*Machado & al.*, 1972); this model was based primarily on the geological features of the islands. Another simpler model (Fig. 1b) had been proposed by *Krause and Watkins* (1970; see also *White & al.*, 1976). In this latter model, based on oceanographic and geomagnetic data, it is assumed that the mid-Atlantic rift passes undisturbed between Fayal and Flores, whereas the model of *Machado & al.* assumes that the rift has been shifted to the east and crosses all the islands having present-day volcanic activity.

Both models can probably be reconciled if the rift is divided into two branches: one passing between Fayal and Flores and the other through the active volcanoes (Fig. 1c). The existence of parallel branches of the mid-Atlantic rift has been considered also in Iceland (see, for instance, *Walker*, 1965).

The main purpose of the present research was to monitor the behaviour of Pico volcano and find out if any actual displacements could be attributed to the transform faults of Fig. 1a (or 1c). The measurements revealed that the trans-current movement is complicated by general horizontal expansion of the volcano and by a vertical pulsation.

These phenomena seem to represent a remarkable feature of some of the active volcanoes of the Azores.

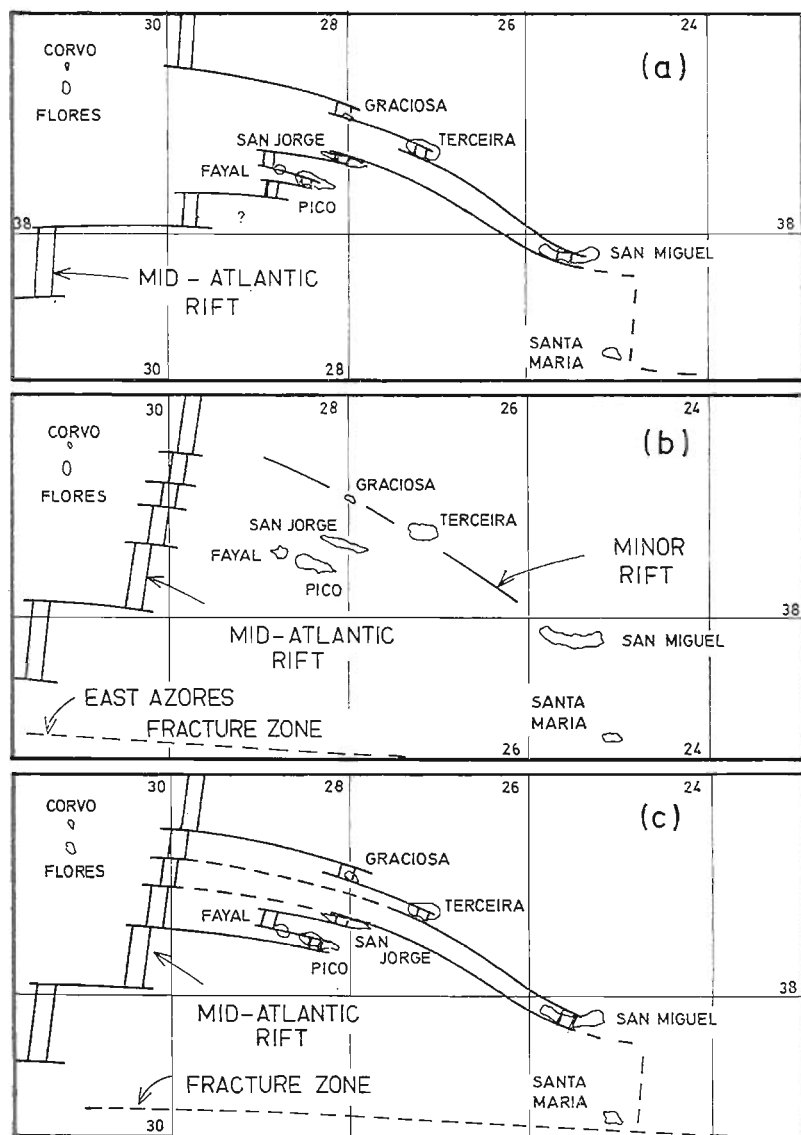


Fig. 1—Tectonic models of the Azores : (a) after MACHADO & al., 1972 (with slight changes) ; (b) after WHITE & al., 1976 ; (c) superposition of models (a) and (b).

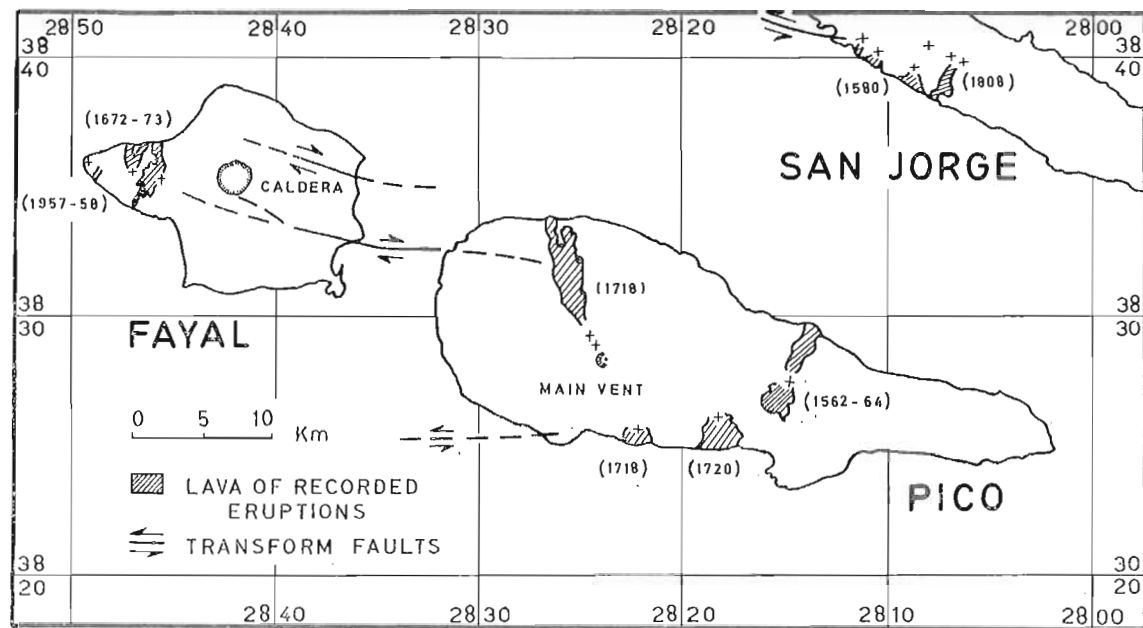


Fig. 2 — Islands of Fayal and Pico (with indication of the recorded eruptions and active transform faults).

LONG TERM HORIZONTAL DISPLACEMENTS

Measurements, using geodetic methods, were made in Fayal and Pico (Fig. 2), where a transform fault has considerable seismic activity. In fact, along this fault have been located the epicentre of the great earthquake of 1926 (*Agostinho*, 1927) and those of some strong shocks of the 1973 swarm (*Machado & al.*, 1974). Pico is a very young volcano having a regular big central cone with steep slopes ; Fayal is older and exhibits at present a summit caldera.

The work of 1975 consisted of a simple triangulation (using a Wild T2 theodolite) with the likely assumption that the length of one of the lines (conveniently selected) had not changed. The results, when compared with the geodetic survey of 1935 (made by Instituto Geográfico e Cadastral from Lisbon), showed some transcurrent movement superposed on a general horizontal expansion (*Machado and Possolo*, 1976).

As there were some doubts about the general expansion, the field work was repeated in 1979, a small base being then measured with a Wild DI3S (which gives a precision of 1 part in 10^5). The results (Fig. 3) are not significantly different from the previous ones.

A radial expansion of Pico volcano is apparent, whereas Fayal seems to move practically as a rigid block. If we subtract the E-W expansion (and the small N-S expansion found for Fayal), the residuals (Fig. 4) show some movement along two transform faults. In addition there seems to exist a minor N-S fault which makes compatible the N-S components of Pico expansion with the simple translation of Fayal, where radial expansion is practically non-existent.

In any case, the E-W extension is much bigger than expected. The average floor spreading of the Atlantic at this latitude is only about 1.3 cm per year on either direction.

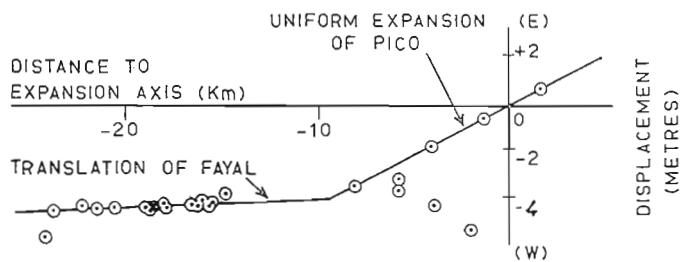
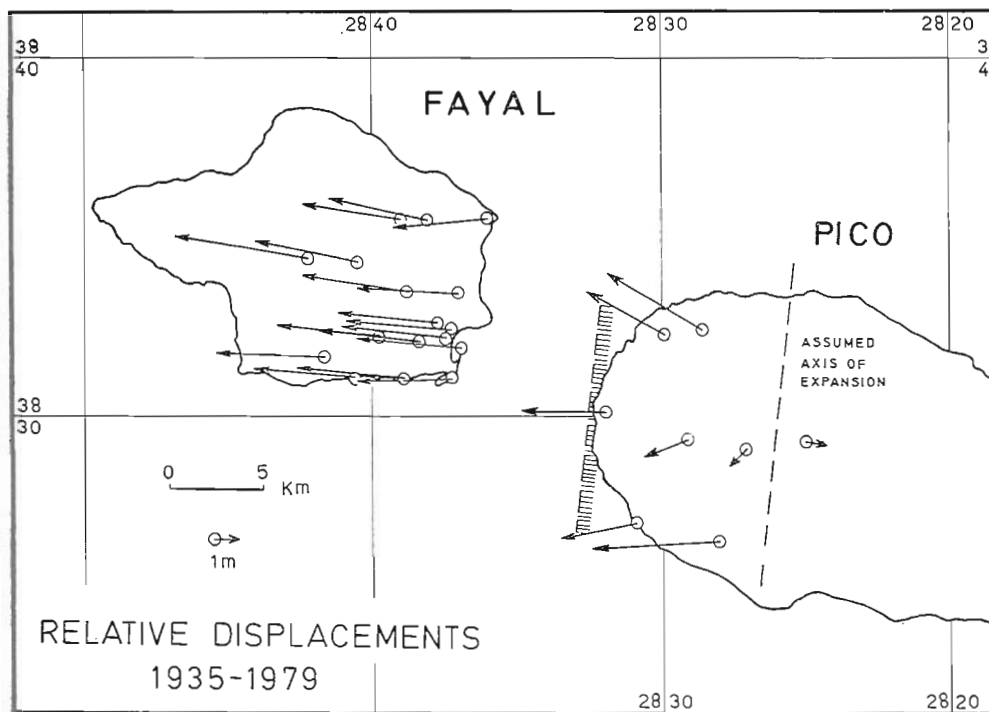


Fig. 3 — Horizontal displacements in Fayal and Pico, during the interval 1935-1979.

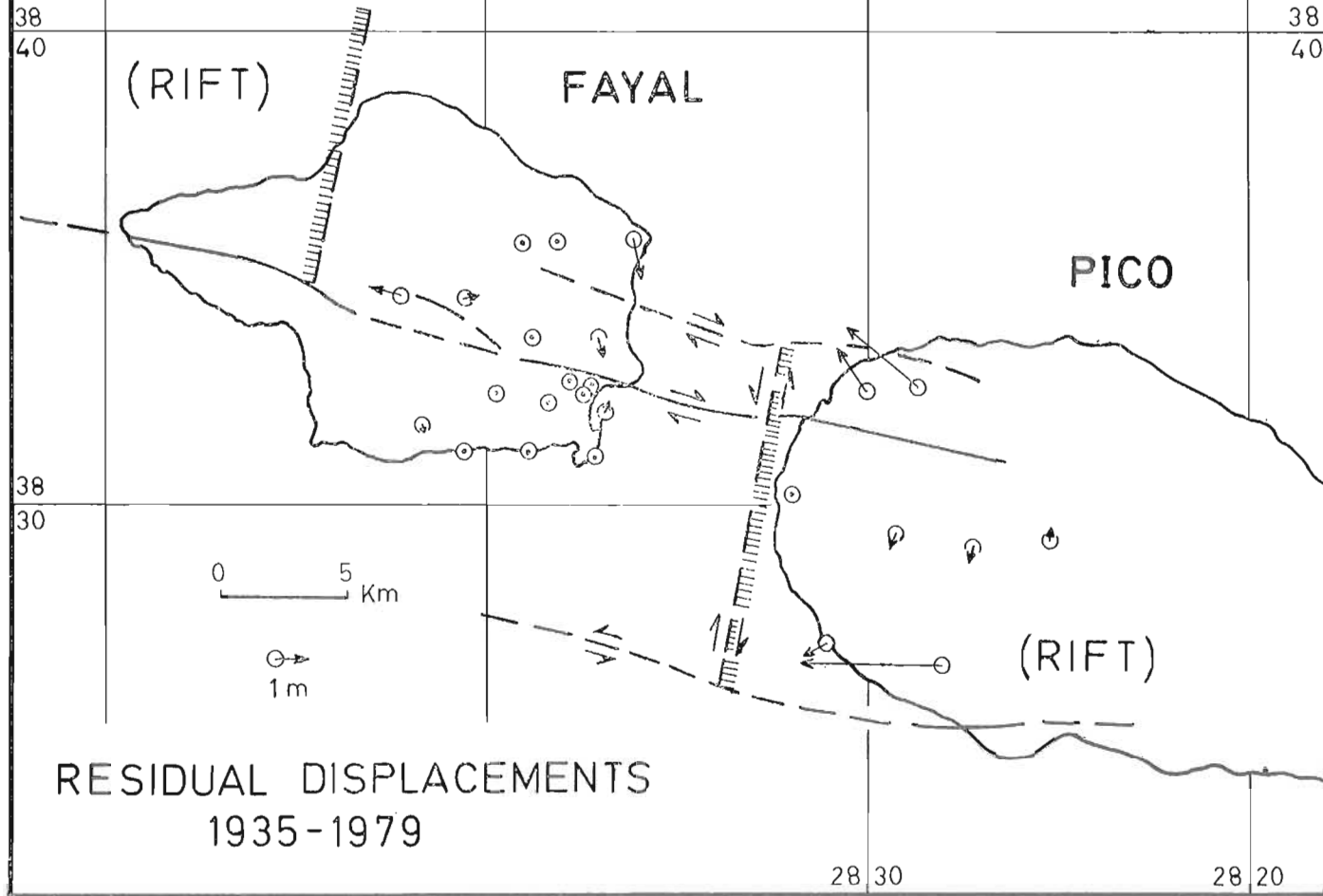


Fig. 4—Residuals of horizontal displacements after subtraction of E-W expansion (and small N-S expansion of Fayal).

Displacement in Fayal amounts to 4 or 5 meters in 44 years (relatively to a central line close to the main vent of Pico volcano), instead of the 50 or 60 cm which would correspond to the average spreading.

Also in a N-S direction the average movement assumed for the Azores-Gibraltar belt (Machado & al., 1972) is several times smaller than the N-S expansion of Pico.

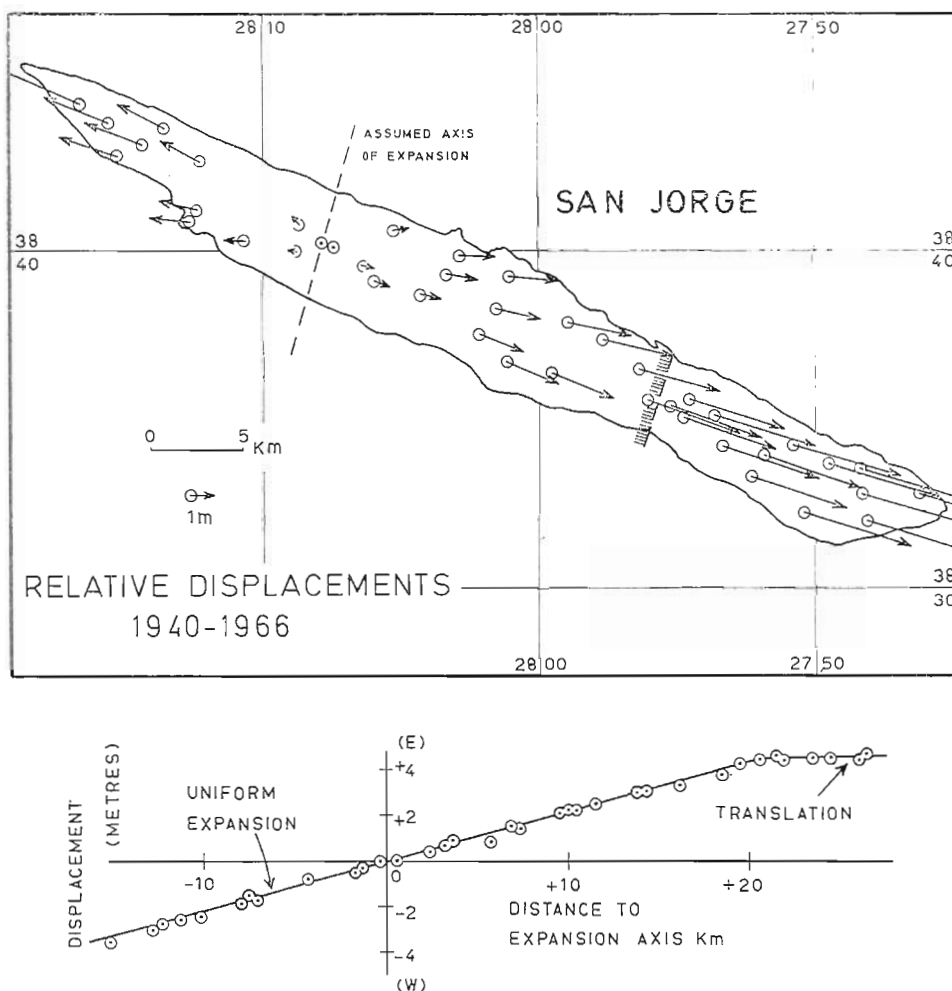


Fig. 5—Horizontal displacements in San Jorge, during the interval 1940-1946.

Very similar results were found in San Jorge (Fig. 5), where Instituto Geográfico e Cadastral repeated in 1966 the geodetic survey of the island, the first survey having been made in 1940.

These results suggest that sea floor spreading is not a continuous phenomenon. Strain could, in fact, accumulate during several centuries, being released in a quick way probably in connection with earthquake swarms or volcanic eruptions.

After 1936, there occurred the eruption of Capelinhos, Fayal, in 1957-1958 with all its seismic activity (Zbyszewski, 1960) and also a severe earthquake swarm in 1973 which was felt mainly in Pico (Machado & al., 1974). In San Jorge there was a seismic swarm in 1964 (Machado and Forjaz, 1964).

On the other hand, the last eruptions in Pico occurred in 1718 and 1720, and in San Jorge in 1808. Fayal had another eruption in 1672.

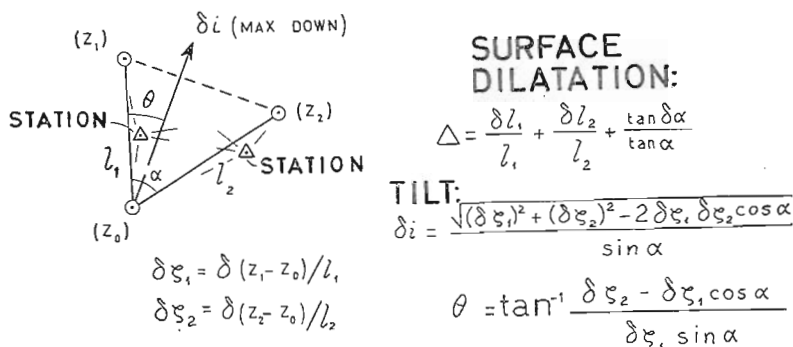


Fig. 6 — Diagram of the triangles for measurement of surface dilation and tilt.

VERTICAL DISPLACEMENT OF PICO

In addition to the horizontal expansion, the volcano of Pico shows a vertical oscillation which is being measured since 1975.

Until October, 1977 the measurements were made on small triangles with sides about 10 m long (Fig. 6) materialized by three stainless steel marks cemented to selected rock exposures.

At intervals of about 3 months we measured the level difference and length between every two marks of the triangles. This was performed by triangulation (with the Wild T2 theodolite) from two adequate stations about 5 meters apart, whose exact distance was measured with an invar stadia. The changes found between successive measurements were adjusted to Mogi's model (*Mogi*, 1958; see also *Machado*, 1974). In this model, changes of pressure in a small spherical magma chamber (Fig. 7), at depth h , produce at the surface radial and vertical displacements, respectively u_r and u_z , given by the equations

$$u_r = C \frac{r}{R^3} \quad (1)$$

$$u_z = C \frac{h}{R^3} \quad (2)$$

where C is a factor proportional to the change of pressure, and R is given by the equation

$$R^2 = r^2 + h^2 \quad (3)$$

r being the radial coordinate.

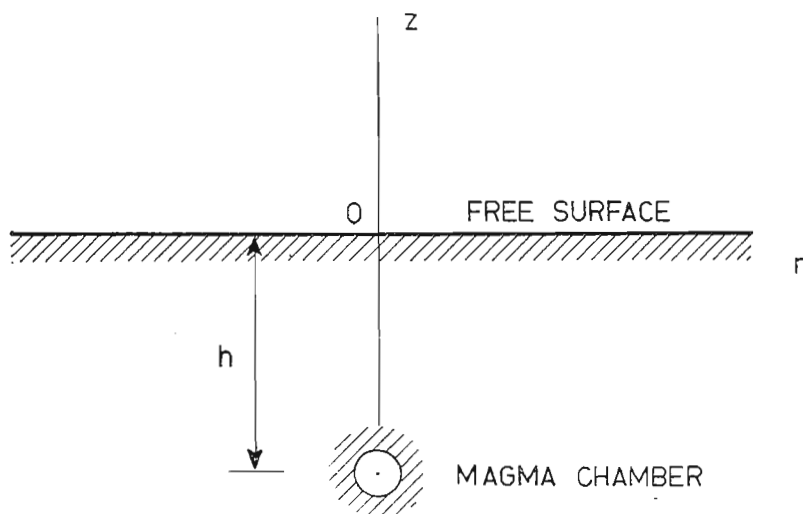


Fig. 7—Diagram of Mogi's model (the size of the magma chamber is supposed to be small when compared to the depth).

The most convenient quantities to consider in each triangle are the surface dilatation Δ and the radial tilt δi , which are easily deduced from the field measurements. For Mogi's model these quantities are

$$\Delta = \frac{du_r}{dr} + \frac{u_r}{r} = C \frac{2h^2 - r^2}{R^5} \quad (4)$$

$$\delta i = - \frac{du_z}{dr} = C \frac{3hr}{R^5} \quad (5)$$

Comparison of the measured values to equations (4) and (5) allowed to estimate the depth h (which was always about 4.7 km) and the factor C , and therefore the maximum vertical displacement

$$u_z (r=0) = \frac{C}{h^2} \quad (6)$$

The field work used to last for several days and the subsequent computation was slightly involved. As said before, this method was used only until Oct. 1977. Afterwards the heights of several existing geodetic monuments are being simply determined by triangulation (using the Wild T2 theodolite), but the atmospheric refraction has to be obtained for each set of observations (which lasts for only 3 or 4 hours).

The level changes are again adjusted to equation (2) and the maximum vertical displacement is then computed using equation (6).

Fig. 8 shows the small triangles and the geodetic monuments which have been used, and Fig. 9 and 10 are examples of the changes using either method.

The final results are fairly consistent. An area centred close to Pico's main vent seems to be pulsating with a period of about 1 year. Vertical displacements at the centre of the oscillating area are given in Fig. 11. Apparently the epochs of maximum height occurred at the beginning of summer until 1976, but are occurring at the beginning of winter since 1978. The amplitude is about 1 m.

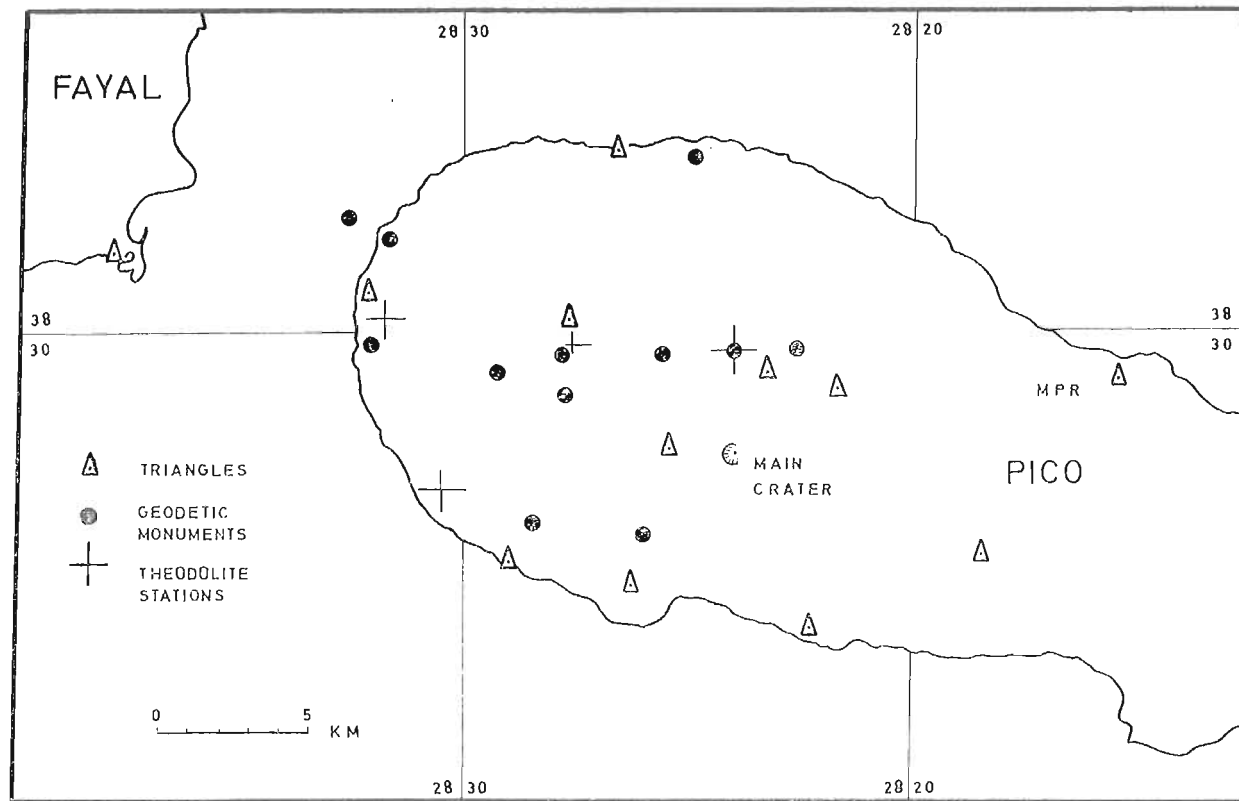


Fig. 8—Location of the small triangles and of geodetic monuments in Pico Island.

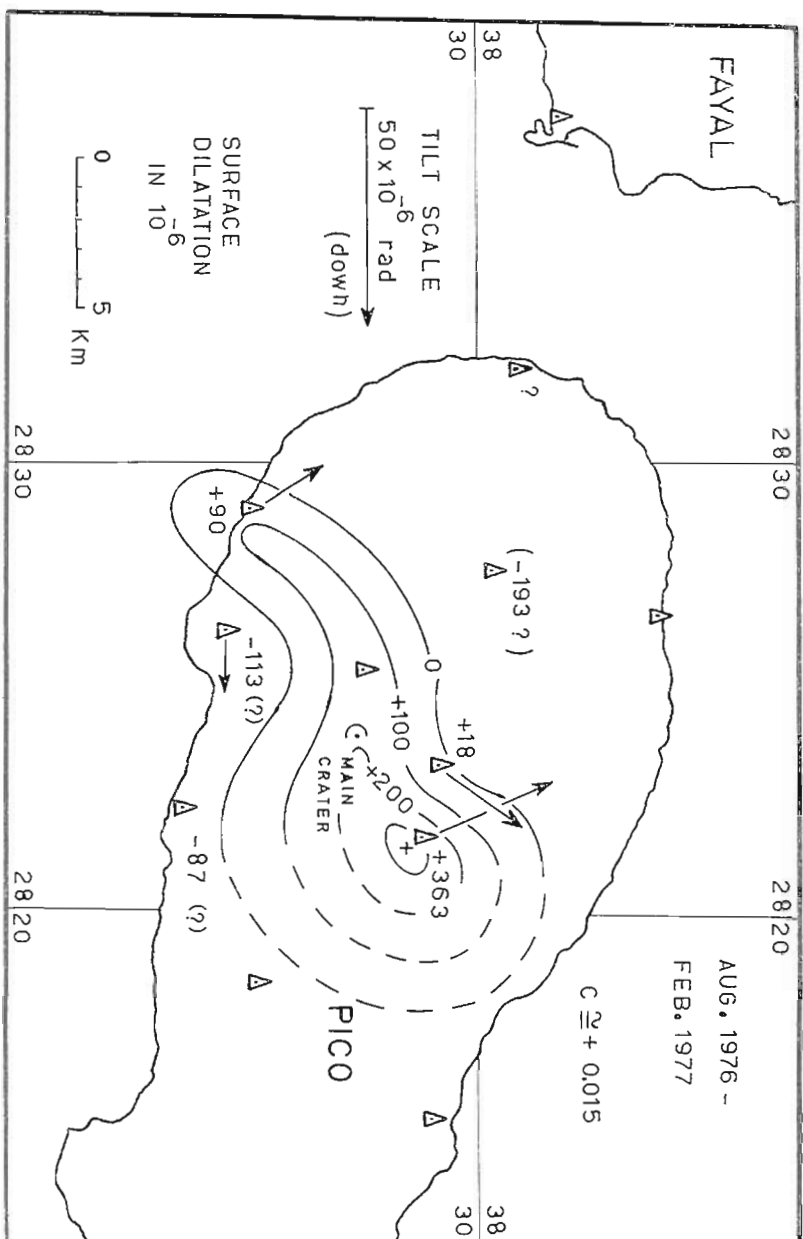


Fig. 9 — Surface dilatation and tilt in Pico for the interval Aug. 1976 — Feb. 1977.

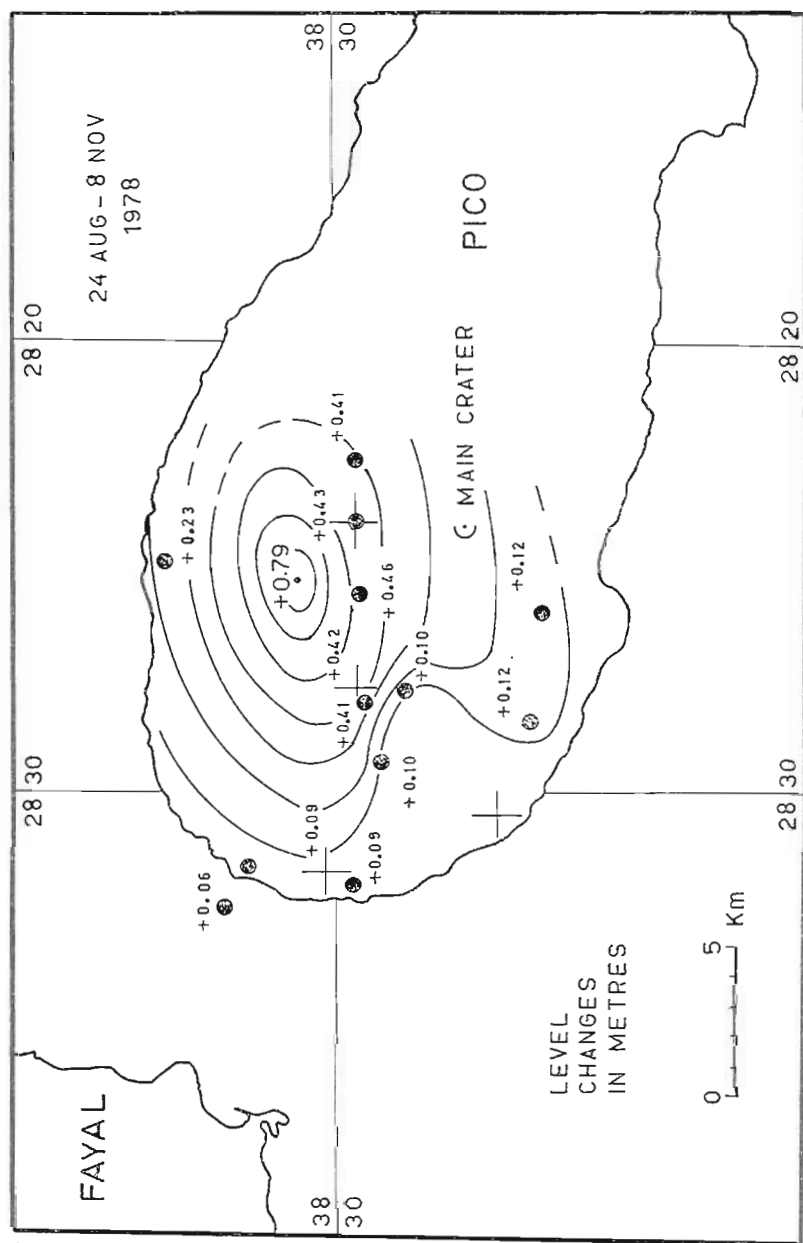


Fig. 10—Level changes in Pico for the interval 24 Aug. — 8 Nov. 1978.

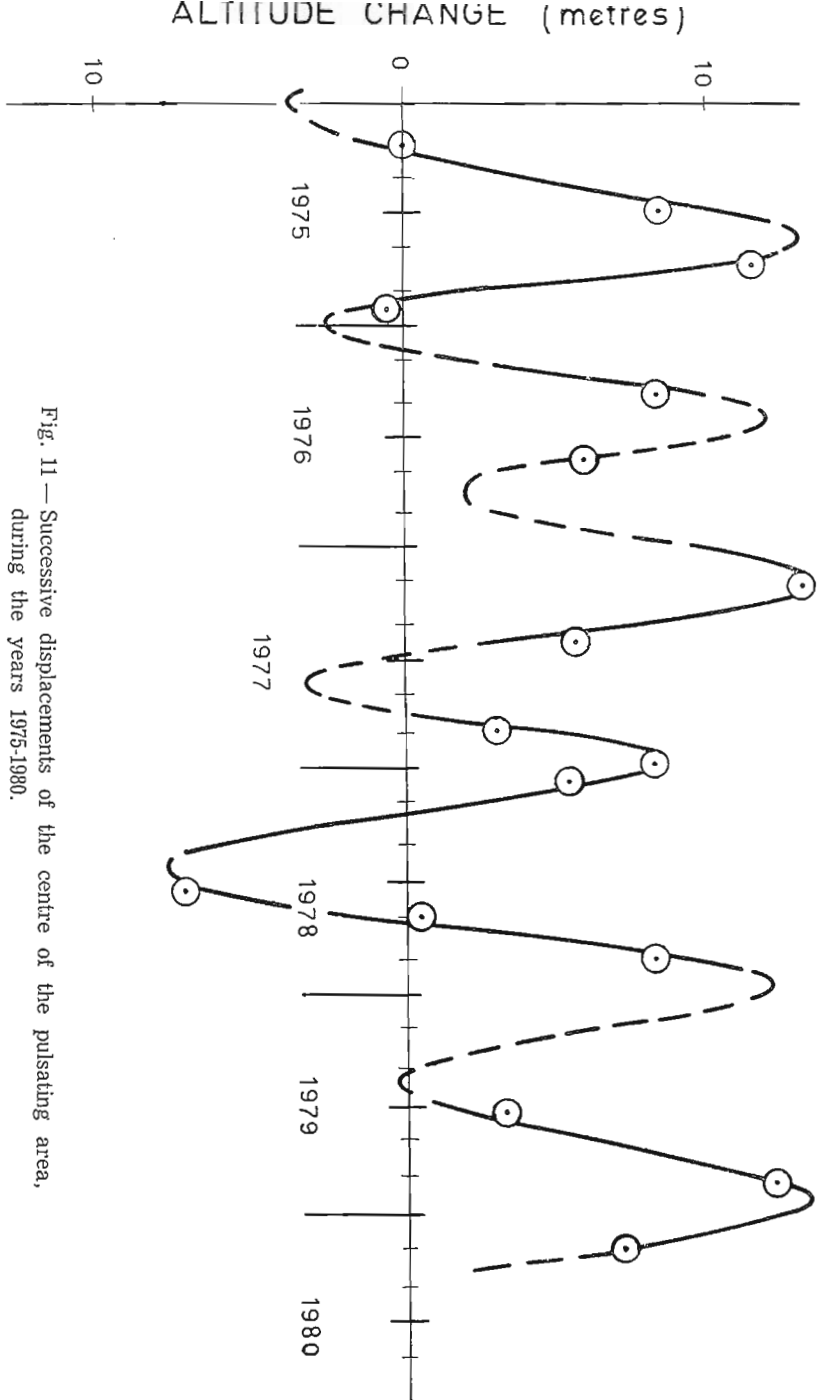


Fig. 11—Successive displacements of the centre of the pulsating area, during the years 1975-1980.

CONCLUSIONS

In Pico Island the expected transcurrent movement along transform faults is superposed on a radial expansion (which is probably an irreversible displacement) and a vertical pulsation with period close to 1 year. The two latter phenomena seem to occur on the same area (centred near the main vent of the volcano).

A detailed interpretation is uncertain; the expansion could well have increased the effect of some pulsation of pressure in the magma chamber of the volcano. This could be a usual feature of the active volcanoes of the Azores, but, except for the expansion of San Jorge, no similar measurements have been made on the other islands.

The phenomena could also be related to gravitational pulsations as proposed elsewhere (*Machado*, 1973; 1975); this is a speculative interpretation which has, however, to be worked out.

ACKNOWLEDGEMENTS

In addition to A. Possolo, the field work had the collaboration of several persons. We wish to thank especially J. Augusto from the Public Works Office in Pico, and M. Serpa from the Department of Oceanography of the Azores University.

REFERENCES

- AGOSTINHO, J., 1927 : *The earthquake in the Azores Islands, on 31st August 1926*. Zs. f. Vulkanol., 10, p. 268-272.
- KRAUSE, D. C., and WATKINS, N. D., 1970 : *North Atlantic crustal genesis in the vicinity of the Azores*. Geophys. J., 19, p. 261-283.
- MACHADO, F., 1973 : *A hipótese duma pulsação de gravitação com período de 11 anos*. Garcia de Orta, Sér. Geol., 1, p. 27-36.
- 1975 : *Pulsation of tectonic phenomena and tectonophysical mechanism*. Geol. Rundschau, 64, p. 74-84.
- MACHADO, F., and POSSOLO, A., 1976 : *Displacement in transform faults of the Azores (Abstract)*. Eos, 57, p. 675.
- MACHADO, F., QUINTINO, J. and MONTEIRO, J. H., 1972 . *Geology of the Azores and the mid-Atlantic rift*. Proc. 24th Int. Congr. (Montreal), 3, p. 134-142.
- MACHADO, F., TREPA, M. V., FERIN, C. and NUNES, J. C., 1974 : *Crise sísmica do Pico (Açores), Nov. 1973*. Com. Serv. Geol. Port., 57, p. 229-242.
- MOGI, K., 1958 : *Relations between the eruptions of various volcanoes and the deformations of the ground surface around them*. Bull. Earthq. Res. Inst., 36, p. 99-134.
- WALKER, G. P. L., 1965 : *Evidence of crustal drift from Iceland geology*. Phil. Trans. Roy. Soc., 258, p. 199-204.
- WHITE, W. M., SCHILLING, J.-G. and HART, S. R., 1976 : *Evidence for the Azores mantle plume from strontium isotope geochemistry of the Central North Atlantic*. Nature, 263, p. 659-663.
- ZBYSZEWSKI, G., 1960 : *L'éruption du volcan de Capelinhos (Ile de Faial, Açores)*. Bull. Volc., 23, p. 77-100.

THE IGNIMBRITES OF TERCEIRA, AZORES

by

STEPHEN SELF

Arizona State University,
Tempe, Arizona 85281 USA

ABSTRACT

On the oceanic islands of the Atlantic ignimbrites are known to occur on Iceland, Gran Canaria, Tenerife and on two of the Azores islands, Terceira and São Miguel. Ignimbrite may form a larger proportion of the total volume of rock on the island of Terceira than it does on any of the other islands, yet the individual eruption volumes are small ($\leq 1 \text{ km}^3$). There have been at least 6 major ignimbrite forming eruptions during the islands' history, and older ash-fall sequences may provide evidence of several others.

The two youngest ignimbrites, the Lajes (23,000yBP) and the Angra (19,000yBP) resulted from caldera-forming eruptions of Pico Alto volcano, a «parasitic» caldera on the north flanks of the older Guilherme Moniz volcano. Both pyroclastic flows reached the edge of the island and much of the deposition was probably submarine. The cooling units are relatively thin (1-20 m), but the Lajes is densely welded, a feature typical of highly alkaline ignimbrites.

The transport mechanism of the two young ignimbrites has been investigated using grain-size relationships between the various layers. Ground surge, ash-cloud surge and co-ignimbrite ash-fall deposits are associated with the main ignimbrite bodies. Nearly all the pyroclastic flows have been constrained by topography to flow along valleys between the main composite volcanoes, and have all reached the sea in the same regions. Sequences exposing to older ignimbrites are hence restricted to these narrow coastal regions and little is known of their and sources.

THE GEOLOGY, VOLCANIC ACTIVITY, AND AGE OF BOUVETØYA, SOUTH ATLANTIC *

by
TORE PRESTVIK
Geologisk Institutt
N-7034 Trondheim-NTH, Norway

ABSTRACT

The volcanic cone of Bouvetøya is built up of two formations. The older formation, of predominantly pyroclastic rocks which are typically hydrothermally altered, is overlain by a formation of mainly lava flows. The rocks present on the island constitute a transitional volcanic series. K/Ar dating indicates that surface rocks were formed as early as 1.4 Ma B.P. The rocks of Nyrøysa are 0.4-0.5 Ma, showing that this platform, which formed between 1955 and 1958, represents a landslide type deposit rather than a recent eruption as has previously been suggested by several workers. Various kinds of evidence suggest that the volcanic activity, or a cycle of activity, of Bouvetøya is now in a declining phase.

* Publication No. 58 of the Norwegian Antarctic Research Expeditions (1978/79).

INTRODUCTION

Bouvetøya is an oceanic island located close to the Bouvet triple junction in the South Atlantic Ocean (Fig. 1). On the basis of distance from the central part of the Bouvet ridge and half spreading rates for this region given by SCLATER *et al.* (1976), the age of the oceanic crust beneath the islands is estimated to 4.5-5.0 Ma. This is thus a maximum age of the volcanic activity responsible for the Bouvetøya cone.

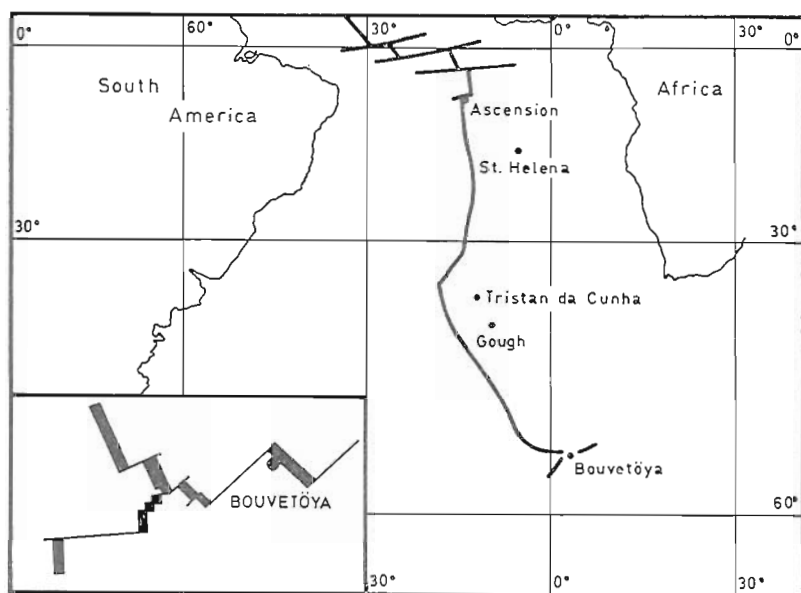


Fig. 1—Location of Bouvetøya. The tectonic relations of the Bouvet triple junction as interpreted by SCLATER *et al.* (1976) are inserted.

The island covers an area of only 55 km² of which about 95 % is capped by permanent ice. The assymetrical shape of Bouvetøya, with the crater/caldera area in the northwest (Fig. 2), is influenced by the wave action caused by almost permanent westerly winds in the area. Rocks are, however, well exposed in outcrops along the coast (Fig. 2).

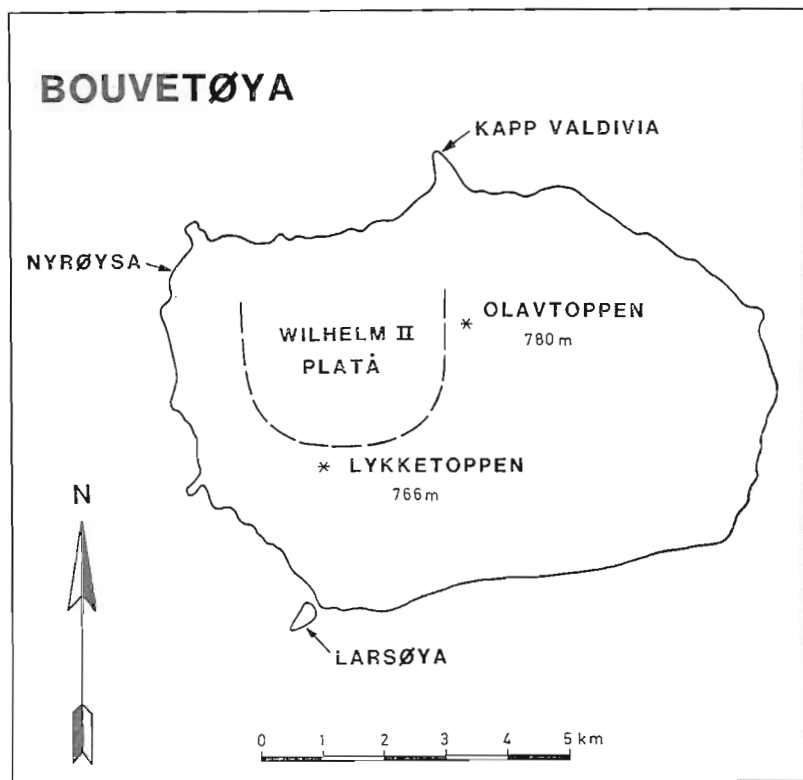


Fig. 2 — Sketch map of Bouvetøya.

The Norvegia expeditions in 1927-28 and 1928-29, a joint British/South African expedition in 1964, a South African expedition in 1966, and a Norwegian expedition in 1977, resulted in several papers concerning the geology and petrology of Bou-

vetøya. It was established that the volcanic rocks of the island belong to a transitional volcanic series (BROCH, 1946; VERWOERD *et al.*, 1976; IMSLAND *et al.*, 1977). The origin of the platform, now called Nyrøysa, that appeared on the northwestern coast between 1955 and 1958 has, however, been a matter of discussion. This platform was interpreted as the result of a recent eruption by some workers, whereas others thought it represented an avalanche type deposit. To resolve this question was therefore one of the goals set by the members of the Norwegian Antarctic Expedition 1978-79 that visited Bouvetøya for about two weeks. Even though the weather conditions were bad during this expedition, much new information was obtained. Detailed descriptions of these findings and a discussion of the geology of Bouvetøya have been given by PRESTVIK and WINSSENS (1981).

In the present paper the new information is summarized and briefly discussed together with what is known about petrology from previous work.

GEOLOGY AND PETROLOGY

The volcanic cone of Bouvetøya is built up of two major formations. The older formation consists of various kinds of volcanoclastic rocks as seen in the steep cliffs of the southern, western, and northern coasts (Fig. 2) where deep sections are well exposed. This formation is frequently cut by dikes and hydrothermal alteration is conspicuous, especially in areas close to the crater/caldera area. Fumarolic activity, which has been observed by all recent expeditions, is also concentrated in this area. The volcanoclastic rocks of this formation were probably formed as hyaloclastites (Surtsey type eruption).

The pyroclastic rocks are overlain by a formation of mainly lava flows which are characteristically less altered than the

underlying rocks. In some places the lava flows are mixed with minor occurrences of palagonitized hyaloclastites. These features indicate that the island had only a thin icecap that gave rise to small amounts of hyaloclastites before it melted locally so that subaerial lava flows could be formed.

The transition between the two formations is interpreted to represent the time when the volcanic structure of the island became rigid and compact enough to prevent seawater from entering conduits or vents.

Transitional basalt (hawaiite) is the predominant rock type of Bouvetøya. Intermediate rocks (trachytic icelandite or benmoreite) occur on the Nyrøysa platform, and peralkaline rhyolite (comendite) is found in a lava flow on the islet of Larsøya, in a dome at Kapp Valdivia (Fig. 2), and in a few dikes cutting the lowermost pyroclastic formation.

The petrology of these rocks has recently been discussed by VERWOERD *et al.* (1976), IMSLAND *et al.* (1977), and PRESTVIK (in press). The different members of the transitional suite are thought to be interrelated by fractional crystallization (VERWOERD *et al.*, 1976). However, IMSLAND *et al.*, (1977) proposed a model for the evolution of the series involving fractional crystallization under varying conditions of P, T, and P_{H_2O} either from slightly different magma types or from one parental magma type in an open system. The whole series displays a chondrite-normalized pattern characterized by light REE enrichment (Fig. 3). A strong negative Eu anomaly of the comendites suggests strong feldspar fractionation under low P_{O_2} at a late stage in the evolution of the series (PRESTVIK, in press).

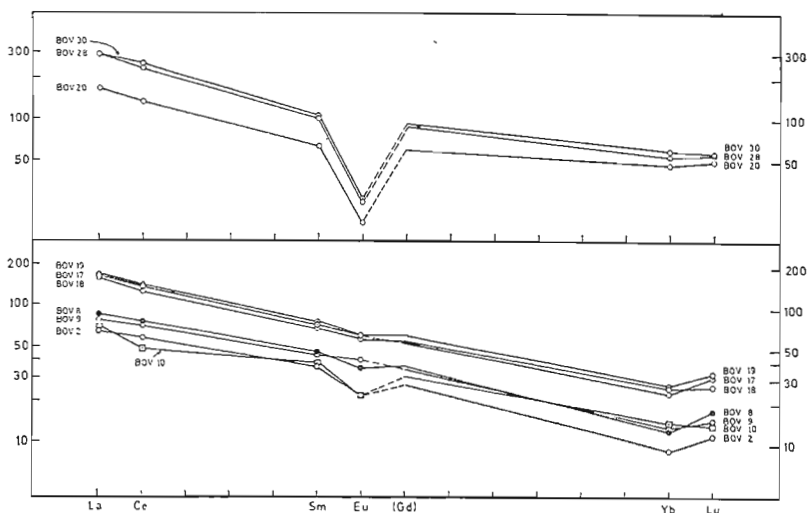


Fig. 3 — Chondrite normalized REE patterns of volcanic rocks from Bouvetøya. From PRESTVIK (in press). Vertical scale : Sample/chondrite ratios.

THE VOLCANIC ACTIVITY AND AGE OF BOUVETØYA AND THE FORMATION OF NYRØYSA

It was mentioned above that the maximum age of the Bouvetøya volcanic cone is estimated to some 4.0-5.0 Ma. The real age of the initiation of this volcanism is unknown because no dating of rocks from the submarine structure has been performed. VERWOERD *et al.* (1976) discussed the age of Bouvetøya (super-surface part) and concluded that an age in excess of 1 Ma was unlikely.

K-Ar datings on new material from Bouvetøya (PRESTVIK and WINSNES, 1981) show that two subaerial lava flows from a low stratigraphic position in the upper formation have apparent ages of 1.39 Ma and 1.06 Ma. The lowermost formation

of pyroclastic rocks is probably only slightly older than the overlying lava flows, because it is unlikely that the predominantly pyroclastic cone could survive the intense marine abrasion of the area for any long time. However, K-Ar dating of the lowermost formation gave lower apparent K-Ar ages than what were found in the overlying lava flows. This is interpreted as due to loss of radiogenic argon, probably as a result of the extensive hydrothermal alteration of these rocks.

Furthermore, field and petrographic evidences (PRESTVIK and WINSNES, 1981) indicate that the silicic rocks of Larsøya and Kapp Valdivia are relatively young. This impression is further substantiated by the K-Ar apparent ages (< 0.2 Ma) obtained for these rocks.

The Nyrøysa platform on the northeastern coast (Fig. 2) has been interpreted as the result of a recent eruption (BAKER and TOMBLIN, 1964; LUNDE, 1965; BAKER, 1967; FURNES and LØVLIE, 1978) whereas WINSNES (1966) and VERWOERD *et al.* (1976) interpreted this platform as an avalanche type deposit.

Recent K-Ar analyses of rocks from the Nyrøysa platform (PRESTVIK and WINSNES, 1981) gave ages of 0.4-0.5 Ma showing that the platform does not represent a recent eruption. Several features, such as consistent magnetic polarity directions (FURNES and LØVLIE, 1978) and stratigraphy, indicate that the platform represents a landslide rather than an avalanche.

The real stratigraphic position of the intermediate rocks of Nyrøysa in the Bouvetøya cone is as yet not known in detail. They must however, represent a position above the basic lavas occurring immediately above pyroclastic rocks of the lower formation. The relatively big age difference between these basic lavas and the rocks now present at Nyrøysa indicates that the volcanic activity was low at this stage of construction of the Bouvetøya cone. The size and shape of the plateau surrounded by a few peaks in the summit area (Fig. 2) indicates that the island has evolved beyond the stage of caldera collapse (IMSLAND *et al.*, 1977). Even though very little is

known about the continuity of the magmatic activity of Bouvetøya, several features such as a general trend from basic to intermediate and silicic rocks with time and the scarcity of apparently very young rocks, indicate that the volcanic activity or a cycle of activity is now in a declining phase.

ACKNOWLEDGEMENTS

Thanks are due to Prof. Chr. Oftedahl and Prof. F. M. Vokes who made comments on the paper.

REFERENCES

- BAKER, P. E., 1967 : *Historical and geological notes on Bouvetøya*. Br. Antarct. Surv. Bull., 13, p. 71-84.
- BAKER, P. E. and TOMBLIN, J. F., 1964 : *A recent volcanic eruption on Bouvetøya, South Atlantic Ocean*. Nature, 203, p. 1055-1056.
- BROCH, O. A., 1946 : *Lavas of Bouvet Island*. Sci. Res. Norw. Antarctic Exped. 1927-1928, No. 25, p. 3-26.
- FURNES, H. and LØVLIE, R., 1978 : *An eruptional model for recent lava flow on Bouvetøya, South Atlantic Ocean*. Norsk Polarinst. Skrifter, Nr. 169, p. 103-108.
- IMSLAND, P., LARSEN, J. G., PRESTVIK, T. and SIGMOND, E., 1977 : *The geology and petrology of Bouvetøya, South Atlantic Ocean*, Lithos, 10, p. 213-234.
- LUNDE, T., 1965 : *Fra et besøk på Bouvetøya* (in Norwegian). Norsk Polarinst. Årbok, 1963, p. 197-203.
- PRESTVIK, T. : *Trace element geochemistry of Bouvetøyt, South Atlantic*. In : Craddock, C. (ed.). Antarctic Geosciences, University of Wisconsin Press, Madison, Wisconsin, U.S.A. (in press).
- PRESTVIK, T. and WINSNES, T., 1981 : *Geology of Bouvetøya, South Atlantic*. Norsk Polarinst. Skrifter 175, p. 41-69.
- SCLATER, J. G., BOWIN, C., HEY, R., HOSKINS, H., PEIRCE, J., PHILLIPS, J. and TAPSCOTT, C., 1976 : *The Bouvet triple junction*. J. Geophys. Res., 81, p. 1857-1869.
- VERWOERD, W. J., ERLANK, A. J. and KABLE, E. J. D., 1976 : *Geology and geochemistry of Bouvet Island*. In : Gonzales Ferran, O. (ed.). Proceedings of the Symposium on « Andean and Antarctic Volcanology Problems », p. 203-237. Stabilimento Tipografico Francesco & Figli. Napoli, Italy.
- WINSNES, T., 1966 : *Besøk på Bouvetøya i 1958 og 1966*. (in Norwegian). Norsk Polarinst. Årbok, 1965, p. 143-149.

LOWER CRETACEOUS SUBMARINE RIFT VOLCANISM IN THE SOUTHERN TRANSDANUBIA IN HUNGARY

by

I. BILIK

Petrological and Geochemical Department
Faculty of Nature Sciences,
Eötvös University
1088. Budapest, Muzeum-ktr 4/A. Hungary

ABSTRACT

In the southern part of Transdanubia during the Triassic and Jurassic times, more than 6,000 metres terrigenous, shelf and pelagic sedimentary sequences formed in a basin which was sinking with different speed.

The oldest volcanic rocks in this territory are intercalations of basic and intermediate tuffs, found in the Lower-Liassic formation, which is characterised by very rapid sinking of the basin. These volcanoclastics are encountered as patches with thicknesses approximating some few metres. The chronology of these rocks is discussed.

The sediments following the Lower-Liassic formation are characterised by continuity till the Lower-Cretaceous, and they are free from volcanic materials.

The main volcanic activity in the area began during the Berriasian age under open pelagic conditions, and continued to the Upper-Valanginian. In the pause of the volcanic activity, during the Upper-Valanginian and Hauterivian shelf and shallow-water clastic sediments were deposited.

After that the volcanic activity commenced again but during this period, the volcanic activity covered a lesser space and time. During this volcanic activity dykes and subvolcanic masses were intruded in the Permian-Triassic and Jurassic sedimentary sequences of the miogeosinclinal basin. The volcanic suite was formed in the fault system of NEE-SWW direction. The Lower-Cretaceous volcanic rocks are highly differentiated, from the basic diabase and dolerite to the albitic diabase and keratophyre and phonolite, the origin of which is explained by magmatic differentiation and contamination with sedimentary rocks. Among the subvolcanic and volcanic rocks spilitic rocks are common. The submarine lava-flows show in situ brecciation, as follows: different kinds of pillow-lavas, lavabreccias and hyaloclastics.

About the origin of the Lower-Cretaceous volcanic suite and its tectonical environment — to our knowledge — we have not a uniform point of view. If the territory is supposed as a tectonical micro-unit, the volcanism seems to be as a result of a continental-rifting. And if we take into consideration the ultrabasic and basic volcanic rock-suite, which can be found North from the northern border of the area, and which can be considered an ophiolite-like-suite, we might regard the Lower-Cretaceous volcanism as an analogy of the island-arc calcalkaline volcanism. The paper tries to give an answer to the two possibilities of the origin of this volcanism on the basis of several petrochemical discriminative diagrams.

THE LARGE TOLBACHIK FISSURE ERUPTION IN 1975-1976, KAMCHATKA

by

S. A. FEDOTOV, S. T. BALESTA
V. I. GORELCHIK, G. B. FLEROV
and V. B. ENMAN

Institute of Volcanology
Petropavlovsk-Kamchatsky, 683006, USSR

ABSTRACT

This paper discusses the results of seismological, geodetic, geophysical, geologic and petrologic investigations of the large Tolbachik fissure eruption which occurred from July 6, 1975 to Dec. 10, 1976 in Kamchatka. The eruption proceeded first in the region of North vents (July 6-Sept. 15, 1975) where the character of its activity was predominantly explosive and then after a short repose period started again in the region of South vent where its activity was predominantly effusive with outpouring of fluid lava. The total volume of erupted products is 2.3 km^3 . The time and locality of eruption were successfully predicted according to the character of a swarm of earthquakes recorded at depths from 0 to 20 (30) km. Furthermore, based on a characteristic decrease of the number of earthquakes

at the end of swarms preceding the formation of eruptive centers, the time of formation of some cones in the regions of North and South vents was predicted also. The dimensions of the feeder dykes in the region of North vents were evaluated based on horizontal crustal deformations. Their thickness is more than 1 m, the visible length is not more than 400 m. The overpressure of magma and gas is 100 to 300 kg/cm². In the course of eruption during a short period of time there was observed a successive change in basalt composition from high-magnesian of moderate alkalinity (North vents) through intermediate to aluminous subalkaline (South vent). Geologic and petrologic data testify to independence of two contrasting in composition basalt magmas. Basalts of intermediate composition are a product of mixing of these magmas. The main feeding source of basalts ejected from the North and South vents was a system of connected intermediate chambers located in the lower crustal layers or in the crust-mantle transition layer. It has been established that during the eruption in the region of North vents a system of peripheral magma chambers formed in the upper crustal layers at depths of 2-3 and 7-8 km. These chambers are considered to be the possible places of magma mixing.

A PROVISIONAL GEOLOGICAL MAP OF ASCENSION ISLAND, SOUTH ATLANTIC

by

J. D. BELL, F. B. ATKINS & C. HARRIS
University of Oxford

ABSTRACT

A new (provisional) geological map of Ascension Island will be exhibited accompanied by representative rock specimens, thin sections, photomicrographs, photographs of volcanic features, and an overlay indicating the distribution of coarse-grained blocks.

(This paper was not read at the Symposium)

