LINEAR VOLCANISM IN FRENCH POLYNESIA

by

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Analyses and arguments presented in this thesis are my own unless otherwise indicated. No part of this work has been submitted to any other university or similar institution.

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The island of Moorea at sunset as viewed from the west coast of Tahiti. The thin black band just beneath the island is the coral reef which circles Tahiti.

ABSTRACT

The islands of French Polynesia are the peaks of individual shield volcanoes which rise 4,000 m from the ocean floor in the eastern Pacific Basin. The arrangement of these islands is in lineaments which are subparallel to one another and whose extensions to the southeast are approximately perpendicular to the East Pacific Rise, the site of production of new oceanic crust for the central Pacific. This geometry suggests at least an indirect relationship between volcanism in French Polynesia and sea floor spreading at the East Pacific Rise.

Potassium-argon whole rock ages for lavas from islands in the Marquesas Islands, the Society Islands, the Pitcairn-Gambier Islands and the Austral Islands are considerably younger than estimates for the age of the ocean floor below. Volcanic activity in the Society and Marquesas Islands began about 8 and 4 Myr B.P., respectively, while the volcanism which produced the Austral Islands is as old as 12 Myr and may have earlier produced members of the Cook Islands. The suggestion from geomorphologic considerations that island ages increase to the northwest (within each island chain) has been confirmed for the four volcanic lineaments in French Polynesia. The rates of migration of volcanism are surprisingly uniform, at 11 ± 1 cm/yr. The observed, or inferred, sites of active volcanism (one for each of the four lineaments) have remained fixed with respect to each other and to other Pacific island chains for at least the last 4 Myr.

Major element chemistry reveals that the lavas from French Polynesian islands belong to the alkali olivine basalt association. The bulk of the shield building lavas are olivine basalts, basanites and ankaramites, but there is also minor development of the hawaiite-mugearite-trachyte differentiation trend at most islands. The

variations in major and minor elements among shield building lavas from three French Polynesian island chains indicate that no single parental magma could have produced the differentiated igneous suites exhibited in French Polynesia. Instead, a variety of primary liquids, segregating under different conditions from a heterogeneous peridotitic source, were available for construction of the shield volcanoes. A study of the isotopic composition of strontium indicates that the source region for French Polynesia lavas has maintained heterogeneities in Rb/Sr for a period of approximately 1.4 billion years.

The angular dispersion of virtual geomagnetic poles (for French Polynesian lavas) about the geographic pole is not anomalous when compared with models of dipole and nondipole field activity and with similar studies of palaeosecular variation. Anomalous deep mantle material, proposed to shield the nondipole field and thus explain low estimates of palaeosecular variation at Hawaii (Doell and Cox, 1971), is not required beneath French Polynesia. A kinematic study of plate to hot spot movements, using the rates of migration of volcanism in French Polynesia, suggests that a world wide distribution of hot spots has maintained a stable configuration for the last 10 Myr.

The data reported in this investigation, although circumstantial, favour deep, dynamic sources for linear volcanism (the Wilson-Morgan hot spot/plume hypothesis) over any of the shallow, passive mechanisms of lithosphere to asthenosphere interaction.

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Chapter 1: Introduction

1.1 Nonorogenic linear volcanism

That there should be an intimate relationship between volcanism and crustal interactions is assured by plate tectonic theory: crustal rifting is the site of plate divergence and of the volcanism responsible for the bulk of production of new crust, while plate convergence is the site of crustal consumption and of the volcanism associated with the subducting plate. Alkalic and tholeiitic flood basalts are produced in the first, whereas the subduction environment is the characteristic location of the calc-alkaline suite, in addition to basalts. The significant portion of igneous rocks forming on the earth today and (at least) since the dispersion of Gondwanaland can be associated, conceptually, with these two magma producing tectonic environments.

The volcanism associated with these two broad igneous provinces marks the sites of relative motion between crustal units, or plates in the present terminology. Plate boundaries are not fixed in space, however, nor is the volcanism associated with plate interaction, but both will react passively to the conditions imposed by plate motions.

Volcanic and plutonic activity found interior to, and crossing the borders of, coherent crustal plates (both continental and oceanic) cannot be accounted for by plate margin interaction. In addition to kimberlites and carbonatites, conspicuous and classic examples of this igneous phenomenon are the White Mountain Magma Series of New England, the Younger Granite Ring Complexes of Nigeria, Iceland and the Tertiary Volcanic Province of northwest Scotland, and the Hawaiian Islands - Emperor Seamounts of the north central Pacific Basin. In a volumetric comparison with the aforementioned magma producing environments, these igneous provinces are insignificant. The cause, or causes, of these

apparently isolated igneous centers or chains is unknown (Gilluly, 1971).

This third class of world-wide volcanism, hereafter called nonorogenic linear volcanism, is characterised by a geometry generally consistent with plate motion. Wilson (1963 a,b,c) first drew attention to lines of oceanic islands as evidence for sea floor spreading. In his early model, volcanic sources, stabilized in the stagnant centres of sublithospheric convection cells, marked an igneous trail on the moving crust above them. Morgan (1971, 1972) has pointed out the geometric symmetry of four volcanic chains in the Pacific: the Hawaiian Islands - Emperor Seamounts, the Tuamotu-Line Islands, the Austral-Marshall-Gilbert Islands, and the seamount chains of the Gulf of Alaska. (see Figure 1-1). A monotonous age-distance relationship is also observed in these provinces. That is, the ages of members of such a volcanic line increase in the direction of plate motion. This relationship has been shown to be remarkably consistent for island chains in the Pacific Basin (Krummenacher and Noetzlin, 1966; McDougall, 1964, 1971; Turner et al., 1973; Jackson et al., 1972; Duncan and McDougall, 1974, 1975; Duncan et al., 1974). But nonorogenic linear volcanism is not restricted to the interiors of crustal plates. In the Atlantic, as noted by Wilson (1963b) aseismic ridges extend from either side of presently active mid-ocean ridge islands (as mirror images of one another), across the ocean floor and onto continental borders. Note for instance the symmetric ridges trailing out from Iceland, St. Helena, and Tristan da Cunha (see Figure 1-1).

Topographically these igneous provinces are prominent, comprising many of the highest volcanoes in the world. Also remarkable in these provinces is the unusually alkaline character of some of the volcanism. Compared with the abyssal, or ocean floor basalts, the chemistry of this

Figure 1-1 The world pattern of hot spots and associated volcanic traces (Mercator projection). Most recently active centres are denoted by solid circles and volcanic traces, identified topographically and geochronologically, are dotted lines. A variety of ages are represented in these commonly proposed linear volcanic provinces:

- (1) Hawaii and the Hawaiian Islands - Emperor Seamounts
- (2) The Marquesas Islands
- (3) Mehetia and the Society Islands
- (4) The Pitcairn-Gambier Islands
- (5) Macdonald Seamount and the Austral Islands
- (6) Easter Island and the Tuamotu Islands (W),
Sala y Gomez Ridge (E)
- (7) Galapagos Islands and Cocos Rise, Carnegie Rise
- (8) The Pratt-Welker Seamounts
- (9) Yellowstone and the Snake River Volcanic Province
- (10) Raton and associated volcanic province
- (11) Iceland and the Wyville-Thompson Ridge (E), Iceland-Greenland Ridge (W)
- (12) Eifel and the Central European Volcanic Province
- (13) Massif Central
- (14) Azores and the New England Seamounts (W)
- (15) Canary Islands
- (16) Cape Verde Islands
- (17) Fernando Po
- (18) Ascension Island
- (19) St. Helena and the Gulf of Guinea Islands
- (20) Trindade - Martin Vas Islands
- (21) Tristan da Cunha and the Walvis Ridge (E),
Rio Grande Rise (W)
- (22) Bouvet and rise to Agulhas Plateau
- (23) Prince Edward Island
- (24) Tibesti and volcanic line to Garian plateau
- (25) Comores Islands
- (26) Reunion and Chagos Plateau to Maldive-Laccadive Islands
- (27) Amsterdam Island and Ninety East Ridge (N),
Kerguelen to Gaussberg Ridge (S)
- (28) Tasmanid Seamounts
- (29) Newer Volcanics of Victoria and SE Australian
Tertiary Volcanics
- (30) Balleny and rise to Tasmania

volcanism suggests a mantle source region which is less depleted in the incompatible elements (Green, 1971). Geophysical measurements (gravity, seismicity, heat flow) are at present ambiguous and do not distinguish these provinces from other stable regions.

Various mechanisms, both dynamic and passive, have been proposed to explain the position and anomalous character of these provinces. Of the dynamic mechanisms, Wilson (1963c) first suggested that linear volcanic features such as the Hawaiian Islands are the result of plate motion over a 'hot spot', a thermal anomaly which is fixed in the mantle and hence produces a volcanic trail on the plate passing above. These volcanic trails, then, record the absolute motion of plates with respect to the Earth's mantle (presumed fixed). Minister et al., (1974) have shown that a collection of presently active hot spots have remained fixed with respect to one another (and presumably to the mantle) for the last 10 Myr. Morgan (1971, 1972) augmented Wilson's hot spot concept by proposing that hot spots are the tops of narrow, upwardly convecting diapirs of mantle material which, owing to their abnormal heat and buoyancy, thermally and mechanically pierce the overlying crust, regardless of the local plate tectonic environment. According to Morgan, the greater part of the upwelling diapir or 'plume' impinges on the bottom side of the lithosphere and spreads out radially in the asthenosphere. It is this feature of the model which allows the possibility that viscous forces exerted on the underside of the lithosphere, by a combination of several such plumes, may initiate rifting and drive the spreading of crustal plates. Modifications of this dynamic model have been proposed by McDougall (1971) who suggests that hot spots may be mobile and travel in an asthenospheric convection pattern at plate velocities, and Anderson (1975) who describes the hot spots as chemical channels through which the Earth, at an earlier stage, 'degassed' and stabilized chemically.

Quite a different dynamic mechanism has been suggested by Shaw and Jackson (1973). They propose a mechanical rather than thermal anomaly to account for such features as the Hawaiian Islands. In their model shear melting, caused by opposed plate and asthenosphere motions, produces a magma reservoir which feeds melt to the lithosphere, leaving a dense residuum which sinks. As this residuum sinks, new material flows in and the process continues. Because the sinking residuum is proposed to stabilize the melting anomaly, Shaw and Jackson term their mechanism a 'gravitational anchor'. (It should be noted, however, that this model requires that the size of the sinking residual bodies be on the order of 100 km radius).

Another class of models suggests that nonorogenic linear volcanism is the result of a passive response of the crust to forces operating from great distances. Betz and Hess (1942) first proposed that the Hawaiian Islands are due to eruption of magma from a major fault in the oceanic crust. Green (1971) first suggested that a zone of tension could be produced interior to a plate which is moving over an irregularly shaped upper mantle. As the lithosphere fails, the short-lived release of pressure will allow partial melting and upwelling from the low velocity zone. Members of such a volcanic province are still expected to exhibit an age-distance relationship and a geometry consistent with relative plate motion. Deep mantle convection is not involved in this model which relies on the persistence of topographic irregularities in the upper mantle to stabilize the point of opening of a propagating fracture. More recently Turcotte and Oxburgh (1973) have explained the mid-plate tensional regime, again using the Hawaiian Islands as an example, in terms of a combination of thermal stresses, due to the cooling of the lithosphere, and membrane stresses, due to changes in the radii of curvature of moving plates. They do not suggest, however, that the opening of propagating fractures produced

in this way should remain stationary relative to the mantle. The tension, then, is purely a product of relative plate motion.

These passive models do not, however, explain the igneous provinces which span either side of the mid-ocean ridges and exhibit the chemical, geometrical and geochronological characteristics of the mid-plate provinces. (See, for example, a discussion of the formation of the Brito-Arctic Province from the Iceland hot spot (Duncan et al., 1972). But how much of the explanation of nonorogenic linear volcanism can be accommodated in a single model? Adherents to the hot spot / plume model propose that mantle diapiric upwelling is responsible for all igneous activity which is or was clearly independent of plate divergence or convergence. Those who advance the propagating fracture model are much less encompassing yet more willing to fit these provinces into the existing structure of plate tectonic theory.

1.2 The aims and extent of this investigation

The present study has been undertaken in an attempt to investigate the character of nonorogenic linear volcanism with regard to the island chains of French Polynesia, in the southeast Pacific Basin. Volcanism in the Pacific Basin is far removed from the borders of the Pacific plate and not associated with the orogenic volcanism of the East Pacific Rise or the circum-Pacific subduction system (see Figure 1-2). Volcanic cones in various stages of evolution (shield volcano to coral atoll to seamount) are arranged in sub-parallel lineaments which are geometrically compatible with the present and past direction of movement of the Pacific plate. Physiographically, at least five island groups can be identified: the Marquesas Islands, the Tuamotu Islands, the Society Islands, the Pitcairn-Gambier Islands, and the Austral Islands (see Figure 1-3). Together with the generally alkaline chemistry reported for shield lavas and suggestions

Figure 1-2: The Pacific Basin. The Pacific plate is the largest and fastest moving contiguous lithospheric plate. In the plate tectonic model, new oceanic crust is produced at the Pacific-Antarctic Ridge, and East Pacific Rise and the Juan de Fuca Ridge (just west of Vancouver, Canada), while older oceanic crust is consumed in the trenches bordering the Pacific plate's western margins. Sea-level is denoted by stippling and bathymetry is indicated by selected 4000 m contour. The present direction of motion of the Pacific plate is shown by blackened arrows. Features in the eastern Pacific enclose the area generated by the present ridge configuration (Herron, 1972). Solid and dashed line segments are transform faults associated with present and past spreading ridges. Sites of active nonorogenic volcanism are marked with stars. (Mercator projection after Pitman et al., 1974).

of increasing volcano ages from the southeast to the northwest, these island groups appear to be good candidates for 'plume' provinces. Indeed, the geometric symmetry of the Tuamotu Islands and the Austral Islands to the Hawaiian Islands and the seamounts of the Gulf of Alaska has been explained by the passage of the Pacific plate over four stationary plumes (Morgan, 1972). The Hawaiian Islands have previously received much attention, both geochronologically and geochemically, but this information has not been available for other island chains in the Pacific Basin.

Because the lithosphere in an oceanic basin is more uniform (both chemically and mechanically) than in continental cratons, the surface manifestation of sub-lithospheric volcanic sources should be far less susceptible to crustal complications. In addition, the Pacific is the largest and among the fastest moving plates (Minster et al., 1974) so that anomalies in geometry will be more clearly defined than on a slower moving plate. These advantages, plus the comprehensive investigations in the Hawaiian Island chain for comparison, make the island chains of French Polynesia an attractive area to study nonorogenic linear volcanism.

Volcanic and plutonic samples were collected in two sessions of field work, comprising a total of 14 weeks. Members of three of the island groups were visited (see Figure 1-3) - the Marquesas Islands, the Society Islands, and the Austral Islands; samples from Pitcairn Island (in a fourth island group) have been kindly provided by Drs. Carter and Coombs of Otago University. (All members of the Tuamotu Islands are coralline at sea level and volcanic material could not be sampled). Chapter 2 reviews geological aspects of the region.

To test the predictions of the stationary source models, a major part of this study has been to determine the rates of migration of volcanism along the four island chains. This was accomplished by the

Figure 1-3 The island chains of French Polynesia (Mercator projection). Bathymetry is illustrated with 2,000 m and 4,000 m contours. The islands are peaks of enormous shield volcanoes which rise more than 4,000 m from the ocean floor. Their linear arrangement suggests a relationship to sea floor spreading at the East Pacific Rise. Structure within island chains appears to be influenced by ancient fracture zones (dotted lines) which are transform faults associated with an earlier orientation of the East Pacific Rise. After Mammerickx et al. (1975).

potassium-argon (K-Ar) whole rock analysis method. Results are presented in Chapter 3, along with a discussion of the compatibility of rates of migration of volcanism with Pacific plate motion.

To characterise the chemistry and differentiation of the shield building lavas and to test the proposition that all members of a given island chain have a common parental magma, both major elements and isotopic compositions (strontium $^{87}/^{86}$) have been analyzed. Chapter 4 discusses this information with regard to vertical and lateral heterogeneities in the upper mantle.

Geophysical techniques have not been successful in unambiguously determining anomalies beneath proposed hot spots. The suggestion, however, that anomalous geomagnetic behaviour may be due to chemical or thermal variations deep in the mantle under Hawaii (Doell and Cox, 1971) could be checked for the French Polynesian region by collection of oriented samples in the field. Chapter 5 reviews some of the attempts (particularly in oceanic regions) at geophysical constraints on the source modelling, and discusses the implications of a study of palaeosecular variation in the French Polynesian region as determined from analysis of magnetic directions from Society Islands' sites.

Chapter 6 integrates the conclusions of previous chapters and discusses the oceanic island volcanism in French Polynesia and its setting in the tectonics of the Pacific Basin. Using this information about motion of the Pacific plate with respect to underlying volcanic sources, whatever their nature, the world-wide pattern of plate motion and attending nonorogenic linear volcanism is tested for consistency. The nature of the anomalies leading to the volcanism in French Polynesia is still unresolved, though evidence presented here favours a dynamic mechanism.

The analytical techniques used and the reduction of data are described in Appendix 1. Geographic and petrographic descriptions of samples selected for geochronology and/or geochemistry are presented in Appendix 2. References will be made often to articles previously published on the work discussed here and these are available in Appendix 3.

Before proceeding with discussion of the analytical work of this investigation, a brief account of the geologic and tectonic setting of the island chains of French Polynesia is in order.

Chapter 2: The Geologic and Tectonic Setting

2.1 The Pacific Basin

The theory of plate tectonics proposes that the Earth's outer shell (the lithosphere) is a zone of strength and elasticity, which is resistant to shearing stresses, and overlies a malleable zone of weakness (the asthenosphere). All earthquakes occur within the lithosphere, which behaves as a stress-guide and deforms only along the boundaries of its constituent pieces, called 'plates'. This difference in the rheological properties of the lithosphere and asthenosphere allows horizontal movement of plates with respect to one another and to the mantle below the asthenosphere.

The Pacific plate has the largest area of the dozen or so commonly proposed contiguous lithospheric units (Morgan, 1968; Le Pichon, 1968). The boundaries of the plate (see Figure 1-2) can be identified bathymetrically but a more precise demarkation is provided by seismicity (Barazangi and Dorman, 1969). To the east, the Pacific plate is spreading away from the Nazca plate and the Cocos plate at the East Pacific Rise and sliding northward against North America along the San Andreas - Queen Charlotte Fault Systems. To the north and west, convergence of this oceanic plate with other oceanic and continental plates is marked by a system of trenches and calc-alkaline volcanism from the Aleutian subduction zone through the Japan and Philippines trenches to the island arcs of the New Hebrides, Fiji and the Tonga-Kermadec Islands. Strike-slip faulting along the Alpine Fault of New Zealand separates the Pacific from the Austral-Indian plate and the Pacific-Antarctic spreading ridge marks the southern extent of this immense area of crust. Shallow focus earthquakes at the spreading ridges indicate a tensional environment, whereas the deep earthquakes associated with subduction zones suggest compression (Isacks et al.,

1968). Seismicity at the plate's borders, then, implies motion of the Pacific crust from the spreading centres to the trenches, SE to NW.

Seismic refraction studies have led to a rather uniform picture of the vertical structure of an 'average' oceanic plate, of which the Pacific is certainly an example. Combining surface wave and shear wave data, Kanamori and Press (1970) have shown that the base of the oceanic lithosphere is marked by a rapid decrease in seismic velocities at about 70 km depth, indicating the boundary between the lithosphere and the underlying asthenosphere (low velocity zone). (Attenuation of body waves, p- and s-waves, determines the asthenosphere to lie between 70 and 200 km depth.) The uppermost section of the lithosphere, above the Mohorovicic discontinuity, is called the crust.

The seismic structure of 'average' Pacific Basin crust has been investigated by Shor et al. (1970) from refraction studies. Below the sedimentary blanket the igneous crust is made up of two seismically distinct layers: a 'transitional' layer with p-wave velocity about 5.1 km/sec and average thickness of 1.7 km, and an 'oceanic' layer of velocity 6.7 km/sec and thickness between 3 and 6 km. The Mohorovicic discontinuity (or Moho) separates the oceanic layer from underlying material with p-wave velocity about 8.1 km/sec. Maynard (1970) suggests a more gradual increase in seismic velocities through the oceanic layer. The Moho, except near the ridges and the trenches, is very clearly defined seismically. This discontinuity has been alternatively explained as a phase transition between basalt and eclogite (revived by Lovering, 1958; Kennedy, 1959), or as a chemical discontinuity between either serpentinite and peridotite (Hess, 1962), or basalt and some related ultramafic assemblage (Ringwood and Green, 1966; Green, 1971).

Heat flow, away from the trenches and spreading ridges, is about the same as average continental heat flow, ~ 1.5 H.F.U. Analysis of

satellite gravimetry indicates that the Pacific Basin is close to isostatic equilibrium (Kaula, 1970). Seismicity within the plate is almost unknown but rare events indicate a compressional environment (Sykes and Sbar, 1973).

Ocean depths over the basin range from 4 to 6 km; depths shallow to 2.5 km over the East Pacific Rise and deepen to 8 to 10 km in the trenches of the western Pacific. Sedimentary cover increases in thickness from the spreading ridges to the trenches, in keeping with the increasing age of the oceanic crust. Pelagic sediments are derived from many sources but the common types are clays and calcareous or siliceous organic oozes. Rates of deep-sea sedimentation lie in the range of 0.3 to 1.2 mm/10³ yr (Opdyke, 1969). The assumption that new oceanic crust accumulates sediment from the moment of its production at the spreading ridge allows determination of the age of that crust from the palaeontology of the oldest overlying sediment. Fischer et al., (1970) report Jurassic ages for the oldest sediments in the northwest Pacific Basin (collected by The Deep Sea Drilling Project), and progressively younger ages to the east.

Since the discovery of linear magnetic anomalies recorded in the oceanic crust (Mason, 1958) and their interpretation as evidence of sea-floor spreading (Vine and Matthews, 1963; Vine, 1966; Heirtzler et al., 1968) the history of the accretion of the Pacific plate has begun to unravel. By assigning a geochronological age to each of the distinctive reversals in the magnetic record (Cox, 1969), rates of sea-floor spreading from the East Pacific Rise may be determined.

Herron (1972) and more recently Mammerickx et al. (1975) have studied the magnetic anomaly patterns recorded in the young oceanic crust on either side of the East Pacific Rise. From correlation of

Figure 2-1

The East Pacific Rise - transform faults and magnetic anomalies (Mercator projection). After Mammickx et al. (1975). The site of present spreading is indicated by dotted line segments. Present transform faults intersect these spreading sections at right angles. Between 6 and 16 Myr B.P. the East Pacific Rise had a more north-northwesterly orientation. Associated transform faults and magnetic anomalies from this previous episode can be correlated. (Primed and unprimed letters indicate matching transform faults.)

identified anomalies, Herron (1972) finds that spreading rates vary along the ridge, which has changed both its orientation and position during the past 50 Myr. The East Pacific Rise is broad and smooth along most of its length. A central narrow continuous ridge is offset in places by transform faults (Wilson, 1965).

Present half spreading rates (Pacific plate relative to the East Pacific Rise) increase southward from 4.5 cm/yr at 15°N to 9.0 cm/yr at 30°S . This variation, however, is in accord with the clockwise rotation of the Pacific plate about the instantaneous Pacific pole of rotation. (The most recent estimate of its position is 67°N , 59°W by Minster et al., 1974.) Herron (1972) notes that the depth to the Rise increases northward as the spreading rate decreases.

The present spreading geometry at the East Pacific Rise has not persisted through much of the history of Pacific plate accretion. Both Herron (1972) and Mammerickx et al. (1975) identify an older spreading ridge, oriented north-northwest, which produced Pacific crust prior to about 10 Myr ago. The very prominent fracture zones of the Eastern Pacific Basin (Clarion, Clipperton, Marquesas, etc. Fracture Zones) trend west-southwest and are perpendicular to this older spreading centre. These are older transform faults which were generated prior to 10 Myr by relative motion between the Pacific plate and the deceased Farallon plate (McKenzie and Morgan, 1969). According to Mammerickx et al. (1975), the present East Pacific Rise grew northward, opening at 30°S about 16 Myr B.P. and at 5°S about 6 Myr B.P. Fracture zones associated with this present spreading ridge intersect the older fracture zones at about 35° , the same angle as separates the two ridge axes (see Figure 2-1). The eastern spreading centre, the site of Pacific plate accretion, may, then, vary both its orientation and rate of spreading.

The composition of the igneous crust of the Pacific plate is

inferred from dredged and cored samples (most comprehensively by the Deep Sea Drilling Project's Glomar Challenger). These samples are largely basalts but some rhyolite, tuffaceous rocks and gabbroic rocks are also found (Friesen, 1973). The predominant basalt is an olivine tholeiite, although alkali olivine basalt and high alumina tholeiites are not rare. With wider and more intensive sampling, the initial belief of great uniformity in chemistry of ocean floor basalts is diminishing. Nevertheless the petrologic association of the ocean ridges, dominated by olivine tholeiite, is characterised by very low K_2O and TiO_2 , low total iron and P_2O_5 , and high CaO ; trace element abundances are low, K/Rb is very high, Th/U is low and Sr^{87}/Sr^{86} is low. These chemical characteristics can be explained by Green's (1971) petrologic model of oceanic crust creation (Figure 2-2).

Beginning with rifting of the lithosphere, the asthenosphere (low velocity zone) of pyrolitic* composition is tapped. This is a region of <5% partial melting, the magma being highly undersaturated and volatile rich. Small amounts of olivine nephelinites and olivine melilitites result from tapping this source (80 to 120 km), leaving a relatively depleted pyrolitic source region. As crustal thinning progresses, more and more of the mobile low velocity zone material upwells, partially melting in greater percentages (>5%) at shallower depths, yielding basanites, alkali olivine basalts, olivine basalts and olivine tholeiites. In the 'mature' stage of a mid-ocean spreading ridge, Green proposes that partially depleted pyrolite from the low

* pyrolite (Green and Ringwood, 1967) is an hypothetical assemblage of 3 parts harzburgite (olivine, enstatite, chrome spinel) and 1 part Hawaiian olivine tholeiite.

Figure 2-2 A model for the creation of oceanic crust, after Green (1971). The lithosphere is composed, in part, of residual peridotite. The pyrolitic low velocity zone (LVZ) underlies the lithosphere which is chemically partitioned into an upper volatile-rich zone and a lower zone depleted in volatile and incompatible elements by partial melting. (webbed region in each of the three drawings).

(1) The crust and lithosphere begin to rift and undersaturated magmas, characteristic of small degrees of partial melting, reach the surface, leaving a region of (relatively) depleted pyrolite in the upper LVZ.

(2) The crust and lithosphere thin further and material from the LVZ upwells. In addition to continued volcanism of stage (1), magmas produced by >5% partial melting at <60 km depth of upwelling LVZ material will yield alkali olivine basalts, olivine basalts and olivine tholeiites.

(3) In the 'mature' stage of new oceanic crust production, pyrolite from the lower LVZ and residual pyrolite from stage (1) will upwell to shallow depths (< 30 km) and 20-25% partial melting will produce the chemistry of typical ocean floor basalts.

velocity zone upwells to depths of 30 km or less and 20 to 25% melting produces high alumina olivine tholeiites. The two-stage melting (<5% melting of the pyrolite parent, followed by 20-25% melting of the more refractory portion) will lead to the very low concentrations of the incompatible elements (K, Ti, P, Ba, Sr, Rb, Zr, U, Th, Pb) found in abyssal tholeiites.

Scattered across this chemically and structurally uniform lithospheric plate, like sentences in Braille, are the conspicuously linear chains of volcanic islands and seamounts.

2-2 The volcanic lineaments of French Polynesia

Four island groups in French Polynesia are predominantly volcanic at sea level: the Marquesas Islands, the Society Islands, the Austral Islands and the Gambier Islands (of which Pitcairn Island is physiographically, but not politically, related). Another group of islands, the Tuamotus, are all coral atolls, some of which have been uplifted (e.g. Makatea). The region is administered from Tahiti and covers an area slightly less than that of western Europe (see Figure 2-3). The islands themselves are only the peaks of enormous volcanic edifices which rise as distinct cones from depths exceeding 4,000 m to a maximum subaerial elevation of 2,200 m (on Tahiti). Thus very little of the total volcanic pile is exposed and one very big assumption in petrologic studies is that subaerial samples are representative of the entire volcano.

So curious is the arrangement of island members into nearby parallel chains that from earliest geologic studies (Chubb, 1930) the position of these islands was thought to be tectonically controlled, either by ancient crustal fractures or by compression into fold belts. From the earliest visits of naturalists to the islands, geomorphologic

Figure 2-3 French Polynesia superimposed on Europe (equal area projection).

progression from high volcanic cone to coral atoll has been thought to indicate an increase in age from the southeast to the northwest ends of the island chains. Confirmation of this position has come recently from geochronology (discussion in Chapter 3).

The age of the ocean floor beneath French Polynesia has been determined, from palaeontology of basal sediments (Tracey et al. 1971) and magnetic anomaly correlation (Herron, 1972), to be between 45 and 65 Myr. As will be discussed in Chapter 3, these islands are considerably younger than the crust on which they sit. The extensions of these linear features are approximately perpendicular to the present axis of the East Pacific Rise, suggesting that the geometry of the island chains is partially controlled by the recent spreading history of that ridge.

The Marquesas, Tuamotu, Austral and other unnamed fracture zones (Figure 1-3) are transform faults generated by spreading at the ancient East Pacific Rise (axis north-northwest). These lines of lithospheric weakness intersect the trend of the French Polynesian island chains at an oblique angle ($\sim 35^\circ$) and may exert a structural control on the distribution of islands within the various groups.

This, then, is the tectonic setting of the island lineaments of French Polynesia: volcanic spires planted upon a relatively old and fractured basaltic crust which has endured a number of changes in direction and rate of drift from an inconstant spreading ridge to the east.

Rather than review the geology of each and every island in French Polynesia, the geology of Tahiti (which has been most completely studied) will be discussed in detail. Differences with other islands will then be noted in general remarks about each of the four island groups.

2-3 The Geology of Tahiti

a. Geomorphology

Apart from the Hawaiian Islands, no oceanic island of the Pacific Basin has recieved more scientific attention than Tahiti, the administrative centre of French Polynesia. Following Cook, the island has been visited by many of the eminent naturalists of the last century: Darwin, Dana, Murray and Agassiz. In the early part of the 20th century, the first petrologic studies were undertaken by Lacroix (1904, 1910), Marshall (1911, 1913, 1915) and Iddings (1916). These were of very restricted coverage and it was not until 1933 that Williams published the first comprehensive account of the geology and petrography of Tahiti. Recent contributions by French government geologists (Obellainne, 1955; DeNeufbourg, 1965) have concentrated on the structure of the volcanic edifice, while McBirney and Aoki (1968) have studied the petrology of the shield lavas and their plutonic equivalents.

By far the largest island of French Polynesia, Tahiti consists of two adjacent shield volcanoes, Tahiti-Nui to the northwest and Tahiti-Iti (Taïarapu) to the southeast (see Figure 2-4). Tahiti makes up more than half of the total area of the Society Islands. The larger of the two cones, Tahiti-Nui, is roughly circular in plan, with diameter 30 km and a maximum elevation of 2241 m (Mt. Orohena). Taïarapu is elliptical, with major axis along the line of the two volcano centres. Its highest point is Mt. Roniu, 1332 m, and its northeast to southwest minor axis 14 km across. A narrow isthmus joins the two volcanoes which rise together from oceanic depths of 4,000 m. Williams (1933) has estimated that the original shield of Tahiti-Nui rose to a height of about 2,800 m and, together with its submarine base, occupied a volume of $60,000 \text{ km}^3$, loaded on the ocean floor.

Figure 2-4 The Island of Tahiti (Mercator projection). The Island has been built by two shield volcanoes, Tahiti-Nui (the larger) and Tahiti-Iti (also called Taïarapu). Caldera regions are indicated by a dashed and notched symbol. Plutonic rocks, which outcrop in the floors of the two calderas, are blackened regions. The dashed line intermittantly following the perimeter of the island is the seaward extent of coral growth. Sample localities are marked.

A great deal of the original volcano survives at Tahiti-Nui, cut like wedges from a wheel of cheese by radiating rivers, flowing from the rugged centre. Three morphologic divisions have been suggested by DeNeufbourg (1965): the caldera, whose 2000m walls bristle with spires and pyramids, the 'planeze' or conical surface of the shield volcano, and the littoral plain, separated from the planeze by coastal cliffs.

The caldera remains intact except for a break in the northern wall, forming a great central amphitheatre which the Papenoo River drains. It is in the eroded caldera floor that Tahiti's remarkable plutonic rocks outcrop. Before caldera collapse these rocks would have cooled under a volcanic pile of 2000 - 2500 m thickness. The slope of the planeze, 8-10 degrees, is quite steep compared to the Hawaiian volcanoes.

As mentioned previously the original cone is dissected by deep and narrow valleys, creating extremely precipitous topography, especially close to the caldera rim. Between two such river valleys, the planeze narrows from a broad coastal apron to a knife-edge arête, often no more than a metre across. The intercalation of resistant and porous flows allows, through differential erosion, the very steep topography of the upper river valleys (Williams, 1933). Weathering has reduced the upper surface of the planeze to laterite (called 'mamu'), yet this is bound by a dense vegetation which prevents severe erosion and maintains the constructional shape of the original shield.

An almost continuous line of coastal cliffs surrounds the island of Tahiti. According to Crossland (1928), cliffs were formed before the establishment of the barrier reef which today encompasses the island and its peninsula, with intermittent breaks. (Along the north coast cliffing

is again taking place because of the inundation of the coral reef there). The coastal cliffs reach heights of 300 m but are not generally higher than 150 m. Where cliffs are now protected by the barrier reef, a littoral plain, generally less than 1 km in width, has covered older parts of the fringing reef.

Both in the river valleys, whose rock floors are likely to be much deeper than present river level, and in the benches cut in the coastal cliffs (e.g. a conspicuous shelf 2 m above present sea level) the island has recorded changes in the relative level of the ocean.

The coral reef surrounding Tahiti has been a focus of interest from Darwin (1842) to the present day. Darwin proposed that the evolution from fringing reef to barrier reef to coral atoll could be easily explained by gradual subsidence of the volcanic pile on which the coral grows. The Society Islands provide an excellent example of this evolution from Mehetia, which has no development of coral, through Tahiti, whose reef is a combination of fringing and barrier, to Raiatea-Tahaa and Bora Bora, which have well developed barrier reefs, finally to Motu One, a coral atoll.

The presence of drowned river valleys and ancient coastal cliffs is interpreted as evidence of the subsidence of Tahiti. Other observers do not believe that the island has subsided but that the coral have grown radially outward on a foundation of its own talus (Murray, 1895), or on a base of calcareous algae which grow at suitable depth (Setchell, 1926). Rapid loading of the thin oceanic crust by such a tremendous volcanic pile, however, will lead to subsidence, as has been shown for the island of Hawaii (Moore, 1970). The reef around Tahiti is at most 2 km wide and less than 300 m thick, which is consistent with a subsidence rate of $1/5$ to $1/3$ mm/yr and the age of the island (Chapter 3).

b. Volcanic and Plutonic Rocks

The predominant part of both Tahiti-Nui and Tahiti-Iti is lava flows; pyroclastic material forms a very small fraction of the total volume. The thickness of these flows varies considerably. In sections exposed along the coast, flows range from less than 1 m to 20 m. Thickness increases toward the calderas and there is no doubt that the bulk of lavas and pyroclastics were erupted adjacent to or at the islands' summits. Although rare, vertical, radial dykes, more resistant to weathering than lava flows, are exposed in the narrow ridges approaching the caldera. A few subsidiary cones have been located (DeNeufbourg, 1965) but these cannot have been a significant source of volcanic material.

Compact and scoriaceous zones alternate in the exposed cliff sections, the former often exhibiting columnar jointing while the latter are usually heavily weathered. This alternation of compact and rubbly zones has been explained (Macdonald, 1972) by the observation that a lava flow, when erupted, will freeze on both top and bottom so that as it moves these two surfaces will constantly be broken up into scoriaceous rubble, but the centre of the flow will eventually freeze into a more compact and massive texture. In addition to this predominant 'aa' type of flow, ropy, undulating lava flows of the 'pahoehoe' type can also be seen in cliff sections. No detailed mapping of individual flows has been undertaken (due principally to the very dense vegetation), but exposures in coastal sections indicate a range in lateral extent of from about 50 m to many kilometres.

Apart from the shield building lavas, small volumes of valley-filling lavas ('formations de remplissage des vallées') have been identified (DeNeufbourg, 1965) which overlie stream deposits of mud and cobbles. These formations, which post-date the shield lavas, reach thicknesses of several hundred metres and must have had much greater

Figure 2-5 A palaeomagnetic polarity study at Tahiti (Mercator projection). Both normal and reversely magnetised flows were discovered using a field magnetometer. From geochronological study (Chapt. 3) it was found that Tahitian volcanism spanned the Brunhes (N) - Matuyama (R) magnetic boundary. The shaded region indicates the located or inferred (dashed line) reversal boundary (age ~ 0.69 Myr). Material younger than 0.69 Myr lies above or outside the shaded region. (No reversely magnetised material was found at Taiarapu.) The shaded region also gives a rough indication of the size of Tahiti-Nui at 0.69 Myr B.P.

Cliff outcropping in the Punaruu Valley. Layers of 'aa' flows vary in thickness between 0.5 and 2 metres. Brunhes-Matuyama reversal boundary occurs across the thin layer on which the field assistant is standing.

geographic extent before erosion. Field palaeomagnetic measurements (discussed in Chapter 5) have determined the coastal distribution of normally and reversely magnetized flows (the Brunhes-Matuyama reversal boundary) and so give one time contour on the structure of the larger volcano (see Figure 2-5).

Nomenclature is always confusing in petrologic discussion, especially in alkalic provinces. Table 2-1 presents the system to be used in this discussion, based on Irving (1971) and McBirney and Aoki (1968). This alkalic spectrum is divided using chemistry, normative mineralogy and a number of indices derived by Coombs and Wilkinson (1969).

Petrologically the most abundant lavas are alkali olivine basalts, basanites and ankaramites (alkali olivine basalts unusually rich in olivine and clinopyroxene phenocrysts). Williams (1933) estimates that these lavas account for over 95 percent of the Tahiti-Nui volcano. Members of the differentiation series hawaiite-mugearite-trachyte-phonolite are also represented but are, volumetrically, subordinate to the more mafic shield-building lavas. The common lava types are ne- to hy- normative, but vary texturally and mineralogically. Olivine and titanaugite are abundant phenocrysts in the shield lavas; aegirine augite occurs in the highly differentiated members (trachytes and phonolites). Feldspar is rarely seen as a phenocryst phase, while orthopyroxene is entirely absent. Biotite has been observed occasionally but hornblende is very scarce. Noted by previous researchers is the absence of melilite, for the reason that silica deficiency has not been great enough to precipitate melilite in place of calcium-rich pyroxenes (McBirney and Aoki, 1968). The groundmass phases are commonly labradorite, titanaugite, olivine, ilmenite and apatite.

Lacroix (1910), Williams (1933) and McBirney and Aoki (1968) have

Table 2.1: Nomenclature of basaltic rocks and their plutonic equivalents (from Irving, 1971; McBirney and Aoki, 1968).

$\frac{an}{an+ab} \times 100$	D.I.*	hy	hy >2%	hy<2%, ne<0.5%	0.5<ne<5%	ne>5%
44	<65	tholeiite	olivine basalt	(alkali gabbro)	alkali olivine basalt	basanite
30-44	<65	(gabbro)			(theralite)	(essexite)
					hawaiiite	ne-hawaiiite
					(monzonite)	(ne-monzonite)
10-30	<65				mugearite	ne-mugearite
	65-75				benmoreite	ne-benmoreite
	<75				trachyte	phonolite
					(syenite)	(ne=syenite)

* D.I. = Σqz , or, ab, ne, lc

all remarked on the absence of ultramafic xenoliths (or 'basic enclaves' in the older literature) in the Tahitian lavas. This, however, is not the case, as several alkali olivine basalt flows on Tahiti-Iti contain large quantities of peridotitic nodules. Xenolith-bearing lavas were found on other islands of the Society group.

The differentiated members of the petrologic association exhibited at Tahiti are most common on the smaller volcano, Tahiti-Iti, and rare on Tahiti-Nui. Trachytes are pale gray with a satiny sheen and characteristically weather to a chalky white. These are exceptionally fine-grained, holocrystalline, feldspathic rocks with small aegirine augite phenocrysts. Outcrop is restricted to single flows which, because of their great thickness and small lateral extent, must have been very viscous lavas. Their cleavage is due to flow-banding of the alkali feldspar groundmass. Phonolites are found on both volcanoes and differ from trachytes in containing nepheline and sodalite.

Rocks of plutonic aspect outcrop, with very little topographic expression, in the floors of both the Tahiti-Nui and Tahiti-Iti calderas. Marshall (1913) was first to describe the field occurrence of these rocks, although their presence was known by Dana (1849). Now new chemistry is exhibited by the plutonic specimens at Tahiti - all are direct equivalents of the surface lavas (Williams, 1933). There are, of course, extreme differences in mineralogy due to the very different conditions of crystallization. Excellent petrographic descriptions exist for the plutonic spectrum (Lacroix, 1910, Williams, 1933; McBirney and Aoki, 1968). The main plug of the Tahiti-Nui caldera consists of nepheline monzonite and theralite with marginal nepheline syenite. Specimens from the complete range of syenite to peridotite have been described. In addition to olivine and titanaugite, plagioclase

and orthoclase are common, as well as biotite and amphibole (kaersutite). Apatite is unusually abundant in some specimens and common throughout, as are the opaque minerals. Nepheline is also present, but not universally. The range of plutonic rocks described from Tahiti-Iti plug is more restricted and generally more mafic (perhaps inadequate sampling).

Aside from the great difference in size of phenocrysts between the volcanic and plutonic chemical equivalent, amphibole and biotite are much more abundant in the plutonic rocks, at the expense of olivine. That water and other volatiles have been present under pressure during the cooling of the plutonic core is also evidenced by significant amounts of extraneous argon trapped in such plutonic specimens, yielding erroneously high whole rock K-Ar ages (Krummenacher and Noetzlin, 1966). The plutonic core of the volcano is, of course, contemporaneous with or younger than the shield lavas.

Stark and Howland (1941), from work on the dissected volcano at Bora Bora, suggest the following sequence of events for the evolution of a Society Island:

- (1) Shield formation by vast outpourings of vesicular basaltic lavas;
- (2) Subsidence and resurgence of the magma supply, leading first to cooling and jointing of the shield, followed by radial dyke emplacement along vertical cracks. Those authors identify at least four episodes of resurgence at Bora Bora;
- (3) Intrusion of non-porphyritic, basaltic dykes, cutting across the radial pattern in places. (Trachytic intrusions and flows succeed this phase);

CCopra freighter in the port of Papeete. Accommodation and transport during field work were provided by these ships in the Marquesas and Austral Islands.

A view of the interior of Tahiti from the Punaruu Valley.

- (4) Regional faulting which often removes large portions of the shield edifice;
- (5) Erosion, subsidence and reef building.

With this model (and specific evidence from Tahiti) in mind, then, a brief review of the other volcanic islands of French Polynesia is in order.

2.4 The Society Islands

Petrologically Tahiti is the most highly developed member of the Society Islands, but is generally representative of the geology of her sister islands. The volcanic members of this island group stretch from Mehetia ($17^{\circ}53'S$, $148^{\circ}04'W$) north-westward to Maupiti ($16^{\circ}26'S$, $152^{\circ}16'W$), a distance of 500 km. The lineament continues another 400 km to the northwest as scattered coral atolls and seamounts (see Figure 2-6).

a. Mehetia

Mehetia is the southeasternmost island of the Society Island Chain. While no geochronological work has been done on rocks from this island, Williams (1933, p. 11) states that "there can be no question that the youngest volcano in the Society Islands is ... Mehetia". The island is small and roughly circular (2 km diameter), reaching a respectable elevation of 435 m. A fringing reef is developing but is still patchy, so that coastal cliffs, 10 to 30 m bound the island. Lavas and pyroclastics are present. All specimens which Williams (1933) reports are basaltic - no trachytes or phonolites have been found. A seismic array on Tahiti has located shallow earthquakes south of Mehetia, at $18.4^{\circ}S$, $148.6^{\circ}W$, a known topographic high (Talandier, personal communication). The southeastern end of the Society Islands, then, is still very active. To the northwest of Mehetia lie the paired cones of Tahiti. Next in line is Moorea, 16 km northwest of Tahiti.

Figure 2-6

The Society Islands - spacing and structural control. Volcanic centres, denoted by open circles, appear to lie along orthogonal lineaments within the broader orientation of the island chain. These lineaments correspond to transform faults and the spreading axis associated with an earlier orientation of the East Pacific Rise. This ancient basement structure appears to have influenced the distribution and erosion of shield volcanoes within the Society Islands. (Mercator projection).

b. Moorea

The volcanic centres of Moorea, Tahiti-Nui and Taiarapu are approximately equidistant from one another. Moorea is a more heavily dissected cone than Tahiti-Nui, but still partly preserves the constructional form of a shield volcano. A significant feature of the northern coast is a normal fault which has down-faulted approximately one third of the cone. The strike of this fault is about 260° (west-southwest) which is very close to parallel to the ancient transform faults which underlie the islands. Coastal cliffs are prominent on the north coast but are rarely significant on the south coast, an observation ascribed by Williams (1933) to tilting of the island toward the south following the massive faulting of the northern part of the island. Fringing to barrier reefs circle the island, broken only at 'passes' opposite to two great bays opening to the north at the mouths of fresh water streams. The central caldera region opens to the north and is surrounded by a wall of castellated peaks, the highest of which is 1207 m.

Petrologically Moorea does not differ significantly from Tahiti. All varieties of Tahitian lavas have been found on Moorea. Differences exist in the percentages of rock types, however. Trachyte, for example, forms a greater part of Moorea than of Tahiti (see Figure 2-7), yet the common shield lavas appear to be less silica deficient on the smaller island. Gabbroic rocks occur as boulders at the center of the caldera floor, the main plutonic body presumably buried beneath some thickness of laterised soil.

c. Maiao

To the southwest of Moorea lies the small island of Maiao, consisting of a volcanic core (maximum elevation 154 m) surrounded by

Figure 2-7 The island of Moorea (Mercator projection after DeNeufbourg, 1965). Moorea is the eroded remnant of a single shield volcano whose northern half has been removed through faulting. The remains of a caldera wall are denoted by the dashed and notched line. Intrusive trachytic bodies are indicated by horizontal lines, and the perimeter of the volcanic shield is stippled. The solid line outside the shield is the littoral beach, sand and laterised soil. The seaward extent of the coral reef is a dashed line. Sample localities are marked.

shallow lagoons and a barrier reef. The very minor outcrop of volcanic material does not allow definitive statements about the petrology of the volcano (Williams, 1933). No trachytes or phonolites have been reported and the predominant lava type is an olivine basalt. Samples were not collected from this island for this study.

d. Tetiaroa

To the northeast of Moorea lies the coral atoll of Tetiaroa ($17^{\circ}00'S$, $149^{\circ}33'W$). No volcanic rock appears at the surface although the edifice on which the coral has grown is surely volcanic. The island's geomorphologic state suggests great age, making it anomalous with respect to its young neighbours but another explanation for its very low subaerial relief is that the volcanic cone never grew much above sea level so that a little erosion and coral growth have produced a coral atoll. Or, possibly, Tetiaroa is a much more ancient feature of the Pacific Basin which has been 'overrun' by the young Society Islands lineament.

e. Huahine

The island of Huahine ($16^{\circ}47'S$, $151^{\circ}00'S$) consists of a pair of twinned islets encircled by a common barrier reef. Actually, the two islets are halves of the original cone which has been severed by faulting (DeNeufbourg, 1965). The fault strike is again west-southwest, Figure 2-8). In spite of extensive erosion the dislocated caldera region can be identified, an oblong basin central to the deeply embayed shield remnant. None of the original constructional surface (the planèze) of the volcano is preserved. Numerous short valleys drain the island's interior peaks, the highest of which is 670 m.

The shield building lavas on Huahine are similar to those of Tahiti: alkali basalts, basanites and ankaramites (DeNeufbourg, 1965,

Figure 2-8 The island of Huahine (Mercator projection after DeNeufbourg, 1965). Two islets within a common barrier reef form Huahine. The position of trachytic intrusions (horizontal lines) and the fragmentation of the original shield volcano may be fault controlled. The dashed and dotted straight lines are the structural lineaments prevalent in the underlying ocean floor. Other symbols as in Fig. 2-7.

see Chapter 4). Trachytic rocks are far more conspicuous here, however, forming four massive hills on the northern islet. The position of these bodies (intrusive to flows) is fault controlled and postdates the construction of the shield. These trachytes contain phenocrystic aegirine and nepheline and grade into phonolites.

Gabbroic rocks are reported by DeNeufbourg (1965) from the caldera basin on the northern islet, but he remarks that the main plutonic core is probably submerged beneath the bay which separates the two islets.

f. Raiatea - Tahaa

The twinned islands of Raiatea and Tahaa lie between $16^{\circ}33'S$ and $16^{\circ}55'S$ and $151^{\circ}19'W$ and $151^{\circ}34'W$. A diminutive version of Tahiti, these islands are bounded by a common barrier reef. Raiatea is second only to Tahiti in size among the Society Islands. Its triangular shape suggests structural control by faulting (Figure 2-9). The direction of elongation of Raiatea is coincident with the strike defined by the calderas of Tahaa and Raiatea, about 350° . The calderas are now only mild depressions in the undulating topography. Slopes are gentle and rise to a maximum elevation of 1033 m. Both islands are surrounded by narrow littoral plains which continue inland as floors of numerous drowned stream valleys, evidence of the islands' recent subsidence.

The shield building lavas are the Tahitian types and are, at the surface, in an advanced stage of weathering to lateritic soil ('mamu'). Fresh material is found in road cuttings and in stream-cut cliffs. Vertical dykes radiating from the caldera region are seen much more often on these islands than on Tahiti. Trachytic flows and intrusives are again conspicuous on Raiatea, but almost entirely lacking on Tahaa.

Figure 2-9 The islands of Raiatea and Tahaa (Mercator projection after DeNeufbourg, 1965). These two islands are separate shield volcanoes enclosed in a common barrier reef. The irregular shape of Raiatea suggests structural control as does the alignment of the islands along an old line of crustal weakness. Symbols as in Fig. 2-7.

DeNeufbourg (1965) divides the shield lavas into two groups on the basis of trachytic flows outcropping on the northern slopes of Raiatea. The two episodes of basalt volcanism do not differ petrologically and cannot be distinguished without the trachyte "time contour". Plutonic rocks outcrop in the caldera basins of both islands. These are biotite gabbros containing interstitial nepheline.

g. Bora Bora

Situated at $16^{\circ}30' S$, $151^{\circ}45' W$ is the small island of Bora Bora. Very little of the original volcano remains (Figure 2-10). The outline of the caldera can still be traced in the high arcuate ridge surrounding Vaitape and Teavanui Harbour, completed by the islet of Toopua. Narrow ridges radiate out from this caldera rim to the northern reaches of the island. Darwin (1842) chose this island as the type barrier reef island, intermediate between the fringing reef and atoll stages.

The location of the caldera is confirmed by the pattern of vertical dykes, radiating away from that area. The ridges forming the backbone of the island are supported by these closely packed dykes (Stark and Howland, 1941). Faulting and submergence are believed to have led to the very irregular form of the volcanic coastline. The western half of the island has been downfaulted approximately 600 m (Stark and Howland, 1941). The strike of the predominant faulting is again $\sim 350^{\circ}$ and 260° , ancient lines of weakness in the Pacific crust.

Stark and Howland (1941) describe the lavas of Bora Bora as all olivine-bearing grading from alkali basalt to ankaramite (some varieties called oceanites contain as much as 50% olivine). Dyke rocks exhibit the same narrow range in petrology. No trachytes or phonolites are found on the island. It may be that erosion has removed these differentiates or that differentiation never progressed to the acid members of the

Figure 2-10 The island of Bora Bora (Mercator projection after DeNeufbourg, 1965). The embayed remnants of the shield volcano of Bora Bora comprises two islets. The location of the caldera is indicated by the radial pattern of closely packed vertical dykes (Stark and Howland, 1941), which surrounds a deep lagoon harbour at Vaitape. Symbols as in Fig. 2-7.

Tahitian sequence. Coarse grained doleritic equivalents of the common basalts are present as dykes in the wall surrounding the caldera, but no rocks of a plutonic core have been discovered (due to the fact that their probable location is submerged beneath Teavanui Harbour).

h. Tupai

The coral atoll Tupai ($16^{\circ}14' S$, $151^{\circ}50' W$) lies 25 km northwest of Bora Bora. No volcanic material outcrops on the island which resembles Tetiaroa.

i. Maupiti

The island of Maupiti is the westernmost volcanic member of the Society Island Chain, located at $16^{\circ}26' S$, $152^{\circ}15' W$. No geological study, previous to this investigation, has been reported. This eroded remnant is roughly circular, with diameter ~ 1.5 km, and is surrounded by a barrier reef of diameter 4 km. The most prominent feature of the island is a high precipitous ridge which runs from the center to the southeast corner, reaching a maximum elevation of 380 m. In addition to this central ridge, several spires of the horizontally layered flow sequence emerge from the heavily vegetated hillsides as conspicuous features. The lavas are uniform in thickness, 1-2 m, and are cut by vertical dykes of similar thickness.

From a study of the numerous dykes around the perimeter of Maupiti, the general area of the central caldera was determined to be on the south side of the island (Figure 2-11). Lavas are of the types common to other Society Islands and trachytes are found, though rare. Dolerite outcrops in the lateritic soil on the northwest side of the island.

j. Maupihaa, Manuae, Motu One

To the west and northwest of Maupiti, only coral atolls and

Figure 2-11 The island of Maupiti (Mercator projection). Very little of the original shield volcano is preserved at Maupiti. This sketch map records the orientation of vertical dykes (dashed and dotted lines), which suggest a caldera toward the southern perimeter of the island. No intrusive bodies were found.

seamounts occur. These are thought to be older features than the volcanic islands to the southeast, but this proposition cannot be geochronologically tested without dredging or drilling down to the volcanic basement.

2.5 The Marquesas Islands

Eleven volcanic islands located approximately 1350 km northeast of Tahiti form the Marquesas Island chain. This group stretches northwestwardly from Fatu Hiva ($10^{\circ}35'S$, $138^{\circ}35'W$) to Eiao ($8^{\circ}00'S$, $141^{\circ}27'W$), a distance of 355 km. The geometry of the islands (Figure 2-12) is nearly linear, striking approximately 308° , and sub-parallel to the trend of the Society Islands, $\sim 302^{\circ}$. Other prominent lineaments occur within the group, 353° and 263° , which surely reflect structural control by the underlying Pacific crust. All the islands are small but reach heights in excess of 1000 m, producing a very rugged and precipitous topography. As for the Society Islands, bathymetric surveys (Mammerickx et al, 1975) indicate that the Marquesas Islands rise as discrete volcanic cones from depths exceeding 4000 m, rather than as the peaks of a prominent submarine ridge.

Geomorphologically the islands appear to be extremely young. The absence of fringing or barrier coral reefs has led to precipitous sea cliffs which rise from the ocean to heights of over 650 m. Chubb (1930) believes that no age progression is evident within the group. The lack of geomorphologic evidence for island evolution is, however, probably due to this absence of coral growth which exposes each island to the same rapid wave erosion and, hence, constant 'youth'. Various explanations have been offered for the failure of reefs to become established. The most plausible is that cold water from the Humboldt Current, which delivers Antarctic water northward along the west coast

Figure 2-12 The Marquesas Islands - spacing and structural control (Mercator projection after Chubb, 1930). The distribution of volcanic centres (both subaerial and submarine) are denoted by open circles. An orthogonality in structural trends, similar to those in the Society Islands, is evident. Chubb (1930) ascribed this structural control to uniaxial compression (northeast-southwest) but it is more likely due to ancient lines of weakness in the oceanic crust.

of South America, then westward along the equator, periodically spills out into the region of the Marquesas Islands and kills incipient coral growth (Davis, 1923; Crossland, 1925).

Chubb (1930) calls attention to the remarkably faulted coastlines of nearly all the islands (see Figure 2-12). He suggests that the predominant directions are oriented at 45° to the direction of horizontal compression (drawing an analogy to laboratory tests of uniaxial compression). It is clear from the discussion in section 2.1 of this chapter, however, that the orientations of faulting are those of the ancient transform faults, $\sim 260^{\circ}$, and of the previous spreading axis of the East Pacific Rise, $\sim 350^{\circ}$. These perpendicular zones of weakness influence the position of islands within the chain and exert structural control on the islands in their erosional stages.

The only published geological accounts of the Marquesas Islands are due to Lacroix (1928), Chubb (1930) and Obellianne (1955). Recently Bishop and Woolley (1973) reported on the petrology of Ua Pou. General comments apply to the island group:

- (1) The islands are very similar petrologically, although some islands exhibit extreme differentiates which have not been observed throughout.
- (2) The predominant rock types are olivine basalt (containing phenocrysts of olivine, clinopyroxene and plagioclase) and mugearite (usually very fine-grained, with small augite and plagioclase phenocrysts and variable amounts of larger olivine phenocrysts). Chemically the basaltic lavas range from slightly nepheline normative to hypersthene-olivine normative to occasionally quartz-normative and are thus generally alkaline to transitional in character. This petrologic suite is very similar to

yet slightly more silica saturated and less alkaline than that represented by Tahitian lavas.

- (3) A three-stage volcanic cycle can be identified on each island: shield-building, followed by intermittent violent eruptions, with caldera collapse, followed by a phase of minor intrusions, including dykes and sills. The three phases are not sharply distinct in time but do describe the general trend in the evolution of these islands. Regional faulting, subsidence and erosion has followed the cessation of volcanism.
- (4) Differentiation proceeded from the olivine basalt through the more silica enriched mugearite to trachytes and phonolites (Ua Pou only). As in the Society Islands, the extreme differentiates appear to have been erupted only in the waning stages of volcanic activity. Often these later eruptions contain ultramafic xenoliths.
- (5) No plutonic specimens have been described, but they may exist, still buried beneath the volcanic piles.

It is appropriate at this point to mention specific features of some of the Marquesas Islands.

a. Fatu Hiva

At the southwestern end of this island chain is Fatu Hiva ($10^{\circ}35' \text{ S}$, $138^{\circ}35' \text{ W}$). One immediately notices the semi-circular plan of the island, the western halves of both the shield and inner caldera submerged by normal faulting. High cliffs line the western side of the island, whereas large areas of planeze form the eastern remnant of the shield, evidence of the island's youth. Caldera subsidence has left a downfaulted central cone within the inner rim of the shield.

Typical of all of the volcanic islands in this group, Fatu Hiva

is composed mainly of lava flows, rare pyroclastic rocks, and numerous vertical dykes which radiate from the caldera region. More scoriaceous flows and ashes occur in the central cone. The typical flow is an olivine basalt containing phenocrysts of olivine and augite in a glassy groundmass of labradorite and augite.

b. Hiva Oa - Tahuata

Small islets and submerged banks occur between Fatu Hiva and the paired islands of Hiva Oa and Tahuata, separated from one another by a narrow channel. Here again, evidence of regional faulting of the original shields into irregular island shapes is clear. Chubb (1930) believes the central caldera at Tahuata to have been on the eastern side of the island, whereas at Hiva Oa, he describes a north-south trending series of three craters. Hiva Oa has an elongated ridge, running approximately east-west which may be ascribed to a fault of the 260° orientation. The northern half of Tahuata consists of long volcanic slopes, rising to a central ridge, which is concave to the east. The eastern and southern borders of the island are precipitous. The western and northern slopes of Hiva Oa rise gently to the main ridge, reaching heights in excess of 1,000 m. Drowned river valleys and elevated wave-cut platforms indicate a combination of subsidence and eustatic change in sea level.

In addition to the common shield building lavas (alkali basalts and ankaramites), fine-grained olivine hawaiites are abundant on both of these islands. Gabbroic dykes are also found (Chubb, 1930) which consist of feldspar, olivine and augite with accessory magnetite, biotite and apatite. On Tahuata, the faulted coastlines are commonly plastered with brick-red palagonite tuff, associated with late-stage faulting. Chubb also described an occurrence of nepheline-normative

lavas on Tahuata, similar to the Tahitian basanites (ne >5%). These flows, except at Ua Pou, are the exception rather than the rule, and appear to have been intruded into ash beds during the last stages of volcanism.

c. Ua Pou

The next substantial volcanic island to the northwest (there are intermediate seamounts and shoals) is Ua Pou ($9^{\circ}25'S$, $140^{\circ}5'W$) which has been treated in some detail by Bishop and Woolley (1973). The island is roughly circular but with a deeply embayed coastline and an extremely craggy topography. A central ridge traverses Ua Pou from north to south, reaching an elevation of 1232 m. No central caldera feature is preserved.

The soil of Ua Pou is whitish, in contrast to the red soil of most of the other islands (Geographical Handbook, N.I.D., 1943), which is the characteristic weathering style of trachytic and phonolitic rocks. Indeed, lavas of this type form a large fraction of the shield at Ua Pou. Bishop and Woolley (1973) state that the Ua Pou rocks form a strongly alkaline series from alkali basalt through hawaiiite, mugearite and trachyte to phonolite, the most completely developed petrologic suite of the Marquesas Islands. The rocks at Ua Pou are unusually potassic and are similar to lavas from Tristan da Cunha and Gough Islands in the South Atlantic.

d. Ua Huka

Half of the original island of Ua Huka ($8^{\circ}55'S$, $131^{\circ}32'W$) remains, its missing portion faulted away to the south. A semicircular ridge forms the backbone of the island, with steep but uniform slopes cut by numerous small valleys to the north, and abrupt slopes to the south. Several long valleys trace the caldera region which is a broad plateau in the centre of the island.

The shield lavas are on the order of 1 m thickness. The petrology is similar to Fatu Hiva and Hiva Oa. Conspicuous are large phenocrysts of augite (~ 1 cm) in these basalts. Along the south coast, the proposed fault scarp, outcrop numerous islets which have not been sampled but whose brick-red (Hemeni) and beige-white (Motu Hanai) weathering suggests minor development of palagonite tuff and trachyte.

e. Nuku Hiva

Nuku Hiva ($8^{\circ}55'$ S, $140^{\circ}04'$ W at Taiohae) is the largest of the Marquesas Islands. The structure appears to be essentially that of Fatu Hiva and Ua Huka - approximately one half of original shield remaining. The outer and inner cones are separated by a deep semi-circular valley which identifies the ring fracture along which the caldera collapsed. High peaks surround the central depression, some reaching elevations well over 1,000 m. The outer shield is dissected, with radial dyke-supported ridges leading down from the high caldera rim to the sea. The north coast is deeply embayed suggesting subsidence. Cliffs dominate the south coast which is frequently plastered with palagonite tuff. A conspicuous trachyte sill intruded ash layers of the central depression.

The common shield building lavas are found on this island as well as differentiated members of the suite (Chubb (1930) and Chapter 4). Again the hawaiites, mugearites and trachytes (some containing biotite and sanidine) are erupted in the late stages of volcanism. Xenoliths of peridotite are found in the shield lavas here and suggest that discovery of similar material of deep origin found in Tahitian lavas is not anomalous.

f. Eiao, Hatutu

Several small islands lie at the northwestern end of the Marquesas Islands. Eiao ($8^{\circ}00'S$, $140^{\circ}40'W$) and Hatutu ($7^{\circ}56'S$, $140^{\circ}33'W$) are elongated in the same direction and Chubb (1930) thinks it probable that they are emergent remnants of a long ridge, most of which is submerged. The outcrop consists of stacks of sub-horizontal lavas.

2.6 The Tuamotu Islands

Between the Marquesas Islands and the Society Islands lies the Tuamotu Archipelago (Figure 1-3), composed solely of coral atolls which stretch from the southeast to the northwest, closely parallel to the volcanic lineaments of French Polynesia. Whether or not the volcanic bases on which these atolls are built also become younger to the southeast is unknown. The answer must wait for dredging or coring of volcanic material. Interestingly, this lineament is the longest and the only one of the five island groups in French Polynesia to change direction at its western end and continue, as the Line Islands, along a north-northwestward strike. (As has been noted by Wilson (1963) and Morgan (1971), the Tuamotu Islands - Line Islands and the Hawaiian Islands - Emperor Seamounts are geometrically congruent.)

2.7 Pitcairn and the Gambier Islands

South of the Tuamotu Archipelago and southeast of the Society Islands lie Pitcairn Island and the Gambier Islands, and, to the northwest of them, the Duke of Gloucester Islands, a line of coral atolls. This lineament is subparallel to the Society Islands and geomorphology suggests an age progression from a young volcanic island at its southeast end (Pitcairn) through dissected volcanic remnants surrounded by barrier

reefs (the Gambier Islands) to coral atolls (the Duke of Gloucester Islands) at its northwest termination.

Pitcairn ($24^{\circ}04'S$, $130^{\circ}06'W$) lies about 2100 km from the East Pacific Rise. Carter (1967) has produced a comprehensive account of the island's geology (see Figure 3-6). This elliptical, entirely volcanic island rises 3,500 m from the ocean floor. Volcanic rocks of two types - lavas and pyroclastics - predominate. Structurally, Pitcairn Island is the remnant of a single shield volcano, its dissected caldera rim forming a prominent annular ridge which opens to the north. Four volcanic units have been distinguished: the Tedside Volcanics, gently dipping flows exposed near the west coast; the Christian's Cave Formation, an agglomeratic tuff generally overlying the Tedside Volcanics with erosional unconformity; the Adamstown Volcanics, horizontal flows which fill the central basin (overlying the Christian's Cave Formation), and the Pulawana Volcanics, a sequence of lava flows at the western extremity of the island, which may have been contemporary with the Adamstown unit. Dykes are rarely seen and occur mainly in the Tedside unit.

Petrologically, the lavas range from alkali olivine basalts (Tedside Volcanics) through hawaiites and mugearites to minor trachytes (Pulawana Volcanics). The basalts contain phenocrysts of plagioclase feldspar, olivine and augite in a fine grained to glassy basalt matrix. The hawaiites and mugearites exhibit a fluidal texture and generally contain fewer phenocrysts than the basalts. A common additional groundmass phase in these samples is biotite. Almost all the rocks contain interstitial alkali feldspar, the amount increasing in the sequence alkali olivine basalt to trachyte. Chemical analyses of Pitcairn samples have been reported by Lacroix (1936) who finds them comparable to Marquesan lavas.

Bathymetry of the region to the northwest of Pitcairn is poorly known but several seamounts have been identified. Further on lie the Gambier Islands, 600 km from Pitcairn. These comprise remnants of older volcanoes as well as coral atolls. The islands of the Mangareva group lie between latitudes $25^{\circ}00'$ and $23^{\circ}15'$ S, longitude $134^{\circ}50'$ and $135^{\circ}00'$ W and consist of four high islands: Mangareva, Taravai, Akamaru and Aukena, together with several smaller islands of very low relief. This archipelago is encompassed by a common barrier reef which suggests that the volcanic islands are in fact remnants of a single volcanic edifice.

Brousse et al. (1972) report that the lavas from this group exhibit a very restricted petrological range, from slightly alkaline tholeiite to olivine-rich basalts. Without further petrographic description it is difficult to decide whether these lavas are similar to those of Pitcairn or indicate a significantly more siliceous suite.

Northwest of Mangareva lie Morane, Fangataufa and Mururoa Atolls, all coralline at sea level. Mururoa, the site of recent nuclear explosions by the French Government, has been drilled down to a basaltic basement (Winterer, 1973). Five hundred kilometres further along this lineament lie the four coral atolls of the Duke of Gloucester Islands.

2.8 The Austral Islands

The most southerly island group of French Polynesia is the Austral Islands chain, stretching 1,500 km northwestwardly from Marotiri Rocks and Rapa Island ($27^{\circ}35'$ S, $144^{\circ}20'$ W) to Maria Island ($21^{\circ}48'$ S, $154^{\circ}41'$ W), a coral atoll. Physiographically, the chain continues into the Cook Islands (Figure 2-13). Chubb (1927) notes that this island chain parallels the Tuamotu and Society Islands. There is possibly a break in the linearity between Rapa and Raivavae where members of the

group northwest of Raivavae appear to lie north of the extension of the Rapa-Marotiri Island line (see Figure 2-13). This offset may be an expression of the structural control exerted by the ocean floor.

Johnson and Malahoff (1971) first reported active submarine volcanism at Macdonald Seamount ($29^{\circ}00'S$, $140^{\circ}13'W$) which lies 500 m beneath sea level at the southeast end of the Austral Island chain. This is the only site of active volcanism in French Polynesia and, naturally, is the youngest member of this most southerly lineament. No significant topographic features interrupt the abyssal deep between Macdonald Seamount and Marotiri Rocks, a group of stark pinnacles of low elevation, unprotected by reefs.

a. Rapa

Ninety kilometres to the northwest is the island of Rapa, whose eroded caldera is surrounded by a horseshoe of high craggy peaks. Development of coral reef is insignificant and this leads to the high coastal cliffs. The coastline is extremely embayed, much like Nuku Hiva in the Marquesas Islands.

The lavas at Rapa are predominantly olivine basalts, but grade through hawaiites and mugearites to trachytes and phonolites. Porphyritic basalts (ankaramites) are also common. Smith and Chubb (1927) remark on the close resemblance between the Tahitian and Rapan petrologic suites, particularly at the basic end of the series. Plutonic rocks outcrop in the partly submerged caldera region. These are nepheline-normative gabbros (essexites).

Numerous banks and seamounts lie between Rapa and Raivavae. Although these have not been sampled, their morphology indicates volcanic cones (Johnson and Malahoff, 1971).

Figure 2-13 The Austral and Cook Islands (Mercator projection). The southernmost island chain in French Polynesia is composed of the Austral Islands, whose northwestward physiographic extension is the Cook Islands group. Bathymetry is denoted by the 1000, 2000 and 4000 m contours.

Palm trees mark an elevated section of the barrier reef surrounding Raivavae in the Austral Islands following the daily afternoon rainstorm.

Horizontal lavas from these cliffs on the north coast of Raivavae in the Austral Islands.

b. Raivavae

Raivavae ($23^{\circ}52'S$, $147^{\circ}41'W$) is an elongate island (ENE-WSW) surrounded by a barrier reef. The Naval Intelligence Division Geographical Handbook (1943) advises that Raivavae is one of the most beautiful in the Eastern Pacific. The volcanic remnant is high (460 m) and rugged, and cliffed on the northern coast. No central caldera is apparent. None of the original shield structure (planeze) is preserved.

The lavas, which are of nearly horizontal attitude, range from olivine basalts to phonolites, as at Rapa, but no plutonic specimens have been reported. Basalt containing feldspar phenocrysts, uncommon on other islands, forms a prominent cliff on the north coast.

c. Tubuai

Northwest of Raivavae is Tubuai ($23^{\circ}23'S$, $149^{\circ}30'W$), the administrative centre of the Austral Islands. This island is elliptical, its major axis striking roughly east-west. A barrier reef surrounds the rather low and unaccentuated relief of the volcanic remnant. Outcrop is restricted to ridge crests and stream beds. The arcuate ridge at the western end of the island possibly marks an ancient caldera. Again, differentiated members of the olivine basalt to phonolite series are represented in the lavas of this island. Porphyritic basalts (olivine and augite are the phenocryst phases) are also common.

d. Rurutu

One hundred and eighty kilometres northwest of Tubuai lies Rurutu ($22^{\circ}25'S$, $151^{\circ}20'W$). Structurally Rurutu is again the remnant of a single shield volcano, but it seems to have had a considerably more complicated geological history than its siblings to the southeast.

Figure 2-14 The island of Raivavae (mercator projection after Obellianne, 1955). This volcanic stack of lavas records very little of the original form of the shield volcano.

Figure 2-15 The island of Tubuai (Mercator projection after Obellianne, 1955). This island is deeply weathered with an undulating topography.

Figure 2-16 The island of Rurutu (Mercator projection after Obellianne, 1955). A complex geological history is recorded here. The production and gradual subsidence of the original shield volcano was followed, after a hiatus of 10 Myr, by uplift and volcanism. Ancient coral reefs (diagonal lines) demonstrate a minimum 100 m uplift. Young lavas are confined to the island's centre. Other symbols as in Fig. 2-7.

An ancient raised coral reef surrounds the island, in places 100 m high, and is evidence for uplift of the island well after construction of the underlying shield. The contact between the basal lavas and the reef is a conglomerate of basalt cobbles and limestone.

A subsequent episode of volcanism, perhaps coincident with the island's uplift, produced small volumes of lavas which cap the older shield. Some of these later lavas have flowed down to the sea through breaks in the raised reef. Pyroclastic material fills much of the central region, which is probably the source of (at least) the second phase of volcanism. The island is now surrounded by a fringing reef of the Tahitian stage of development. Surface lavas are heavily weathered to laterite.

A restricted range of chemistry is exhibited at Rurutu. The predominant lavas of both ages of volcanism are olivine basalts and ankaramites, which are nepheline normative. No trachytes or phonolites are found here, but these may be buried beneath the later lavas.

e. Rimatara and Maria

At the northwest end of the Austral Islands are Rimatara and Maria. No fresh volcanic material is exposed on Rimatara, a hill of laterite 90 m high, or on Maria, a coral atoll. A raised coral reef girdles Rimatara, suggesting uplift of at least 11 m at some time in the past (Obellianne, 1955).

Continuing along the extension of the Austral Islands lineament one encounters the Cook Islands, all of which are volcanic and of various erosional stages (Wood and Hay, 1970). These islands show petrologic affinity to the islands of French Polynesia and their possible genetic relationships to one another will be discussed in the following chapter.

Chapter 3: Geochronology - The Time-Space Evolution of French Polynesia

3.1 Methods and previous work

Only recently has it been possible to directly determine the ages of volcanic rocks from French Polynesia. Previously ages were estimates from the islands' geomorphologic state. Erosional rates, however, are dependent on climate, petrology of the lavas, state of reef development and other poorly defined variables. Thus, geomorphology gives a general impression of the age progression within island groups but is imprecise and subject to error.

Ages of rocks from young volcanic islands are now commonly determined using the potassium-argon whole rock method. This is a chemical method which depends upon the accumulation of an isotope of argon, Ar^{40} , produced by the radioactive decay of an isotope of potassium, K^{40} , with time. Employing various assumptions and knowledge of the rate of decay of K^{40} to Ar^{40} , measurement of the absolute amounts of K^{40} and Ar^{40} in a specimen allows calculation of its age. A discussion of the potassium-argon method and specific analytical techniques used in this study appear in Appendix 1.

It is essential that the assumptions of the method be met before determined ages are geologically meaningful. These assumptions are:

- (1) That the decay constants for K^{40} are accurately known.
- (2) That K^{40} forms a constant fraction of total K in natural systems, and that the K content of the specimen is measured accurately.
- (3) That the total Ar^{40} in the rock, the fraction which has accumulated from K^{40} decay (henceforth denoted *Ar^{40}), can be determined accurately. (Other contributions of Ar^{40} may be from the atmosphere and/or

inherited from the specimen's source region.)

- (4) That at the time of crystallization of the rock from a magma, no ^{40}Ar is present. (Any ^{40}Ar that is retained by the crystallizing magma will lead to an erroneously old age for the rock, unless a correction is made for this 'excess' ^{40}Ar)
- (5) That no loss or gain of ^{40}K or ^{40}Ar has occurred since the crystallization of the rock, other than the changes expected from radioactive decay.

These assumptions cannot be checked for each specimen. Instead, the reliability of the method in determining ages is assessed by the consistency of the results from a series of measurements on related rocks. In addition, geological constraints may provide an independent estimate of age. In the determination of ages from the volcanic rocks of French Polynesia, reliability has been assessed by the consistency of ages for a wide variety of lava types at each volcanic centre. This will be discussed in more detail in the presentation of age determinations for each island group.

One of the more critical factors in the successful determination of ages by the K-Ar method is the condition of the rock sample. Potassium is not uniformly distributed through the various minerals of a volcanic rock, but rather resides predominantly in the last phases to solidify - particularly the interstitial material. Biotite and potassium feldspar contain significant percentages of potassium but these minerals are usually too small (if present at all) to be separated from fine grained basaltic rocks, so whole rock samples are analysed. Oceanic island basalts commonly contain from between 0.5 to 1.5 percent potassium which is sufficient to have generated measurable amounts of radiogenic argon for rocks older than a few hundred thousand years. Thus whole

rock samples will yield meaningful ages provided care is taken in sample selection. It is particularly important that the potassium-bearing phases have not undergone alteration into secondary minerals (clays, zeolites). During weathering, radiogenic argon may escape from the rock, leading to an erroneously young estimate of age. If glass is prevalent then it must not be devitrified. There is also a danger of inherited argon in rocks which contain large phenocrysts and/or xenolithic fragments. Generally, fine-grained holocrystalline samples are most suitable and yield ages with a precision of $\pm 2\%$. (A more comprehensive discussion of errors appears in Appendix 1).

During field sampling of volcanic material for K-Ar geochronology, only rocks devoid of macroscopic weathering were collected. Later, microscopic examination of thin sections of these samples for secondary mineralization and state of crystallization allowed further selection. A second consideration was the geological situation of the samples. Where possible samples were collected in situ, and of the widest possible petrological spectrum. As many of the shield volcanoes (described in Chapter 2) are deeply weathered, outcrop of fresh material was often limited and samples were also collected from stream and beach cobbles.

Before proceeding with the geochronological results of this study, it is useful to review previous contributions to the age distribution of volcanic rocks in French Polynesia. The first age determinations for this region were published by Krummenacher and Noetzlin (1966). These authors report K-Ar whole rock ages from Tahiti, Moorea, Huahine, Raiatea, Tahaa and Bora Bora in the Society Islands, and Rapa, Raivavae, Tubuai and Rurutu in the Austral Islands. Unfortunately their results are very scattered and cannot be used with

any confidence. Ages from Tahiti, for example, range from 0.3 to 156×10^6 yrs. Undoubtedly some of their samples contained inherited argon but no information is given regarding sample chemistry or petrology, so there may have been other problems. Dymond (1975) reports ages at Tahiti (41 determinations) of from 0.2 to 1.0 Myr and at Moorea (23 determinations) of from 1.5 to 2.0 Myr. These ages appear to fulfill the requirements of consistency. Results of Brousse and Bellon (1974) from the island of Eiao in the Marquesas Islands are scattered between 4.9 and 8.7 Myr. These authors suspect problems with inherited argon and propose that the island is younger than the ages they determine. And, most recently, Dalrymple et al. (1975) report four young ages (1.05 ± 0.04 Myr) from Rurutu (Austral Islands).

Measurements which will be discussed in following sections were made using the techniques previously described by McDougall (1964, 1966) and reviewed in Appendix 1. In all determinations whole rock samples were crushed to a size of 3-1.5 mm: a representative split of this crushing was used in the potassium analysis, the remainder saved for argon extraction. Potassium was measured in duplicate by flame photometry (see Appendix 1).

Argon was extracted by fusing 15-30 g of sample in a high vacuum line. The method of isotope dilution, adding a known volume of Ar^{38} , was used to determine the absolute amount of argon liberated from the fused sample. The isotopic composition of the argon extracted from rock samples was found using a modified AEI MS-10 gas source mass spectrometer with digital readout and 4.2 kilogauss permanent magnet. Data reduction and age calculations were by computer, using the method described by McDougall et al. (1969). Constants used in the calculations are: $\lambda_e = 0.585 \times 10^{-10} \text{ yr}^{-1}$, $\lambda_\beta = 4.72 \times 10^{-10} \text{ yr}^{-1}$, $K^{40}/K = 1.19 \times 10^{-4}$ mole/mole.

3.2 Age determinations on volcanic rocks from the Marquesas Islands

Potassium-argon whole rock ages were determined on samples from five of the Marquesas Islands: Fatu Hiva, Tahuata, Hiva Oa, Ua Huka and Nuku Hiva. The results of this study have been previously published (Duncan and McDougall, 1974). No new analyses will be reported here. Of the 50 samples collected, 29 were selected for geochronology. Table 3-1 presents the results of 33 age determinations on 29 samples, 4 run in duplicate. Samples are grouped according to island and the islands arranged in geographical order, from southeast to northwest. Average ages for the five islands are: Fatu Hiva - 1.35 Myr, Tahuata - 1.91 Myr, Hiva Oa - 1.85 Myr, Ua Huka - 2.75 Myr and Nuku Hiva - 3.85 Myr. Figure 3.1 illustrates the distribution of samples and ages within the islands.

Commercial transport in the Marquesas Islands is restricted to small trading vessels which stop at each of the many copra plantations dotted around the coasts of the inhabited islands. Consequently, sampling was confined to the coastal areas. At Ua Huka sampling was restricted to a single valley (Hanai). The four samples give concordant ages, leaving little doubt that this is their age of crystallization. Sample 73-5 was collected from a dyke that intrudes the lava sequence from which samples 73-8 and 73-9 were taken. A similar consistency in ages is found at Fatu Hiva and Tahuata, but over a wider sampling area.

Nuku Hiva and Hiva Oa have been more thoroughly sampled than the other islands and exhibit the greatest variation in measured age. Except for a dyke (73-18) from Taipai Vai all samples from Nuku Hiva have ages which lie within a relatively short (0.5 Myr) period. This suggests an extremely rapid episode of shield formation, compatible with similar observations from geochronology of the Hawaiian Islands (McDougall, 1964). No extreme differentiates were found suitable for

Table 3-1: Geochronology of Basalts from the Marquesas Islands

Sample	Location and Description	% K	Radiogenic ^{40}Ar ($\times 10^{-12}$ mol/g)	$\frac{^{40}\text{Ar}}{\text{Total } ^{40}\text{Ar}} \times 100$	Age (m.y.) ± 2 s.d.
<u>Nuku Hiva</u>					
73-11	cobble from beach at Anaho	0.609, 0.599	4.188	57.2	3.89 $\pm$ 0.06
73-12	cobble from beach at Anaho	{ 0.706, 0.698	{ 5.242 5.270	{ 49.0 46.3	{ 4.21 $\pm$ 0.06 4.23 $\pm$ 0.07
73-13	cobble from beach at Anaho	0.381, 0.382	2.674	18.5	3.93 $\pm$ 0.14
73-14	cobble from beach at Anaho	0.649, 0.636	4.238	56.0	3.70 $\pm$ 0.09
73-16	cobble from hillside at Taipi Vai	1.383, 1.398	9.218	65.3	3.72 $\pm$ 0.06
73-18	dyke from hillside east of Taipi Vai	{ 1.629, 1.624	{ 8.658 8.828	{ 50.2 45.4	{ 2.99 $\pm$ 0.05 3.05 $\pm$ 0.05
73-19	cobble from cliff base at Hakatea	0.585, 0.586	3.954	25.8	3.79 $\pm$ 0.09
73-20	dyke from beach platform at Taiohae	1.437, 1.441	10.433	61.7	4.07 $\pm$ 0.06
73-22	dyke from cliff at Taiohae	1.990, 1.990	13.313	31.8	3.76 $\pm$ 0.08
73-23	dyke from cliff at Taiohae	1.974, 1.987	13.622	47.7	3.86 $\pm$ 0.07
<u>Ua Huka</u>					
73-5	dyke from cliff east of Hanai	1.091, 1.092	5.338	45.4	2.75 $\pm$ 0.04
73-6	cobble from cliff east of Hanai	1.103, 1.103	5.327	81.2	2.71 $\pm$ 0.03
73-8	flow from cliff east of Hanai	1.198, 1.199	5.936	83.1	2.78 $\pm$ 0.03
73-9	flow from cliff east of Hanai	1.172, 1.162	5.779	74.4	2.78 $\pm$ 0.05
<u>Hiva Oa</u>					
73-27	dyke from cliff west of Puamau	{ 1.979, 1.996	{ 6.533 6.694	{ 14.0 10.5	{ 1.85 $\pm$ 0.15 1.89 $\pm$ 0.11
73-28	flow from cliff at Nahoe	1.960, 1.943	5.514	69.4	1.59 $\pm$ 0.03
73-29	cobble from beach at Hanaupe	1.640, 1.630	5.351	75.1	1.62 $\pm$ 0.03
73-31	cobble from beach at Hanaupe	2.133, 2.131	6.011	68.0	1.58 $\pm$ 0.03
73-32	flow from cliff at Taaoa	{ 1.030, 1.045	{ 4.592 4.572	{ 63.2 62.3	{ 2.48 $\pm$ 0.06 2.47 $\pm$ 0.06
73-33	cobble from beach at Taaoa	2.366, 2.341	8.352	78.4	1.99 $\pm$ 0.04
73-34	dyke from hillside at Atuona	1.102, 1.099	3.379	38.3	1.72 $\pm$ 0.04
<u>Tahuata</u>					
73-40	cobble from beach at Hanateio	0.938, 0.937	3.406	39.7	2.04 $\pm$ 0.04
73-41	cobble from beach at Hapatoni	0.949, 0.949	3.251	79.6	1.92 $\pm$ 0.03
73-42	cobble from beach at Vaitahu	1.074, 0.074	3.613	51.5	1.89 $\pm$ 0.03
73-43	cobble from beach at Vaitahu	1.101, 1.103	3.484	31.9	1.78 $\pm$ 0.04
<u>Fatu Hiva</u>					
73-44	flow from cliff at Omoa	0.599, 0.595	1.473	15.1	1.39 $\pm$ 0.05
73-45	cobble from beach at Omoa	1.482, 1.484	3.574	64.2	1.35 $\pm$ 0.02
73-47	cobble from beach at Omoa	0.860, 0.860	1.983	45.9	1.30 $\pm$ 0.02
73-49	cobble from beach at Hanavave	1.404, 1.418	3.469	54.1	1.38 $\pm$ 0.03

Figure 3-1 The Marquesas Islands are located in the eastern equatorial Pacific. On the locality inset sea level is denoted by stippling while the single bathymetric contour marks the 4000 m depth. The four oceanic plates involved in sea-floor spreading at the East Pacific Rise are named as in Herron (1972). At each of five Marquesas islands, sample localities are indicated by sample number followed by the measured age (underlined). The general structure at the five islands is schematically shown by the arcuate segments representing caldera rims. Incomplete bathymetry about the islands is indicated by the 2000 m and 3000 m contours.

dating but their field relationships suggest that they are younger than the shield lavas. Sample 73-18 may be associated with a late stage of intrusion.

Ages of rocks from Tahuata and Hiva Oa are mixed, suggesting simultaneous volcanic activity at about 1.9 Myr ago. Although Hiva Oa and Tahuata are now separate islands, their physical proximity permits the possibility that lavas on one were produced from vents on the other, leading to the observed overlapping ages. Indeed, bathymetry (Mammerickx et al. 1975) suggests that the two islands are part of the same volcanic event.

There is close agreement amongst the ages of samples within each island. In addition, very similar ages are determined from samples of quite different potassium contents (%K in samples from Nuku Hiva varies by as much as a factor of 5). This, together with the similarity in ages from 4 duplicate analyses, argues for the reproducibility of the results and against the possibility of either argon loss or excess argon. $^*Ar^{40}$ is usually a large fraction of the total Ar^{40} extracted from these samples, indicating that contamination by atmospheric Ar^{40} does not constitute a significant problem in age determinations. Except for the intermixed ages at Hiva Oa and Tahuata, volcanic activity proceeded systematically from northwest to southeast (Nuku Hiva to Fatu Hiva) over a span of nearly 3 Myr. (4.2 to 1.3 Myr).

The K-Ar age determinations for Marquesan rocks are plotted against distance from Fatu Hiva, the southeasternmost member of the chain (Figure 3-2). Average ages as well as extreme ages at each island are illustrated. Age decreases systematically from Nuku Hiva to Fatu Hiva, a distance of 235 Km. A rate of migration of volcanic activity of 10.9 cm/yr best fits the average ages of the five islands. This estimate was determined by least squares, giving equal weight to each

Figure 3-2 Ages at each of five Marquesas islands are plotted against island position as projected onto the great circle through Fatu Hiva and Eiao Islands (see Fig. 3-1). Distance is measured from Fatu Hiva, the southeasternmost member of the group. Average ages are shown with heavy dots while extreme ages are indicated by vertical bars. A rate of migration of volcanism of 10.9 cm/yr best fits the age-distance relationship exhibited by the five islands. The dashed lines are 11.9 and 9.9 cm/yr, the suggested confidence limits for the best fitting rate of migration of volcanism.

average age. The standard error of this rate (effectively the slope of the age-distance relationship pictured in Figure 3-2) is 0.4 cm/yr. This uncertainty is, however, unrealistically small because it is not clear that this relationship is linear or that average age necessarily records identical points within the volcanic cycle at each island. Instead, a more realistic estimate of the uncertainty in this volcanic migration rate parameter derives from the extreme rates allowed by the total range of ages. Following this reasoning, an uncertainty of ± 1 cm/yr is proposed. The best fitting volcanic migration rate is, of course, an average rate of movement and does not dismiss the possibility of a somewhat episodic migration history (hence the fact that the Ua Huka and Fatu Hiva ages fall off this best-fit rate). Using the 10.9 cm/yr rate of migration of volcanism, an average age of 4.3 Myr is predicted for Eiao, 355 km northwest of Fatu Hiva. (This predicted age is somewhat younger than the scattered ages reported by Brousse and Bellon (1974).) There is no topographic evidence for volcanism northwest of Eiao. The Marquesas Islands Chain, then, is a very young linear feature in the Pacific Basin. The compatibility of these data with those from other young Pacific Basin island chains will be discussed in a later section of this chapter.

3.3 Age determinations on volcanic rocks from the Society Islands

Fifty-six whole rock samples from eight of the shield volcanoes of the Society Islands have been dated by the K-Ar method. The samples selected for geochronology come from Taiarapu, Tahiti-Nui, Moorea, Huahine, Raiatea, Tahaa, Bora Bora and Maupiti, in geographical order from southeast to northwest. (See Figures 2-5 and 2-8 to 2-12 for sample localities). Seven duplicate analyses were made as well. Table 3-2 presents these results.

Table 3-2: Geochronology of samples from the Society Islands

Sample	Location and Description	% K	Radiogenic ^{40}Ar (X 10 ⁻¹² mol/g)	Rad ^{40}Ar Total ^{40}Ar X100	Age (m.y.) ± 2 s.d.
<u>Maupiti</u>					
73-199	flow from cliff; S coast	1.663,1.652	12.809	36.8	4.34 ± 0.08
73-200	dyke; S coast	0.601,0.596	4.621	77.1	4.33 ± 0.07
73-201	dyke; SW point	1.326,1.345	10.688	50.7	4.49 ± 0.09
73-204	dyke; N coast	1.386,1.386	9.736	61.3	3.94 ± 0.06
73-206	flow from cliff, N coast	1.649,1.651	11.960	64.4	4.07 ± 0.06
73-207	dyke; N coast	0.650,0.649	4.998	77.0	4.32 ± 0.06
73-209	flow from cliff, N coast	{1.956,1.955 1.956,1.955	{14.934 14.956	{78.5 78.3	{4.29 ± 0.06 4.29 ± 0.06
<u>Bora Bora</u>					
73-257	dyke from central ridge	1.202,1.200	6.685	55.7	3.12 ± 0.05
73-295	dyke; NE point	1.115,1.105	6.282	58.8	3.18 ± 0.08
73-308	dyke; islet of Toopua	1.545,1.530	8.651	75.9	3.16 ± 0.05
73-318	dyke; E coast	{1.476,1.484 1.476,1.484	{8.757 8.815	{86.9 89.0	{3.32 ± 0.04 3.34 ± 0.05
73-319	dyke; SE point	1.376,1.364	8.002	87.1	3.28 ± 0.04
73-332	flow; SE point	{1.232,1.236 1.232,1.236	{7.441 7.451	{79.9 78.7	{3.38 ± 0.09 3.39 ± 0.06
73-336	dyke; E coast	1.024,1.030	5.904	72.8	3.23 ± 0.05
<u>Tahaa</u>					
73-185	dyke; N coast	1.935,1.933	9.750	83.3	2.83 ± 0.04
73-186	flow from cliff; N coast	1.488,1.490	7.563	63.7	2.85 ± 0.04
73-189	flow from cliff; N coast	0.732,0.730	3.330	63.7	2.56 ± 0.04
73-190	dyke; Baie Raai	1.317,1.313	6.785	77.3	2.90 ± 0.04
73-192	cobble; stream at Hipu	1.677,1.682	9.467	81.1	3.16 ± 0.04
73-193	dyke; Baie Raai	1.267,1.267	6.622	76.7	2.93 ± 0.04
73-194	flow from cliff; Hipu	1.685,1.693	8.694	75.1	2.89 ± 0.04
73-196	flow from cliff; NW coast	{1.674,1.670 1.674,1.670	{8.589 8.621	{76.0 77.1	{2.88 ± 0.04 2.89 ± 0.05
<u>Raiatea</u>					
73-123	intrusive; Mt. Tauopa	4.229,4.225	18.646	68.8	2.48 ± 0.03
73-124	intrusive; Mt. Tauopa	4.199,4.193	18.504	69.2	2.48 ± 0.03
73-135	dyke; NE coast	1.310,1.310	5.991	58.7	2.57 ± 0.04
73-140	flow from cliff; Baie Faaroa	1.902,1.900	8.016	8.1	2.38 ± 0.16
73-141	flow from cliff; Baie Faaroa	1.549,1.555	6.741	59.5	2.44 ± 0.03
73-153	intrusive; Pt. Opehia	4.198,4.161	18.010	79.4	2.42 ± 0.04
73-161	intrusive; Baie Fafao	4.459,4.453	18.504	89.9	2.43 ± 0.03
<u>Huahine</u>					
73-56	dyke; Lac Maeva	2.114,2.112	9.723	82.4	2.58 ± 0.03
73-72	intrusive; Mt. Tiva	4.506,4.522	16.161	88.2	2.01 ± 0.03
73-83	flow; Te Fareeri	{0.975,0.990 0.975,0.990	{3.835 3.765	{54.7 61.2	{2.19 ± 0.05 2.15 ± 0.04
73-95	flow; Baie Maroe	1.655,1.678	7.438	68.8	2.51 ± 0.04
73-111	dyke; Baie Faie	1.722,1.731	7.802	62.6	2.54 ± 0.04
73-117	intrusive; Mt. Vaihi	4.171,4.188	14.969	79.9	2.01 ± 0.02
<u>Moorea</u>					
73-211	dyke; Oponohu	3.095,3.063	8.273	81.9	1.51 ± 0.03
73-215	flow; Haapiti	3.252,3.266	8.981	66.4	1.55 ± 0.02
73-225	intrusive; Cook's Bay	{4.449,4.440 4.449,4.440	{11.976 12.212	{85.6 43.3	{1.50 ± 0.03 1.54 ± 0.03
73-233	flow; W coast	1.173,1.177	3.420	79.9	1.49 ± 0.02
73-234	flow; Haapiti	3.319,3.307	8.782	68.1	1.64 ± 0.02
73-239	flow; S coast	3.388,3.370	9.127	72.4	1.52 ± 0.02
73-244	dyke; SW coast	1.370,1.369	3.926	54.1	1.61 ± 0.03
73-248	dyke; W coast	1.660,1.678	4.492	53.3	1.51 ± 0.02
74-414	cobble; Afareaitu	2.719,2.741	7.419	24.5	1.53 ± 0.04
<u>Tahiti</u>					
73-367	flow; Faataua	0.812,0.811	1.894	37.9	0.51 ± 0.01
73-373	flow; Punaruu R.	1.150,1.160	2.469	60.3	0.92 ± 0.01
73-396	flow; W coast	1.661,1.690	2.461	51.4	0.82 ± 0.02
73-429	cobble; Pirae	{1.203,1.213 1.203,1.213	{2.802 2.763	{42.6 40.3	{1.30 ± 0.03 1.28 ± 0.03
74-416	cobble, Punaruu R.	0.625,0.625	1.369	17.8	1.23 ± 0.04
74-421	cobble, Orofero R.	1.340,1.337	1.853	48.2	0.78 ± 0.01
74-422	cobble; Arahurahu	1.200,1.201	1.624	36.4	0.76 ± 0.01
74-423	cobble; Mairipeha R.	4.116,4.100	3.474	64.6	0.48 ± 0.01
<u>Taiarapu</u>					
73-362	flow from cliff, S coast	4.722,4.759	3.574	74.5	0.42 ± 0.01
73-419	flow from cliff, Tautira	1.201,1.200	0.804	13.3	0.38 ± 0.02
73-424	intrusive; Pt. Riri	4.674,4.672	3.727	74.4	0.45 ± 0.01
74-417	cobble; Vaitepiha R.	1.594,1.593	1.355	5.2	0.46 ± 0.06

Beginning at the southeast end of this chain, the recent volcanism at Mehetia and nearby seismic activity has been noted. Unfortunately, no samples were collected from Mehetia to confirm its youth geochronologically. If, as Williams (1933) suggests, the island is no more than a few thousand years old, its age cannot be precisely determined by the K-Ar method.

Ages at Taiarapu and Tahiti-Nui span the period 0.4 to 1.3 Myr, but activity at the larger centre clearly began earlier than at Taiarapu and continued over a longer period of time. Most of the surface lavas on Tahiti are weathered and unsuitable for geochronology. Field magnetic measurements (Figure 2-5), however, indicate that the surface flows are normally magnetized and, from geochronology of stratigraphically older samples, must be younger than the age of the Brunhes-Matuyama boundary (~0.70 Myr). Fresher material, more suitable for geochronology, is found at the base of cliffs which are being rapidly eroded (e.g. coastal wave-cut cliffs and valley stream-cut cliffs) or as cobbles in stream beds. Normal polarity only is found in lavas from Taiarapu. This is confirmed by the four samples from that volcano, whose ages are younger than the reversal boundary.

Reversely magnetized lavas outcrop only in deeply eroded river valleys, or at the base of high coastal cliffs. (A 300 m veneer of normally magnetized lavas overlies the 0.70 Myr reversal boundary in the Punaruu valley.) Two flows collected from the western side of Tahiti-Nui (73-373, 73-396) are older than the Brunhes-Matuyama boundary and, indeed, lie within a reversely magnetized section of the volcano. Sample 73-367 is a flow outcropping very near the coast and yields an age of 0.51 Myr, compatible with its normal magnetic polarity. Older specimens (73-429, 74-416) are cobbles from river valleys on the western side of the island which have presumably been uncovered by

Figure 3-3 The Society Islands rise as individual shield volcanoes from the ocean floor, rather than as the peaks of a prominent submarine ridge. The vertical exaggeration is about 100 times. Sea-level plan views of the islands are blackened and submarine bathymetry is shaded. Average ages at eight shield volcanoes are marked. Ages increase monotonically to the northwest.

erosion. Petrologically they are indistinguishable from younger shield building lavas. Thus the construction of the shield at Tahiti-Nui has taken at least 0.8 Myr. No lavas of the late valley-filling stage were found suitable for dating. Sample 74-423, a biotite mugearite, is distinctly younger than the shield lavas and its differentiated chemistry also suggests a late stage of eruption. Volcanic activity, then, spanned a period of about 0.9 Myr. at Tahiti-Nui and a much shorter period at Taiarapu. Sampling has not been as exhaustive at Taiarapu but younger ages at this smaller volcano indicate that volcanism has migrated southeast from Tahiti-Nui.

Ages from samples at Moorea span a very narrow period, 1.49 to 1.64 Myr and agree with ages reported by Dymond (1975). No significant age differences can be seen amongst various rock types, suggesting that shield construction was very rapid and that a variety of magma types were available for eruption toward the end of the volcanic activity. All oriented samples collected from Moorea are reversely magnetized, which is compatible with eruption during the Matuyama reversed epoch (see Figure 3-4). All ages at Moorea are significantly older than those from Tahiti.

At Huahine, the next island to the northwest, ages are a bit more scattered and probably reflect at least two distinct volcanic episodes. Two of the conspicuous trachytic intrusions which lie along the orthogonal lines of weakness (260° , 350°), Mt. Tiva (sample 73-72) and Mt. Vaihi (sample 73-117), have been dated and yield the youngest ages of the six samples from this island, 2.01 Myr. As proposed in Chapter 2, these trachytic intrusions appear to be associated with the final episode of regional faulting and the cessation of volcanic activity. Older samples, both flows and dykes, indicate a shield building phase lasting at least 0.4 Myr (2.58 to 2.19 Myr). All oriented samples proved

Figure 3-4 The magnetic polarity time-scale (after Cox, 1969) has been derived from potassium-argon geochronology of a world-wide distribution of subaerial volcanic provinces for which magnetic polarity is also known. Oriented samples from the Society Islands yield ages (Chapt. 3) and polarities (Chapt. 5) which agree well with this time-scale.

to be normally magnetized (further discussion in Chapter 5). The determined ages, however, span both normal and reversed epochs and events. Thus only those samples which have been oriented and dated provide a test of geochronological reliability. These are 73-56, 73-95 and 73-111. The ages of all three samples are within the Gauss normal epoch, correlating well with the polarity time scale (Cox, 1969).

At Raiatea great uniformity in age is again the rule. Trachytic intrusions mark the cessation of volcanic activity, as at other islands, at about 2.45 Myr here at Raiatea. Only three samples from the shield building stage could be dated. Sample 73-135 predates the trachytic intrusions by about 100,000 years, while 73-141 is probably a late stage flow. Sample 73-140 was difficult to measure because of its low percentage of radiogenic argon, hence its high error limits. Ages from Raiatea and Huahine overlap but average ages for the two centres are significantly different. Comparing the ages of similar petrologic types (the late stage trachytic intrusions, for example) one again sees that Huahine is the younger island. Raiatea exhibits both normal and reversely magnetized rocks. Sample 73-135 is normally magnetized and its age places it in the Gauss normal epoch. Sample 73-153 is magnetized in an intermediate direction and it is not surprising that its age places it directly on the Gauss-Matuyama boundary, suggesting this lava cooled during a reversal of the Earth's magnetic field.

Samples from Raiatea's older sister, Tahaa, range in age from 2.56 to 3.16 Myr, a period of 0.5 Myr. Dykes and flows of the common petrologic types appear to be contemporaneous. Samples were collected only on the north and northwestern slopes of the island and it is not known whether the southeastern portion of the island is younger (as at Tahiti-Nui). As mentioned in section 2.4, Tahaa exhibits no significant

outcrop of trachyte (DeNeufbourg, 1965), which, on other islands, are the youngest episodes of volcanic activity. The ages of samples from Tahaa are significantly older than from islands to the southeast.

At Bora Bora determined ages ranged between 3.12 and 3.51 Myr. The samples are all of the shield building type, lavas and dykes. The relatively short time spanned by these ages, 0.4 Myr, is consistent with the shield building episodes recorded at other islands of this chain and is, again, a minimum estimate. The dated samples from Bora Bora which have also been oriented for magnetic measurement (73-295, 73-336) are normally magnetized and consistent with eruption during the Gauss normal epoch. Comparison of ages from Bora Bora with those from Tahaa suggests that the waning stages of volcanism at Bora Bora may have coincided with the initiation of activity at Tahaa.

The northwesternmost volcanic member of this island chain is Maupiti. Ages have been determined on a wide range of petrologic types, but again the time spanned in construction of the island was short - ≥ 0.5 Myr. No trachytic intrusions were observed here but the youngest age was found for a biotite-bearing olivine dolerite which intruded the shield lavas. Ages are consistent over a wide range of potassium content (0.6 to 2.0%K) and the fraction of $^{40}\text{Ar}^*$ in total Ar^{40} is uniformly high, implying reliability. No magnetic measurements were recorded. Maupiti is, then, the oldest volcanic island in the Society Islands Chain.

Figure 3-5 presents the age-distance relationship for the Society Islands, applying the same analysis as for the Marquesas Islands' geochronology. Average ages, extreme ages and ages of trachytic intrusion (if this is found) at each of the eight shield volcanoes are pictured. Mehetia is assumed to be in the shield-building stage - roughly age 'zero'. Average ages decrease without exception from northwest to

Figure 3-5 Ages at each of eight shield volcanoes of the Society Islands are plotted against island position as projected onto the great circle through Mehetia and Maupiti (see Fig. 3-3). Distance is measured from Mehetia, the southeasternmost member of the group. Average ages are shown with heavy dots, trachytic volcanism by open circles, and extreme ages by vertical bars. A rate of migration of volcanism of 11.1 cm/yr best fits the age-distance relationship exhibited by the average ages at the eight volcanoes. The dashed lines are 12.1 and 10.1 cm/yr, the suggested confidence limits on the best fitting rate of migration of volcanism.

southeast, confirming geomorphologic evidence of the age progression. Over the 500 km from Maupiti to Mehetia volcanism has migrated at varying rates. An average rate of migration of volcanism of 11.1 ± 1.0 cm/yr characterizes the age-distance relationship in the Society Islands. Using this rate to extrapolate to the northwest, the oldest (northwesternmost) topographic feature, a seamount (15.5°S , 156.0°W) is predicted to be 7.8 Myr, which dates the initiation of the volcanism which produced the Society Islands. These data will be treated further in section 3.6.

3.4 Age determinations on volcanic rocks from Pitcairn Island

Southeast of Mehetia lie the islands and seamounts of the Pitcairn-Gambier Islands lineament. Volcanic samples from Pitcairn Island were generously provided by Drs. R.M. Carter and D.S. Coombs of Otago University, Dunedin, N.Z. Age determinations on 12 of these samples have been reported earlier (Duncan et al., 1974) and no new analyses will be discussed here.

As described in section 2.7, four lithologic units (the Tedside Volcanics, the Christians Cave Formation, the Adamstown Volcanics and the Pulawana Volcanics) have been distinguished and all are represented in the age determinations. Table 3-3 presents the analytical data for 20 K-Ar age determinations on the 12 selected samples. Generally the duplicate analyses on 8 of the samples show agreement to within experimental error. The samples are grouped according to the formations defined by Carter (1967). Localities and ages are pictured in Figure 3-6.

The calculated ages fall into at least two groups. Ages for four samples from the Tedside Volcanics, stratigraphically the oldest unit exposed on Pitcairn, form a group distinctly older than the rest,

Figure 3-6 Generalised geology of Pitcairn Island (Mercator projection after Carter, 1967). Arrows indicate the general direction of lava flows. Observed geological contacts are indicated by solid lines; inferred contacts are dashed lines. Sample numbers are followed by K-Ar ages ($\times 10^6$ yr). Contours are at 50 m intervals.

- PULAWANA VOLCANICS
- ADAMSTOWN VOLCANICS
- CHRISTIANS CAVE FORMATION
- TEDSIDE VOLCANICS

— GENERALIZED DIRECTION OF LAVA FLOW

— GEOLGICAL CONTACTS:

— — — — — SEEN

— — — — — INFERRED

Table 3-3: Geochronology of whole rock samples from Pitcairn Island

Sample	Location and Description	% K	Radiogenic ^{40}Ar ($\times 10^{-12}$ mol/g)	Rad ^{40}Ar Total	Age (m.y.) ± 2 s.d.
<u>Pulawana Volcanics</u>					
73-647	Hawaiiite; Cabin Cave, NW coast	1.694, 1.696	1.90	32.5	0.63 ± 0.01
73-642	Hawaiiite; Jinser Valley, SW coast	$\begin{Bmatrix} 2.183, 2.169 \\ 3.826, 3.828 \end{Bmatrix}$	$\begin{Bmatrix} 2.42 \\ 4.56 \end{Bmatrix}$	$\begin{Bmatrix} 28.8 \\ 28.9 \\ 34.8 \end{Bmatrix}$	$\begin{Bmatrix} 0.62 \pm 0.01 \\ 0.62 \pm 0.02 \\ 0.67 \pm 0.01 \end{Bmatrix}$
73-648	Trachyte; Dyke, W coast				
<u>Adamstown Volcanics</u>					
73-652	Hawaiiite; St Pauls Point, Eastern tip of island	$\begin{Bmatrix} 1.537, 1.535 \\ 1.718, 1.736 \end{Bmatrix}$	$\begin{Bmatrix} 1.25 \\ 1.24 \\ 1.67 \\ 1.58 \end{Bmatrix}$	$\begin{Bmatrix} 20.8 \\ 16.7 \\ 9.0 \\ 18.0 \end{Bmatrix}$	$\begin{Bmatrix} 0.46 \pm 0.01 \\ 0.45 \pm 0.02 \\ 0.54 \pm 0.04 \\ 0.52 \pm 0.02 \end{Bmatrix}$
73-654	Biotite Artefact site, Tautama, adjacent mugearite: to south coast				
73-650	Olivine Western side of John Mills Valley, mugearite; centre of island	$\begin{Bmatrix} 2.315, 2.311 \\ 2.024, 2.036 \end{Bmatrix}$	$\begin{Bmatrix} 2.53 \\ 2.66 \end{Bmatrix}$	$\begin{Bmatrix} 26.0 \\ 36.7 \end{Bmatrix}$	$\begin{Bmatrix} 0.61 \pm 0.02 \\ 0.64 \pm 0.02 \end{Bmatrix}$
73-640	Mugearite; Western end Tautama, southern coast, possibly Adamstown Volcanics		2.28	31.7	0.63 ± 0.01
<u>Christians Cave Formation</u>					
73-649	Hawaiiite; Block in agglomerate, Palva Ridge	1.274, 1.286	1.36	19.5	0.60 ± 0.02
<u>Tedside Volcanics</u>					
74-1	Basalt; Water Valley	$\begin{Bmatrix} 1.429, 1.418 \\ 1.960, 1.964 \end{Bmatrix}$	$\begin{Bmatrix} 2.41 \\ 2.34 \\ 3.14 \\ 2.97 \end{Bmatrix}$	$\begin{Bmatrix} 40.5 \\ 38.3 \\ 16.2 \\ 51.3 \end{Bmatrix}$	$\begin{Bmatrix} 0.95 \pm 0.02 \\ 0.92 \pm 0.02 \\ 0.90 \pm 0.03 \\ 0.85 \pm 0.01 \end{Bmatrix}$
74-2	Hawaiiite; Jinser Valley				
73-645	Hawaiiite; Jinser Valley	$\begin{Bmatrix} 1.491, 1.496 \\ 1.065, 1.067 \end{Bmatrix}$	$\begin{Bmatrix} 2.24 \\ 2.24 \\ 1.44 \\ 1.47 \end{Bmatrix}$	$\begin{Bmatrix} 46.0 \\ 47.3 \\ 42.1 \\ 34.5 \end{Bmatrix}$	$\begin{Bmatrix} 0.84 \pm 0.01 \\ 0.84 \pm 0.01 \\ 0.76 \pm 0.01 \\ 0.77 \pm 0.02 \end{Bmatrix}$
73-643	Basalt; Jinser Valley				

averaging 0.85 Myr and spanning the short period 0.76 - 0.93 Myr.

The samples from the Christians Cave Formation and the Adamstown and Pulawana Volcanics all yield younger K-Ar ages - from 0.46 to 0.63 Myr - consistent with their stratigraphically younger age. The ages from the Adamstown and Pulawana Volcanics are indistinguishable and support Carter's (1967) contention that the two units are coeval.

These age determinations outline the geological history of Pitcairn. The Tedside Volcanics, regarded as a relic of a dome building phase, were erupted between 0.8 and 0.9 Myr ago. Following a hiatus of 0.2 Myr, during which time the main caldera collapsed, the Christians Cave Formation, the Adamstown Volcanics (which filled the caldera) and the extracaldera flows of the Pulawana Volcanics, were erupted about 0.62 Myr ago, with volcanism continuing until at least 0.45 Myr ago. The volcano was therefore active for a period of 0.5 Myr during the younger part of the Pleistocene.

In addition to demonstrated reproducibility and the consistency of results with varying potassium content, field magnetic measurements (Carter, 1967) confirm the geochronological placing of the Adamstown Volcanics and Christians Cave Formation in the Brunhes normal epoch.

Northwest of Pitcairn lie the Gambier Islands, which are partly volcanic, and the Duke of Gloucester Islands, which are coral atolls (Figure 3-7). Potassium-argon age determinations on volcanic samples from the Gambier Islands (Brousse et al., 1972) suggest that the tholeiitic and alkali olivine basalt lavas which form the islands of Mangareva, Aukena and Makapu were erupted between 5.2 Myr and 7.2 Myr ago. Winterer (1973) reports an age of 8 Myr (D. Krummenacher, personal communication) for basalt from a drill hole at Mururoa atoll at the northwestern end of the Gambier Islands. There is, then, an indication of a southeastwardly migration of volcanism, though not as completely

Figure 3-7 Pitcairn Island is situated in the Pacific Basin, possibly the southeasternmost member of the Duke of Gloucester-Gambier Islands volcanic lineament which is parallel to other Pacific island chains. The Duke of Gloucester Islands are all coral atolls at sea level. Numbers indicate K-Ar ages. Bathymetry in the Pacific Basin is denoted by the 4000 m contour; in the inset, by the 4000 m and 2000 m contours (dashed lines). (Mercator projection).

documented as for either the Marquesas or the Society Chains. The rate of migration of volcanism which characterizes this lineament is about 11 cm/yr and the predicted age for the Duke of Gloucester Island group is about 14 Myr (The volcanic migration rate cannot be precisely constrained because of the small number of dated islands in the lineament and because of the wide range in ages in the Gambier Islands.)

3.5 Age determinations on volcanic rocks from the Austral Islands

The southernmost island chain in French Polynesia is formed by seven Austral Islands (Figure 2-13), stretching 1500 km from Macdonald Seamount to Maria Island. Although the longest chain of volcanic islands, the Austral Islands are much less closely spaced than those of their northerly neighbours. Samples were collected from three islands: Rurutu, Tubuai and Raivavae. Ages for selected samples are presented in Table 3-4.

Beginning with the southeastern end of this lineament, it is recalled that active volcanism has been observed at Macdonald Seamount (Johnson and Malahoff, 1971), making this the youngest member. K-Ar ages have been determined for samples from Rapa Island by Krummenacher and Noetzlin (1966). Although their analyses are generally poor, the three ages which they report from Rapa are consistent at 5.1 ± 0.3 Myr and may be acceptable.

A very uniform chemistry is reflected in the analysed samples from Raivavae. The percentage of $^{40}\text{Ar}^*$ is nevertheless uniformly high and sample 74-410 reproduces well, so that the reasonably large spread in determined ages is probably real. The three samples which were collected in situ (74-407, 74-410 and 74-412) are from the stratigraphically oldest exposed parts of the island and yield an average age of 6.8 Myr. Samples 74-406 and 74-409 are cobbles and their stratigraphic position

Table 3-4: Geochronology of Austral Islands samples

Sample	Location and Description	% K	Radiogenic ^{40}Ar ($\times 10^{-12}$ mol/g)	$\frac{\text{Rad } ^{40}\text{Ar}}{\text{Total } ^{40}\text{Ar}} \times 100$	Age (m.y.) ± 2 s.d.
<u>Rurutu</u>					
74-386	ol. basalt; flow, SW coast	0.614, 0.613	9.224	49.7	8.43 ± 0.14
74-388	ol. mugearite; cobble, SW coast	1.396, 1.394	2.654	50.1	1.07 ± 0.02
74-390	basalt; flow, N coast	0.610, 0.614	{13.343 13.021}	{47.8 37.3}	{ 12.21 ± 0.21 11.94 ± 0.22 }
74-392	ol. basalt; flow, NW coast	1.197, 1.197	2.235	37.0	1.05 ± 0.02
74-394	ol. basalt; coral inclusion, W. coast	0.902, 0.901	2.972	55.3	1.85 ± 0.08
74-396	porph. basalt; agglomeratic inclusion, center of island	0.860, 0.859	18.475	63.6	12.04 ± 0.20
<u>Tubuai</u>					
74-397	trachyte; cobble, E of Mataura	{2.658, 2.669 0.633, 0.635 0.442, 0.437}	{41.543 40.757 9.714 8.319 7.970}	{69.0 41.5 47.0 43.2 25.2}	{ 8.74 ± 0.14 8.53 ± 0.15 8.59 ± 0.16 10.60 ± 0.21 10.20 ± 0.25 }
74-399	basalt; cobble, E of Mataura		7.871	53.1	8.63 ± 0.16
74-401	basalt; cobble, E of Mataura		14.287	47.1	9.49 ± 0.17
74-402	basalt; cobble, E of Mataura		19.578	65.8	9.48 ± 0.16
74-403	basalt; cobble, island center				
74-404	basalt; cobble, W coast				
<u>Raiavavae</u>					
74-406	porph. basalt; cobble, N coast	0.546, 0.547	5.381	68.5	5.52 ± 0.09
74-407	basalt; flow, N coast	0.693, 0.688	8.350	52.9	6.78 ± 0.13
74-409	basalt; cobble, N coast	0.653, 0.654	6.882	78.8	5.91 ± 0.09
74-410	porph. basalt; dyke Mahanatoa	{0.523, 0.519 0.650, 0.649}	{6.744 6.841 7.187}	{69.7 70.7 79.2}	{ 7.26 ± 0.13 7.37 ± 0.12 6.21 ± 0.10 }
74-412	basalt; dyke, SW coast				

is unknown, though they most likely come from the prominent ridge facing the northern coast of the island. Whether volcanic activity was continuous from 7.3 to 5.5 Myr ago or was broken by a hiatus of unknown duration cannot be deduced from these five ages.

Age determinations on six samples (two of them duplicated) from Tubuai range between 8.6 and 10.4 Myr., as prolonged an episode of volcanism as at Raivavae. Because of the extreme paucity of outcrop and generally high degree of weathering, only stream cobbles exhibited the prerequisite standards of freshness for geochronology. The chemistry of these samples varies between basalt and trachyte. The trachytic sample, 74-397, is young whereas the remaining five basaltic samples vary in age over the 1.8 Myr shield building phase(s).

From section 2.8 it is recalled that Rurutu has some geomorphologic features of a young volcanic island: narrow fringing reef and 'accentuated' volcanic topography. In addition, the restricted range of exposed petrologic types (no trachytes or phonolites) may also be an indication of its youth. Previous geochronology (Krummenacher and Noetzlin, 1966; Dalrymple et al., 1975) has confirmed Rurutu's youth (1.05 Myr) and presents an anomaly in the orderly age progression demonstrated for other volcanic island chains in French Polynesia.

Samples were collected from Rurutu in an effort to illuminate the relationship of the young lavas on that island to the older islands to the southeast. Age determinations for these samples, reported by Duncan (1975b), appear to partially resolve the quandry. While very young ages compatible with earlier reports are found, much older material has also been sampled (Table 3-4). Two samples give concordant ages of 12.0 ± 0.2 Myr. One of these, 74-390, is from a fine grained basalt flow on the north side of the island and the other, 74-396, is a cobble included in an agglomeratic tuff. A third sample, 74-386,

aged 8.4 Myr, is suspected of having lost $^{40}\text{Ar}^*$ due to its weathered state. (An age was attempted due to the paucity of fresh material and in an effort to provide information about the basal flows on the island).

Samples 74-388 and 74-392 are from the young lavas which cap the centre of the island. The ages determined from them (1.06 ± 0.03 Myr) are within experimental error of four ages reported by Dalrymple *et al.* (1975). Sample 74-394 is an olivine basalt collected from the basal conglomerate of the ancient upraised coral reef. The age of this sample, 1.85 Myr, places a maximum age on the episode of uplift, which raised the island at least 100 m. The uplift, commencing at or after 1.85 Myr age, culminated in the 1.06 Myr volcanism which covers the centre of the island and obscures the much older volcanic base. This appears to have been very much a local event as the islands southeast of Rurutu (Tubuai, Raivavae, Rapa) show no evidence of similar secondary episodes. To the northwest lie the coral and laterite islands of Maria and Rimatara. At Rimatara an ancient reef has been upraised at least 11 m at some time in the past (Obellianne, 1955) and may be evidence of the same phenomenon seen at Rurutu.

The northwest extension of the Austral Islands lineament is occupied by the Southern Cook Islands. Wood and Hay (1970) have described the geology and petrology of these volcanic and coral islands (Figure 2-13) which are very similar to the islands of French Polynesia. Dalrymple *et al.* (1975) report ages for basalts and their differentiates from Rarotonga, Mangaia and Aitutaki. Volcanism on Rarotonga appears to have been episodic, between 1.2 and 1.8 Myr ago, on Mangaia from 16.6 to 18.9 Myr and on Aitutaki from 0.66 to 0.77 Myr. This great confusion of ages in the Cook Islands appears irreconcilable with a simple north-westerly increasing age progression. The data from Rurutu, however,

Figure 3-8 Ages at each of six shield volcanoes of the Austral and Cook Island Chains are plotted against island position as projected onto the great circle through Macdonald Seamount and Rarotonga (see Fig. 2-13). Distance is measured from Macdonald Seamount, the southeasternmost member of the group. Average ages are shown with heavy dots while extreme ages are indicated by vertical bars. Rurutu exhibits lavas of two widely separate ages, one of which appears to fit the trend of the age progression from the south-east, the other much younger. Only young ages are found at Rarotonga, which may not, then, belong to this lineament. A rate of migration of volcanism of 10.7 cm/yr best fits the age-distance relationship exhibited by the average ages at five islands (older age at Rurutu). The dashed lines are 11.7 and 9.7 cm/yr, the suggested confidence limits on the best fitting rate of migration of volcanism. Ages at Rapa are from Krummenacher and Noetzlin, while those at Rarotonga are from Dalrymple et al. (1975).

suggest that the islands at the northwest end of the Austral Islands may not have had a simple volcanic history and that detailed work is necessary to resolve the complex age-distance relationships of these islands.

Figure 3-8 plots ages of the Austral Islands and Cook Islands against distance from Macdonald Seamount. The older ages at Rurutu fall in line with the age-distance relationship described by islands to the southeast. An average rate of migration of volcanism of 10.7 ± 1.0 cm/yr best fits the progressive younging of island ages to the southeast. Ages for Mangaia fall precisely on this trend, but ages for Rarotonga and Aitutaki are far too young for their northwesterly position, as are the younger ages at Rurutu. One immediate explanation for these anomalously young islands is that they are the most recent sites of independent sources of volcanism and, therefore, the source beneath Macdonald Seamount is not responsible for all volcanism in the Austral-Cook Islands Chain. By this reasoning additional sources are proposed to lie beneath Rurutu (uplift and young volcanism), Mangaia (uplifted coral reef and young volcanism at Rarotonga, only 200 km to the northwest) and Aitutaki (young volcanism). Other possibilities exist, such as renewed volcanism after periods as long as 20 Myr., but require mechanisms for which there is no evidence from other volcanic lineaments of French Polynesia.

3.6 A comparison of rates of migration of volcanism in the Pacific Basin

It has been demonstrated that, with a qualification for the northwest end of the Austral Islands, the volcanic chains of islands and seamounts of French Polynesia become younger from the northwest to the southeast, in agreement with geomorphologic evidence. Rates

of migration of volcanism, determined simply from the distance between centres and their average ages, appear to be surprisingly uniform. Data exist for two other Pacific island and seamount chains.

The Pratt-Welker Seamounts stretch northwestwardly from Cobb Seamount to the Aleutian Trench, subparallel to the young volcanic island lineaments to the south. The northwesternmost edifices of this chain, Kodiak and Giacomini Seamounts, are ages 22.6 ± 1.1 Myr. and 19.9 ± 1.0 Myr, respectively (Turner et al., 1973). These ages have been determined from both mineral separates and whole rock specimens. The rocks are alkali olivine basalts and trachytes. Fission-track ages of samples from both seamounts are concordant with K-Ar ages so the authors conclude that inherited argon is not significant. The trend of the lineament is oblique to the magnetic anomaly pattern recorded in the ocean floor on which these seamounts sit. The age of the ocean floor beneath the two dated volcanoes is between 45 and 50 Myr and it is clear that the volcanism which produced the Pratt-Welker Chain postdated the production of oceanic crust by at least 20 Myr. Turner et al. (1973) use the two average ages which they determine to calculate a rate of migration of volcanism - from northwest to southeast - of 6.6 cm/yr. Because only two members of the lineament have been sampled, there is considerable uncertainty in the average rate of migration of volcanism for the entire chain of seamounts, particularly for its southeastern section.

The Hawaiian Islands have received considerable geochronological attention, beginning with McDougall (1964), who confirmed that this west-northwest trending line of shield volcanoes and seamounts increases in age northwestward from the active volcano of Kilauea on the island of Hawaii. Funkhouser et al. (1968), Dalrymple (1971), McDougall and

Figure 3-9 Ages at each of eight islands of the Hawaiian Islands are plotted against distance from Hawaii. Average ages are shown with heavy dots while extreme ages are indicated by vertical bars. A rate of migration of volcanism of 11 cm/yr is taken to characterize the age-distance relationship exhibited by the Hawaiian Islands. The dashed lines illustrate 10 and 12 cm/yr rates for comparison. Shaw (1973) has proposed a non-linear volcanic migration pattern for the volcanism extending southeastwards from Kauai. For simplicity in comparison with French Polynesian volcanic migration rates, a linear model is fitted to these Hawaiian data. Ages at Niihau are omitted as they are anomalously young for their position along the chain. The islands of Oahu, Molokai and Maui each have two shield volcanoes but only a mean position is pictured here. Separate ages were used, however, in the regression of average ages on distance.

Swanson (1972), Jackson et al. (1972), Doell and Dalrymple (1973) and Dalrymple et al. (1974) have contributed to knowledge of the age distribution of the volcanic centres in this island chain. Figure 3-9 illustrates the combined results of these studies. Except for Niihau, shield volcano ages increase monotonically to the northwest, from Hawaii ('zero' age) to French Frigate Shoals (11.7 ± 0.4 Myr). Much controversy has attended the interpretation of the geochronology of the southeast end of the Hawaiian lineament. The simplest model describing the relationship between island age and distance is a linear one and the one preferred by McDougall (1971) and Morgan (1972), Jackson et al. (1972) and Shaw (1973) believe that an accelerating (non-linear) rate of migration of volcanism can be resolved in the ages of islands between Kauai and Hawaii. Departures from linearity are observed in the age-distance relationships for the island chains of French Polynesia but these are not systematic as are those proposed by Shaw (1973). For the purposes of comparison with the French Polynesian data, a linear rate of migration of volcanism of 11 cm/yr will be taken for the Hawaiian Islands from French Frigate Shoals to Hawaii (Fig. 3-9). Geochronology for older members of this lineament (e.g. Midway Island) is compatible with increasing age north-westward but possible changes in the rates of migration of volcanism have not been clearly resolved (Clague and Dalrymple, 1973).

Figure 3-10 compares the age-distance relationships exhibited by five island and seamount chains in the Pacific Basin with the intent of revealing volcanic episodes which are simultaneous in all or some of the lineaments. No such pattern emerges and it is concluded that volcanism within each chain is irregular and that there is no chronological correlation amongst the five island groups.

To test the proposition that these subparallel volcanic

Figure 3-10 The age-distance relationships in five Pacific island chains are compared. Age is plotted against west longitude, measured from the site or inferred site of present volcanism. Longitude 'zero' is the great circle through the Pacific plate-mantle rotation pole (from Minster et al., 1974) and the site of present volcanism, calculated for each chain following the suggestion of Dymond (1975). Island members for which geochronology is known are blackened, whereas inferred ages for unsampled volcanoes (submarine as well) are open circles. There seem to be no great similarities among the patterns of volcanism for the five island chains.

lineaments owe their origin to the northwestward movement of the Pacific plate over fixed sources (be they hot spots or irregular topography in the upper mantle) a rotation pole for Pacific plate over mantle movement must be defined. If these sources are assumed to be fixed in the mantle, their traces on the Pacific plate will be small circles of latitude about the rotation pole. This supposes that Pacific plate over mantle motion has been fairly constant over the time considered so that the rotation pole has not moved appreciably. There is, however, geometrical evidence from the Hawaiian and Marquesas Islands that the Pacific plate has had a slightly greater northerly component of motion over the past 4 Myr (McDougall, 1971; Duncan and McDougall, 1974). By fixing the rate of rotation of the Pacific plate about this pole with the geochronological data of one lineament, all ages in other lineaments will be predicted. Because the Hawaiian Islands chain is the longest and most completely documented of the Pacific Basin volcanic lineaments, it is most reasonable to use its 11 cm/yr rate of migration of volcanism to fix the rate of rotation about the geometrically defined pole of rotation for the Pacific-mantle motion.

This pole of rotation must lie along the great circle path which is perpendicular to each of the island lineaments and at such a distance from each as to explain any curvature of those volcanic line segments. The six island and seamount chains in question are all reasonably straight, especially those of French Polynesia, so that they must lie close to the equator of the pole of rotation, making its precise location difficult to estimate. Morgan (1972), Clague and Jarrard (1973) and Winterer (1973) have proposed Pacific plate rotation poles based on the trends of island chains on the Pacific plate. Minster et al. (1974) have determined this rotation pole by first

finding the best set of relative motion poles of rotation for all the plates and then defining the absolute motion poles of rotation (plate-mantle) by fitting to trends of nonorogenic linear volcanism. Their Pacific-mantle rotation pole is thus mainly determined by volcanic traces on other plates. All proposed Pacific-mantle rotation poles are close to one another (Table 3-5). Figure 3-11 illustrates the agreement between observed and predicted rates for the Minster et al. (1974) Pacific plate rotation pole.

The 11 cm/yr rate of migration of volcanism in the Hawaiian Islands fixes the rate of rotation of the Pacific plate, with respect to the mantle, about each of the four proposed rotation poles. Predicted rates of migration of volcanism for the remaining five lineaments are calculated from the rotation rate and the great circle distance from the pole to each lineament. The four rotation poles lead to very similar predicted rates for the Pacific Basin volcanic chains. All fit the observed rates (calculated from age-distance relationships within each lineament) to within the limits of discrimination (± 1 cm/yr). It is concluded, then, that the sources of volcanism responsible for six separate volcanic lineaments on the Pacific plate have been fixed with respect to one another for at least the last 4 Myr, and probably much longer.

In Section 2.1 present spreading rates at the East Pacific Rise were reviewed. It will be recalled that half-spreading rates (Pacific plate relative to the East Pacific Rise) increase southward from 4.5 cm/yr at 15°N to 9.0 cm/yr at 30°S . Moreover, this spreading geometry at the East Pacific Rise has emerged only relatively recently, between 6 and 16 Myr ago. The former spreading axis was oriented north-northwest and was responsible for the ancient transform faults which appear to exert structural control on the young island chains of the Pacific Basin.

Table 3-5: Rates of migration of volcanism predicted from Pacific plate-mantle rotation poles with rate for Hawaiian Islands accepted as 11 cm/yr.

Volcanic Chain	Morgan (1972) [67°N 73°W]	Clague and Jarrard (1973) [72°N 83°W]	Winterer (1973) [67°N 45°W]	Minster et al. (1974) [67.3°N, 59.4°W]	Observed rate cm/yr
Pratt-Welker Seamounts	36°* 6.9 ⁺	34°* 6.7	46° 7.3	41° 7.5	6.6
Hawaiian Islands	69° 11.0	66° 11.0	79° 11.0	74° 11.0	[11.0]
Marquesas Islands	89° 11.8	89° 12.0	79° 11.0	84° 11.3	10.9 +1.0
Society Islands	79° 11.6	78° 11.8	68° 10.3	73° 10.9	11.1 +1.0
Pitcairn-Gambier Islands	79° 11.6	79° 11.8	70° 10.6	74° 11.0	11.0 +1.0
Austral Islands	72° 11.2	72° 11.4	62° 9.9	67° 10.6	10.7 +1.0

* Great circle distance from the rotation pole to the volcanic chain

+ Predicted rate of migration of volcanism (cm/yr).

Figure 3-11 The Pacific plate to mantle pole of rotation predicts the position and age of the island members of the Pacific island chains, assuming fixed sources of volcanism. Actual island positions are shown as dots for the Hawaiian Islands (H), the Marquesas Islands (M), the Society Islands (S), the Pitcairn-Gambier Islands (G-P), the Austral Islands (A) and the Easter Island-Tuamotu Islands (E). Predicted volcanic traces are solid lines and predicted ages are strokes of 2 Myr intervals.

Using any of the four proposed Pacific plate-mantle rotation poles and the Hawaiian rate of migration of volcanism, the half-spreading rates of the Pacific plate from the East Pacific Rise can be calculated, assuming that the Rise is stationary. At 15°N , the half-spreading rate should be 10 cm/yr (4.5 cm/yr observed) and at 30°S a rate of 11 cm/yr (9.0 cm/yr observed) is predicted. (Both points are reasonably close to the equator of the pole of rotation). These predicted half-spreading rates are clearly incompatible with rates observed from magnetic anomaly correlation (Herron, 1972). This implies that there is movement of the Rise toward the volcanic sources, or vice versa. The closing rate of movement is not constant, however, varying from about 5.5 cm/yr at 15°N to 2 cm/yr at 30°S . Movement of spreading ridges must happen elsewhere (the ridge system surrounding Africa, for example) so it is concluded for the moment that all differential motion is taken up by movement of the East Pacific Rise to the west at between 2 and 5.5 cm/yr.

The change in orientation of the axis of the East Pacific Rise at between 6 and 16 Myr ago is possibly reflected in the geometry of the older lineaments. The kinks in the trends of the Hawaiian Islands between Kauai and Nihoa (McDougall, 1971) and the Austral Islands between Rapa and Raivavae occurred roughly at this time. Nothing conclusive can be said about the Tuamotu Islands. The volcanism which produced the Marquesas, the Society, the Austral and the Pitcairn-Gambier Island chains began during or shortly after this interval, whereas volcanism in the Hawaiian, Austral, and, presumably, the Tuamotu Island chains was well established by this time. It is, then, possible that the initiation of volcanism in French Polynesia led to the reorientation of spreading at the East Pacific or that the two events are related to some larger readjustment of plate motions.

Chapter 4: Geochemistry and Petrology of French Polynesian lavas.

4.1 Oceanic Volcanism

The petrology and geochemistry of igneous rocks from oceanic islands has, deservedly, received attention out of proportion to the small percentage of the Earth's crust which they represent. Until the advent of the Deep Sea Drilling Project, islands presented virtually the only opportunity to sample over 70% of the Earth's surface - that covered by the oceans. With the recent sampling of igneous material from the ocean floors, however, the previously held assumption that oceanic island chemistry was representative of the whole of the oceanic crust has had to be abandoned. For, as noted in Chapter 2, the compositions of ocean floor and oceanic island rocks are significantly different, implying that the source regions and/or the conditions of magma genesis for the two provinces are distinct.

Compared to the tholeiitic basalts commonly dredged or drilled from the ocean floor, the rock types generally found on oceanic islands (excluding those associated with subduction environments) are enriched in incompatible elements - K,Ti,P,Ba,Sr,Rb,Zr,Hf,U and Pb and the lighter rare earth elements - and are characteristically less silica saturated. Experimental petrology (see Green and Ringwood, 1967; Green, 1970a,b for a review) supports the view that the major element characteristics of basaltic magmas of both oceanic provinces can be derived by appropriate degrees of partial melting of source material of peridotitic composition (pyrolite) at differing conditions of pressure and temperature. The predominant olivine tholeiite of the abyssal oceanic crust may result from a 25-30% partial melt of pyrolite at shallow depth (10-30 km), whereas the more alkali-rich shield building basalts, predominant in the majority of oceanic provinces, may result from a 20-30% melting (Hawaiian

olivine tholeiite) or 10-15% melting (alkali olivine basalt) of the same pyrolite parent at greater depth (50-70 km). Studies of epicentral locations (Eaton, 1962, 1967) give empirical evidence that olivine tholeiitic magmas erupted at Hawaii have segregated from a source at a depth of at least 60 km.

Varying P, T, P_{H_2O} and percent partial melting can account for differences in the major element chemical compositions of the two oceanic provinces but does not explain differences in the incompatible elements or the differences evident in studies of the isotopic composition of strontium (section 4.3). To complete the model, mantle source heterogeneities must exist. Abyssal (ocean floor) basalts must be generated from a mantle previously depleted in large radius cations (Kay et al., 1970; Gast, 1968; Armstrong, 1968), while oceanic island basalts are derived from an enriched (or relatively undepleted) mantle source region. This argument will be resumed after discussion of some of the specific aspects of French Polynesian geochemistry.

4.2 Chemical variation in French Polynesian lavas

As reviewed in Chapter 2, the petrology and geochemistry of French Polynesian igneous rocks, particularly those exhibited at Tahiti, have been studied since the beginning of this century. Previous investigations (Lacroix, 1904, 1910, 1923, 1928; Chubb, 1927, 1930; Williams, 1933; Stark and Howland, 1941; McBirney and Aoki, 1968; Bishop and Woolley, 1973) have concentrated on the petrogenesis of the igneous suites at individual islands. In spite of this activity, many islands have received little or no attention.

Results of the geochronological investigation (Chapter 3) suggest that the island members of a particular chain are manifestations of a common volcanic mechanism. The variation in chemistry along a

Table 4-1 Chemical analyses for volcanic samples from the Marquesas Islands

OXIDE	73-5	73-9	73-11	73-12	73-13	73-14	73-16	73-18	73-19	73-20	73-22	73-23	73-25	73-27	73-28
wt%															
SiO2	45.03	46.10	49.27	49.06	47.20	47.38	46.24	46.70	49.23	48.63	47.63	45.72	62.68	47.46	46.98
TiO2	3.44	2.95	3.18	3.52	2.96	2.56	3.96	2.98	3.61	3.30	3.83	3.41	0.58	3.35	4.27
Al2O3	14.91	12.35	13.10	13.40	11.68	10.45	14.62	15.06	13.88	15.19	15.95	14.53	17.91	14.78	15.56
FE2O3	2.00	2.02	1.82	1.97	2.04	2.02	2.04	1.78	2.06	1.97	1.77	1.98	0.54	1.95	1.80
FeO	10.00	10.12	9.09	9.85	10.21	10.12	10.19	8.92	10.28	9.84	8.83	9.89	2.70	9.74	9.02
MNO	0.17	0.24	0.19	0.22	0.17	0.26	0.20	0.19	0.20	0.14	0.24	0.17	0.16	0.14	0.23
MGO	6.88	10.27	7.02	5.64	10.90	13.80	5.03	6.15	5.33	5.45	3.73	6.05	0.84	6.04	3.68
CAO	10.33	10.03	11.09	11.26	10.14	9.60	9.65	8.93	11.16	8.57	9.77	9.01	0.18	8.47	9.11
NA2O	3.14	2.19	2.46	2.52	2.08	2.02	2.88	3.21	2.66	3.21	2.79	3.08	6.32	3.36	3.42
K2O	1.32	1.27	0.70	0.83	0.45	0.77	1.68	1.94	0.71	1.73	2.42	2.35	5.54	2.33	2.32
P2O5	0.52	0.38	0.31	0.35	0.29	0.26	0.53	0.67	0.41	0.61	0.57	0.56	0.12	0.67	0.72
LOSS	1.63	1.18	0.72	0.28	0.68	-0.17	1.32	1.88	0.34	0.64	1.78	2.25	1.71	1.86	1.63
MG NO.	55.7	64.4	57.9	50.5	65.5	70.3	52.7	55.1	48.0	49.7	42.9	52.2	35.7	52.5	42.1
CIPW NORMS															
Q			0.81	1.31					1.30				1.71		
OR	7.98	7.66	4.21	4.97	2.71	4.58	10.23	11.87	4.21	10.36	14.66	14.35	33.55	14.00	14.11
AB	18.36	18.92	21.19	21.62	17.93	17.22	25.11	24.67	22.61	27.53	24.20	18.85	54.81	24.07	25.83
AN	23.21	20.54	23.04	23.11	21.61	17.30	22.67	21.70	23.94	22.22	24.45	19.51	0.11	18.68	20.85
NE	4.77							1.87				4.38		2.63	2.15
DI	21.01	22.44	25.14	25.61	22.39	23.29	19.01	16.05	23.87	13.83	17.69	18.76		16.06	17.19
HY		1.91	16.04	12.87	16.35	9.62	0.29		13.20	7.08	1.04		6.09		
OL	13.76	18.90			9.57	19.52	10.60	13.66		8.27	6.49	13.13		13.61	7.09
MT	2.97	2.99	2.69	2.90	3.01	2.95	3.05	2.67	3.00	2.89	2.63	2.97	0.80	2.88	2.69
IL	6.68	5.72	6.15	6.78	5.73	4.90	7.75	5.86	6.89	6.35	7.46	6.69	1.13	6.47	8.35
AP	1.26	0.92	0.75	0.84	0.70	0.62	1.29	1.64	0.98	1.46	1.38	1.37	0.29	1.61	1.76
SR	618	480	381	414	346	344	551	736	419	611	771	611	195	682	795
RB	56	38	14	20	9	16	36	78	12	47	53	74	142	65	54
K/RB	195	277	414	344	414	398	230	206	490	305	378	263	323	297	356
ppm															

Table 4-1 Chemical analyses for volcanic samples from the Marquesas Islands

OXIDE	73-29	73-31	73-32	73-33	73-34	73-40	73-41	73-42	73-43	73-45	73-47
wgt %											
SiO2	46.95	47.40	48.33	50.69	45.94	46.08	47.93	48.97	49.22	49.38	48.87
TiO2	4.23	4.66	3.36	3.04	3.69	3.29	4.40	3.87	4.03	3.32	3.66
Al2O3	15.98	15.16	13.60	16.07	13.43	12.17	14.72	13.94	14.55	15.64	15.67
Fe2O3	2.02	2.16	1.92	1.68	2.05	2.28	2.26	1.95	2.12	1.93	1.95
FeO	10.12	10.83	9.61	8.42	10.26	11.39	11.31	9.73	10.58	9.63	9.73
MnO	0.22	0.22	0.20	0.22	0.21	0.15	0.16	0.24	0.15	0.19	0.20
MgO	3.78	3.68	7.62	2.71	8.85	10.41	3.97	6.47	5.27	4.73	5.54
CaO	9.54	9.31	8.87	7.38	10.61	10.99	9.25	10.26	10.37	8.05	10.16
Na2O	3.64	3.74	3.23	3.90	2.22	1.96	3.12	2.81	3.00	3.98	2.85
K2O	2.18	2.58	1.25	2.81	1.32	1.13	1.14	1.29	1.34	1.77	1.25
P2O5	0.66	0.79	0.57	0.72	0.41	0.38	0.58	0.54	0.50	0.80	0.78
LOSS	-0.19	0.14	0.25	1.09	1.10	0.95	0.79	-0.21	-0.03	0.20	0.53
MG NO.	39.9	37.7	58.6	36.4	60.6	61.9	38.5	54.2	47.0	46.7	50.4
CIPW NORMS											
Q							0.05				
OR	12.97	15.16	7.49	17.00	7.88	6.66	6.81	7.62	7.83	10.52	7.34
AB	22.39	21.58	27.72	33.78	18.97	16.54	26.70	23.75	25.09	33.86	23.95
AN	20.96	16.86	19.19	18.47	23.01	21.02	23.05	21.59	22.02	19.69	26.09
NE	4.67	5.35									
DI	18.54	19.91	17.61	11.88	22.25	25.06	16.26	20.99	21.08	12.55	15.50
HY			5.60	3.15	0.48	1.10	13.96	10.45	7.90	0.29	12.79
OL	7.88	7.36	11.72	5.57	16.36	19.19		4.16	4.30	12.04	2.79
MT	2.95	3.11	2.82	2.49	3.00	3.30	3.31	2.82	3.04	2.81	2.81
IL	8.09	8.80	6.47	5.91	7.08	6.23	8.45	7.34	7.57	6.34	6.90
AP	1.57	1.86	1.37	1.75	0.98	0.90	1.39	1.28	1.17	1.91	1.83
ppm											
SR	854	870	651	659	578	500	550	534	548	847	511
RB	59	67	22	75	31	28	14	28	30	42	15
K/RB	306	319	470	310	352	334	674	381	370	349	690

Table 4-2 Chemical analyses for volcanic samples from the Society Islands

OXIDE	73-56	-72	-83	-95	-101	-111	-124	-140	-141	-153	-185	-186	-189	-190	-192
wt %															
SI02	47.40	59.56	43.24	47.80	46.12	48.73	61.96	46.14	45.15	60.48	48.28	45.67	46.77	47.50	48.72
TI02	3.00	0.30	2.58	2.75	2.40	2.62	0.49	3.20	3.82	0.79	3.68	3.28	3.29	2.66	3.02
AL203	13.46	17.64	9.76	12.80	11.45	13.09	18.24	16.81	14.29	17.90	13.35	11.83	12.11	12.55	12.96
FE203	1.84	0.88	2.20	1.98	2.03	1.79	0.50	1.75	2.11	0.67	1.79	2.11	2.08	2.07	1.96
FE0	9.18	4.40	11.02	9.89	10.13	8.96	2.51	8.76	10.54	3.35	8.93	10.53	10.42	10.33	9.78
MNO	0.14	0.33	0.17	0.15	0.15	0.14	0.23	0.17	0.16	0.20	0.14	0.16	0.15	0.15	0.14
MG0	8.72	1.19	15.06	9.60	14.04	8.84	1.28	4.92	8.18	1.53	7.60	10.45	10.09	10.76	9.21
CA0	9.02	1.40	9.71	9.22	9.19	8.84	1.22	8.86	9.28	1.67	7.82	9.73	9.12	9.36	9.32
NA20	2.87	6.49	2.80	2.45	2.36	2.70	7.43	4.01	3.22	7.19	3.70	2.01	3.41	2.40	2.59
K20	2.52	5.49	1.15	1.98	1.51	2.03	5.05	2.28	1.86	4.94	2.30	1.76	0.84	1.54	1.98
P205	0.65	0.22	0.38	0.54	0.41	0.52	0.21	1.56	0.80	0.41	0.81	0.50	0.57	0.37	0.46
LOSS	1.49	1.82	1.43	1.01	0.54	0.35	0.93	1.95	0.80	0.76	1.98	1.02	1.09	0.83	-0.09
MG NO.	62.9	32.5	70.9	63.4	70.9	63.7	47.6	50.0	58.0	44.9	60.3	63.9	63.3	65.0	62.7
CIPW NORMS															
OR	15.07	33.13	6.93	11.80	8.94	12.20	30.10	13.67	11.05	29.44	13.81	10.61	5.02	9.13	11.68
AB	18.50	47.45	6.53	20.90	14.51	23.24	54.43	23.36	17.05	51.50	27.28	17.34	25.45	20.37	21.88
AN	16.59	2.85	10.87	18.23	16.22	17.91	1.53	21.45	19.15	2.00	13.23	18.42	15.43	18.98	17.86
NE	3.29	4.68	9.54		2.98		4.87	6.00	5.60	5.34	2.45		2.02		
DI	19.73	2.37	28.71	19.71	21.52	18.82	2.67	10.52	17.79	3.05	16.82	22.22	21.49	20.37	20.49
HY				1.46		4.34						0.31		1.20	2.29
OL	16.80	7.10	28.24	18.46	27.34	14.52	4.24	12.51	17.08	5.20	14.72	20.42	19.85	21.00	16.14
MT	2.70	1.30	3.25	2.89	2.95	2.64	0.73	2.57	3.08	0.98	2.64	3.12	3.05	3.01	2.84
IL	5.76	0.58	5.00	5.27	4.57	5.06	0.94	6.17	7.29	1.51	7.10	6.35	6.32	5.07	5.73
AP	1.56	0.53	0.92	1.29	0.97	1.25	0.50	3.75	1.91	0.98	1.95	1.21	1.37	0.88	1.09
SR	733	128		579	508	577	197	1935	979	334	837	613	677	519	536
RB	53	152		31	38	43	178	79	49	116	112	36	84	34	44
K/RB	394	299		529	329	391	235	239	314	353	170	405	83	375	372

Table 4-2 Chemical analyses for volcanic samples from the Society Islands

OXIDE	73-196	-199	-201	-204	-206	-207	-209	-211	-215	-225	-233	-234	-248	74-414	73-257
wgt % SiO ₂	47.04	47.50	44.92	45.74	48.41	46.62	48.90	55.00	52.32	55.45	52.28	46.79	47.89	49.90	44.80
TiO ₂	3.88	2.98	3.06	2.76	3.08	2.13	3.52	1.79	1.65	0.87	1.64	3.04	3.69	2.13	2.90
Al ₂ O ₃	12.52	14.40	11.20	11.18	14.69	11.08	15.23	17.35	15.28	17.26	15.17	13.23	15.75	14.57	11.33
Fe ₂ O ₃	1.99	1.89	2.15	2.15	1.86	2.69	1.88	1.38	1.77	1.46	1.74	2.10	2.08	1.79	2.09
FeO	9.97	9.47	10.74	10.75	9.28	13.43	9.41	6.89	8.83	7.31	8.71	10.48	10.40	8.93	10.47
MnO	0.13	0.15	0.17	0.17	0.15	0.22	0.15	0.17	0.16	0.16	0.16	0.16	0.16	0.16	0.17
MgO	8.75	8.32	12.35	12.47	6.13	11.62	4.82	2.80	5.06	1.99	5.25	9.42	5.47	6.78	11.87
CaO	8.60	8.36	9.41	9.21	8.34	8.87	8.56	4.79	5.78	2.99	5.69	9.84	8.46	7.82	9.46
Na ₂ O	2.81	3.23	1.63	2.29	3.40	2.38	3.30	4.96	3.73	4.79	4.04	2.69	3.23	3.13	2.15
K ₂ O	1.98	1.91	1.59	1.62	1.96	0.76	2.32	3.66	3.85	5.28	3.93	1.42	1.94	3.25	1.44
P ₂ O ₅	0.61	0.56	0.44	0.42	0.67	0.23	0.69	1.07	0.55	0.64	0.54	0.48	0.53	0.52	0.43
LOSS	1.09	1.16	1.85	1.21	2.21	-0.43	0.90	0.98	1.46	0.93	0.73	0.02	0.13	2.19	1.50
MG NO.	60.9	61.0	67.2	67.4	54.1	60.7	47.7	42.0	50.5	32.7	51.8	61.6	48.4	57.5	66.9
CIPW NORMS															
OR	11.90	11.42	9.62	9.69	11.82	4.49	13.87	21.65	22.98	31.76	23.42	8.42	11.51	19.60	8.76
AB	24.14	23.99	14.12	15.69	29.36	20.13	28.26	42.00	31.26	38.70	29.63	20.29	27.43	26.89	16.12
AN	15.97	19.38	18.99	15.63	19.42	17.30	20.13	14.28	13.72	10.18	11.75	19.90	22.83	16.43	17.51
NE	0.02	1.99		2.13					0.33	1.39	2.62	1.38		0.07	1.41
DI	18.90	15.33	20.85	22.47	14.99	20.57	15.05	2.09	9.60	0.56	10.86	21.05	13.02	11.95	22.52
HY			3.49		0.33	4.32	1.78	4.10					0.12		
OL	17.16	18.04	22.73	24.92	13.75	24.71	9.74	7.94	15.03	12.03	14.75	18.98	13.77	17.02	23.83
MT	2.93	2.77	3.19	3.16	2.75	3.90	2.76	2.00	2.59	2.15	2.54	3.05	3.03	2.65	3.12
IL	7.50	5.73	5.95	5.31	5.97	4.04	6.77	3.40	3.17	1.68	3.14	5.79	7.03	4.13	5.67
AP	1.47	1.34	1.07	1.01	1.62	0.54	1.65	2.54	1.32	1.54	1.29	1.14	1.26	1.26	1.05
SR	698	667	519	575	723	242	679	838	553	305	361	574	622	402	453
RB	27	57	41	40	35	20	47	72	125	120	135	30	49	84	32
K/RB	607	277	321	335	464	315	409	421	255	364	241	392	328	320	372

Table 4-2 Chemical analyses for volcanic samples from the Society Islands

OXIDE	73-295	-318	-332	-362	-367	-396	74-419	-429	-416	-417	-422	-423
wt %												
SiO ₂	47.51	48.01	44.91	60.11	44.53	46.04	43.48	44.87	42.37	42.98	42.78	52.23
TiO ₂	2.74	3.69	3.00	0.72	3.59	3.80	4.87	4.12	3.15	4.43	3.74	1.43
Al ₂ O ₃	13.66	14.94	10.91	18.49	13.82	15.98	14.34	12.07	9.72	15.63	11.83	20.60
Fe ₂ O ₃	2.04	1.95	2.26	0.57	2.24	2.06	2.54	2.30	2.29	2.14	2.34	0.83
FeO	10.18	9.73	11.29	2.87	11.18	10.31	12.68	11.50	11.46	10.68	11.69	4.14
MnO	0.15	0.15	0.18	0.22	0.18	0.17	0.18	0.18	0.18	0.20	0.19	0.17
MgO	8.40	5.28	13.57	1.30	7.73	5.71	6.37	8.78	15.07	6.04	10.70	2.09
CaO	9.02	10.78	9.54	1.73	11.25	9.82	11.33	12.27	10.54	10.18	11.73	3.63
Na ₂ O	2.92	2.93	2.24	6.64	2.88	3.46	2.88	2.66	1.41	4.01	2.73	7.80
K ₂ O	1.29	1.78	1.47	5.68	0.98	2.02	1.45	1.50	0.73	1.91	1.43	4.84
P ₂ O ₅	0.52	0.54	0.38	0.17	0.58	0.52	0.60	0.56	0.45	1.90	0.65	0.33
LOSS	1.06	0.41	-0.38	1.60	0.76	0.19	-0.40	-0.30	1.97	1.09	0.53	1.20
MG NO.	59.5	49.2	68.2	44.7	55.2	49.7	47.2	57.6	70.1	50.1	61.9	47.4
CIPW NORMS												
OR	7.74	10.54	8.71	34.07	5.85	11.95	8.50	8.80	4.43	11.26	8.46	29.15
AB	25.09	23.19	11.80	46.02	15.96	16.84	11.31	9.29	9.91	14.20	3.79	22.32
AN	20.67	22.40	15.41	3.93	22.11	22.12	21.75	16.70	18.52	18.97	15.83	7.04
NE		0.89	3.90	5.97	4.69	6.75	6.97	6.79	1.27	10.65	10.48	24.35
DI	17.37	22.70	23.81	3.05	24.87	19.08	24.81	32.58	25.84	15.59	30.96	7.53
HY	3.09											
OL	16.49	9.14	26.48	4.33	14.96	11.82	12.40	13.45	29.38	13.34	18.43	4.82
MT	3.00	2.83	3.28	0.84	3.28	2.99	3.66	3.31	3.41	3.10	3.40	1.23
IL	5.29	7.02	5.71	1.39	6.89	7.22	9.18	7.76	6.14	8.40	7.11	2.77
AP	1.25	1.28	0.90	0.41	1.39	1.23	1.41	1.32	1.09	4.49	1.54	0.80
ppm												
SR	539	640	446	578	675	854	669	601	782	1480	750	1338
RB	26	38	36	133	50	56	29	31	22	43	35	158
K/RB	411	388	338	354	162	299	414	401	275	368	338	254

Table 4-3 Chemical analyses for volcanic samples from the Austral Islands

	OXIDE	74-386	-388	-390	-392	-394	-396	-397	-399	-401	-402	-403	-404	-406	-409	-410	-412
wt%																	
	SiO2	44.01	41.99	43.93	45.38	43.01	44.29	55.05	44.77	43.44	43.89	44.51	41.15	46.22	47.90	43.02	44.81
	TiO2	2.80	4.52	3.17	3.53	3.55	2.91	0.53	2.59	2.43	2.27	2.80	2.59	1.84	2.34	2.60	2.15
	Al2O3	14.45	14.75	15.15	15.26	12.70	14.77	17.03	12.67	9.57	11.08	14.64	11.96	10.60	14.70	11.79	11.02
	Fe2O3	2.39	2.68	2.42	2.57	2.48	2.23	1.41	2.31	2.27	2.25	2.25	2.41	2.13	2.12	2.29	2.26
	FeO	11.94	13.41	12.08	12.86	12.38	11.13	7.03	11.54	11.34	11.24	11.23	12.06	10.58	10.58	11.46	11.28
	MnO	0.25	0.27	0.26	0.29	0.27	0.24	0.32	0.28	0.26	0.27	0.28	0.32	0.24	0.24	0.28	0.25
	MgO	7.24	7.64	6.29	5.93	10.33	7.89	1.38	8.90	14.45	11.97	7.45	9.98	14.33	7.89	11.32	13.61
	CaO	12.64	8.54	11.68	7.38	9.80	10.72	2.59	12.93	12.08	12.14	12.06	10.73	9.97	10.02	12.70	10.14
	Na2O	2.39	4.46	3.04	4.80	3.50	2.88	8.58	2.63	1.80	1.88	3.03	5.32	1.99	2.85	2.38	2.63
	K2O	0.75	1.66	0.73	1.45	1.08	1.03	3.11	0.75	0.52	0.60	1.01	1.37	0.65	0.79	0.61	0.75
	P2O5	0.34	0.92	0.44	0.90	0.67	0.54	0.34	0.45	0.51	0.39	0.49	1.00	0.31	0.37	0.35	0.36
	LOSS	1.58	-0.24	1.35	0.58	0.72	1.06	3.21	0.81	0.70	1.56	1.09	0.62	0.39	0.35	0.76	-0.12
	MG NO.	51.9	50.4	48.1	45.1	59.8	55.8	25.9	57.9	69.4	65.5	54.2	59.6	70.0	57.1	63.8	68.2
CIPW NORMS																	
	OR	4.47	9.72	4.35	8.53	6.39	6.11	18.87	4.44	3.11	3.62	5.98	2.41	3.88	4.68	3.65	4.46
	AB	11.13	9.40	14.17	23.37	10.65	11.99	42.05	10.02	7.88	10.34	10.96		17.03	24.16	5.16	12.84
	AN	26.69	15.19	25.74	15.75	15.78	19.91		20.58	16.71	20.43	23.41	4.76	18.28	25.03	19.92	16.16
	LC												4.53				
	NE	5.01	15.17	6.37	9.26	10.31	11.35	16.32	6.65	4.09	3.19	7.98	24.65			8.24	5.19
	DI	28.23	17.14	24.62	12.34	23.23	24.49	9.39	33.30	32.76	31.15	27.37	34.65	23.94	18.30	33.70	25.97
	HY													2.64	4.29		
	OL	14.81	18.87	14.10	18.25	21.69	16.08	8.37	15.66	26.21	22.61	14.53	18.12	26.83	15.14	20.13	27.11
	MT	3.49	3.85	3.54	3.71	3.60	3.24	1.05	3.35	3.33	3.33	3.27	3.53	3.12	3.08	3.36	3.30
	IL	5.36	8.51	6.07	6.68	6.76	5.55	1.03	4.93	4.68	4.40	5.33	4.97	3.53	4.45	5.00	4.11
	AP	0.81	2.16	1.05	2.12	1.59	1.28	0.83	1.07	1.22	0.94	1.16	2.39	0.74	0.88	0.84	0.86
ppm	SR	434	907	520	1044	576	599	1402	537	441	487	571	1172	409	422	486	478
"	RB	15	36	17	31	18	21	160	17	14	14	23	44	16	17	12	20
	K/RB	414	382	355	587	497	409	161	365	307	355	363	258	336	385	421	310

Olivine mugearite, Moorea, Society Islands. Olivine phenocrysts in a feldspathic basalt matrix. This rock type results from the removal of olivine (primarily) from olivine basaltic melts represented by the common shield lavas. (Magnification $\approx$ X140). Crossed nicols.

Olivine basalt, Maupiti, Society Islands. Phenocrysts of olivine and clinopyroxene in a basaltic matrix. These basalts, together with ankaramites and basanites make up roughly 95% of the subaerial edifice of a typical French Polynesian island. (Magnification $\approx$ X140). Crossed nicols.

Kaersutite essexite, Tahiti, Society Islands. Euhedral amphibole with potassium feldspar, titanite, plagioclase, opaque mineral, apatite and sphene. Plutonic rocks outcrop at the centres of both Tahiti-Nui and Tahiti-Iti and are direct equivalents of the surface lavas. (Magnification $\approx$ X140). Crossed nicols.

Trachyte, Tubuai, Austral Islands. Potassium feldspar, plagioclase, aegirine-augite and opaque minerals in a typical fluidal texture. These highly differentiated rocks occur at most of the French Polynesian volcanic islands. (Magnification $\approx$ X140). Crossed nicols.

Xenolith-bearing basalt, Taiohae, Nuku Hiva, Marquesas Islands. Peridotitic nodule in a fine-grained basaltic matrix. Such xenolith-bearing flows have been found at most French Polynesian Islands. (Magnification $\approx$ X140). Crossed nicols.

Palagonite tuff, Tahuata, Marquesas Islands. Shards of olivine, clinopyroxene and plagioclase in a matrix of calcite and glass. These tuffs commonly outcrop along fault scarps in the Marquesas Islands. (Magnification $\approx$ X140). Crossed nicols.

volcanic lineament provides a measure of the variability of the source region and may test the proposition that magma from a single source is responsible for all members of the associated island chain.

A total of 85 samples from those previously selected for age determination were chosen for chemical analysis. All islands sampled in the three island chains are represented in the analyses. Major elements for all samples were analyzed by X-ray fluorescence and flame photometry. Methods used are described in Appendix 1. Analyses appear in Tables 4-1, 2, 3 with normative compositions (C.I.P.W.) calculated after Kelsey (1965) by computer programs written by N. Ware. Brief petrographic descriptions are given in Appendix 2.

The French Polynesian lavas are predominantly basaltic and may be classified using a variety of proposed criteria. Macdonald and Katsura (1964) and Macdonald (1968) have shown that Hawaiian alkalic and tholeiitic basalts plot in separate fields on an alkali-silica diagram. Figure 4-1 presents such a plot for the samples analyzed in this study. The straight line boundary between the tholeiitic and alkalic fields is from Macdonald (1968), and the curved boundary is that proposed by Irvine and Baragar (1971). A few samples plot just into the tholeiitic field but do not exhibit the petrographic characteristics of true tholeiites (of the Hawaiian type). Thus the lavas of French Polynesia form a distinctly alkalic igneous suite. The bulk of the analyses plotted in Figure 4-1 are for shield-building lavas and cluster about an alkali basalt composition. The classic shallow level fractionation pattern from alkali basalt through hawaiite and mugearite to trachyte is displayed by differentiated lavas in all three island groups. The considerable scatter in alkali basalt composition, however, suggests that shallow fractionation from a single magma or simple group of magmas will not explain the variation

Figure 4-1 Weight percent SiO_2 plotted against weight percent $\text{Na}_2\text{O} + \text{K}_2\text{O}$ for French Polynesian samples. The straight line (from Macdonald, 1968) and the curved line (Irvine and Baragar, 1971) are proposed boundaries between alkalic (above) and tholeiitic (below) basalts. Solid triangles are samples from the Austral Islands, solid circles are samples from the Society Islands and open circles are samples from the Marquesas Islands. Data from Tables 4-1, 2, 3 are not normalized to 100%. Bold letters indicate the shallow level differentiation trend from alkalic basalt to hawaiite to mugearite to benmoreite to trachyte. Bold cross is an average pyroxene composition (McBirney and Aoki, 1968).

in chemistry seen in this province.

A useful criterion for following differentiation from a parental magma is the Mg number = $100 \times \text{Mg} / (\text{Mg} + \text{Fe}^{+2})$. Green et al. (1974) report that magmas with Mg numbers in the range 63-73 could have suffered little or no crystal fractionation since segregation as primary melts from an ultramafic source in the upper mantle. Figure 4-2 is a histogram of Mg numbers calculated for samples in this study (Tables 4-1,2,3). As can be seen, many samples are candidates for primary magmas, having coexisted, before segregation from the source region, with olivine of composition Fo_{85-90} . The variation in composition of these samples indicate that the French Polynesian lavas have been produced from a heterogeneous source, and/or a spectrum of conditions of P,T and % partial melting.

These primitive basalts have high normative nepheline and olivine contents, implying an origin at a depth of between 80 and 100 km (Green, 1973). Candidates for parental magmas for the Marquesas Islands tend to be more siliceous and less alkaline than corresponding magmas for the Society and Austral Islands, but this is a small distinction. Removal of olivine and pyroxene phenocrysts from these primary melts at shallow levels can account for main differentiation trends seen in Figure 4-1. The average composition of pyroxenes in Tahitian lavas and plutonic rocks determined by McBirney and Aoki (1968) is plotted. Because the silica contents of the pyroxenes and of the parental magmas are very close, the main effect of removal of pyroxenes is to increase the alkali content of the liquid. Olivine removal increases both silica and alkali content in the liquid, which follows the trend from alkali basalt (AB) to trachyte (T).

Yoder and Tilley (1962) suggested that basalts could be

Figure 4-2 The Mg-number of a specimen is a useful criterion for determining degree of differentiation. Specimens with Mg-numbers between 63 and 73 are unlikely to have had significant shallow fractionation and can be considered as candidates for primary melts from the upper mantle. Samples with lower Mg-numbers have more likely resulted from crystal fractionation from a primary melt at shallow levels. This histogram shows the distribution of Mg-numbers within samples from French Polynesia.

classified in terms of their normative quartz, hypersthene, diopside, olivine and nepheline. Figure 4-3 illustrates the normative compositions of the samples analyzed for this study, plotted in the system Q-Hy-Di-Ol-Ne. It is apparent from this figure that large variations exist in normative chemistry among the members of any of the three island chains and that there are differences between the average compositions of rocks from separate island chains as well. Most conspicuous is the less siliceous nature of samples from the Austral and Society Island chains compared with samples from the Marquesas Islands, most of the former being ne-normative to hy-normative, while a few of the latter are qtz-normative.

McBirney and Aoki (1968) have summarized the conclusions of petrochemical studies of Tahitian lavas. These may be generalized to the other volcanic islands of French Polynesia: most rocks are undersaturated in silica; calcium and titanium are both high, whereas magnesium is low (compared with lavas from similarly undersaturated provinces); the alkali elements content is high, with Na usually in greater abundance than K. No samples of plutonic aspect were analyzed in this investigation but previous studies (Williams, 1933; McBirney and Aoki, 1968) have revealed that the chemical features of the plutonic series are identical to those of the volcanic series.

Ternary diagrams such as Figures 4-4 and 4-5 display the variation of common chemical components through the French Polynesian lavas. In an AFM diagram (Figure 4-4) the differentiation trend from alkali basalt through hawaiiite and mugearite to trachyte may be seen for the three island chains. The trend exhibited by lavas from the Society Islands is broader than for either of the other two groups, reflecting a wider variety of shallow level fractionation environments. The trend towards iron enrichment is stronger for lavas from the

Figure 4-3. French Polynesian samples plotted in the system Ne-Ol-Di-Hy-Q. Normative compositions have been used (Yoder and Tilley, 1962) to illustrate chemical variations in basalts. Solid circles are samples from the Society Islands, open circles are samples from the Marquesas Islands and solid triangles are samples from the Austral Islands. All samples from the Society and Austral Islands are Ne-to Hy-normative, while Marquesas Islands' samples grade into Q-normative varieties. A wide variety in normative chemistry is exhibited by samples in all three groups.

Figure 4-4 French Polynesian samples plotted in the system $\text{Na}_2\text{O} + \text{K}_2\text{O} - \Sigma\text{FeO} - \text{MgO}$ (all expressed in weight percentages). Symbols are as in Fig. 4-1. The Hawaiian trend is shown as a curved line for comparison. Society Islands' samples (solid circles) do not attain the same degree of iron enrichment as those from the other two island groups.

Figure 4-5 French Polynesian samples plotted in the system Na_2O - K_2O - CaO (all expressed in weight percentages). The predominant lavas in all three island groups have approximately 10% CaO and 3 to 5% $\text{Na}_2\text{O} + \text{K}_2\text{O}$. Samples from the Austral Islands (solid triangle) may be distinguished in having higher concentration of Na_2O relative to K_2O than samples from either the Marquesas (open circles) or Society Islands (closed circles).

Austral and Marquesas Islands than for the Society Islands suggesting a lower oxygen fugacity during differentiation in a shallow magma chamber (Osborn, 1959). Indeed, the presence of abundant amphibole in Tahitian plutonic species (McBirney and Aoki, 1968) and of biotite in surface lavas (sample 73-423) suggests that the Society Islands' magmas were not dry and higher P_{H_2O} was accompanied by a relatively high oxygen fugacity.

Figure 4-5 illustrates the variation of the alkali elements, Na and K, with CaO in French Polynesian lavas. The common shield building lavas contain about 10 wgt% CaO and about 3.5 wgt% $Na_2O + K_2O$. As can be seen in this figure, these lavas are far more abundant than the differentiated species, which may contain virtually no CaO and up to 12% alkalis (sample 73-25). Bishop and Woolley distinguish two separate alkalic trends at Ua Pou in the Marquesas Islands: one of about equal concentrations of K_2O and Na_2O , and the other distinctly enriched in Na_2O . It is clear from Figure 4-5 that the proportions of K_2O and Na_2O vary and that this variation may distinguish lavas from separate island groups. Samples from the Austral Islands define a distinctly more sodic trend than do samples from the Marquesas and Society Islands in which sodium more nearly equals potassium content. In some samples from Moorea (in the Society Islands) the concentration of K_2O slightly exceeds that of Na_2O , but in general $\%K_2O < \%Na_2O < 2\%K_2O$.

In Figures 4-6, 4-7 and 4-8 major element concentrations have been plotted against that of MgO, an oxide which shows great variation and which is sensitive to the removal of olivine and pyroxene from the liquid. Differentiation in all three island groups has produced much the same variation in oxide concentrations. Alumina increases nearly linearly with decreasing weight percent MgO from about 9% Al_2O_3 in the ankaramitic lavas to 18-20% in trachytic samples.

Figure 4-6 The variation of major element oxide concentrations with the concentration of MgO (all expressed as weight percentages) in samples from the Austral Islands.

Figure 4-7 The variation of major element oxide concentrations with the concentration of MgO (all expressed as weight percentages) in samples from the Society Islands.

Figure 4-8 The variation of major element oxide concentrations with the concentration of MgO (all expressed as weight percentages) in samples from the Marquesas Islands.

Williams (1933) noted that the progressive enrichment in alumina may have been produced by the fractionation of olivine and pyroxene without significant removal of plagioclase. As evidence he has remarked on the paucity of plagioclase as a phenocryst phase in the common shield lavas at Tahiti.

As for the other major element concentrations, very little variation is observed until the MgO content decreases to a very low value. Silica and the alkalis increase dramatically in the final differentiation products, whereas titanium, iron ($\text{FeO} + \text{Fe}_2\text{O}_3$) and calcium decrease. In the Austral and Society Islands samples this inflection occurs at about 6 wgt%MgO, whereas for the Marquesas Islands samples, oxide concentrations do not change significantly until about 4 wgt%MgO, which again reflects small differences in the composition of the primitive magmas which formed these island chains.

Rb and Sr have been measured by X-ray fluorescence (description of equipment and technique in Appendix 1) and are reported in the chemical analyses in parts per million. Figure 4-9 illustrates the strong correlation between K_2O and Rb, which is evidence of the similar chemical behaviour of K and Rb in melting/crystallization processes. Strontium varies between 100 and 200 ppm whereas rubidium varies between 10 and 200 ppm, both ranges common in oceanic island basalts. The variations in major and minor elements among shield building lavas from the three island chains with Mg numbers in the range 63 to 73 indicate that no single parental magma is adequate to produce the differentiated igneous suites exhibited in French Polynesia. Instead, a variety of primary liquids, segregating under different conditions from a heterogeneous peridotitic source, were available for construction of the shield volcanoes. Alkali basalts, basanites and ankaramites are all common and comprise the major part of the shield

Figure 4-9 A log-log plot of the variation of K₂O (weight percent) with Rb (parts per million). The high correlation between K and Rb is not surprising in view of their similar chemical behaviour.

building lavas. The presence of ankaramite suggests that olivine and pyroxene fractionation was important in the differentiation of parental magmas at shallow levels, yielding high wgt%MgO cumulate lavas and residual liquids trending towards silica enrichment.

From this investigation of major element chemistry and the variation of oxide concentrations within the three French Polynesian island chains it may be concluded that all islands exhibit lavas which are of the alkali olivine basalt association. Lavas from the Marquesas Islands are generally more siliceous than at either the Society or Austral Islands. The Austral Islands samples may be distinguished by the predominance of Na_2O over K_2O . Variation, however, in major element chemistries for lavas from within island groups and, indeed, from individual islands require multiple sources from a heterogeneous peridotitic source region.

4.3 The isotopic composition of strontium in French Polynesian Lavas

While major element chemistry may be very similar for separate igneous provinces, reflecting common melting conditions from a peridotitic parent, isotopic chemistry can be diagnostic in exposing differences between the source regions. This proposition relies on the assumption that isotopes are not fractionated during melting so that the isotopic composition of a liquid will be the same as that of the source region. The isotopic composition of strontium is particularly useful. Because a certain isotope of strontium, Sr^{87} , accumulates with time from the radioactive decay of Rb^{87} , the isotopic composition of a melt is the time-integrated history of the concentration of rubidium in the source region.

In the case of the decay of Rb^{87} to Sr^{87} , the amount of Sr^{87} relative to Sr^{86} (an isotope whose amount does not change with time) in a chemical system is determined by an initial ratio, $[\text{Sr}^{87}/\text{Sr}^{86}]_0$,

Figure 4-10 The relationship which governs the time accumulation of Sr^{87} from the radioactive decay of Rb^{87} implies that all closed systems of age t which had a common initial $\text{Sr}^{87}/\text{Sr}^{86}$ will plot along a straight line (isochron) on a diagram of $\text{Sr}^{87}/\text{Sr}^{86}$ vs $\text{Rb}^{87}/\text{Sr}^{86}$. The slope of this line yields the common age of the chemical systems and the intercept of the line gives the initial ratio common to those systems. If the systems had different initial ratios but are of the same age, they will plot on parallel lines in such a diagram.

which characterized the system at the last homogenization of the strontium isotopes, plus an amount of radiogenic Sr^{87} which has accumulated from the decay of Rb^{87} , also normalized to the reference amount of Sr^{87} . The increase in $\text{Sr}^{87}/\text{Sr}^{86}$ is the following function of time, t , and of the concentration of $\text{Rb}^{87}/\text{Sr}^{86}$ in the system:

$$[\text{Sr}^{87}/\text{Sr}^{86}]_t = [\text{Sr}^{87}/\text{Sr}^{86}]_0 + [\text{Rb}^{87}/\text{Sr}^{86}]_t (e^{\lambda t} - 1).$$

This relationship is valid only for a chemical system which has remained closed to gain or loss of Rb and Sr since the system last equilibrated. Considered over comparable periods of time, a region which has been enriched in Rb (relative to Sr) will produce melts with higher $\text{Sr}^{87}/\text{Sr}^{86}$ than a region which has been depleted in Rb (/Sr). Figure 4-10 illustrates the Rb-Sr systematics.

To characterize the source region for the island chains of French Polynesia, volcanic rocks from the Austral, Marquesas and Society Islands have been analyzed for the isotopic composition of strontium using a modified Nuclide Analysis Associates (30 cm radius of curvature, thermal ionization) mass spectrometer. Particulars of the equipment and methods used are given in Appendix 1.

Sixty samples were selected for determination of strontium isotopic composition. The results of this investigation are presented in Table 4-4 and illustrated in Figure 4-11. Each island sampled for geochronology (Chapter 3) is represented in this study. In addition the concentrations of Rb and Sr have been determined by X-ray fluorescence (see Appendix 1 for description). Swainbank (1968) has determined the isotopic composition of strontium from 27 French Polynesian samples, but does not report Rb/Sr as well. His values for $\text{Sr}^{87}/\text{Sr}^{86}$ are compatible with data reported in Table 4-4.

The isotopic composition of strontium has been used to distinguish abyssal tholeiites from the more alkalic lavas of oceanic

Table 4-4: Isotopic composition of strontium from lavas of French
Polynesia

<u>Sample No.</u>	<u>Rb (ppm)</u>	<u>Sr (ppm)</u>	<u>Rb⁸⁷/Sr⁸⁶</u>	<u>Sr⁸⁷/Sr⁸⁶*</u>
<u>Marquesas Islands</u>				
Nuku Hiva				
73-11	14	381	0.11	0.70392
73-13	9	346	0.08	0.70815
				0.70823
73-14	15	344	0.13	0.70445
73-16	36	551	0.19	0.70422
73-22	53	771	0.20	0.70431
73-23	74	611	0.35	0.70472
73-25	142	195	2.11	0.70440
Ua Huka				
73-5	56	618	0.26	0.70570
				0.70555
73-9	38	480	0.23	0.70656
Hiva Oa				
73-28	54	795	0.20	0.70478
73-29	59	854	0.20	0.70476
73-31	67	870	0.22	0.70483
73-33	75	658	0.33	0.70523
73-34	31	578	0.15	0.70470
Tahuata				
73-40	28	500	0.16	0.70411
73-41	14	550	0.07	0.70379

* $\text{Sr}^{87}/\text{Sr}^{86}$ corrected to $\text{Sr}^{88}/\text{Sr}^{86}=8.375$ and normalized using standard NBS-987 (see Appendix 1).

<u>Sample No.</u>	<u>Rb (ppm)</u>	<u>Sr (ppm)</u>	<u>Rb⁸⁷/Sr⁸⁶</u>	<u>Sr⁸⁷/Sr⁸⁶*</u>
73-43	28	534	0.15	0.70388
Fatū Hiva				
73-45	42	847	0.14	0.70331
73-47	15	511	0.09	0.70362
73-49	39	764	0.15	0.70406
<u>Austral Islands</u>				
Rurutu				
74-386	15	434	0.10	0.70299
74-388	36	910	0.12	0.70313
74-390	17	520	0.09	0.70269
74-392	31	1044	0.09	0.70318
74-396	21	599	0.10	0.70277
Tubuai				
74-399	16	537	0.09	0.70273
74-401	14	441	0.09	0.70262
74-402	14	487	0.08	0.70277
74-403	23	571	0.12	0.70276
74-404	44	1172	0.11	0.70266
Raivavae				
74-406	16	409	0.11	0.70284
74-409	17	422	0.12	0.70295
74-410	12	486	0.07	0.70267
74-412	20	478	0.12	0.70284
<u>Society Islands</u>				0.70295
Maupiti				
73-201	41	519	0.23	0.70507
73-204	40	575	0.20	0.70515
73-207	20	243	0.23	0.70565
				0.70551

<u>Sample No.</u>	<u>Rb(ppm)</u>	<u>Sr (ppm)</u>	<u>Rb⁸⁷/Sr⁸⁶</u>	<u>Sr⁸⁷/Sr⁸⁶</u>
Bora Bora				
73-295	26	539	0.14	0.70400
73-318	38	640	0.17	0.70467
73-332	36	446	0.23	0.70507
Tahaa				
73-185	111	838	0.38	0.70670
73-189	84	677	0.36	0.70540
73-190	34	519	0.19	0.70604
73-196	27	698	0.11	0.70638
Raiatea				
73-124	178	197	2.60	0.70402
73-140	79	1935	0.12	0.70410
				0.70397
73-153	116	334	1.01	0.70409
Huahine				
73-56	53	733	0.21	0.70562
73-72	152	128	3.41	0.70486
73-111	51	529	0.28	0.70597
Moorea				
73-211	72	838	0.25	0.70500
73-215	125	553	0.65	0.70470
73-225	120	305	1.14	0.70447
73-248	49	623	0.23	0.70485
Tahiti				
73-362	133	578	0.66	0.70399
73-396	56	854	0.19	0.70454
74-423	158	1337	0.34	0.70398
73-367	50	675	0.21	0.70482

<u>Sample No.</u>	<u>Rb (ppm)</u>	<u>Sr (ppm)</u>	<u>Rb⁸⁷/Sr⁸⁶</u>	<u>Sr⁸⁷/Sr⁸⁶</u>
74-422	35	750	0.14	0.70373
73-419	29	669	0.13	0.70360

islands (Gast, 1968). The majority of ocean floor basalts exhibit values for initial $\text{Sr}^{87}/\text{Sr}^{86}$ between 0.7015 and 0.7030, averaging about 0.7025, and $\text{Rb}^{87}/\text{Sr}^{86}$ between 0.01 and 0.1 (Faure and Powell, 1972). As most estimates of the initial $\text{Sr}^{87}/\text{Sr}^{86}$ for the Earth (e.g. Faure and Powell, 1972) are close to 0.6990, $\text{Sr}^{87}/\text{Sr}^{86}$ in the upper mantle source regions for ocean floor basalts has not increased dramatically over the 4.5 billion years of Earth's history. An average $\text{Rb}^{87}/\text{Sr}^{86}$ of only 0.05 for the ocean floor basalt source region could have generated the small amount of radiogenic strontium from an initial $\text{Sr}^{87}/\text{Sr}^{86}$ of 0.6990 in 4.5 billion years.

Values of initial $\text{Sr}^{87}/\text{Sr}^{86}$ for shield lavas from oceanic islands, on the other hand, are significantly greater on average than for those from the ocean floor. Faure and Powell (1972) calculate a mean initial $\text{Sr}^{87}/\text{Sr}^{86}$ of 0.7037 for oceanic island basalts but it must be mentioned that reported $\text{Sr}^{87}/\text{Sr}^{86}$ for these lavas vary from 0.7025 to greater than 0.7070.

Gast (1968) suggested that the ocean floor basalts, with their relatively unradiogenic strontium, were derived from regions of the upper mantle which had previously produced alkali basalts in earlier melting episodes and were, therefore, depleted in Rb/Sr (assuming Rb is partitioned in melts at a greater rate than Sr). Peterman and Hedge (1971) propose that the upper mantle is progressively losing rubidium to the crust, so that rubidium-rich regions of the mantle have suffered least loss and are in that sense the most 'primitive'. By this reasoning oceanic island basalts have sampled more rubidium-rich (relative to strontium) regions (more 'primitive') than ocean floor basalts (less 'primitive'). Regardless of whether oceanic island basalts have come from a more or less 'primitive' source than that for ocean floor basalts

Figure 4-11 The variation of $\text{Sr}^{87}/\text{Sr}^{86}$ with $\text{Rb}^{87}/\text{Sr}^{86}$ for samples from French Polynesia. Triangles are samples from the Austral Islands, circles are samples from the Society Islands and squares are samples from the Marquesas Islands. Filled symbols are samples with high Mg-numbers and likely to be primary melts. If the correlation between $\text{Sr}^{87}/\text{Sr}^{86}$ and $\text{Rb}^{87}/\text{Sr}^{86}$ is interpreted as an isochron, then an age of 1.2 ± 0.18 billion years best fits this data. An initial ratio of 0.7015 would have then been common to all parts of the source region for French Polynesian lavas. A 1.4 billion year isochron is shown for reference.

○ average seawater

·7090

·7080

·7070

·7060

·7050

·7040

·7030

·7020

·7010

·7000

$\text{Sr}^{87}/\text{Sr}^{86}$

1.4 Billion Years

1.2 Billion Years

$\text{Rb}^{87}/\text{Sr}^{86}$

0.05

0.10

0.15

0.20

0.25

two separate source regions must exist and must have maintained differences in Rb/Sr (i.e. have not mixed) for a long period of time.

Values for initial $\text{Sr}^{87}/\text{Sr}^{86}$ determined for lavas from three island chains of French Polynesia show the same degree of variability as is exhibited by a world-wide distribution of oceanic islands. Table 4-4 and Figure 4-11 show that the initial ratios range from 0.7026 for some lavas from the Austral Islands, to nearly 0.7070 for samples from the Marquesas and Society Islands. From replicate analysis of five samples (Table 4-4) and multiple analysis of standard NBS 987 (e.g. Roddick, 1974), the uncertainty in determination of $\text{Sr}^{87}/\text{Sr}^{86}$ is ± 0.00006 (1σ).

Initial ratios for samples from the Austral Islands vary between 0.7026 and 0.7032 and are significantly lower than those analyzed from the Society and Marquesas Islands. In addition, the Austral Islands' samples exhibit a more narrow range than those from either the Marquesas or Society Islands. Klerkx et al. (1974) have suggested that low initial $\text{Sr}^{87}/\text{Sr}^{86}$ for lavas from the Cape Verde Islands might be explained by an enrichment of strontium in their source region, rather than a depletion in rubidium. Both potassium and rubidium are at relatively lower concentrations in samples from the Austral Islands whereas strontium concentrations are not significantly higher than in samples from the Society and Marquesas Islands. It is concluded that the source region for the French Polynesian lavas is heterogeneous in Rb/Sr, due to variations in Rb and Sr concentrations.

More surprising, however, is the strong correlation between $\text{Sr}^{87}/\text{Sr}^{86}$ and $\text{Rb}^{87}/\text{Sr}^{86}$ of shield lavas from islands in the three separate lineaments. Recall that such a relationship is to be expected in a isolated region which has maintained heterogeneities in

Rb/Sr over a length of time. Components which are enriched in Rb/Sr will increase in $\text{Sr}^{87}/\text{Sr}^{86}$ at a greater rate than those which are depleted in Rb/Sr. The slope of the linear relationship between $\text{Sr}^{87}/\text{Sr}^{86}$ and $\text{Rb}^{87}/\text{Sr}^{86}$ is proportional to the time since the last homogenization of the strontium isotopes and the intercept is the starting $\text{Sr}^{87}/\text{Sr}^{86}$, common to all components of the system at that time. Figure 4-11 plots the data of Table 4-4, less those analyses with $\text{Rb}^{87}/\text{Sr}^{86} > 0.30$ (likely products of shallow level fractionation). It is clear that the data do not lie uniquely on one line but may define a number of parallel lines, each produced from isolated chemical systems of the same age but of varying initial $\text{Sr}^{87}/\text{Sr}^{86}$. Assuming this model to be valid, the best fitting isochron to this data yields an age of 1.18 ± 0.18 (2 σ) billion years. This, providing the assumption that the lavas have been formed from a set of closed systems is a valid one, is the last time that the source region was homogenized (by melting or some wide scale diffusion process). The initial $\text{Sr}^{87}/\text{Sr}^{86}$ for the multiple sources vary between 0.7005 and 0.7020 and average 0.7015, somewhat below the present range in ocean floor basalts. Excluding the Austral Islands data which, it may be argued, form a distinct subset, a best fitting age of 0.88 ± 0.22 billion years (mean initial $\text{Sr}^{87}/\text{Sr}^{86} = 0.7024$) is obtained for the Marquesas and Society Islands data only.

Implicit in this isochron interpretation is the assumption that the ratio of rubidium to strontium in partial melts is the same as the rubidium to strontium ratio in the regions of the upper mantle from which the melts were derived. If this assumption is not valid and the $\text{Rb}^{87}/\text{Sr}^{86}$ of melts and source regions differ, then an erroneous 'age' will result in interpretation of the data by the isochron method. Figure 4-12 illustrates such a failing. The distribution coefficients for rubidium

Figure 4-12 An example of the calculation of an erroneously young age by the isochron method. Because Rb and Sr have different partition coefficients, a higher Rb/Sr is expected in liquids resulting from small degrees of partial melting than in the parental material. Measured Rb/Sr in lavas will be systematically greater than the Rb/Sr in the source region from which they come (lower diagram). The result is an isochron which underestimates the age of the heterogeneity in the source region (upper diagram).

Source Region

and strontium for melting at pressures up to 30 kbars (Shimizu, 1974) favour a higher Rb/Sr in the melt than in the source region for less than about 30% partial melting. The study of major element chemistry, discussed in the preceeding section, indicates that many of the shield lavas at islands in each of the three lineaments could be primary melts from a heterogeneous peridotitic parent. These melts are likely to have been formed under similar melting conditions (P,T,% partial melting) and so should exhibit values of Rb/Sr which are systematically greater than those of the regions from which they came.

The work of Shimizu (1974) indicates an 18% increase in Rb/Sr in melt over parent for 20% partial melting at 25 Kbars. Lavas with high Mg numbers are indicated in Figure 4-11 and define virtually the same isochron and exhibit the same range in $\text{Sr}^{87}/\text{Sr}^{86}$ as does the entire data set. As is illustrated in Figure 4-12, the effect of such a systematic increase in melt Rb/Sr compared with parent Rb/Sr is to produce an 'isochron' of younger age than the true age. Hence 1.2 billion years may be considered a minimum age for the last homogenization of the mantle parental material for French Polynesian lavas, the true age being closer to 1.4 billion years.

Another source of departure of the data in Figure 4-11 from the 'true age' isochron is the effect of crystal fractionation. The mineral plagioclase preferentially stores Sr at the expense of Rb so that the Rb/Sr for plagioclase is less than that for the whole rock. Hence a rock enriched (by crystal accumulation) or depleted (by crystal flotation) in plagioclase will lie to the left or right, respectively, of the 'true age' isochron. As noted previously, plagioclase is rare as a phenocryst mineral in French Polynesian lavas and so has probably not been important in disrupting the relationship seen in Figure 4-11. The presence of olivine and pyroxene does not seriously affect the whole

rock Rb/Sr because Rb and Sr concentrations are very low in these phases.

The strong correlation between initial $\text{Sr}^{87}/\text{Sr}^{86}$ and $\text{Rb}^{87}/\text{Sr}^{86}$ might be the result of mixing varying proportions of low $\text{Sr}^{87}/\text{Sr}^{86}$ (~ 0.7025), low $\text{Rb}^{87}/\text{Sr}^{86}$ (~ 0.05) with high $\text{Sr}^{87}/\text{Sr}^{86}$ (> 0.7070), high $\text{Rb}^{87}/\text{Sr}^{86}$ (> 0.30). Ocean floor basalt could well be the unradiogenic end member. It is difficult, however, to imagine a probable candidate for the more radiogenic end member. For isolated islands such as those of French Polynesia, varying contamination with older, more radiogenic continental material is not likely in this purely oceanic, youthful environment. Contamination with sediments will not significantly change the $\text{Sr}^{87}/\text{Sr}^{86}$ without also drastically influencing the major element chemistry. As noted previously, all of the major element chemical variation in French Polynesian lavas can be accounted for by igneous processes. Using a $\text{Sr}^{87}/\text{Sr}^{86}$ of 0.7095 for pelagic sediments (Faure and Powell, 1972), a rock with $\text{Sr}^{87}/\text{Sr}^{86} = 0.7060$ must be a 1 to 1 mix of ocean floor basalt and calcareous to argillaceous sediments. No such gross addition of Ca and Si to olivine tholeiite would be allowed by the chemistries of any of the basalts reported in the preceeding section.

Seawater has a very different isotopic composition to any of those along the trend defined by the French Polynesian initial ratios and, except for the possible exception of sample 73-13, cannot have been an important source of contamination. It is most likely, then, that contamination has not been a significant influence in the chemistry of French Polynesian lavas and that the correlation of initial $\text{Sr}^{87}/\text{Sr}^{86}$ with $\text{Rb}^{87}/\text{Sr}^{86}$ is instead characteristic of a source region which has maintained heterogeneities in Rb/Sr over a very long time.

Graham and Ringwood (1971) and Compston (1974) discuss the

possibility of disequilibrium melting with reference to variation in $\text{Sr}^{87}/\text{Sr}^{86}$ in lunar mare basalts. In their proposal, interstitial material and accessory minerals, such as phlogopite and amphibole, which preferentially store rubidium relative to strontium will have significantly higher $\text{Sr}^{87}/\text{Sr}^{86}$ than the more common mantle minerals such as olivine and pyroxene. In a partial melting event these minor phases will enter the melt first, producing higher $\text{Sr}^{87}/\text{Sr}^{86}$ and Rb/Sr in the melt than in the source material - a disequilibrium in isotopic composition. Successive partial melts from such a source region will yield a range of Rb/Sr and $\text{Sr}^{87}/\text{Sr}^{86}$, depending on the degree of heterogeneity (corresponding to age) in the source and the percentage partial melting. Such a relationship will define an isochron which measures the duration of the Rb/Sr heterogeneity in the source.

Variation in major element chemistry, however, does not support the prediction of this model that higher $\text{Sr}^{87}/\text{Sr}^{86}$ attend smaller degrees of partial melting. From Figure 4-11 it may be seen that shield lavas (with high Mg numbers) vary in initial $\text{Sr}^{87}/\text{Sr}^{86}$ between 0.7026 and 0.7066, or virtually the entire spectrum of the whole data set. A more tenable model for the correlation between $\text{Sr}^{87}/\text{Sr}^{86}$ and $\text{Rb}^{87}/\text{Sr}^{86}$ is that French Polynesian shield lavas have been generated by partial melting under roughly uniform conditions from a source region which is heterogeneous in Rb/Sr on a scale greater than crystal-matrix differences. The gross difference between the initial $\text{Sr}^{87}/\text{Sr}^{86}$ for Austral Islands samples and those for the Marquesas and Society Islands implies significant regional variation in Rb/Sr in the upper mantle.

In a study of the isotopic composition of strontium and lead in alkali basalts from oceanic islands, Sun and Hanson (1975) have

concluded that a high correlation between initial $\text{Sr}^{87}/\text{Sr}^{86}$ and $\text{Rb}^{87}/\text{Sr}^{86}$ involves preservation of chemical heterogeneities in the upper mantle source regions for these basalts over periods of from 1,000 to 3,000 Myr. They propose from their study that at present there is not mantle-wide convection and that the source regions for these oceanic basalts lie deeper than the presently convecting mantle.

The data from French Polynesia imply that the source region from which partial melts have been derived has maintained its Rb/Sr heterogeneity for a period of from 1.2 to 1.4 billion years. Such a source region must not have been involved in melting events during this time and so must be effectively isolated from the production of new crust at the mid-ocean ridges and in the subduction zones. The location of the source region could possibly be in the lowermost lithosphere or in the upper low velocity zone. If neither of these possibilities are acceptable, then parental material for French Polynesian lavas must have risen, plume-like, from below the low velocity zone. The constraints placed on source region location and volcanic mechanism by this study of the isotopic composition of strontium in French Polynesian lavas will be discussed in the concluding chapter.

Chapter 5: Geophysical Constraints

5.1 Defining an anomalous source region

Attempts at geophysically defining the sub-lithospheric structure of proposed sources of nonorogenic linear volcanism have been ambiguous at best. It has been demonstrated from gravity, seismicity and geochemistry that the earth's mantle is inhomogeneous, both laterally and radially and this heterogeneity undoubtedly bears on the origin of this planet and on its evolution as a chemical system. With Morgan's (1971, 1972) suggestion that hot spots are the surface expression of thin, upwardly convecting plumes of (relatively) warmer mantle material has come the possibility of a driving mechanism for plate motion and a method of mixing the mantle. Morgan proposes, on the basis of an assumed 2.0 billion year mantle turnover rate, an average plume diameter of 150 km with an upward velocity of 2 m/yr, extending from the core-mantle boundary up to the asthenosphere.

If, as Morgan suggests, this material rises because it is hotter and lighter than the surrounding mantle, it should be geophysically identifiable. Morgan (1972) remarks on the topographic prominence of proposed hot spots and attributes this to the buoyancy of the rising plume. Neglecting volcanic products, he estimates 1 km extra height as typical of the region near a hot spot and from this calculates the buoyancy force produced by the rising anomalous material. In addition Morgan estimates the density of plume material to be about 5 percent less than the surrounding mantle material (down to at least 1,000 km depth). It should be possible, then, to delineate the deep structure of plumes through gravity and seismology.

Referring to the external isostatic gravity field (Kaula, 1970, Gaposchkin and Lambeck, 1971; Khan, 1974) both positive and negative ($\geq + 10$ mgal) anomalies are evident. Morgan (1972) remarks that isolated gravity highs are apparent over Iceland, Hawaii, and most other

hot spots, with some exceptions (see Figure 5-1) and uses this as evidence for rising plumes in the mantle. Khan (1974), however, takes exception to this interpretation, pointing out that some of the proposed hot spots are associated with negative gravity anomalies and the correlation is not a valid one. In addition, some areas which are not associated with hot spots also exhibit high gravity anomalies. It is not clear, then, that the isostatic gravity field provides any conclusive evidence for anomalous behaviour beneath proposed sources of nonorogenic volcanism.

To investigate the character of the deep mantle beneath Hawaii, Kanasewich et al. (1972) and Kanasewich and Gutowski (1975) have recorded arrivals of p-waves at three arrays of seismometers in Canada. P-wave arrivals from events in the vicinity of the Tonga Islands have anomalously high values of phase velocity compared to values predicted by the Jeffreys-Bullen earth model. These waves pass directly beneath Hawaii and outline an anomalous region extending from the base of the mantle to at least 400 km above the core. The authors dismiss the possibility that the anomalies are due to crustal structure beneath the receivers because events from other areas yield normal phase velocities. (The observed anomalies can be associated with a cylindrical body with a high velocity nucleus having a radius of 150 km surrounded by a low velocity region. No conclusions can be drawn from this study about the character of the upper mantle beneath Hawaii.) Kanasewich et al. ascribe the higher velocities at the base of the mantle beneath Hawaii to refractory material, denser than the surrounding mantle. It will be remembered, however, that Morgan has proposed that plumes are hotter zones of less dense material which should produce delays in waves passing through them.

In a similar study of p-wave arrivals, Davies and Sheppard

Figure 5-1 The isostatic gravity field of the Earth (after Khan, 1974). Regions ≥ 10 mgals are vertically striped while regions ≤ -10 mgals are horizontally striped. Proposed hot spots are shown as solid circles. There is no clear correlation of hot spot location with positive gravity anomalies.

(1972) find lateral heterogeneities in the lower mantle which are conveniently described by azimuthal variations of incoming seismic rays. That is, departures from the great circle route from the earthquake epicentre to the receiver can be interpreted as heterogeneities along that path. (The first arrivals will 'avoid' seismically slow regions and will be channeled by seismically fast regions.) These authors note that the heterogeneities leading to azimuthal anomalies may lie anywhere along the path of the seismic rays, so that it is difficult to identify the anomalous regions. Three directions of incoming rays to LASA (Large Aperture Seismic Array, Montana, U.S.A.) exhibit particularly rapid azimuthal variation - 20° , 155° and 265° , passing beneath Iceland, the Galapagos Islands and Hawaii, respectively. The incoming rays from these three regions have avoided the great circle path by 5° which, when traced back to the bottoming points imply lateral anomalies of about 200 km (radius). In order for these rays of longer path to arrive before rays along the great circle paths, the great circle rays must be delayed about 2 seconds. The authors calculate that anomalous regions of horizontal radius 200 km and 6% lower velocity than surrounding mantle beneath Iceland, the Galapagos Islands and Hawaii will produce the azimuthal anomalies observed at LASA.

Best et al. (1974) have recorded arrivals of S-waves at Oahu which have been reflected by the core-mantle boundary (ScS) from an earthquake at Hawaii, only 300 km to the southeast. These arrivals at near vertical incidence provide an accurate estimate of total crust-mantle travel time and attenuation under Hawaii and Oahu. These authors find that the S-waves are delayed under Hawaii, relative to the surrounding mantle, by an amount depending on the reference model used. This delay implies less dense material beneath Hawaii. Okal and Anderson

(1975) have used multiply reflected ScS phases to investigate travel-time anomalies in oceanic areas. They found that the mantle under Hawaii, Iceland and Trindade-Martin Vas Islands (S. Atlantic) is anomalously slow when compared to other oceanic regions and the average Earth. This is in general agreement with Davies and Sheppard (1972) and Best et al. (1974) but in contradiction to Kanasewich et al. (1972) and Kanasewich and Gutowski (1975). So again, the geophysical character of the mantle beneath these volcanic lineaments cannot be uniquely described. Seismic anomalies do appear to exist, however, but their nature will be revealed only by further investigation.

A more indirect geophysical investigation of the lower mantle beneath Hawaii has been reported by Doell and Cox (1972). They have studied the secular variation of the geomagnetic field recorded palaeomagnetically in the lavas of the young Hawaiian Islands. Study of the long period variation of the geomagnetic field in the earth's core may reveal the conductivity structure of the lower mantle. Because the geomagnetic source function is uncertain, solutions are non-unique. In fact, it is possible that all variation in the geomagnetic field occurs in the source and not because of heterogeneities in the lower mantle. Doell and Cox (1972) propose that the geomagnetic field recorded during the last few million years in the lavas at the Hawaiian Islands has been significantly less variable than at other areas. They conclude that a lateral heterogeneity exists in the lower mantle (beneath Hawaii) which is coupled to the core in such a way as to partially prevent the generation of nondipole magnetic fluctuations with periods in the range of 200 to 2000 years.

5.2 Palaeosecular Variation

Although the geomagnetic field is dipolar and axially symmetric when averaged over tens of thousands of years, at any given instant the position of the north magnetic pole measured from any other point on the earth's surface may be offset from the rotational axis because of a combination of perturbing effects. These short time and small amplitude departures in direction (and intensity) of the geomagnetic field are termed secular variation, distinguished from the longer episodes of polar reversal and true polar wander. Perturbations from the dipole field have been observed directly over the past 400 years. These features form foci commonly of 10,000 km diameter which grow and diminish in intensity and generally drift westward over the regular dipole field. Whether or not these anomalies persist over very long periods of time is an extremely important question bearing on the nature and origin of the geomagnetic field. Another source of short term field perturbation is wobble of the dipole field about the rotational axis and its variation in intensity. These effects together contribute to the geomagnetic secular variation. In palaeomagnetic studies, which provide information about the geomagnetic field back beyond the last 400 years, they have contributed to the scatter in measured directions.

The variability of the geomagnetic field is commonly measured by either of two sets of data, one derived from the other. Palaeomagnetic field directions (I, inclination, and D, declination) may be determined directly from samples. Fluctuations in I and D are easily analysed by reference to the Fisher (1953) distribution and provide an estimate of the secular variation at a given locality over the time sampled. A more useful variable for palaeosecular variation analyses, however, is the virtual geomagnetic pole (VGP) (Cox and Doell, 1960) which is simply

that north magnetic pole which would produce the observed inclination and declination at a given latitude and longitude (assuming a perfect dipole field). The coordinates of the VGP are derived directly from the palaeomagnetic field directions, I and D , and the coordinates of the sampling, λ and ϕ , by

$$\lambda' = \sin^{-1} (\sin \lambda \cos p + \cos \lambda \sin p \cos D) \quad (90^\circ \leq \lambda \leq 90^\circ)$$

$$\phi' = \begin{cases} \phi + \beta & \text{for } \cos p \geq \sin \lambda \sin \lambda \\ \phi + 180 - \beta & \text{for } \cos p < \sin \lambda \sin \lambda \end{cases}$$

$$\text{where } \beta = \sin^{-1} (\sin p \sin D / \cos \lambda') \quad (-90^\circ \leq \beta \leq 90^\circ)$$

$$\text{and } p = \cot^{-1} (\frac{1}{2} \tan I) \quad (-0^\circ \leq p \leq 180^\circ)$$

Virtual geomagnetic poles provide a convenient method for comparing secular variation, or angular dispersion, at widely separated localities. Scatter of the VGP's can be analysed by Fisher's method or by a method proposed by Doell and Cox (1971), with very little difference. One great advantage in using VGP's is that angular dispersion resulting from wobble of the dipole field is constant with latitude whereas it is strongly latitude dependent when magnetic field directions are analysed. Significant differences in the estimated magnitude of secular variation arise between analyses employing VGP's and those using magnetic field directions so it is important to use the same type of data when comparing secular variation from separate localities and/or separate intervals of time.

The magnitude of secular variation as measured by the angular dispersion of palaeomagnetic data is latitude dependent, in accord with the latitude dependence of dipole field intensity. Models which attempt to explain this latitude dependence differ only in the relative contributions of a wobbling main dipole field of variable

intensity and a westward drifting non-dipole field (Irving and Ward, 1964; Creer et al., 1959; Cox, 1962 and Cox, 1970). Combining the two effects, angular dispersion S_F can be written as

$$S_F = (S_D^2 + S_N^2)^{1/2}$$

where S_D is the angular dispersion of VGP's produced by the dipole wobble only and S_N is the angular dispersion due to the drifting non-dipole field. Reference to such a model is necessary when comparing magnitudes of secular variation from localities of differing latitude. Studies of palaeosecular variation have not been able to distinguish, at the 95% confidence level (Cox, 1969), one model from another, and in most cases model dispersions fall within the estimated confidence limits.

In a study of palaeosecular variation of Brunhes age (< 0.7 Myr) lavas from the island of Hawaii in the central Pacific basin, Doell and Cox (1965) found a much lower dispersion of VGP's than determined in similar studies of Brunhes age samples at other localities and lower than the model predicted dispersion for that latitude (20°). The present non-dipole field in the Hawaiian area is negligible and has been so over the few centuries of direct measurement of the geomagnetic field. From these observations and the low value of palaeosecular variation determined from Hawaiian lavas, Doell and Cox concluded that the central Pacific has been an area of persistently low non-dipole activity over the last 700,000 years. Because of this virtual absence of non-dipole perturbation, they believed the measured dispersion (11°) to be a good estimate of dipole wobble. To explain the attenuation of the non-dipole field in the central Pacific they suggested that a lateral inhomogeneity exists in the lower mantle beneath the region, which effectively shields or prevents generation of the non-dipole fluctuations. More recently Doell (1972 a,b,c)

and Doell and Dalrymple (1973) proposed that low values of dispersion of VGP's from lavas at Oahu, Kauai, Niihau and Nihoa (all Hawaiian Islands) indicate that this central Pacific low in non-dipole field activity has persisted for at least the last 5 Myr. While similarly persistent low anomalies in the non-dipole field have not been discovered elsewhere, much importance has been attached to this result because of its potential power in indicating heterogeneities deep in the Earth (Doell and Cox, 1971).

The Society Islands of French Polynesia are recent volcanic features (Chapter 3), due south of the Hawaiian Islands and at about the same latitude, though in the southern hemisphere. Duncan (1975^a) has estimated the magnitude of recent secular variation in the French Polynesian area to establish the southern extent of the central Pacific low angular dispersion proposed by Doell and Cox. Results of that study are summarized in the next sections.

5.3 Palaeomagnetic results from the Society Islands

Previous palaeomagnetic work in two of the Society Islands by Tarling (1963) proved inconclusive. Samples from Tahiti and Bora Bora did not respond to magnetic cleaning and he concluded that a very intense and rapidly fluctuating non-dipole field was present in this region during the cooling of those lavas.

For the present study sixty dykes and flows at five of the Society Islands (Bora Bora, Raiatea, Huahine, Moorea and Tahiti) were sampled (see Figure 5-2 for localities). Hand specimens were oriented in the field using a sun-compass and later cored for laboratory measurement. Sites were selected to give the widest coverage possible, but confined to coastal cliffs and stream cut valleys. On average four samples per site were collected. The palaeomagnetic field directions, or natural remanent magnetizations (NRM's) of 4 cores from each oriented

Figure 5-2 Palaeomagnetic sampling in the Society Islands, French Polynesia. Site localities are shown with determined polarities. (Mercator projection from Duncan, 1975).

sample were measured using either a PAR fast spinner magnetometer or a Digico slow spinner magnetometer (Further description of methods is found in Appendix 1). NRM's were sufficiently scattered that alternating field (AF) demagnetization proved necessary. Sites were treated individually and varied in the strength of peak magnetic field needed to clean away secondary magnetic components (after McElhinny and Gough, 1963). In general, however, 15 milliteslas (mT) AF was sufficient to produce stable directions. Results of the magnetic cleaning and final directions appear in Table 5-1. In addition Fisher statistics and virtual geomagnetic poles have been calculated for each site. Figure 5-3 illustrates the distribution of VGP's of cleaned field directions for 53 normal and reversed sites. (Seven intermediate directions* are omitted in this discussion.) The average of 53 VGP's is indistinguishable from the rotational axis at the 95% confidence level and uniform distribution of the individual VGP's about the average insures that Fisher's method is appropriate for analysis of these data.

Geochronology (K-Ar) of volcanic material from these islands (Chapter 3) suggests that surface lavas and dykes at Bora Bora range in age from 3.1 to 3.4 Myr. At Raiatea ages vary from 2.4 to 2.6 Myr while at Huahine from 2.0 to 2.5 Myr. Moorean ages lie between 1.4 and 1.6 Myr and at Tahiti the range is 0.4 to 1.3 Myr. The time control provided

* Wilson et al. (1972) defined intermediate directions as those for which the VGP latitude $< 50^{\circ}$. Their study employed a large number of Icelandic sites to show empirically that dipole moment (intensity) decreases dramatically for flows with VGP latitude $< 50^{\circ}$. Others (Cox and Dalrymple, 1967 and Dagley et al., 1967) have arbitrarily chosen narrower bounds for 'intermediate' directions. Estimated angular dispersion is, however, extremely sensitive to inclusion of widely separated VGP's and in this study Wilson et al.'s more selective bounds have been used, leading to a minimum angular dispersion.

Table 5-1: Site-mean palaeomagnetic results

Dyke or flow number	N	Peak cleaning field (mT)	Mean declination	Mean inclination	k	α_{95}	R	Polarity	θ'	VGP ϕ'
Bora Bora (Mean site location is 16° 30' S, 151° 44' W)										
B2	4	20	1.7	-37.7	322.1	5.1	3.99	N	85.0	9.1
B4	4	20	166.0	39.6	79.7	10.3	3.96	R	-75.5	91.3
B5	4	40	347.9	-36.1	13.4	25.9	3.78	N	78.0	99.0
B6	5	20	194.0	42.6	67.0	9.4	4.94	R	-74.5	152.6
B7	4	NRM	164.3	32.7	531.7	3.9	3.99	R	-74.9	290.8
B8	4	20	182.4	26.3	753.1	3.3	3.99	R	-86.5	70.3
B9	4	20	183.1	16.5	58.1	12.1	3.95	R	-81.3	49.5
B10	4	10	359.5	-21.4	56.3	12.3	3.95	N	-84.6	203.1
B11	4	NRM	6.7	-35.7	392.4	4.6	3.99	N	82.7	326.3
B13	3	15	351.3	-53.4	13.5	34.9	2.85	N	70.8	50.5
B19	3	15	14.8	-37.9	88.6	13.1	2.98	N	75.1	319.2
B23	4	15	5.1	-28.8	230.7	6.0	3.98	N	84.9	286.0
B24	4	15	179.3	37.7	176.7	6.9	3.98	R	-85.2	215.7
B26	4	15	18.5	-37.6	143.7	7.6	3.98	N	71.8	315.6
B27	3	15	354.4	-37.3	544.1	5.2	2.99	N	78.6	103.4
Average	15		0.7	-35.1	44.1	5.8	14.68		87.0	15.1
Raiatea (Mean site location is 16° 50' S, 151° 27' W)										
R4	5	5	7.9	-40.1	452.3	3.6	4.99	N	80.2	339.2
R6	5	10	21.7	-17.1	323.1	4.2	4.98	N	67.3	280.8
R9	5	20	352.1	-20.0	353.1	4.0	4.98	N	80.0	157.2
R10	3	15	188.5	-29.7	13.5	34.8	2.85	I	-56.3	43.5
R11	5	15	7.7	-29.8	241.4	4.9	4.98	N	82.5	294.4
R13	4	15	179.0	37.1	1548.6	2.3	3.99	R	-85.8	220.3
R14	4	15	10.0	-39.3	122.2	8.3	3.97	N	79.0	330.8
R15	5	15	0.3	-36.8	260.1	4.7	4.98	N	86.1	23.4
R16	3	10	351.5	-29.3	243.6	7.9	2.99	N	81.8	124.3
R20	4	10	4.5	-31.6	329.1	5.0	3.99	N	85.6	304.9
Average	9		3.8	-31.5	48.1	7.4	8.83		86.2	305.1
Huahine (Mean site location is 16° 45' S, 151° 00' W)										
H1	5	15	319.0	-61.9	500.7	3.4	4.99	I	46.3	72.7
H2	4	15	344.1	36.9	52.1	12.8	3.94	I	49.7	186.1
H6	5	15	334.1	-27.5	344.0	4.1	4.99	N	65.0	120.7
H7	5	15	15.0	-44.5	21.5	16.8	4.81	N	73.0	336.0
H9	5	15	18.5	-29.4	19.4	17.7	4.79	N	72.1	298.9
H11	5	10	242.9	-21.1	155.3	6.1	4.97	I	-21.8	99.4
H12	4	10	350.6	-27.9	436.9	4.4	3.99	N	80.8	128.6
H13	5	15	337.3	-31.4	81.6	8.5	4.95	N	68.3	114.6
H14	5	15	8.2	-26.9	752.4	2.7	4.99	N	81.7	283.8
H15	5	10	33.5	-35.6	323.7	4.2	4.98	N	58.0	309.7
H19	4	10	29.4	-36.8	257.9	5.7	3.98	N	61.8	311.6
Average	8		2.2	-37.9	12.6	15.0	8.37		84.7	5.4
Moorea (Mean site location is 17° 32' S, 149° 51' W)										
M6	3	15	181.3	21.1	84.0	13.5	2.97	R	-83.3	220.0
M8	4	10	192.0	22.5	918.8	3.0	3.99	R	-76.9	95.5
M13	5	15	192.3	37.3	720.6	2.8	4.99	R	-77.8	138.2
M16	4	15	175.5	52.5	117.9	8.4	3.97	R	-73.8	223.8
M19	4	20	183.5	34.5	323.0	5.1	3.99	R	-86.3	144.7
M20	3	15	178.9	40.9	101.6	12.2	2.98	R	-83.9	219.0
M23	4	20	201.6	22.5	19.7	21.1	3.84	R	-53.7	247.9
Average	7		4.4	-34.9	38.7	10.8	5.87		85.4	323.4
Tahiti (Mean site location is 17° 40' S, 149° 25' W)										
T1	4	20	355.9	-34.9	467.8	4.2	3.99	N	85.7	95.6
T2	5	15	356.3	-13.2	50.4	10.8	4.92	N	78.6	192.0
T3	5	20	11.7	-32.4	379.4	3.9	4.99	N	78.8	302.9
T4	1	15	7.6	-25.8				N	81.6	273.7
T5	4	15	354.7	-31.2	48.5	13.3	3.94	N	84.9	126.5
T6	5	15	358.9	-28.6	97.9	7.7	4.96	N	87.3	188.7
T9	4	15	178.0	26.3	1380.0	2.4	3.99	R	-86.0	1.4
T10	5	NRM	338.2	-35.7	59.9	9.9	4.93	N	69.2	109.9
T11	3	15	1.5	-23.9	408.5	6.1	2.99	N	84.8	215.7
T13	4	NRM	1.0	-45.7	46.5	13.6	3.94	N	80.1	25.1
T14	4	15	332.6	40.4	32.9	16.2	3.91	I	41.4	176.0
T15	4	20	305.2	36.2	198.7	6.5	3.98	I	24.4	153.0
T16	3	20	196.3	72.3	90.1	13.0	2.98	I	-48.1	197.2
T17	4	20	178.2	64.6	20.1	20.9	3.85	R	-60.9	212.8
T19	3	15	169.2	5.7	545.3	5.2	2.99	R	-72.0	353.2
T22	5	NRM	186.5	35.9	96.0	7.8	4.96	R	-83.3	142.9
T23	3	NRM	357.3	-11.3	253.9	7.7	2.99	N	77.9	197.4
Average	14		358.6	-32.5	17.9	9.2	13.21		88.6	111.1
Average of normal poles	35		1.6	-33.4	27.2	4.6	33.71		87.9	338.1
Average of reversed poles	17		180.6	36.0	21.7	7.5	17.19		-86.9	198.9
Average of all sites	53		1.3	-34.3	25.4	3.9	51.92		87.7	355.2

N: number of samples per site

 α_{95} : radius of 95 per cent confidence circle

k: precision parameter

R: sum of N unit vectors

Figure 5-3 Stereographic projection of site mean VGP's of remanent magnetization after magnetic cleaning (usually 15 mT or less). Solid circles are Northern Hemisphere VGP's (normal polarity epoch), open circles are Southern Hemisphere VGP's (reversed polarity epoch). The mean VGP is not distinguishable from the geographic axis at the 95 per cent confidence level. (After Duncan, 1975).

by the geochronology indicates that the palaeomagnetic sites have sampled the geomagnetic field in the French Polynesian region at discrete points in time over the period 0.4 to 3.4 Myr.

No significant differences exist amongst mean VGP's (or dispersion about these means) for each of the 5 islands. For the purposes of this study the 5 data sets can be combined to characterize the palaeosecular variation of the region over the time length of the geomagnetic field sampled by the lavas.

5-4 Dispersion and Discussion

Several methods have been proposed for the estimation of angular dispersion of VGP's. Fisher's method (1953) determined the variance of the poles from their mean assuming each is an independent estimate of the position of the mean. Because there are two sources of dispersion, within site dispersion and between site dispersion, Watson and Irving (1957) proposed an analysis of variance technique which they termed a two-tier analysis of dispersion. Effectively, this method estimates the between site dispersion, which is a measure of the palaeosecular variation, and the within site dispersion, which is assumed to be constant for all sites. Doell and Cox (1971) use a similar expression for S_F , the between site angular dispersion.

$$S_F = \left(\frac{1}{(N-1)} \sum_{i=1}^N \Delta_i^2 - S_W^2 \right)^{1/2}$$

where N is the number of sites, Δ_i is the angular displacement of each VGP from the mean VGP and S_W is a correction for the within site dispersion. All methods yield very similar estimates of angular dispersion of VGP's. An important question to be resolved is whether the angular dispersion at a given locality should be measured about the mean VGP or the rotational axis, if these are different. For young rocks which have experienced negligible trans-latitudinal tectonic movement, we expect the VGP's to average to the rotational axis if the

geomagnetic field has been, on average, an axially centred dipole and a sufficiently long period of geomagnetic time has been recorded to sample the complete cycle of secular variation (both dipole wobble and non-dipole drift). Indeed, wide separation of the mean VGP from the rotational axis is evidence for insufficient sampling of the secular variation (Aziz-ur-Rahman and McDougall, 1973). (The mean VGP for Society Islands sites is indistinguishable from the rotation axis at the 95% level of confidence). The oldest sites sampled (Bora Bora) are younger than 3.5 Myr, implying maximum trans-latitudinal movement of 165 km, or 1.5 degrees. For younger sites, the movement has been correspondingly less. Hence a small correction may be applied to the position of the rotational axis about which dispersion is measured, for each island. For such a calculation, this 'tectonic correction' made little difference so that in this study angular dispersion of VGP's has been measured about the present rotational axis. For comparative purposes, all studies from the Hawaiian Islands have been recalculated to this form if necessary.

Table 5-2 presents the results of palaeosecular variation calculations for the 53 Society Islands sites of this study as well as data from other studies in the Pacific region. S_F , the angular standard deviation of site VGP's, has been calculated by the Doell and Cox method, with respect to the rotational axis. S_u and S_l are the upper and lower 95% confidence limits on S , determined from Cox (1969), and N is the number of site VGP's employed in each calculation. Figure 5-4 presents these results graphically.

An angular standard deviation of VGP's about the rotational axis of 13.8° characterises palaeosecular variation in the region of the Society Islands for the period 0.4 - 3.4 Myr. This estimate is well-constrained by the 95% confidence limits, S_u , S_l due to the size

Table 5-2: Pacific palaeosecular variation

	N	S_T	S_W	$\bar{n}$	S_F	S_u	S_I	Ref.
Hawaiian Islands								
Recent	72	8.7	5	7.5	8.5	9.6	7.6	Doell and Cox (1965)
Brunhes only	130	13.5	5	7.0	13.4	14.7	12.4	Doell and Cox (1965) Doell (1972a)
Pre-Brunhes only	197	13.1	7	7.0	12.8	13.9	12.1	Doell (1972b,c), Doell and Dalrymple (1973)
All Hawaiian data	399	12.6	6	7.0	12.4	12.9	11.8	
Society Islands	53	14.6	9.6	4.0	13.8	15.9	12.2	this study
Easter Island	70	-	-	-	13.2	15.0	11.8	Isaacson et al. (1973)

- N : number of sites
 S_T : total angular dispersion
 S_W : within-site angular dispersion
 $\bar{n}$: average number of samples per site
 S_F : between-site angular dispersion
 S_u : upper and lower 95% confidence limits on S_F , from Cox (1969)
 S_I :

Figure 5-4 Graphical comparison of results from palaeosecular variation studies on lavas from volcanic islands in the central Pacific Basin. Angular dispersion, measured about the geographic axis, is expressed in degrees along the abscissa. Error bars are the upper and lower 95 per cent confidence bounds from Cox (1969b). The Hawaiian data have been divided chronologically, as explained in the text. (After Duncan, 1975).

of the sample, 53. The predicted dispersion, from reference to secular variation models^{*}, lies within this range so that non-dipole field activity during the time interval recorded has not been anomalous with regard to other localities. Palaeosecular variation recorded in Brunhes age lavas at Easter Island ($27.1^{\circ}\text{S } 109.4^{\circ}\text{W}$) is likewise in accord with model dispersion (Isaacson et al., 1973), so that the southern extent of the Pacific non-dipole field anomaly recorded at Hawaii lies north of the Society Islands and Easter Island.

Combining all Hawaiian data, 399 site VGP's yield an angular standard deviation of 12.4° with very narrow limits. A t-test indicates that palaeosecular variation has been significantly lower in the Hawaiian area over the last 4 Myr than in the French Polynesian and Easter Island areas. Closer inspection reveals that the angular dispersions resulting from various Hawaiian studies (Doell and Cox, 1965, Doell 1972 a,b,c, and Doell and Dalrymple, 1973), each calculated about the rotational pole rather than the mean VGP, fall into two groups. The groups are best separated chronologically: VGP for sites from Recent eruptions (Kau and Puna series on Hawaii) show a very low angular dispersion, 8.5° . Indeed, this palaeosecular variation is lower than the 11° attributed to dipole wobble alone (Doell and Cox, 1971), suggesting a considerably smaller component from that source of dispersion. Alternatively, as suggested by Aziz-ur-Rahman and McDougall (1973), these Recent eruptions have not completely sampled the cycle of secular variation. Excluding the Kau and Puna series, Brunhes age site VGP's are more dispersed, with an angular

* The model angular standard deviation at latitude 20° equals 14.2° from Cox (1970).

standard deviation of 13.4° , indistinguishable from older pre-Brunhes Hawaiian palaeosecular variation, 12.8° . Those studies of Recent lavas from the Hawaiian area, then, show significantly lower angular dispersion of VGP's and suggest a fundamentally lower level of non-dipole field activity during the last several thousand years.

The magnitude of palaeosecular variation in the Hawaiian region from sites older than the Recent Kau and Puna eruptions is comparable to that calculated from the Society Islands and at Easter Island sites and not anomalous when referred to model dispersions based on globally distributed studies. And, while the Central Pacific Basin is presently an area of subdued non-dipole field activity, the anomaly has not persisted for times older than Recent nor is there evidence for its extension south to the French Polynesian region for any time over the last 4 Myr. So that if the mantle beneath Hawaii produces anomalous geomagnetic behaviour, this is only a recent development (within the last few hundred thousand years). And, for the mantle beneath the hot spots of French Polynesia, no anomaly is required by the palaeosecular variation data.

5.5 The kinematical approach

Perhaps the only conclusive geophysical constraint which has been applied to nonorogenic linear volcanism is a kinematical test of the proposition that hot spots (or the sources of these volcanic trends) remain fixed relative to one another over long periods of time. This problem was first attacked by Morgan (1972) in a discussion of the congruence of volcanic trends on the Pacific plate. If, as Morgan reasons, the Pacific plate behaves rigidly and overrides several volcanic sources fixed in the mantle beneath, then the Pacific plate-mantle pole of rotation defined by one of the ensuing volcanic traces

will also describe the geometry of the others. The extension of his argument to encompass the world-wide system of plates and nonorogenic volcanic traces has been performed by Minster et al. (1974).

This test consists first of constructing the best fitting model of relative motions among ten major plates calculated from a least squares fit to spreading rates and azimuths of transform faults at active spreading ridges. From this procedure the authors arrive at an internally consistent set of poles of relative motion for each plate pair. By next fixing the motion of one of the plates to the volcanic sources beneath it, the geometry and migration rates at volcanic traces on other plates will be predicted from the relative motion model. This results in an absolute motion model. It is then an easy matter to check the predictions with the observed.

Using the volcanic traces on the Pacific plate to fix its motion with respect to six or more sources of volcanism, and the derived relative motion model, Minster et al. predict volcanic traces on other plates. The overall fit of this model to the data is quite good. (The Iceland to Wyville-Thomson Ridge is the exception but the authors suggest that the present orientation of that trace may be more north-south, along the Mid-Atlantic Ridge, thereby matching the predicted azimuth.) This method is appropriate only as long as the relative motion model is valid (the last 10 Myr only). We may conclude, however, that over the past 10 Myr, the hot spots examined by Minster et al. have remained fixed relative to one another, in agreement with the Wilson-Morgan hypothesis.

The rates of migration of volcanism used by Minster et al. to fix the rotation of the Pacific plate with respect to the Pacific hot spots (8.9 cm/yr at Hawaii) are smaller than the best fitting rates determined in this study. Using the preferred French Polynesian

volcanic migration rates, Minster et al.'s analysis may be repeated to produce a modified absolute motion model. The result is presented in Table 5-3. Figure 5-5 compares Minster et al.'s absolute motion model with the modification required from the new French Polynesian geochronology.

At most hot spots, the rates and azimuths predicted by the two models are very close. There are, however, significant differences in the predicted motions of Africa and Eurasia. In the model of Minster et al., the mid-Atlantic Ridge is virtually stationary while Africa and Eurasia move to the east at between 1 and 2 cm/yr. The increased rate of movement of the Pacific plate over the French Polynesian hot spots requires that the mid-Atlantic Ridge move westwards at about 1.6 cm/yr while Africa remains virtually stationary. The recent offset of the Mid-Atlantic Ridge to the west of most of the Mid-Atlantic hot spots (e.g. St. Helena, Tristan da Cunha) supports this prediction. Eurasia is predicted to be moving slowly to the west in the modified model, contrasting the slow northward movement predicted by Minster et al. A slightly slower rotation rate for the Nazca plate allows the East Pacific Rise to move slowly westwards. Improved geochronology will, perhaps, distinguish the two models. The essential conclusion regarding the fixed nature of hot spots during the past 10 Myr remains.

For times earlier than 10 Myr B.P., tests of the fixed source hypothesis are less conclusive due to increasing uncertainty in the relative motion model used. In addition, errors in determining geometry and rates of migration of volcanism increase with age. Nevertheless, Molnar and Atwater (1973) have attempted such a test for 21 and 38 Myr ago, both the age of prominent magnetic anomalies. They constrain Pacific plate motion over the mantle with the geometry and ages along the Hawaiian Islands chain. The positions of the 21 and 38 Myr ages

Table 5-3: Absolute Motion Model (Minster et al., 1974, model
AM1 modified by French Polynesian geochronology).

Plate	ϕ (Lat. °N)	θ (Long. °E)	α^* (deg/Myr)
PCFC	-67.3	120.6	1.03
NOAM	-76.0	-82.3	0.38
EURA	-59.6	-11.7	0.14
INDI	14.5	36.6	0.55
COCO	16.5	-121.5	1.35
NAZC	36.1	-76.6	0.71
ANTA	3.9	169.1	0.14
SOAM	-70.2	-156.0	0.38
AFRC	33.8	-59.2	0.09

* rotation is anti-clockwise.

Each absolute motion pole of rotation (ϕ, θ, α) is determined by vector addition,

$$\Omega_1 + \Omega_2 + \Omega_3 = 0,$$

using the pole of rotation for absolute Pacific plate motion and the best fitting relative motion model (Minster et al. 1974).

Figure 5-5 Plate motions with respect to hot spots computed from model AM1 of Minster et al. (1974) and rates of migration of volcanism for Pacific island chains (Chapt. 3). Velocities are in centimetres per year. Dashed lines are the azimuths and velocities (indicated by lengths) predicted by model AM1. Arrow-headed solid lines are the azimuths and velocities predicted using the slightly greater velocity of the Pacific plate with respect to the hot spots (mantle). Poles of rotation for the absolute motion models are shown (open squares for model AM1, open circles for the modified version). Agreement in general is quite close, although some difference exist in predicted motions of slow-moving continents (Eurasia, Africa). These hot spots have maintained a stable frame of reference over the last 10 Myr, to within ± 1 cm/yr (Mercator projection).

along that island chain must be extrapolated from known ages at the southeastern end. By then closing up spreading ridges around the world to their 21 and 38 Myr positions, presently active hot spots should underlie the 21 and 38 Myr ages along their respective traces. This, however, is not the case, according to Molnar and Atwater, who report discrepancies on the order of several hundred kilometres. They conclude that motion among the hot spots has been comparable to plate motion (~ 2.0 cm/yr) over the period investigated and that the sources do not, then, form a fixed reference frame with which to describe the motions of plates. In view of the large uncertainties in both age determination along these volcanic traces and in accurate reconstruction of the plates at times earlier than 10 Myr B.P., and of the results of Minster et al. (1974), Molnar and Atwater's test is inconclusive. Nevertheless, investigations of this type are crucial to establish or deny the Wilson-Morgan hypothesis of fixed hot spots.

Whether or not these volcanic sources are fixed over long periods of time bears circumstantially on their vertical extent. For if the origin of hot spots is deep in the mantle, then their surface expressions should not move appreciably. Motion comparable with the velocities of plates would, however, suggest a much shallower source or mechanism, such as propagating fracture, which is dependent upon the transient kinematics of plate tectonics.

Chapter 6

6.1 Conclusions

A) The Geological Setting

1. The islands of French Polynesia are the peaks of individual shield volcanoes which rise some 4,000 m from the ocean floor.
2. The members of four French Polynesian islands chains (the Marquesas, the Society, the Pitcairn-Gambier and the Austral Islands) postdate the ocean floor on which they sit. A fifth group, the Tuamotu Islands, may be contemporaneous with the underlying crust.
3. The French Polynesian island chains form line segments which are nearly parallel to one another and to the Hawaiian Islands' lineament. Their extensions are approximately perpendicular to the East Pacific Rise.
4. There is geomorphologic evidence (erosional state, development of coral reef) that island ages increase to the northwest.
5. Within each island chain the positions of individual volcanoes have been influenced by the prevailing lines of weakness in the underlying oceanic crust. These are the orientation of the ancient East Pacific Rise spreading zone and the attendant transform faults, orthogonal to one another. Faulting and the distribution of late stage intrusions can also be attributed to the structural control exerted by these ancient lines of weakness.
6. A volcanic cycle beginning with a shield building stage of alkalic basalt, progressing through caldera collapse and

dyke emplacement to intrusion of highly differentiated rock types is common to all island groups, but variable in extent at individual islands.

B) Development with Time

1. The subaerial shield building stage of the volcanic cycle is completed within about 0.5 Myr. Dyke emplacement and post-caldera collapse flows follow. Volcanic activity finishes with the intrusion of differentiates (trachytes, phonolites) at about 0.4 Myr after the shield building stage. Thus the subaerial volcanic development of most island members of French Polynesia occurs within about 1.0 Myr. Volcanic activity producing the Austral Islands, however, appears to have continued at each island for about twice as long, 2.0 Myr. and at Rurutu shield building volcanism has recurred after a hiatus of about 11 Myr, but this phenomenon is atypical for French Polynesian volcanism.
2. Within each of the four volcanic island chains, average ages for island members increase monotonically from the southeast to the northwest. Ages at adjacent islands may overlap but this may be attributed to sampling from different stages in the volcanic cycle.
3. The progression in age from the southeast to the northwest within each chain has been irregular. However, a rate of migration of volcanism may be calculated which best fits the inter-island distances and average ages for each island member of an island chain. Such calculations yield for the Marquesas Islands a rate of migration of volcanism of 10.9 cm/yr, for the Society Islands, 11.1 cm/yr, for the Pitcairn-

Gambier Islands, 11.0 cm/yr and for the Austral Islands, 10.7 cm/yr.

4. There is close agreement between these observed rates and rates predicted for French Polynesia using the Pacific Plate rotation pole proposed by Minster et al. (1974) and a volcanic migration rate of 11 cm/yr for the Hawaiian Island Chain. This supports the contention that the independent sources responsible for the volcanism in the separate island chains have been fixed relative to one another, within the limits of detectibility (± 1 cm/yr).
5. Some lineaments have begun only recently (Marquesas, Societies) while others have a much longer history (Australis, Tuamotus?).

C) Chemical Variations

1. Igneous rocks from the islands of French Polynesia belong to the alkali olivine basalt association. Shield lavas consist primarily of alkalic basalt, ankaramite and basanite, which have differentiated at shallow level to produce minor amounts of hawaiiite, mugearite and trachyte. Phonolite is occasionally found as well.
2. Differentiation of no single parental magma can account for the major element chemical variation found in the shield lavas of the Marquesas, Society and Austral Islands. Instead, it appears as though the magmas available for the construction of these islands came from a variety of partial melting conditions from a peridotitic source region of variable composition.
3. Samples from the Austral Islands are distinctly more sodic than samples from either the Marquesas or Society Islands, having $\text{Na}_2\text{O}/\text{K}_2\text{O} \geq 2$, compared to lower ratios of 1.0 - 1.5 for the latter two groups. Marquesan samples are generally more

siliceous than those analyzed from either the Austral or Society Islands.

4. The isotopic composition of Sr in shield building lavas of French Polynesia ranges from 0.7026 to 0.7082 which is the same magnitude of variation as is seen in a world-wide distribution of oceanic island lavas. The strong correlation of $\text{Sr}^{87}/\text{Sr}^{86}$ with $\text{Rb}^{87}/\text{Sr}^{86}$ is interpreted as an isochron, which dates the last homogenization of the source region for French Polynesian lavas at about 1.4 billion years.

D) Geophysical Constraints

1. No conclusions can be drawn from gravimetry or heat flow.
2. Seismic anomalies exist beneath several proposed hot spots. There is, however, no unanimity as to the nature of the velocity anomaly (fast or slow) of such regions or as to their lateral or vertical dimensions.
3. Palaeosecular variation at the Society Islands, between 0.5 and 3.3 Myr. B.P., has not been anomalously low when compared with a world-wide distribution of similar measurements. This suggests that whatever anomaly may exist beneath the southeast corner of this island chain has not masked the activity of the geomagnetic field, as was suggested for anomalous Hawaiian results (Doell and Cox, 1971).
4. Tying Minster et al.'s best fitting relative motion model to the volcanic migration rates in French Polynesia accurately predicts the rates and azimuths of most of the proposed hot spot/plume traces on other plates. This implies that over the past 10 Myr, these volcanic sources

have remained fixed relative to one another, within the limits of detectability.

6.2 Implications for the source regions and volcanic mechanisms

What, then, can be said about the origin and nature of non-orogenic linear volcanism in French Polynesia? Conclusions drawn from geological, geochronological, geochemical and geophysical studies have been outlined in the preceding section and may be summarized as follows:

First, French Polynesian volcanism is far removed from either the East Pacific Rise or the subduction zones of the western and northern margins of the Pacific plate and cannot be directly related to either of those two magma producing environments. Nevertheless, the arrangement of island members into lineaments which parallel the northwesterly motion of the Pacific plate suggests that the surface manifestation of the volcanism is dependent on the direction and velocity of the lithosphere. Thus, an indirect association with crustal production at the East Pacific Rise is implied.

Within each of the four island chains investigated, ages at island members increase monotonically to the northwest, consistent with the direction of movement of the Pacific plate. Moreover, the rates of migration of volcanism exhibited in each of the four lineaments are compatible with production from four sites which have been fixed with respect to one another over at least the last 4 Myr.

Dymond (1975) contends that rates of migration of volcanism increase northwards, from the Austral Islands up to the Hawaiian Islands, but some of his estimates are based on the data of Krummenacher and Noetzlin (1966) which are unsuitable for such calculations.

Comparison of Pacific volcanic migration rates with the best estimates

for the half spreading rates at the East Pacific Rise demonstrates a closing movement of from 2 to 5 cm/yr. Either the sites of volcanic activity in French Polynesia are moving towards the East Pacific Rise or vice versa, or a combination of the two.

McDougall (1971) and, more recently, Forristal (1974) have proposed, for the example of the Hawaiian Islands Chain, that the volcanic source for that lineament resides in the asthenosphere and migrates toward the East Pacific Rise as part of the mass balance flow for lithospheric movement and production.

Forristal (1974) argues that because the northward component of movement of the Pacific plate determined from the sedimentary record (equatorial sedimentary maximum) is significantly less than the same component determined from the geochronology of the Hawaiian Islands, the Hawaiian hot spot must be migrating, at least to the south and possibly to the east. First, this comparison conflicts with Clague and Jarrard (1973) who find the independently determined rates compatible and secondly, Forristal fails to recognize the possibility of true polar wander (movement of the whole of the lithosphere with respect to the spin axis) which could explain such discrepancies. Indeed, a very curious picture results when the East Pacific Rise is assumed to be fixed, but spreading at variable rates. In that case, the sites of volcanism for each of the young island chains of the eastern Pacific Basin (Hawaiian Islands and the Pratt-Walker Seamounts included) must move at varying rates toward the East Pacific Rise to account for the observed volcanic migration rates in the island chains. These hot spots, however, cannot have moved significantly with respect to one another over the past 4 Myr. As spreading centres are elsewhere observed to move at rates comparable to plate motions, it is more

convincing to conclude that it is the East Pacific Rise which is moving northwestwardly toward the sites of volcanism in French Polynesia (and those further north) rather than the opposite.

The test proposed by Minster et al. (1974) using the best fitting model of relative plate motions to compare azimuths and rates of migration of volcanism along proposed hot spot traces from separate plates has demonstrated that a world-wide distribution of selected hot spots have had no significant relative motions in the last 10 Myr. This statement must be qualified to the extent that any movement between hot spots is an order of magnitude less than plate motion, or within the limits of detectability. Also, many proposed hot spots have been ignored in this test (compare Fig. 1-1 with Fig. 5-5) and significant motions may be found (for example, in the Comores Archipelago). With the more accurately known "absolute" motion for the Pacific plate, determined from the geometry and rates of migration of volcanism in French Polynesia, the predicted absolute motions of other plates are modified slightly but do not significantly alter the conclusions of Minster et al. Interestingly, using the French Polynesian data, Africa is predicted to be virtually stationary (with respect to the hot spot frame of reference), as contended by Burke and Wilson (1973) and Briden and Gass (1974), and both the East Pacific Rise and the Mid-Atlantic Ridge are moving slowly to the west.

The fixed nature of the sources for nonorogenic linear volcanism is more easily reconciled to the Wilson-Morgan hypothesis than to models which depend on the interaction of the asthenosphere with the lithosphere. Because their proposed origin lies below the level of crustal production and lithospheric plate motions, which is in a constant process of reorientation, the positions of hot spots, relative to one another, are not expected to be influenced by changes in plate motions.

This is not the case for the propagating fracture hypotheses or the models which involve asthenosphere-lithosphere interaction. For these models reordering of the system of lithospheric motions would also involve changes of a comparable magnitude in the positions of the volcanic sources. The possibility of movement of the hot spots over time is not removed with deep origins. Any movement, however, will be independent of lithospheric motion and of much smaller magnitude.

The parental magmas which produced the volcanic islands of French Polynesia were derived from source material which has maintained heterogeneities in Rb/Sr for a very long period of time. Major element chemistry of shield lavas is consistent with the proposition that parental magmas were formed by variable degrees of partial melting (5-20%) of a peridotitic source material under varying melting conditions (60-100 km depth). This upper mantle source material is heterogeneous, both with regard to major elements (particularly Na,K) and trace elements (Rb,Sr).

The isotopic composition of strontium in shield lavas from French Polynesian Islands correlates strongly with Rb/Sr. The isochron defined by this correlation yields approximately 1.4 billion years since the last homogenization of strontium isotopes. Such an homogenization was most likely a melting event but diffusion processes may have also been important. This ancient age places severe constraints on the origin of the source material. Sun and Hanson (1975) likewise, find a strong linear correlation between isotopic strontium and Rb/Sr for lavas from a world-wide distribution of oceanic islands. They interpret these data to indicate that the source regions for oceanic islands (all products of nonorogenic linear volcanism) have maintained closed chemical systems for from 1.0 to 3.0 billion years. Magma mixing, sediment contamination and isotopic disequilibrium during

melting can be dismissed as controlling factors in determining the isotopic composition of strontium in the subaerial lavas. Instead, these variations appear to be the result of inhomogeneities in Rb/Sr in the source region imposed about 1.4 billion years ago.

There are essentially two possibilities for the origin of the source region for French Polynesian lavas - from the upper part of the asthenosphere or from below the asthenosphere. Major element chemistry suggests that melts segregated from parental material in the region 60-100 km depth, but this does not eliminate a sub-asthenospheric origin as parental material may have risen diapirically before partially melting. Figures 6-1 and 6-2 will aid in the following discussion.

Figure 6-1 illustrates Green's (1971) model for the genesis of ocean floor basalt and oceanic island basalt. Recall that in this model the pyrolitic low velocity zone (= asthenosphere) exhibits a dual nature - the lower LVZ is a region of partial melting which is depleted, relative to the 'frozen' upper LVZ, in incompatible elements and volatiles. The upper LVZ is progressively enriched in incompatible elements (Rb at a greater rate than Sr) by accumulation of partial melts from the lower LVZ. The upper LVZ moves with the overlying lithosphere, while the lower LVZ flows toward the mid-ocean ridges to return material for new crust production. No appreciable amount of material upwells from below the asthenosphere at the mid-ocean ridge in this model. The uppermost part of the lithosphere, above the Moho, is about 5 km thick. Its composition is predominantly olivine tholeiite which has been produced by 20-30% partial melting of depleted (lower LVZ) pyrolite. Thus, another 15 km thickness accounts for the further depleted residuum from the production of new oceanic lithosphere. The remaining 50-60 km, down to the top of the low velocity zone, must comprise material from the low velocity zone which did not melt, but rose to

fill the void left by the spreading lithosphere. This material, in the 'mature' stage of a mid-ocean ridge, such as the East Pacific Rise, must be primarily from the lower LVZ which provides return flow for mass balance (The upper LVZ moves away with the lithosphere.). In addition to being partially melted (and thus likely to be somewhat homogeneous isotopically) prior to incorporation into the lower lithosphere, this lower LVZ material has been depleted in Rb relative to Sr and is not a likely source of $\text{Sr}^{87}/\text{Sr}^{86}$ in the range exhibited by French Polynesian lavas. (Recall that the $\text{Sr}^{87}/\text{Sr}^{86}$ of ocean floor basalts, which, in the model, are produced from this reservoir, average about 0.7025.)

The upper LVZ may be a source of high $\text{Sr}^{87}/\text{Sr}^{86}$, owing to the upward concentration of Rb, producing higher Rb/Sr than in the lower LVZ. Green's model proposes that tensional fractures thin the lithosphere sufficiently to allow diapirism from the upper LVZ and subsequent volcanism of the type seen at Hawaii and in French Polynesia. To generate the relationship illustrated in Figure 4-11, however, this uppermost region of the LVZ must have maintained heterogeneities in Rb/Sr for approximately 1.4 billion years. The age of the crust underlying the islands of French Polynesia is estimated to be only about 50 Myr which was the time when this section of the lithosphere was in the mid-ocean ridge environment. For the upper LVZ material to be considerably older than this it must be relic or have been incorporated from elsewhere, either by lateral transport from beneath old continental cratons or by upwelling from beneath the asthenosphere. It is difficult to imagine, however, a systematic mechanism by which very old material is preserved in a youthful circulating system of lithospheric production. Evidence exists, however, for the preservation of very old material at the mid-ocean ridges (for example Bonatti and Honnorez, 1971; Ozima et al., 1975) but it is not known how universal these occurrences may be.

Another complication emerges in maintaining isolated regions of old material directly beneath the mid-ocean ridges. Although this model, and others proposing propagating fractures or asthenospheric currents, does not attempt to explain anomalous (nonorogenic linear) volcanism on or near mid-ocean ridges (e.g. Iceland, Ascension, Easter, Bouvet) geometrical and geochronological evidence suggests that they and intraplate volcanism (e.g. the Hawaiian Islands, the islands of French Polynesia) owe their origin to a common mechanism drawing on old and enriched (relative to ocean floor basalts) peridotitic source regions. Suppose that, in the initial stages of continental fragmentation and sea-floor spreading (the Red Sea, for a modern example), lithosphere and old upper LVZ material ($\text{Sr}^{87}/\text{Sr}^{86} = 0.7025$ to ≥ 0.7070) from beneath the rifting craton remain as screens between new increments of lithosphere and LVZ, beneath the spreading centre. The subsequent contribution of this material to any mid-ocean ridge islands, however, will progressively diminish owing to the dilution effect of lower LVZ material ($\text{Sr}^{87}/\text{Sr}^{86} = 0.7025$). Thus for any ridge-centred hot spot, a decrease in average $\text{Sr}^{87}/\text{Sr}^{86}$ from older members close to the continents to more recent activity near the mid-ocean ridge should be observed. This test has not yet been performed. There is, at present, no indication that $\text{Sr}^{87}/\text{Sr}^{86}$ for mid-ocean ridge centred islands is significantly lower than for intraplate oceanic islands (see Peterman and Hedge, 1971 and Sun and Hanson, 1975, for summary of existing data).

The second possibility requires that the parental material for French Polynesian lavas originate below the asthenosphere. If this is the case then it follows that the chemistry of French Polynesian lavas is more representative of the bulk chemistry of the Earth than are the typical ocean floor basalts. Consequently the source region for French Polynesian volcanism (and all proposed nonorogenic linear volcanism) can be considered 'primitive', whereas the source regions for mid-ocean

ridge volcanism must be relatively depleted.

A specific model for the upwelling of material from below the asthenosphere is the hot spot, or mantle plume model, suggested originally by Carey (1958), emphasized by Wilson (1963a,b, 1965a,b) and later expanded by Morgan (1971, 1972). In the Carey - Wilson - Morgan hypothesis oceanic and continental nonorogenic linear volcanism is the manifestation of plate motion over fixed sources of diapirically upwelling mantle material. Wilson proposed that these anomalously hot materials originated in nonconvecting areas of the upper mantle - hence their fixed nature. Morgan, however, believes that such plumes are a fundamental part of the convection pattern of the entire mantle, rising from near the core-mantle boundary. Vogt (1975) has further suggested that plumes are initiated by material core-mantle interaction. According to Morgan, plumes are the upward balance of mantle convection, while elsewhere mantle material is slowly descending, producing a distinctly asymmetric convection pattern.

The rubidium-strontium data from French Polynesia require only that the source region for that volcanism has maintained its chemical heterogeneity for about 1.4 billion years. This is satisfied either by a nonconvecting (isolated) region of the mantle, or by material involved in a convection pattern such as that suggested by Morgan, as long as that material does not participate in melting events. Figure 6-2 illustrates this model. The 1.4 billion year age is possibly an average age for the last homogenization of the source material for French Polynesian volcanism. This may be a measure of the turnover rate of the mantle, as proposed by Morgan (1972) from Pb data from South Atlantic Islands (Oversby and Gast, 1970). If so, then the volcanic material erupted at French Polynesia may have been involved in melting (production of lithosphere?) about 1.4 billion years ago and have since descended slowly to some depth in the mantle to then be upwelled

Figure 6-2 A model for the production of ocean floor and oceanic island basalts - the hot spot plume model. New oceanic lithosphere is created at the mid-ocean ridges (after Green, Figs. 2-2, 6-1). Ocean island volcanism, however, is produced by diapiric upwelling of material from beneath the LVZ, either as part of the upward convection of the mantle, or from a nonconvecting region of the mantle. Numbers refer to $\text{Sr}^{87}/\text{Sr}^{86}$ in various regions of the lithosphere and upper mantle.

in the Marquesas, Society and Austral Islands mantle plumes.

Consequently, this 1.4 billion year old material would be the depleted residuum from the melting of material having even higher concentrations of incompatible elements. Evidence supporting or contradicting this implication may possibly be found in the chemistry of Archean mafic rocks.

Both the geochronological evidence for the fixed nature of the volcanic sources beneath French Polynesia and the rubidium-strontium evidence of old parental material are more easily reconciled with the model of sub-asthenospheric sources than with any of the models involving propagating lithospheric fracture with asthenospheric source material. It is unlikely, however, that such circumstantial evidence will ever eliminate either possibility. What is promising, on the other hand, is the probability that this nonorogenic linear volcanism samples regions of the upper mantle untapped by mid-ocean ridge volcanism and, through its fixed nature, provides a record of movement between lithospheric plates and the mantle.

6.3 Further tests and applications

A great attraction lies in discovering or devising a reference frame with which to measure the movements of the lithospheric plates. Such a possibility appears in the network of presently active hot spots. With regard to the "absolute" reference frame defined by the present world-wide distribution of hot spots, the African and Eurasian plates are virtually stationary, the Pacific plate is the most rapidly moving and both the East Pacific Rise and the Mid-Atlantic Ridge are moving slowly to the west.

The test proposed by Minster et al. (1974) to measure the amount of movement among hot spots is applicable only as long as a reliable system of relative motions can be constructed. Before 10 Myr

older volcanic provinces may be more definitive, providing suitable data are available. If hot spots have remained fixed relative to the geomagnetic pole, then every member of a given hot spot trace should have the same palaeomagnetic latitude or, more simply, should have recorded the same geomagnetic field inclination (Figure 6-3). Direct comparison of palaeomagnetic latitude with present hot spot latitude may, however, be complicated by true polar wander, a change in the position of the Earth as a whole or of the outer lithosphere-mantle shell with respect to the spin axis (usually assumed fixed). Such a change could conceivably result from redistributions of mass within the Earth (upward convection and downward subduction) causing a shift in the principle moment of inertia (Goldreich and Toomre, 1969).

McElhinny (1973), following McKenzie (1972), and Jurdy and Van der Voo (1974) find that over the last 50 Myr there has been no significant movement of the lithosphere as a whole with respect to the geomagnetic pole, so that the lithosphere has experienced no true polar wandering in that time. Hargraves and Duncan (1973) have compared 50 Myr palaeomagnetic latitudes with present hot spot latitudes for eight volcanic provinces. They find differences of variable magnitude, some of which they believe to be greater than the errors in defining the locations of the 50 Myr sites of volcanism along the hot spot traces and the 50 Myr palaeomagnetic poles. That study concludes that hot spots do move slowly with respect to the Earth's spin axis, but in a systematic rather than random fashion. If the latitudinal differences are real and hot spots have maintained a stable configuration then the mantle, uncoupled from the lithosphere, must have rotated some 12° about a pole emerging close to the equator during the past 50 Myr. Alternatively, hot spots are moving slowly

Figure 6-3 If a hot spot has remained fixed with respect to the geomagnetic pole, then every member of its volcanic trace will have the same palaeolatitude, θ . If the hot spot moves latitudinally, this movement will be reflected in varying palaeolatitudes for older members of the province.

with respect to the spin axis but in random directions. A decisive test is not possible at present but awaits the acquisition of higher quality palaeomagnetic and geochronologic data.

If it can be shown, through the palaeomagnetic tests outlined above, that hot spots have maintained a fixed configuration (with or without a correction rotation for whole mantle motion) then they may be used as an absolute reference frame for plate movements. Several applications then follow. For example, using the well defined, though poorly documented, volcanic traces stretching from the oceanic islands surrounding Africa (Bouvet, Tristan de Cunha, St. Helena, Ascension and Reunion) a quick test of the expanding earth hypothesis (Carey, 1975) may be performed. Assuming that the five hot spots have been fixed over the last 120 Myr and that the Earth's radius has remained constant, contemporaneous volcanic events along each of the provinces should exhibit the same configuration as the five sites of young oceanic volcanism. If, however, the Earth's radius has increased since early Cretaceous times then inter-hot spot distances would be larger today than previously and reconstruction of African plate motion along any one volcanic trace would reveal significant closing movements of the other four hot spots. Figure 6-4 illustrates this test. The compatibility of both the geometry and ages along each of the hot spot traces with rigid plate motion of Africa would deny significant change in the Earth's radius (providing expansion has been symmetric) and support the proposition that the five hot spots investigated have maintained a fixed configuration since the early Cretaceous. This test could be extended to include a world-wide distribution of hot spots, providing old volcanic sites can be identified.

Another important question which might possibly benefit from the hot spot reference frame is the nature of sea floor spreading prior

Figure 6-4 A test of the expanding Earth hypothesis. Eight hot spots from the Atlantic and Indian Oceans are plotted in their present configuration on the Smith and Hallam (1970) construction of Gondwanaland. Fernando Po, Ascension, St. Helena, Tristan de Cunha, Bouvet and Reunion all then overlie old volcanic provinces. Geochronology is not well known. If contemporaneous ages are confirmed for the old volcanic provinces, a symmetrically expanding Earth is denied and these six hot spots have remained fixed with respect to one another since the age of that volcanism.

A

B

Figure 6-5 A test of pre-Mesozoic plate tectonics. If, as palaeomagnetic studies suggest, large scale continental drift did not begin until the fragmentation of Gondwanaland, pre-Mesozoic hot spot traces (if these can be located) will cross from craton to craton (dashed lines) in an uninterrupted fashion, as in (A). If, on the other hand, intercratonic movement on the scale of present drift were common, then hot spot traces might look more like (B).

to the age of the oldest preserved sea floor (Jurassic). McElhinny et al. (1974) believe that palaeomagnetic polar wander paths for cratonic units of Gondwanaland are sufficiently congruent to exclude major relative movements from at least 750 Myr ago to the Jurassic fragmentation of that supercontinent. This implies that intercratonic movements were small and that pre-Jurassic orogeny was essentially ensialic, reworking older basement in mobile belts. Others, however, offer structural and petrological arguments that a plate tectonic regime may account for most aspects of Phanerozoic tectonics. Dewey and Spall (1975) provide a useful summary.

Figure 6-5 illustrates how ancient hot spot traces, if they can be recognized on old cratons, may provide an independent assessment of pre-Jurassic plate tectonics. Essentially, orderly geometrical and chronological progression of hot spot traces across cratonic boundaries (old mobile belts) would support the hypothesis of Gondwanan coherence in pre-Jurassic times, whereas cessation of such scars at cratonic sutures would imply previous cratonic separation. Anorthosite belts, which Anderson (1975) suggests are the earliest manifestation of hot spots, may chart the course of continental motions in Precambrian times.

To determine the deep structure beneath proposed sources for nonorogenic volcanism, seismological tests are probably the most promising. Work to date (reviewed in Chapter 5) indicates that anomalous material exists beneath proposed hot spots but there is no unanimity about the nature, magnitude, or dimensions of the anomaly. Well designed experiments may resolve this question. The combination of surface wave and body wave (particularly shear wave, such as ScS) data can be employed to investigate lateral and vertical heterogeneities in the mantle and, perhaps, to characterize the sources for nonorogenic linear volcanism.

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Appendix 1: Analytical Techniques

A1.1 Introduction

The equipment and techniques used in the analytical portion of this study were developed and maintained by persons other than myself. I am grateful for the opportunity to use many of the laboratories at the Research School of Earth Sciences, the Australian National University, and for patient instruction.

A1.2 Sample preparation for geochronology and geochemistry

Weathered surfaces, which could not be (easily) chipped away in the field, were removed with a jaw splitter. A 0.2 to 1.0 kg portion of each sample was then processed. For K-Ar geochronology samples were crushed in a steel jaw crusher and sieved for the 1.5 - 3.0 mm size fraction. From this fraction, a 60-80 g split was removed and further crushed to <120 mesh powder in a tungsten carbide siebtechnik, to be used for potassium analyses by the flame photometric method. The remainder of the 1.5 - 3.0 mm fraction was used for argon extraction.

Major element analyses, trace element analyses, and Sr isotope compositions were determined from whole rock crushed powders (<120 mesh) prepared in a tungsten carbide siebtechnik.

A1.3 The K-Ar whole rock method

This geochronological method is based upon the fact that under suitable conditions, radiogenic Ar^{40} , produced by the radioactive decay of K^{40} , accumulates in a rock with time. Figure A1-1 gives a schematic picture of the dual decay of K^{40} to Ar^{40} and Ca^{40} .

Figure A1-1 The decay of ^{40}K . Four branches result in the production of two daughters, ^{40}Ca and ^{40}Ar , by electron capture, positron emission, or beta emission. The energy released in each mode of decay is E and the percentage of ^{40}K that decays in each branch is indicated. (After Dalrymple and Lanphere, 1969).

Basic equations

The basis of the K-Ar isotopic dating method is the age equation which relates time, t , to the absolute amounts of K^{40} , $^{*}Ar^{40}$, in the sample. This is a particular case of a general age equation which is derived as follows: For the radioactive decay of a parent isotope, P , to a daughter, D , the change in the amount of P with time is proportional to P , or

$$\frac{dP}{dt} = -\lambda P \quad \text{where } \lambda \text{ is the decay constant for } P \rightarrow D.$$

Integrating between $t=0$ and $t=t$ (the present), and $P=P_0$, the initial amount of parent and $P=P$ (the present),

$$\int_{P_0}^P \frac{dP}{P} = - \int_0^t \lambda dt$$

$$\text{or} \quad \ln(P/P_0) = -\lambda t$$

$$\Rightarrow \quad P/P_0 = e^{-\lambda t}$$

$$\text{or} \quad P = P_0 e^{-\lambda t}$$

$$\text{but} \quad P_0 = P+D \text{ (no addition or loss to the system),}$$

$$\text{so} \quad P = (P+D)e^{-\lambda t}$$

$$\text{or} \quad e^{\lambda t} = (1 + \frac{D}{P})$$

$$\Rightarrow \quad t = \frac{1}{\lambda} \ln (1 + \frac{D}{P})$$

In the case of radioactive decay of K^{40} which has a dual decay, we are interested in only that fraction of K^{40} which decays to $^{*}Ar^{40}$, which is $\frac{\lambda_e}{\lambda_e + \lambda_\beta}$, so the particular age equation is:

$$t = \frac{1}{\lambda_e + \lambda_\beta} \ln \left(1 + \frac{^{*}Ar^{40}}{\frac{\lambda_e}{(\lambda_e + \lambda_\beta)} K^{40}} \right)$$

Decay constants are becoming more accurately known. Currently most age determinations (including those of this study) are calculated using the Aldrich and Wetherill (1958) constants. Recently Beckinsale and Gale (1969) proposed a slightly different set of decay constants which may be used with greater regularity in the future. The effect of using the newer constants is to increase by about 2.7% the ages calculated in this study using the older constants. (The effect diminishes for older samples and actually reverses for very old rocks.)

The assumptions which must be fulfilled if ages determined by the K-Ar method are to be reliable are:

- (1) That the decay constants, λ_e , λ_β , are known accurately and precisely,
- (2) That K^{40} forms a common fraction of total K throughout natural systems and, hence, it is necessary to measure accurately and precisely only the K content of the sample,
- (3) That the fraction of total Ar^{40} released in the extraction process which has accumulated since the crystallization of the sample, $*Ar^{40}$, can be measured,
- (4) That no $*Ar^{40}$ existed in the sample at $t=0$; that is, there is no inherited $*Ar^{40}$ from the source region,
- (5) That since $t=0$ there has been no loss or gain of $*Ar^{40}$ or K other than that expected from radioactive decay.

The potassium content of the sample is measured in duplicate by the flame photometric method (Cooper, 1963), using a Baird Atomic photometer. All duplicate analyses are within 1% of one another. The fraction of K which is K^{40} is 1.19×10^{-2} atom %.

Figure A1-2 shows the apparatus used for extraction of argon from whole rock samples. A carefully weighed sample (10-30g, depending on %K and expected age) is suspended in a molybdenum crucible within a

glass vacuum line. A mercury diffusion pump establishes a pressure of 5×10^{-7} torr (or lower) overnight. The sample is melted by induction heating from a radio frequency coil and all volatile gases are released. At this stage a known volume of Ar^{38} gas is introduced to the gas sample for later calibration. Spikes were prepared on a separate vacuum line by the pipette method (McDougall, 1966). The gradual depletion of a reservoir of known volume is monitored by calibrating selected spikes against standard samples of pre-determined age. Reproducibility and accuracy of the calibrations are generally better than $\pm 1\%$. Active gases ($\text{O}_2, \text{N}_2, \text{H}_2\text{O}, \text{CO}_2$, etc.) are removed by reaction with gettering agents - zeolite sieve, hot Cu-CuO and Ti-TiO₂ furnaces - leaving purified argon gas. This is collected on activated charcoal by condensation with liquid nitrogen and is loaded on a sample manifold for future mass spectrometry.

Three isotopes of argon are measured by the gas source mass spectrometer: Ar^{36} , Ar^{38} , Ar^{40} . We are really interested in only $^*\text{Ar}^{40}$, the fraction of Ar^{40} which is produced from K^{40} . Ar^{36} and Ar^{38} are both stable isotopes - their abundance does not change with time. Essentially,

$$\text{Ar}^{40} = ^*\text{Ar}^{40} + (\text{Ar}^{40})_{\text{atmospheric}} + (\text{Ar}^{40})_{\text{spike}}.$$

$(\text{Ar}^{40})_{\text{spike}}$ is a trivial amount which is known. McDougall et al. (1969) report $\text{Ar}^{38}/\text{Ar}^{36} = 17,950 \pm 400$ for a measured spike. In order to correct for $(\text{Ar}^{40})_{\text{atmospheric}}$ we use the fact that $(\text{Ar}^{40}/\text{Ar}^{36})_{\text{atmospheric}}$ is ~ 295.5 , so that measurement of Ar^{36} in the gas sample allows simple calculation of the atmospheric component of Ar^{40} . We are now left with $^*\text{Ar}^{40}$ but must know its absolute (molar) amount which necessitates calibration with the known amount of Ar^{38} (the 'spike').

Figure A1-2 Schematic diagram of an argon extraction line, Research School of Earth Sciences, Australian National University. Following the method described by McDougall (1966).

A modified commercial gas source mass spectrometer (AEI MS-10) with 4.2 kilogauss permanent magnet and digital readout was operated in the static mode to measure isotopic argon. Corrections must be made to the raw data because of machine effects. Equilibrium takes place between the walls of the analysing tube and the sample gas during measurement so that in succeeding analyses, isotopic ratios will change with time, reflecting previous isotopic compositions. This so-called 'memory' effect can be monitored by measuring sample isotopic composition over sufficient time to extrapolate back to the moment of introduction of the new sample to the machine. A second correction is for mass discrimination. Each mass spectrometer will fractionate isotopes in its own characteristic manner. As we agree that the atmosphere ratio of Ar^{40} to Ar^{36} is ~ 295.5 , aliquots of atmospheric argon are measured to determine the amount of discrimination of this machine. During the period of measurement for this study, $(\text{Ar}^{40}/\text{Ar}^{36})$ atmospheric was monitored monthly and varied from 284.5 to 288.5. Finally, because only a fraction of the sample gas is let into the machine to be analysed, and this fraction will be enriched in Ar^{36} relative to Ar^{40} (because of their mass differences), we must apply an orifice correction (Baksi and Farrar, 1975). This is calculated by determining the fraction of the sample let in from the volume of the entire sample, measured at the end of each analysis.

Data reduction, age calculation and error analysis were done by computer following the method described by McDougall et al. (1969). Errors are estimated from uncertainties in spike volume, variation in duplicate K analyses and errors in $^*\text{Ar}^{40}$ determination. In the tables in Chapter 3 errors are given as ± 2 standard deviations.

A1.4 Major Element Analysis

Rough weight per cent oxide concentrations of Si, Ti, Al, Fe, Mn, Mg, Ca, K and P were determined relative to U.S.G.S. standards using a modified Phillips PW 1220 semi-automatic X-ray spectrometer operated and maintained by P. Beasley. Na_2O was measured independently by flame photometry. Corrections were then applied for interference effects to achieve the final oxide concentrations.

All samples measured in duplicate from buttons prepared in the standard way (Norrish and Hutton, 1969). The concentrations reported in Chapter 4 are averages of the duplicate analyses. Precision and accuracy were monitored by replicate analyses of international standards, Table A1-1.

Flame photometry

Both K and Na were measured using a modified Baird Atomic flame photometer following the lithium internal standard technique of Cooper (1963). All K analyses were measured in duplicate for K-Ar geochronology and they agreed to within 1%. These also provided a check of the XRF-determined K_2O percentages. The two methods agreed to within 1%.

Other

Fe^{+2} was measured for most of the Marquesas samples using the titration technique developed by Kiss (1967). The proportion of total Fe which now exists as Fe^{+2} in these samples varies from about 40% to 75%. As secondary oxidation to Fe^{+3} may have variably influenced this proportion, it has been assumed for all samples that $\text{FeO}/\text{Fe}_2\text{O}_3 = 0.2$ in calculation of normative compositions.

Loss on ignition (LOI) was measured by heating about 0.5 g of sample at 1050°C for 30 minutes in platinum crucibles. Weights were

Table A1-1: Major Element Analysis of U.S.G.S. Standard Samples
(oxide concentrations by weight percentage)

	<u>AGV1</u>	<u>BCR1</u>	<u>GSP1</u>
FeO	6.687 $\pm$ 0.22%	13.399 $\pm$ 0.05%	4.246 $\pm$ 0.74%
TiO ₂	1.045 $\pm$ 0.39%	2.239 $\pm$ 0.39%	0.662 $\pm$ 0.16%
CaO	4.950 $\pm$ 0.26%	6.992 $\pm$ 0.17%	2.028 $\pm$ 0.08%
K ₂ O	2.949 $\pm$ 0.08%	1.728 $\pm$ 0.06%	5.521 $\pm$ 0.08%
P ₂ O ₅	0.517 $\pm$ 0.37%	0.379 $\pm$ 0.73%	0.302 $\pm$ 0.32%
SiO ₂	59.224 $\pm$ 0.11%	54.369 $\pm$ 0.05%	67.376 $\pm$ 0.02%
Al ₂ O ₃	17.050 $\pm$ 0.09%	13.439 $\pm$ 0.06%	15.131 $\pm$ 0.13%
Number of Analyses	5	6	3

taken before and after heating and the crucible weights were carefully monitored. Normative compositions have been determined on a loss-free basis.

A1.6 Trace elements (Rb,Sr)

Rubidium and strontium concentrations were determined for all samples using a Phillips 1540 X-ray fluorescence spectrometer. Samples consisted of fine powders (<120 mesh) and were analyzed in duplicate, resulting in discrepancies less than 5%. Variation in measured Rb/Sr between the two duplicate measurements was usually less than 3%. To test that the values of Rb/Sr obtained by this method were sufficiently accurate, a dozen samples were selected for precise X-ray fluorescence spectrometry by the boric acid pressed-powder pellet method (Norrish and Chappel, 1967). The two data sets were indistinguishable within the limits reported above.

A1.7 Strontium isotopic analysis

Preparation for the measurement of strontium isotopic compositions involved crushing whole rock samples to powders (<120 mesh) of which approximately 200 μg were treated by the procedure outlined in Table A1-2 to yield 10-20 μg of strontium.

The isotopic composition of strontium was measured by a modified (Nuclide Analysis Associates) 30 cm thermal ionization mass spectrometer equipped with a triple filament source, automatic magnetic field peak switching and digital output. For each analysis at least three sets of $\text{Sr}^{88/86}$ - $\text{Sr}^{87/86}$ - $\text{Sr}^{88/86}$ ratios were measured 'on line' using computing facilities developed by P.A. Arriens. Data collection and reduction are described by Arriens and Compston (1968). All analyses were made with Sr^{88} beam intensities between 1 and 3 volts and with Rb^{87} and Sr^{87} tail effects less than 0.1% and 0.02%, respectively. The mean

Table A1-2

Unspiked Sr Procedure:

- 1) Roughly weigh about 200 mg sample into teflon beaker and moisten sample with water.
- 2) Add 10-15 ml HF (50%), evaporate to dryness.
- 3) Add a further 5 ml HF (50%) + 4 drops HClO_4 , evaporate to dryness.
- 4) Dissolve sample in 3 mls 6N HCl evaporate to dryness.
- 5) Heat residue briefly in ~5 mls 2.5 N HCl, cool.
- 6) Load on large column (20 g wet wt Dowex 50 cation exchange column). Allow to sink in, wash with 1 ml 2.5N HCl.
- 7) Add 95/75 ml 2.5N HCl, depending on the calibration of the column.
- 8) Collect sample in a further 30 ml 2.5 N HCl. Dry sample.
- 9) Add 12 drops 1N HCl.
- 10) Add to small column (2 g wet wt Dowex 50 ...)
- 11) When in resin add 10 drops 1N HCl.
- 12) Add 21 ml 1N HCl. When dry add a further 6 ml.
- 13) Collect sample from the final 4 mls 1N HCl.
- 14) Dry down sample, cover with protective film.

The effect of this technique is to randomize magnetic directions in a specimen due to grains with short relaxation times so as to clean away all but the 'stable' direction due to small grains with long relaxation times, which hopefully record the direction of the geomagnetic field when the rock cooled through its Curie point. A stable direction for a site is attained when, in successive demagnetizations at increasing field strength, similar directions are measured for all specimens while the magnetic intensities, J , decrease. Because of mineralogical and textural variations in the lavas and intrusives, different demagnetization field strengths were needed to produce stable field directions. This varied from no magnetic cleaning (stable NRM) to 20 milliteslas (mT).

Field directions and intensities were reduced by the Digico magnetometer computer and these were then used to derive Fisher statistics by a program written by F. Chamalaun and M. McElhinny. Virtual geomagnetic poles (VGP's) and calculation of palaeosecular variation (see Chapter 5) were computed by a program written by myself.

Appendix 2: Geographic and petrographic descriptions of samples studied.

Each description is preceded by a sample number which identifies that sample in the ANU rock collection. (The prefix refers to the year the sample was added to the collection.) All samples other than those from Pitcairn Island (donated by R. Carter) were collected personally.

The following mineral abbreviations are used: Ap = apatite; Biot = biotite; Calc = calcite; Clay = undifferentiated clay mineral; Cpx = clinopyroxene; Gl = glass; Idd = iddingsite; Ne = nepheline; Ol = olivine; Opq = undifferentiated opaque; Pl = plagioclase; Zeol = undifferentiated zeolite. Modal abundances are visual estimates.

73-5 Ua Huka, Marquesas Is; S $8^{\circ} 56'$, W $139^{\circ} 32'$; dyke in cliff east of Hanai; olivine basalt; porphyritic, Phenocrysts: Ol(10%), Cpx(15%), Pl(5%), Opq, matrix - Pl, Cpx, Opq.

73-9 Ua Huka, Marquesas Is; S $8^{\circ} 56'$, W $139^{\circ} 32'$; flow in cliff east of Hanai; olivine basalt; porphyritic, Phenocrysts: Ol-Idd(15%), Cpx(15%), matrix - Pl, Cpx, Opq.

73-11 Nuku Hiva, Marquesas Is; S $8^{\circ} 50'$, W $140^{\circ} 03'$; cobble from beach at Anaho; olivine basalt; glomeroporphyritic, Phenocrysts: Ol, Cpx, Pl(5%), matrix - Pl, Cpx, Opq.

73-12 Nuku Hiva, Marquesas Is; S $8^{\circ} 50'$; W $140^{\circ} 03'$; cobble from beach at Anaho; basalt, holocrystalline, Phenocrysts: Cpx(<5%), matrix - Pl, Cpx, Ol, Opq.

73-13 Nuku Hiva, Marquesas Is; S $8^{\circ} 50'$, W $140^{\circ} 03'$; cobble from beach at Anaho; basalt; vesicular (15%), porphyritic, Phenocrysts: Ol(<5%), matrix - Pl, Cpx, Opq, Idd, Ap.

- 73-14 Nuku Hiva, Marquesas Is; S $8^{\circ} 50'$, W $140^{\circ} 03'$; cobble from path inland from Anaho; dolerite; porphyritic, Phenocrysts: Ol(15%), Cpx(40%), Pl(40%), Opq(<5%), vesicules.
- 73-16 Nuku Hiva, Marquesas Is; S $8^{\circ} 53'$, W $140^{\circ} 03'$; cobble from streambed at Taipi Vai; basalt; fluidal, porphyritic, Phenocrysts: Ol, Pl, Cpx (all <5%), matrix - Pl, Cpx, Opq, Clay, Gl.
- 73-18 Nuku Hiva, Marquesas Is; S $8^{\circ} 53'$, W $140^{\circ} 03'$; dyke in hillside east of Taipi Vai; dolerite; porphyritic, Phenocrysts: Pl(40%), Ol(15%), Cpx(40%), Opq (<5%).
- 73-19 Nuku Hiva, Marquesas Is; S $8^{\circ} 57'$, W $140^{\circ} 11'$; cobble from cliff at Hakatea; basalt; fine grained, matrix - Pl, Cpx, Gl.
- 73-20 Nuku Hiva, Marquesas Is; S $8^{\circ} 55'$, W $140^{\circ} 07'$; dyke in cliff, east side of Taiohae Bay; hawaiite; porphyritic, Phenocrysts: Pl(5%), Cpx, matrix - Pl, Cpx, Opq, Ap.
- 73-22 Nuku Hiva, Marquesas Is; S $8^{\circ} 55'$, W $140^{\circ} 07'$; dyke in cliff, east side of Taiohae Bay; basalt, porphyritic, Phenocrysts: Pl(20%), Cpx(10%), matrix - Pl, Cpx, Opq, Gl.
- 73-23 Nuku Hiva, Marquesas Is; S $8^{\circ} 55'$, W $140^{\circ} 07'$; dyke in beach platform at Taiohae; olivine basalt, glomeroporphyritic, Phenocrysts: Pl(30%), Cpx(20%), Ol(10%), matrix - Pl, Cpx, Opq.
- 73-25 Nuku Hiva, Marquesas Is; S $8^{\circ} 55'$, W $140^{\circ} 07'$; dyke in beach platform at Taiohae; biotite trachyte, porphyritic, Phenocrysts: Pl(15%), Biot(5%), matrix - Pl, Sanidine, Opq, Biot.
- 73-27 Hiva Oa, Marquesas Is; S $9^{\circ} 46'$, W $138^{\circ} 51'$; dyke in cliff west of Puamau; olivine basalt, porphyritic, Phenocrysts: Pl(15%), Cpx(15%), Ol(10%), matrix - Pl, Opq, Cpx, Gl.

- 73-28 Hiva Oa, Marquesas Is; S $9^{\circ} 44'$, W $138^{\circ} 52'$; flow in cliff east of Nahoe; basalt, glomeroporphyritic, vesicular, Phenocrysts: Pl(5%), Cpx(5%), matrix - Pl, Cpx, Opq, Gl, Calc.
- 73-29 Hiva Oa, Marquesas Is; S $9^{\circ} 48'$, W $138^{\circ} 56'$; cobble from beach at Hanaupe; dolerite, Phenocrysts: Pl(50%), Cpx(40%), Ol, Opq(5%), vesicles.
- 73-31 Hiva Oa, Marquesas Is; S $9^{\circ} 48'$, W $138^{\circ} 56'$; cobble from cliff base west of Hanaupe; dolerite, Phenocrysts: Pl(50%), Cpx(40%), Ol, Opq(10%), vesicles.
- 73-32 Hiva Oa, Marquesas Is; S $9^{\circ} 50'$, W $139^{\circ} 03'$; flow in cliff south of Taaoa; olivine basalt, porphyritic, Phenocrysts: Ol(20%), Cpx, Pl(<5%), matrix - Pl, Cpx, Opq, Ap, Idd.
- 73-33 Hiva Oa, Marquesas Is; S $9^{\circ} 50'$, W $139^{\circ} 03'$; cobble from beach at Taaoa; basalt, glomeroporphyritic, Phenocrysts: Pl, Cpx, Opq(~5%), matrix - Pl, Cpx, Opq, Gl., calc.
- 73-34 Hiva Oa, Marquesas Is; S $9^{\circ} 48'$, W $139^{\circ} 03'$; dyke in hillside east of Atuona (near radio station); olivine basalt, porphyritic, Phenocrysts: Ol(10%), Cpx(10%), matrix - Pl, Cpx, Opq, Caly.
- 73-40 Tahuata, Marquesas Is; S $9^{\circ} 59'$, W $139^{\circ} 06'$; cobble from beach at Hanateio; basalt, porphyritic, Phenocrysts: Cpx(5%), Ol(<5%), matrix - Pl, Cpx, Opq, Ap, Clay.
- 73-41 Tahuata, Marquesas Is; S $9^{\circ} 58'$, W $139^{\circ} 07'$; cobble from cliff base south of Hapatoni; hawaiite, vesicular, fluidal, matrix - Pl, Cpx, Gl, Opq.
- 73-42 Tahuata, Marquesas Is; S $9^{\circ} 55'$, W $139^{\circ} 06'$; cobble from beach at Vaitahu; olivine basalt, glomeroporphyritic, vesicular, Phenocrysts: Ol(10%), Cpx, matrix - Pl, Cpx, Opq, Gl, Idd.

- 73-43 Tahuata, Marquesas Is; S $9^{\circ} 55'$, W $139^{\circ} 06'$; cobble from beach at Vaitahu; basalt, fine grained, holocrystalline, Cpx, Pl, Opq.
- 73-45 Fatu Hiva, Marquesas Is; S $10^{\circ} 31'$, W $138^{\circ} 40'$; cobble from beach at Omoa; hawaiite, fluidal, Phenocrysts: Ol, Opq, matrix - Pl, Cpx, Opq, Idd.
- 73-47 Fatu Hiva, Marquesas Is; S $10^{\circ} 31'$, W $138^{\circ} 40'$; cobble from beach at Omoa; basalt, glomeroporphyritic, Phenocrysts: Cpx, Pl, Opq.
- 74-386 Rurutu, Austral Islands; S $22^{\circ} 28'$, W $151^{\circ} 22'$; flow south of Avera; olivine basalt, glomeroporphyritic, Phenocrysts: Pl(10%), Cpx(5%), Ol(5%), matrix - Cpx, Pl, Gl, Opq.
- 74-388 Rurutu, Austral Islands; S $22^{\circ} 28'$, W $151^{\circ} 22'$; cobble from road close to Avera; basanite, fine grained, Pl, Cpx, Ne, Opq, Ol, Gl.
- 74-390 Rurutu, Austral Islands; S $22^{\circ} 25'$, W $151^{\circ} 21'$; dyke from cliff north of Moerai; basalt, fine grained, Cpx, Pl, Opq, Gl.
- 74-393 Rurutu, Austral Islands; S $22^{\circ} 25'$, W $151^{\circ} 22'$; flow from northern tip of the island; olivine basalt, fine grained, Phenocrysts: Ol, matrix - Pl, Cpx, Opq, Gl.
- 74-394 Rurutu, Austral Islands; S $22^{\circ} 28'$, W $151^{\circ} 22'$; inclusion in ancient coral reef south of Avera; olivine basalt, porphyritic, vesicular, Phenocrysts: Ol(5%), matrix - Ol, Pl, Cpx, Gl.
- 74-396 Rurutu, Austral Islands; S $22^{\circ} 27'$, W $151^{\circ} 21'$; agglomeratic unit at centre of island; ankaramite, porphyritic, Phenocrysts: Ol(10%), Cpx(15%), Pl(5%), matrix - Pl, Cpx, Opq, Gl.

- 74-397 Tubuai, Austral Islands; S $23^{\circ} 21'$, W $149^{\circ} 27'$; cobble from river bed east of Mataura; trachyte, holocrystalline, fluidal, Pl, Cpx, Opq, Clay.
- 74-399 Tubuai, Austral Islands; S $23^{\circ} 22'$, W $149^{\circ} 27'$; cobble from river bed east of Mataura; olivine basalt, fine grained, Phenocrysts: Ol, matrix - Ol, Pl, Cpx, Gl, Clay.
- 74-401 Tubuai, Austral Islands; S $23^{\circ} 21'$, W $149^{\circ} 27'$; cobble from river bed, eastern end of island; ankaramite, porphyritic, Phenocrysts: Ol(15%), Cpx(20%), matrix - Cpx, Pl, Opq, Gl.
- 74-402 Tubuai, Austral Islands; S $23^{\circ} 21'$, W $149^{\circ} 27'$; cobble from river bed, eastern end of island; olivine basalt, porphyritic, Phenocrysts: Ol(5%), matrix - Pl, Cpx, Ol, Opq, Gl, Clay.
- 74-403 Tubuai, Austral Islands; S $23^{\circ} 21'$, W $149^{\circ} 29'$; cobble from roadside, centre of the island; basalt, fine grained, fluidal, Pl, Cpx, Ol, Opq, Gl.
- 74-404 Tubuai, Austral Islands; S $23^{\circ} 22'$, W $149^{\circ} 30'$; cobble from roadside, western end of island; basanite, coarse grained, Phenocrysts: Ol, Cpx, matrix - Cpx, Pl, Ne, Opq, Gl, Clay.
- 74-406 Raivavae, Austral Islands; S $23^{\circ} 52'$, W $147^{\circ} 39'$; cobble from stream bed east of Mahanatoa; olivine basalt, porphyritic, Phenocrysts: Ol(10%), Cpx(20%), matrix - Pl, Cpx, Opq.
- 74-407 Raivavae, Austral Islands; S $23^{\circ} 52'$, W $147^{\circ} 39'$; flow from cliff 2 km east of Mahanatoa; mugearite, porphyritic, fluidal, Phenocrysts: K-feld, matrix - Pl, Cpx, Ol, Idd, Opq.
- 74-409 Raivavae, Austral Islands; S $23^{\circ} 52'$, W $147^{\circ} 39'$; cobble from stream bed east of Mahanatoa; olivine basalt, porphyritic, Phenocrysts: Ol, Cpx, Pl, matrix - Pl, Cpx, Opq, Clay.

- 74-410 Raivavae, Austral Islands; S $23^{\circ} 52'$, W $147^{\circ} 40'$; flow from peninsula near Mahanatoa; ankaramite, porphyritic, Phenocrysts: Ol(20%), Cpx(40%), matrix - Pl, Cpx, Opq, Gl, Clay.
- 74-412 Raivavae, Austral Islands; S $23^{\circ} 52'$, W $147^{\circ} 41'$; dyke from cliff at western end of island olivine basalt, porphyritic, Phenocrysts: Ol(10%), Cpx(15%), matrix - Pl, Cpx, Ol, Opq.
- 73-56 Huahine, Society Is; S $16^{\circ} 43'$, W $151^{\circ} 00'$; dyke cutting flows at Lac Maeva; olivine hawaiiite, porphyritic, Phenocrysts: Ol, Cpx, matrix - Pl, Cpx, Opq.
- 73-72 Huahine, Society Is; S $16^{\circ} 49'$, W $149^{\circ} 59'$; Mt. Tiva intrusion, southern end of Huahine Iti; trachyte, fluidal, Pl, K-feld, Opq, Cpx.
- 73-83 Huahine, Society Is; S $16^{\circ} 47'$, W $149^{\circ} 58'$; flow from cliff, eastern side of Huahine Iti; olivine basalt, porphyritic, vesicular, Phenocrysts: Ol(5%), matrix - Pl, Cpx, Opq, Gl, Clay.
- 73-95 Huahine, Society Is; S $16^{\circ} 46'$, W $149^{\circ} 59'$; flow from cliff northern tip of Huahine Iti; olivine basalt, porphyritic, vesicular, Phenocrysts: Ol(5%), matrix - Pl, Cpx, Opq, Gl.
- 73-101 Huahine, Society Is; S $16^{\circ} 47'$, W $151^{\circ} 01'$; dyke from cliff, western side of Apu Bay; olivine basalt, porphyritic, holocrystalline, Phenocrysts: Ol(20%), matrix - Pl, Cpx, Opq, Clay.
- 73-111 Huahine, Society Is; S $16^{\circ} 44'$, W $151^{\circ} 00'$; dyke from cliff, eastern side of Huahine Nui; olivine hawaiiite, porphyritic, Phenocrysts: Ol(10%), matrix - Pl, Cpx, Opq, Idd.

- 73-117 Huahine, Society Is; S $16^{\circ} 46'$, W $151^{\circ} 02'$; Mt. Vaihi intrusion, southern tip of Huahine Nui; trachyte, fluidal, fine grained, Pl, K-feld, Cpx, Opq.
- 73-124 Raiatea, Society Is; S $16^{\circ} 50'$, W $151^{\circ} 22'$; Mt. Tauopu intrusion, Baie Hotopuhu; biotite trachyte, fluidal, porphyritic, Phenocrysts: K-feld, Biot, matrix - Pl, K-feld, Cpx, Opq.
- 73-135 Raiatea, Society Is; S $16^{\circ} 45'$, W $151^{\circ} 29'$; flow from NW coast: olivine basalt, porphyritic, Phenocrysts: Ol, matrix - Pl, Cpx, Opq, Gl.
- 73-140 Raiatea, Society Is; S $16^{\circ} 50'$, W $151^{\circ} 25'$; cobble from stream bed, head of Baie Faaroa; basalt, fine grained, Cpx, Pl, Opq, Gl.
- 73-141 Raiatea, Society Is; S $16^{\circ} 50'$, W $151^{\circ} 25'$; cobble from stream bed, head of Baie Faaroa; olivine basalt, porphyritic, Phenocrysts: Ol(5%), Cpx(15%), Pl(5%), matrix - Pl, Cpx, Opq.
- 73-153 Raiatea, Society Is; S $16^{\circ} 47'$, W $151^{\circ} 24'$; flow from cliff, Pt. Opeha, east coast; trachyte, fluidal, porphyritic, Phenocrysts: K-feld, Pl, Opq, matrix - K-feld, Pl, Opq, Cpx.
- 73-185 Tahaa, Society Is; S $16^{\circ} 35'$, W $151^{\circ} 32'$; dyke intruding flow sequence at Paripari, north coast; hawaiiite, fine grained, Pl, Cpx, Ol, Opq, Gl.
- 73-186 Tahaa, Society Is; S $16^{\circ} 35'$, W $151^{\circ} 32'$; flow in cliff at Paripari, north coast; olivine basalt, porphyritic, vesicular, Phenocrysts: Ol, Cpx, matrix - Pl, Cpx, Ol, Gl, Opq.
- 73-189 Tahaa, Society Is; S $16^{\circ} 36'$, W $151^{\circ} 32'$; flow in cliff, NW coast; olivine basalt, porphyritic, Phenocrysts: Ol, Cpx, Matrix - Pl, Cpx, Opq, Gl, Idd.

- 73-190 Tahaa, Society Is; S $16^{\circ} 37'$, W $151^{\circ} 26'$; dyke intruding sequence at Baie Raai; dolerite, coarse grained, holocrystalline, Phenocrysts: Ol, Cpx, Pl, Opq, Idd.
- 73-192 Tahaa, Society Is; S $16^{\circ} 35'$, W $151^{\circ} 29'$; cobble from river at Hipu, north coast; xenolithic basalt, clots of twinned anhedral olivine phenocrysts in basaltic matrix.
- 73-196 Tahaa, Society Is; S $16^{\circ} 37'$, W $151^{\circ} 33'$; flow in cliff at Baie Tapuamu, west coast; olivine basalt, porphyritic, Phenocrysts: Ol(5%), Cpx, matrix - Pl, Cpx, Opq, Gl.
- 73-199 Maupiti, Society Is; S $16^{\circ} 26'$, W $152^{\circ} 16'$; flow from hillside, south coast; olivine basalt, porphyritic, Phenocrysts: Ol, Cpx, matrix - Pl, Cpx, Opq.
- 73-200 Maupiti, Society Is; S $16^{\circ} 26'$, W $152^{\circ} 17'$; dyke from cliff south coast; olivine dolerite, porphyritic, Phenocrysts: Ol, Cpx, Pl, Opq, Idd.
- 73-201 Maupiti, Society Is; S $16^{\circ} 26'$, W $152^{\circ} 17'$; cobble from beach, western tip of island; olivine basalt, porphyritic, Phenocrysts: Ol, Cpx, Matrix - Ol, Pl, Cpx, Opq, Clay.
- 73-204 Maupiti, Society Is; S $16^{\circ} 25'$, W $152^{\circ} 16'$; cobble from hillside, north coast; biotite olivine dolerite, porphyritic, Phenocrysts: Ol, Cpx, Pl, Opq, Biot, matrix - Ap, Clay.
- 73-206 Maupiti, Society Is; S $16^{\circ} 25'$, W $152^{\circ} 17'$; flow from hillside, north coast; olivine hawaiiite, porphyritic, vesicular, Phenocrysts: Ol, Cpx, matrix - Pl, Cpx, Opq, Gl, Clay.
- 73-207 Maupiti, Society Is; S $16^{\circ} 25'$, W $152^{\circ} 17'$; flow from hillside, north coast; biotite olivine dolerite, porphyritic, Phenocrysts: Ol, Cpx, Pl, Opq, Biot, matrix - Ap, Idd.
- 73-209 Maupiti, Society Is; S $16^{\circ} 25'$, W $152^{\circ} 16'$; flow from cliff, NE coast; hawaiiite, fine grained, Pl, Cpx, Opq, Clay.

- 73-211 Moorea, Society Is; S $17^{\circ} 32'$, W $149^{\circ} 51'$; dyke from cliff, west side of Oponuhu Baie; mugearite, fine grained, vesicular, Pl, K-feld, Opq, Cpx, Ap, Idd.
- 73-215 Moorea, Society Is; S $17^{\circ} 35'$, W $149^{\circ} 51'$; flow from cliff, SW coast, near Haapiti; olivine mugearite, porphyritic, Phenocrysts: Ol, Cpx, matrix - Pl, K-feld, Opq, Cpx.
- 73-225 Moorea, Society Is; S $17^{\circ} 33'$, W $149^{\circ} 49'$; intrusion into sequence on east side of Cook's Bay; trachyte, fine grained, Pl, K-feld, Ol, Opq, Calc.
- 73-233 Moorea, Society Is; S $17^{\circ} 35'$, W $149^{\circ} 51'$; flow from cliff, SW coast; olivine mugearite, porphyritic, Phenocrysts: Ol, Cpx, matrix - Pl, K-feld, Cpx, Opq, Clay.
- 73-234 Moorea, Society Is; S $17^{\circ} 35'$, W $149^{\circ} 51'$; flow from cliff, SW coast; olivine basalt, porphyritic, vesicular, Phenocrysts: Ol(5%), Cpx, matrix - Pl, Cpx, Opq, Gl.
- 73-239 Moorea, Society Is; S $17^{\circ} 35'$, W $149^{\circ} 50'$; dyke from hillside, south coast; xenolithic hawaiiite, trachyte and olivine basalt inclusions in Pl, Cpx, Opq, Gl matrix, Phenocrysts of Ol and Pl.
- 73-244 Moorea, Society Is; S $17^{\circ} 35'$, W $149^{\circ} 51'$; dyke cutting flows, SW coast; basalt, porphyritic, vesicular, Phenocrysts: Pl, matrix - Pl, Cpx, Opq.
- 73-248 Moorea, Society Is; S $17^{\circ} 34'$, W $149^{\circ} 52'$; dyke cutting flows, W coast; olivine hawaiiite, porphyritic, fluidal, Phenocrysts: Ol, Pl, matrix - Pl, Cpx, Opq, Gl, Idd.
- 74-414 Moorea, Society Is; S $17^{\circ} 34'$, W $149^{\circ} 48'$; cobble from stream bed at Afareaitu; olivine hawaiiite, glomeroporphyritic, Phenocrysts: Ol(5%), Pl(10%), Cpx(10%), matrix - Pl, Cpx, Gl, Clay.

- 73-257 Bora Bora, Society Is; S $16^{\circ} 30'$, W $151^{\circ} 43'$; dyke from central ridge of island; olivine basalt, porphyritic, Phenocrysts: Ol(15%), Cpx(10%), matrix - Pl, Cpx, Opq, Ap.
- 73-295 Bora Bora, Society Is; S $16^{\circ} 29'$, W $151^{\circ} 43'$; dyke from cliff, NE tip of island, olivine basalt, porphyritic, vesicular, Phenocrysts: Ol, matrix - Pl, Cpx, Opq, Gl, Clay.
- 73-308 Bora Bora, Society Is; S $16^{\circ} 32'$, W $151^{\circ} 46'$; dyke from central ridge, islet of Toopua; basalt, glomeroporphyritic, Phenocrysts: Pl(5%), Ol, matrix - Pl, Cpx, Opq, Gl.
- 73-318 Bora Bora, Socetty Is; S $16^{\circ} 31'$, W $151^{\circ} 44'$; dyke from cliff, E coast; basalt, porphyritic, Phenocrysts: Cpx, Pl, matrix - Pl, Cpx, Opq, Gl, Idd.
- 73-319 Bora Bora, Society Is; S $16^{\circ} 32'$, W $151^{\circ} 44'$; dyke from cliff, SE coast; olivine basalt, porphyritic, Phenocrysts: Ol(10%), Cpx(20%), matrix - Pl, Cpx, Opq, Gl.
- 73-332 Bora Bora, Society Is; S $16^{\circ} 32'$, W $151^{\circ} 44'$; flow from cliff, SE coast; olivine basalt, glomeroporphyritic, vesicular, Phenocrysts: Ol(15%), Cpx(25%), matrix - Pl, Cpx, Opq.
- 73-336 Bora Bora, Society Is; S $16^{\circ} 31'$, W $151^{\circ} 44'$; dyke from cliff, E coast; olivine basalt, porphyritic, Phenocrysts: Ol(5%), Cpx, matrix - Pl, Cpx, Opq.
- 73-362 Tahiti, Society Is; S $17^{\circ} 49'$, W $149^{\circ} 20'$; flow from cliff, S coast of Taiarapu; trachyte, fine grained, fluidal, Opq phenocrysts, Pl, K-feld, Opq, Cpx, calc.
- 73-373 Tahiti, Society Is; S $17^{\circ} 41'$, W $149^{\circ} 33'$; flow from cliff, Punaruu R; basalt, fine grained, vesicular, Pl, Cpx, Opq, Gl.
- 73-396 Tahiti, Society Is; S $17^{\circ} 45'$, W $149^{\circ} 32'$; flow from cliff, SW coast; basalt, fine grained, Pl, Cpx, Opq, Gl.

- 73-419 Tahiti, Society Is; S $17^{\circ} 48'$, W $149^{\circ} 17'$; flow from cliff, Tautira, Taiarapu; basalt, fine grained, vesicular, Pl, Cpx, Opq, Gl.
- 73-421 Tahiti, Society Is; S $17^{\circ} 42'$, W $149^{\circ} 33'$; cobble from stream at Pirae, W coast; olivine hawaiiite, porphyritic, fine grained matrix - Pl, Cpx, Opq, Gl.
- 73-424 Tahiti, Society Is; S $17^{\circ} 49'$, W $149^{\circ} 20'$; intrusion, Pt. Riri, Taiarapu; trachyte, fine grained, fluidal, Pl, K-feld, Cpx, Opq.
- 73-429 Tahiti, Society Is; S $17^{\circ} 37'$, W $149^{\circ} 31'$; flow from road cutting, Bain Loti; basalt, fine grained, Pl, Cpx, Opq, Gl.
- 74-416 Tahiti, Society Is; S $17^{\circ} 41'$, W $149^{\circ} 33'$; cobble from Punaruu R; olivine basalt, porphyritic, Phenocrysts: Ol(20%), matrix - Pl, Cpx, Opq, -Gl.
- 74-417 Tahiti, Society Is; S $17^{\circ} 48'$, W $149^{\circ} 15'$; cobble from Vaitepiha R, Taiarapu; basanite, fine grained, Opq, phenocrysts, Pl, Cpx, Opq, Gl.
- 74-422 Tahiti, Society Is; S $17^{\circ} 42'$, W $149^{\circ} 33'$; cobble from Orofero R; olivine basalt, porphyritic, Phenocrysts: Ol(5%), matrix - Pl, Cpx, Opq, Gl.
- 74-423 Tahiti, Society Is; S $17^{\circ} 48'$, W $149^{\circ} 29'$; cobble from Mairipeha R; mugearite containing anhedral amphibole in fine grained fluidal matrix, Phenocrysts: Pl, Amph, matrix - Pl, Opq, Gl.

Pitcairn Island—another Pacific hot spot?

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The volcanic Pitcairn Island, which has surface lavas with apparent ages between 0.45 and 0.93 Myr, may be the youngest member of another NW–SE trending volcanic lineament in the Pacific Ocean.

PITCAIRN Island, in the south-eastern Pacific Ocean (24°04'S, 130°06'W), is built entirely of volcanic rock¹. Unlike the majority of volcanic islands in the Pacific basin, which are grouped in distinctive linear chains (see Fig. 1), Pitcairn is isolated from its nearest neighbours, Oeno, Ducie and Henderson (all coralline islands) by 160 km, and it is 600 km from its closest volcanic neighbour, Mangareva in the Gambier Islands. Easter Island lies approximately 2,100 km to the east, near the summit of the East Pacific Rise. The position and youth of Pitcairn Island beg explanation by one or another of the rapidly proliferating theories of nonorogenic volcanism.

Pitcairn Island is built up from the seafloor, a depth of at least 3,500 m (ref. 2). The age of the oceanic crust on which

(anomalies 7–9). Basal sediments from Deep Sea Drilling Project (DSDP) site 75, 1,200 km NNW of Pitcairn (and roughly the same distance from the East Pacific Rise), have been assigned a lower Oligocene age (35–40 Myr)⁴.

The island is approximately elliptical, measuring 4 km by 2 km and is cliffed on all sides, rising to an altitude of 347 m (Fig. 2). Volcanic rocks of two types—lavas and pyroclastics—predominate. Structurally, the island seems to be the remnant of a single shield volcano, its dissected caldera rim forming a prominent annular ridge which opens to the north. Four units have been discriminated: the Tedside Volcanics—gently dipping flows exposed near the west coast; the Christians Cave Formation—an agglomeratic tuff generally overlaying the Tedside Volcanics with erosional unconformity; the overlying Adamstown Volcanics—horizontal flows which fill the central basin; and the Pulawana Volcanics—a sequence of lava flows at the western extremity of the island, which may be coeval with the Adamstown Volcanics, as their lithology suggests. There are also some dykes, mainly within the Tedside Volcanics.

Petrologically, the lavas range from alkali olivine basalts through hawaiites and mugearites to minor trachytes. The basalts contain phenocrysts of plagioclase feldspar, olivine and augite. The hawaiites and mugearites exhibit trachytic texture and contain fewer phenocrysts than the basalts. A common additional groundmass phase in these samples is biotite. The alkaline character of the trachytes is evident in phenocrysts of alkali feldspar, aegirine augite and fayalitic olivine. Almost all of the rocks contain interstitial alkali feldspar, the amount increasing in the sequence olivine basalt to trachyte. Chemical analyses of hawaiites, mugearites and trachytes from Pitcairn Island have been presented by Lacroix⁵ who finds them comparable to the petrological association in the Marquesas Islands^{5,6}. New analyses from this Pitcairn collection will be published elsewhere.

Geochronology

The samples selected for dating in this study are primarily hawaiites and mugearites. All four lithological units are represented in the age determinations.

Of the 17 selected samples, 12 were chosen for K–Ar age determination, on the basis of thin section examination. Only those specimens which were devoid of any alteration of the K-bearing phases were selected. Generally, they were holocrystalline and nonvesicular.

Measurements were taken following the techniques previously described by McDougall^{7–9}. Constants used in the calculations are: $\lambda_e = 0.585 \times 10^{-10} \text{ yr}^{-1}$; $\lambda_\beta = 4.72 \times 10^{-10} \text{ yr}^{-1}$; $^{40}\text{K}/\text{K} = 1.19 \times 10^{-2} \text{ atom \%}$.

Table 1 presents the analytical data for 20 K–Ar age determinations on the 12 whole rock samples. Generally, the duplicate analyses on 8 of the samples show agreement to within experimental error. The samples are grouped according to the formations defined by Carter¹. Localities and ages are shown in Fig. 2.

The calculated ages fall into at least two groups. Ages for four samples from the Tedside Volcanics, stratigraphically the oldest unit exposed on Pitcairn, form a group distinctly older than the rest, averaging 0.85 Myr and spanning the period

Fig. 1 Pitcairn Island is situated in the Pacific Basin, possibly the SE end of the Duke of Gloucester–Gambier Islands volcanic lineament which is parallel to other well known Pacific Island chains (Hawaiian, Marquesas, Society and Austral). The Duke of Gloucester Islands are all coral atolls at sealevel. Numbers indicate K–Ar ages $\times 10^6$ yr. Bathymetry in the Pacific Basin denoted by the 4,000 m contour; in the inset by the 4,000 m and 2,000 m contours (dotted lines).

the island is constructed is not known with any great precision but magnetic anomaly studies³ predict an age close to 30 Myr

MIGRATION OF VOLCANISM WITH TIME IN THE MARQUESAS ISLANDS, FRENCH POLYNESIA

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Samples from five islands of the Marquesas Island chain (southeast Pacific Ocean) have been dated by the K–Ar method and exhibit a northwest to southeast volcanic migration rate of 9.9 cm/yr. This movement is in the same direction but of intermediate magnitude to results from the Austral Island chain (lower rate) and the Hawaiian Island chain (considerably higher rate). The rate of migration of volcanism in the Marquesas Islands is consistent with the model of rigid Pacific plate movement over a fixed “hot spot” or mantle “plume” provided that the pole of rotation for the Pacific plate for the last 5 my is located near 55°S, 170°E.

1. Introduction

Isolated young volcanic chains of islands and seamounts form nearly parallel line segments in the eastern Pacific Ocean [1]. The extensions of these linear features intersect the East Pacific Rise at approximately right angles, suggesting a connection with the spreading history of that ridge, even though the islands and seamounts are well removed from it and are often considerably younger than the ocean floor on which they sit [2, 3]. Within each island chain, a progressive younging toward the southeastern end has been suggested on geomorphologic evidence [1, 4, 5] and has recently been verified in two cases by potassium–argon geochronology [6, 7].

This paper presents the results of a geochronological investigation of samples from the Marquesas Islands (Fig. 1). While of much shorter length than other island chains of the eastern Pacific, the Marquesas are petrologically representative of non-orogenic Pacific Islands [8, 9] and it was hoped that potassium–argon dating would confirm the youthful age suggested by the appearance of the

islands and lead to an explanation of age distribution within the group consistent with the recent tectonic history of the eastern Pacific.

2. Geology

Eleven volcanic islands located in the eastern equatorial Pacific Ocean form the Marquesas Island chain. The group stretches from Eiao (8°00'S, 141°27'W) on the northwest end to Fatu Hiva (10°35'S, 138°35'W) on the southeast end, a distance of 355 km. The linear arrangement of the islands, striking approximately N 52°W, is nearly parallel to other young volcanic lineaments of the eastern Pacific. Individually the islands are small, the largest being Nuku Hiva, roughly 18 km in diameter, but rise to heights in excess of 1000 m. Apparently the islands are not the peaks of a prominent ridge but instead rise steeply as individual shield volcanoes from profound depths exceeding 4000 m. (Bathymetric data is from chart 6198 of the Service Hydrographique de la Marine.)

Apart from early reports from explorers and

Fig. 1. The Marquesas Islands are located in the eastern equatorial Pacific. On the locality inset sea level is denoted by stippling while the single bathymetric contour marks the 4000 m depth. The four oceanic plates involved in sea-floor spreading at the East Pacific Rise are italicized. At each of five Marquesas islands, sample localities are indicated by sample number followed by the measured age (underlined). The general structure at the five islands is schematically shown by the arcuate segments representing caldera rims. Incomplete bathymetry about the islands indicated by the 2000-m and 3000-m contours.

surveyors the only published geological studies of the Marquesas Islands are due to Lacroix [8], Chubb [9] and Obellianne [10]. Recently Bishop et al. [11] reported on the petrology of Ua Pou. Chubb's observations and data from the other three contributions may be summarized as follows:

(1) The islands are very similar petrologically, although some islands exhibit extreme differentiates which have not been observed throughout.

(2) The predominant rock types are olivine basalt containing phenocrysts of olivine, clinopyroxene and plagioclase, and trachybasalt, usually very fine-grained, with small augite and plagioclase phenocrysts and variable amounts of larger olivine phenocrysts. Chemically the basaltic lavas range from slightly nepheline normative to hypersthene-olivine normative and thus generally are alkaline to transitional in character. (This independence of the 1-atm thermal divide along the plane of silica undersaturation suggests that the liquids which formed these lavas separated from the parent mineral assemblage at considerable depth [12]).

(3) A three-stage volcanic cycle can be identified on each island: shield-building, followed by intermittent violent eruptions, with caldera collapse, followed by a phase of minor intrusions, including dykes and sills. The three phases are not sharply distinct in time but do describe the general trend in the evolution of these islands.

(4) Differentiation proceeded from the olivine basalt through the more silica enriched trachybasalt to trachytes and phonolites (Ua Pou only). Transitional hawaiites and mugearites are also observed. As at other Pacific Islands the extreme types appear to have been erupted only in the waning stages of volcanic activity. Very often these late-stage differentiates contain ultramafic xenoliths.

The samples collected for this study are representative of both the main shield-building lavas and the minor intrusives. Mineralogically they are indistinguishable from the descriptions provided by Chubb and Lacroix.

As reported by Swainbank [13], Marquesas samples are anomalously rich in Pb^{206} with respect to the single stage growth curve and compatible with other South Pacific island samples (with the exception of Rarotonga in the Cook Islands), falling between the -400-my and -600-my isochrons. These lead isotope compositions can be explained only by a multi-stage

model of lead evolution. The $^{87}Sr/^{86}Sr$ ratios [13] are also in accord with world wide non-orogenic ocean islands and significantly higher than the average for ocean ridge samples.

Geomorphologically the islands appear to be extremely young. The absence of fringing or barrier coral reefs has produced precipitous sea cliffs which rise from the ocean to heights of over 650 m. Large "planezes", or remnants of the original surface of the shield, persist, particularly on Fatu Hiva, Tahuata, and Hiva Oa, islands to the southeast end of the chain. No major plutonic bodies have been exposed by erosion.

3. Experimental methods and results

Samples were collected from five of the Marquesas Islands: Nuku Hiva, Ua Huka, Hiva Oa, Tahuata, and Fatu Hiva (see Fig. 1). As the main interest of the study was geochronology, the condition of the samples was of primary importance. Only rocks devoid of macroscopic weathering were collected. A second consideration was the geological situation of the sample. Where possible samples were collected in situ, and of the widest possible petrological spectrum. Where outcrop was limited or heavily weathered, samples were taken from stream and beach cobbles. Due to the extremely precipitous and enclosed nature of the beaches it is unlikely that the cobbles came from any great distance. Specimens representative of the common beach cobbles were collected to reduce this possibility. Of the 50 samples collected, 29 were selected after microscopic examination of thin sections for age determinations. Samples which exhibited deutric alteration were eliminated. Generally the specimens chosen for geochronology were holocrystalline, though some contained glass which showed no evidence of devitrification.

Measurements were made following the techniques previously described by McDougall [6, 14, 15]. In all determinations whole rock samples were crushed to a size of 3–1.5 mm: a representative split of this crushing was used in the potassium analysis, the remainder saved for argon extraction. Potassium was determined in duplicate by flame photometry.

Argon was extracted by fusing a 15–30 g of sample in a high vacuum line. The method of isotope dilution, with an ^{38}Ar tracer, was used to determine the

TABLE 1
Geochronology of basalts from the Marquesas Islands

Sample		% K	Rad. ^{40}Ar ($\times 10^{-8}$ mol/g) [†]	Rad ^{40}Ar Total ^{40}Ar $\times 100$	Age (my) ± 2 s.d.
<i>Nuku Hiva</i>					
73-11	cobble from beach at Anaho	0.609, 0.599	4.188	57.2	3.89 ± 0.06
73-12	cobble from beach at Anaho	0.706, 0.698	5.242	49.0	4.21 ± 0.06
	cobble from beach at Anaho	0.706, 0.698	5.270	46.3	4.23 ± 0.07
73-13	cobble from beach at Anaho	0.381, 0.382	2.674	18.5	3.93 ± 0.14
73-14	cobble from beach at Anaho	0.649, 0.636	4.238	56.0	3.70 ± 0.09
73-16	cobble from hillside at Taipi Vai	1.383, 1.398	9.218	65.3	3.72 ± 0.06
73-18	dyke from hillside east of Taipi Vai	1.629, 1.624	8.658	50.2	2.99 ± 0.05
	dyke from hillside east of Taipi Vai	1.629, 1.624	8.828	45.4	3.05 ± 0.05
73-19	cobble from cliff base at Hakatea	0.585, 0.586	3.954	25.8	3.79 ± 0.09
73-20	dyke from beach platform at Taiohae	1.437, 1.441	10.433	61.7	4.07 ± 0.06
73-22	dyke from cliff at Taiohae	1.990, 1.990	13.313	31.8	3.76 ± 0.08
73-23	dyke from cliff at Taiohae	1.974, 1.987	13.622	47.7	3.86 ± 0.07
<i>Ua Huka</i>					
73-5	dyke from cliff east of Hanai	1.091, 1.092	5.338	45.4	2.75 ± 0.04
73-6	cobble from cliff east of Hanai	1.103, 1.103	5.327	81.2	2.71 ± 0.03
73-8	flow from cliff east of Hanai	1.198, 1.199	5.936	83.1	2.78 ± 0.03
73-9	flow from cliff east of Hanai	1.172, 1.162	5.779	74.4	2.78 ± 0.05
<i>Hiva Oa</i>					
73-27	dyke from cliff west of Puamau	1.979, 1.996	6.533	14.0	1.85 ± 0.15
	dyke from cliff west of Puamau	1.979, 1.996	6.694	10.5	1.89 ± 0.11
73-28	flow from cliff at Nahoe	1.960, 1.943	5.514	69.4	1.59 ± 0.03
73-29	cobble from beach at Hanaupe	1.640, 1.630	5.361	75.1	1.62 ± 0.03
73-31	cobble from beach at Hanaupe	2.133, 2.131	6.011	68.0	1.58 ± 0.03
73-32	flow from cliff at Taaoa	1.030, 1.045	4.592	63.2	2.48 ± 0.06
	flow from cliff at Taaoa	1.030, 1.045	4.572	62.3	2.47 ± 0.06
73-33	cobble from beach at Taaoa	2.366, 2.341	8.352	78.4	1.99 ± 0.04
73-34	dyke from hillside at Atuona	1.102, 1.099	3.379	38.3	1.72 ± 0.04
<i>Tahuata</i>					
73-40	cobble from beach at Hanateio	0.938, 0.937	3.406	39.7	2.04 ± 0.04
73-41	cobble from beach at Hapatoni	0.949, 0.949	3.251	79.6	1.92 ± 0.03
73-42	cobble from beach at Vaitahu	1.074, 0.074	3.613	51.5	1.89 ± 0.03
73-43	cobble from beach at Vaitahu	1.101, 1.103	3.484	31.9	1.78 ± 0.04
<i>Fatu Hiva</i>					
73-44	flow from cliff at Omoa	0.599, 0.595	1.473	15.1	1.39 ± 0.05
73-45	cobble from beach at Omoa	1.482, 1.484	3.574	64.2	1.35 ± 0.02
73-47	cobble from beach at Omoa	0.860, 0.860	1.983	45.9	1.30 ± 0.02
73-49	cobble from beach at Hanavave	1.404, 1.418	3.469	54.1	1.38 ± 0.03

amount of radiogenic argon liberated from the fused sample. Mass spectrometry, using an AEI MS-10 with digital read out system and 4.2-kilogauss permanent magnet, determined the isotopic composition of the

argon extracted from the samples. Data reduction and calculation of ages were by computer, as was an error analysis described previously [15]. Constants used in the calculations were: $\lambda_e = 0.585 \times 10^{-10} \text{ yr}^{-1}$, $\lambda_\beta = 4.72 \times 10^{-10} \text{ yr}^{-1}$, $^{40}\text{K}/\text{K} = 1.19 \times 10^{-2}$ at. %.

Table 1 presents the results of 33 age determinations on 29 samples, 4 run in duplicate. Samples are grouped according to island and the islands are arranged in geographical order from northwest to southeast. Average ages for the five islands are: Nuku Hiva — 3.85 my, Ua Huka — 2.75 my, Hiva Oa — 1.85 my, Tahuata — 1.91 my, and Fatu Hiva — 1.35 my. Fig. 1 illustrates the distribution of samples and ages within the islands.

Sampling on Ua Huka was restricted to a single valley (Hanai). The four samples give concordant ages leaving little doubt that this is their age of crystallization. Sample 73-5 was collected from a dyke that intrudes the lava sequence from which samples 73-8 and 73-9 were taken. At Fatu Hiva and Tahuata a similar consistency is observed, but over a wider sampling area (Fig. 1).

Nuku Hiva and Hiva Oa have been more thoroughly sampled than the other islands and exhibit the greatest variation in measured age. The ages of those samples which from field relations and petrology are likely to be from the shield-building stage of volcanism at Nuku Hiva (73-11, 73-12, 73-13, 73-19) span a relatively short 0.4-my period, which is comparable with observations from geochronology of the Hawaiian Islands [6]. The remainder of the samples from Nuku Hiva are minor intrusives or feldspathic differentiates and are of comparable or younger age than the shield-building lavas. (73-18 is probably a very late stage intrusion). At Hiva Oa 73-32 is the only likely shield-building lava and this is suggested by its older age. 73-28 and 73-33 are differentiates while the remainder are intrusives. Although Hiva Oa and Tahuata are now separate islands, their physical proximity permits the possibility that lavas on one were produced from vents on the other, leading to the observed overlapping ages. Indeed, bathymetry (SHOM chart 6198) suggests that the two islands are part of the same edifice and probably should be treated as parts of the same volcanic event.

There is close agreement amongst samples within each island. In addition, very similar ages are determined from samples of quite different potassium contents (some differ by as much as a factor of 5). This, together with the similarity in ages from the 4 duplicate runs, argues for the reproducibility of the results and against the possibility of either argon loss or excess argon. Except for the intermixed ages at Hiva Oa/Tahuata, volcanic activity proceeded systematically from northwest to southeast over a span of nearly 3 my (4.2–1.3 my).

4. Discussion

The curiously linear and parallel nature of young Pacific island chains has attracted much attention in the past [1, 4, 16, 20] and is a crucial piece of evidence in the "hot spot" model of Wilson [17] and the recent "plume" model of Morgan [18]. Previous geochronology of the Hawaiian Islands [6, 19] and of the Austral Islands [7, 21] has documented a migration of volcanic activity from northwest to southeast. McDougall [6, 22] deduced a rate of movement of 15.2 cm/yr for the Hawaiian Islands while Jackson et al. [19] determined a slightly lower rate of 14.8 cm/yr. Johnson and Malahoff [21] found a much lower rate of 9.0 cm/yr for age determinations from the Austral Islands [7].

The K–Ar determinations for samples from five Marquesan Islands (see Table 1) are plotted against distance from Fatu Hiva, the southeasternmost member of the chain (Fig. 2). Average ages as well as extreme ages at each island are illustrated. Age decreases systematically from Nuku Hiva to Fatu Hiva, a distance of 235 km. A rate of migration of volcanic activity of 9.9 cm/yr is determined from a least-squares fit to the average ages at the five islands. This is, of course, an average rate of movement and does not dismiss the possibility of a somewhat episodic history. (Hence the fact that the Ua Huka ages fall off this best-fit rate).

Fig. 2. Ages at each of the five Marquesan islands are plotted against island position as projected on the great circle through Fatu Hiva and Eiao Islands (see Fig. 1). Distance is measured from Fatu Hiva, the southeasternmost member of the chain. Average ages are shown with heavy dots while extreme ages are indicated by vertical bars.

Using this best-fit rate of migration of volcanism an age of 4.7 my is predicted for Eiao, 355 km northwest of Fatu Hiva. If the Marquesan "hot spot" still exists, but is at present dormant, its position is predicted to be 112 km southeast of Fatu Hiva.

Volcanic activity spans a greater amount of time at Nuku Hiva and Hiva Oa-Tahuata than at the other two islands, but this may be simply a sampling bias. Whether volcanism at Ua Huka and Fatu Hiva spans comparable periods of time is a question that can be resolved only by more comprehensive sampling. The age-distance relationship for the five islands very closely approximates a straight line and suggests that the rate of migration of volcanism over the time interval sampled (1.3–4.2 my) has been nearly constant, and no major discontinuities exist.

The Marquesan rate of 9.9 cm/yr lies between the Hawaiian and Austral Island migration rates, though closer to the 9.0 cm/yr rate for the Austral Islands.

If a particular volcanic province (e.g. linear chain of oceanic islands) is the surface manifestation of plate movement over a mantle-fixed "plume", then the geometry and rate of migration of volcanism in that province should determine the *absolute* movement of that plate with respect to the mantle. Other data from the boundaries of the plate, such as sea-floor spreading rates at ridges and strikes of transform faults, determine the *relative* movement of one plate with respect to its neighbours but cannot be used to evaluate the proposition of fixed volcanic sources unless combined with all other relative movements as Minster et al. [23] have shown.

The constraint of rigid plate motion over two or more "plumes" provides an immediate test of "inter-plume" movement. Geometrical and geochronological data from plume traces on the same plate should define a common pole of rotation (or sequence of poles of rotation) of the plate with respect to the plume system (by the model this is the mantle).

In order to make the rates of volcanic migration in the Marquesas Islands and the Austral Islands compatible with the undeniably greater rate in the Hawaiian Islands under the constraints of rigid Pacific plate movement and mantle-fixed plumes, the rotation pole for the Pacific-plume movement for the last 5 my must be closer to the Austral and Marquesas Island chains than to the Hawaiian chain. We propose such a pole at 55°S, 170°E which is consistent with the distribution of vol-

canic migration rates as well as the geometry of the last 5-my segments of the three island chains. If other data preclude this possibility than the proposition that the sources of volcanism for these island chains are fixed relative to one another needs reconsideration.

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Mantle Plumes, Movement of the European Plate, and Polar Wandering

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The passage of lithospheric plates over thermal plumes or hotspots in the mantle is discussed with respect to the history of two igneous chains in the European plate.

IN plate tectonics, magma is generated at ocean ridges and at subduction zones, the former producing alkalic and tholeiitic flood basalts and the latter basalts and calc-alkaline igneous suites. Volcanic and plutonic activity inside the borders of stable cratons, however, cannot be accounted for by interactions at plate margins¹. In addition to kimberlites and carbonatites, examples of this typically alkalic igneous phenomenon are the White Mountain Magma Series of New Hampshire, the Younger Granite Ring Complexes of Nigeria, and the Tertiary Volcanic Province of North-west Scotland. We examine here the hypothesis that the igneous activity on cratons which is unrelated to plate boundaries can be attributed to the passage of the plates over hotspots in the mantle, which represent narrow, upwardly convecting "plumes" of mantle material which thermally and mechanically pierce the overlying crust²⁻⁴. According to Morgan^{2,3}, the greater part of the upwelling plume impinges on the bottom side of the lithosphere and spreads out radially in the aesthenosphere and it is possible that the combination of several such plumes exerting viscous forces on the underside of the lithosphere may initiate rifting and drive the spreading of crustal plates.

In support of this model, Morgan^{2,3} has pointed out the geometric symmetry of four volcanic chains which could be due to the passage of the Pacific plate over four stationary plumes: the Hawaiian Islands-Emperor Seamounts, the Tuamotu Islands-Line Islands, the Austral (Tubai) Islands-Marshall-Gilbert Islands, and the seamount chains in the Gulf of Alaska. A monotonically increasing age relationship has been shown for the Austral Chain⁵ and also for the Hawaiians⁶ as suggested by Wilson⁴. The Atlantic Ocean, on the other hand, may be interpreted as spreading from a rift which propagated between the Iceland, Azores, and Tristan da Cunha plants. Aseismic ridges which extend from these and other Atlantic islands across the ocean floor to the continents may be manifestations of continued plume activity.

In the plume model, material rising from the lower mantle undergoes some degree of partial melting. The resulting igneous activity is associated with large-scale crustal doming. As magma generation and associated differentiation can occur

at relatively great depth (>100 km), alkalic basalts and associated derivatives may be produced in addition to tholeiitic basalts of shallower origin^{7,8}.

Along the oceanic rifts (ridges) which propagate between plumes, on the other hand, aesthenosphere rises and partial melting occurs at much shallower depth, so that the magmas are characteristically tholeiitic. The chemical differences between ocean ridge (*per se*) or abyssal basalts and the basalts and related differentiates of the oceanic island "plumes" have been noted elsewhere^{9,11}.

Fig. 1 Geographic map showing location of Thulean and Central European volcanic provinces, and the principal centres of which they are composed. 1, Eifel, Eifel Siebengebirge; 2, Westerwald; 3, Vogelsberg; 4, Knüllgebirge; 5, Rhön; 6, Reichsforst; 7, Dupperrauergebirge; 8, Bohemian Mittelgebirge; 9, Lausitz; 10, Bober-Katzbach.

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Plumes and Cratonic Activity

Dunham¹² has suggested that the magma source now beneath Iceland earlier produced the Tertiary igneous centres of Scotland and other members of the Thulean Province. Migration from this stationary centre is attributed to seafloor spreading. If Morgan's model is correct, plumes tend to rupture and drive crystal plates, and normal faulting or rifting is commonly associated with the emplacement of alkalic igneous complexes. The general hypothesis advanced here, however, is that mantle plumes are primarily responsible for all cratonic igneous activity which is or was clearly independent of plate boundaries.

The passage of a lithospheric plate over mantle plumes should leave surface scars in the form of igneous activity. If the plumes are stationary within the mantle (aesthenosphere), the pattern of these scars, on oceanic or continental crust,

oceanic and continental crust; the other (the Eifel plume) is entirely confined within the continent (Fig. 1).

The principal volcanic or igneous components of the two plume provinces are shown in Fig. 1. A summary of the rock types found and the available radiometric and stratigraphic age data are listed in Table 1¹³⁻²³ and Table 2²⁴⁻³⁴.

In the Thulean Province¹³⁻²³, the age data suggest a general increase in age from Iceland to the Faeroe Islands and Northern Ireland. If this is considered to be a single scar-trace, however, the distribution pattern and age data reveal distinct segments. These may represent different orientations and velocities for the motion of plate relative to plume. Interpretation in terms of the plume model is by no means simple, however, and for clarity the history of the Thulean Province will be discussed with reference to three geographic segments: Iceland-Faeroes; Skye-Mourne Mountains; Faeroes-Rockall-Skye.

Table 1

Igneous centre	Form of activity	Age (m.y.)	Reference	Petrology
Iceland ¹³		0-16.3	14	
Faeroes ¹⁵	Flows	55	16	Olivine basalts, tholeiite
Rockall	Intrusive	49, 60	17	Aegirine granite
St Kilda	Central complex	36 ± 6	17	Basic
Skye	Tuffs, fissure eruption, central plutonic complexes	54 ± 3	19	Olivine basalts, mangerites, trachytes, alkalic granites, gabbros and differentiates
Rhum ¹²	Lavas, layered complex in volcanic throat			Olivine basalts, differentiated ultrabasics
Ardnamurchan	Three volcanic centres, stocks, ring dykes	55 ± 6	20	Gabbro, eucrite, granophyre
Mull ¹⁸	Calderas, sheets ring dykes, explosive vents (doming)	60 ± 12	21	Granophyres, olivine, gabbros, eucrite
Arran	Granite stock, central ring complex	60 ± 6	20	Granite, gabbro, ring dykes
Antrim Plateau	Lavas	74	18	Olivine basalts, tholeiites
Mourne Mountain	Ring dykes	75 ± 7	22	Granites, felsites
Slieve Gullion ²³	Ring dykes			
Carlingford				

Table 2

Igneous centre	Form of activity	Age	Reference	Petrology
Eifel ^{24,25}	Strato volcanics	10,000 yr		
Eifel-Siebengebirge ^{24,25}	Tuffs, flows	Miocene, 18-26 m.y.	30	Trachyte, andesite, basalt
Westerwald ²⁵	Tuffs, lavas	Pliocene		Trachyte, basalt
Vogelsberg ^{24,26,27}	Lavas (doming)	Tortonian, Sarmatian, Pliocene		Phonolites, basalts
Knüllgebirge ²⁴	Lavas	Miocene		Trachy-dolerites, feldspathic basalts
Rhön ²⁴	Lavas (doming)	Tortonian		Basalts, phonolite
Reichsforst, Oberpfalz ^{24,28}	Cones, flows	Miocene, 18-23 m.y.	30	Nepheline and melilite basalt
Duppauergebirge ^{25,29}	Tuffs, flows	Oligocene, Miocene		Nepheline basalt
N-Bohemian volcanics ²⁵	Tuffs, flows, laccoliths	Oligocene		Basalts, phonolites, essexite
Lausitz ^{24,30-32}		20-32 m.y.	30	Trachytes
Bober-Katzbach ^{25,33-35}	Flows	29-34 m.y.	34, 35	Basalt

would be a record of the absolute motion of the plate relative to the aesthenosphere.

Therefore if cratonic igneous centres are plume scars, they should demonstrate a temporal sequence, and a systematic geometry. Furthermore if polar wander curves are the result of lithospheric plate motion only, and the Earth's magnetic field is and has been a centred, axially symmetric dipole field, then the magnetic polar wander curve for a particular plate should likewise be compatible with its plume trace. If only the last test is not met, then either the assumptions implicit in it are invalid or true polar wander has occurred.

Here we attempt such an appraisal by reference to the history of two partly contemporaneous igneous chains in the European plate which are inferred to be the result of plume activity. One of these (the Thulean plume) traverses both

The oldest rocks on Iceland are found at the eastern (12.5 m.y. and western (16.0 m.y.) edges of the island¹⁴. Ages decrease toward the active rift zone¹³.

The Wyville-Thompson ridge extends from Iceland to the Faeroes. Basalt has been dredged from this ridge, but no age data have been reported. The Faeroe Islands represent a high-lying part of the Wyville-Thompson ridge composed predominantly of quartz tholeiite flood basalts¹⁵. No substantially alkalic rocks have been described here. The age of the volcanic activity as given by radiometric age determination¹⁶ is between 50 and 60 m.y. (average 54 m.y.). The apparent velocity of plate-plume motion is 1.3 cm yr⁻¹ (Fig. 2).

From Skye to the Mourne Mountains the ages of igneous activity increased from about 54 m.y. (Skye) to 75 m.y. in the Mourne Mountains²² 350 km away. The volcanic centres

Fig. 2 A distance/time plot for the Thulean province showing contrasting rates for the three segments: Iceland-Faeroes, Faeroes-Skye, Skye-Mourne Mts. The rapid Faeroes-Skye segment could be extended back to Arran; in this case the Arran-Mourne Mountains segment would require a rate of 1.0 cm yr^{-1} . Such a choice is not critical for the arguments presented.

include some of the classic central complexes in the annals of igneous petrology. Rocks representing a spectrum of magma types and associated differentiates occur, but alkalic varieties are well represented (Table 1). The apparent plate velocity is between 1.0 and 1.8 cm yr^{-1} (Fig. 2), but because the trace is roughly north-south, the rotation pole is quite different from that associated with the Iceland-Faeroe segment.

We have not found a satisfactory explanation for the plate and plume relation during the formation of the Faeroes-Skye segment. The igneous activity responsible for Skye, the Faeroes, for Rockall and St Kilda is all essentially contemporaneous, around $55 \text{ m.y. ago}^{19,21}$, and a very rapid (about

15 to 20 cm yr^{-1}) southward movement of the plate is demanded (Fig. 2). But the principal opening of the north-east Atlantic is thought to have started about this time^{36,37}, and may be a cause of the complexity.

The Faeroes are located at the northern end of the Rockall Plateau. There are differing views on how this continental fragment became separated from Europe: Le Pichon *et al.*³⁶ suggest that it is the result of a once active NE-SW spreading ridge with its axis in the Rockall Trough, but Laughton³⁷ considers the Rockall Trough to be a NE-SW trending transform fault. In either case, however, if the separation of Rockall from Europe occurred after the plume had produced the Faeroes, the anomalously high plate velocity inferred from the Faeroes-Skye segment would in part be explained.

Fig. 3 Distance/time plot for the Central European volcanic province. The early initiation of volcanism in the E. Eifel-Siebengebirge-Westerwald area is tentatively attributed to the prior existence of the Rhine graben, which provided an easier ascent route for the magmas.

The volcanic province of Central Europe^{24,34,35} (the Eifel Plume) comprises a volcanic chain, over 700 km long, which extends from the Eifel Mountains in the west across central Germany and Czechoslovakia to upper Silesia in Poland (Fig. 1). The available data on age and petrological character

Table 3

Rotation pole	Rotation angle	Site	Reference	Age (m.y.)	Unrotated pole		Rotated pole	
					Latitude	Longitude	Latitude	Longitude
a 72N, 10E	8.2 c c=clockwise	Upper T, mean	39	20	83N	159W	83N	173E
			40	20	83N	143E	84N	172W
			39	20	83N	137E	81N	121E
(a) b	a+5.5 ac ac=anti-clockwise	Lower T, mean	39	60	77N	150E	66N	123E
			40	60	76N	150E	66N	123E
			39	65	78N	153E	67N	125E
4.5N, 116W	6.1 c	British Tertiary (Igneous)	41	65	77N	171E	67N	135E
			42	10	86N	66E	84N	72E
			43	6	68N	132E	67N	128E
a 72N 10E	3.2 4.7 5.8	Iceland Westerwald Vogelsberg	43	11	86N	168E	85N	147E
			44	12	83N	163W	83N	177E
			45	26	72N	137W	73N	160W
	12.4 23.4	N. Bohemia Lausitz	46	28	75N	123E	71N	108E
			41	54	77N	161E	72N	122E
					02S	94W	4.5N	116W
(a) b	a+4.5 ac a+5.0 a+5.2	Skye Rhum Ardnamurchan	40	55	74N	157E	65N	127E
			40		68.5N	151E	58N	125E
			40	55	69N	165E	59N	135E
4.5N 116W	a+6.3 a+7.0	Mull { lavas intrusives	40	60	82N	69W	85N	162E
			40		72N	133E	61N	113E
			40	60	78N	149E	67N	123E
		Arran Antrim igneous non-igneous lavas	40	74	78N	145E	66N	122E
			40		76N	138E	64N	118E
			47		70N	163E	69N	134E

are given in Table 2 and Fig. 3. Igneous activity at individual centres has apparently persisted for up to 25 m.y., which suggests more or less contemporaneous activity throughout the length of the chain. If, however, only the period of most abundant and intense igneous activity at individual centres is considered, a consistent temporal sequence is evident (Fig. 3). The implied plate motion has a velocity of about 2.3 cm yr^{-1} .

Contemporaneous igneous activity over such a wide area can be regarded as evidence against the plume model; however, if these data are interpreted in this way, it requires that minor igneous activity can occur up to 600 km away from the plume site.

Comparison of Thulean and Eifel Plume Traces

The history of the Eifel plume can be traced back only about 35 m.y. In the Thulean province this interval is "represented" by a segment of the Wyville-Thompson ridge extending from Iceland about 400 km toward the Faeroes. From these two segments a rotation pole can be derived at 10° E , 72° N , that simulates well both the geometry of these two contemporaneous segments, and their apparent plate motion velocity (Thulean, 1.9 cm yr^{-1} ; Eifel, 2.3 cm yr^{-1} , see Fig. 4).

Fig. 4 Rotation pole (10° E , 72° N) and small circle arcs for Iceland-Faeroes segment of Thulean province and for the Central European province.

Using this rotation pole, the Eifel equivalent of the remaining Iceland-Faeroes segment can be predicted (Fig. 4). Although the exposed terrain here, however, consists of Mesozoic sediments³⁸, no evidence of an eastward extension of the volcanic chain can be found; it remains uncertain whether the Eifel plume was "born" only about 35 m.y. ago, or whether its earlier extensions are obscured by younger deposits or deformation (as in the Carpathians).

The Faeros-Skye-Mourne Mountains segment of the Thulean plume scar extends almost north-south. Clearly a different rotation pole, located perpendicular to this trend, on

either side, and approaching 90° distant, must be invoked. The pole we favour is located at 4.5° N , 116° W (Fig. 4).

There appear to be no older igneous centres south of the Mourne Mountain, Slieve Gullion, Carlingford group (about 75 m.y.²²). Whether the Thulean plume was born only then, or before the European plate began to move in a direction other than due south, the plume scars being elsewhere, remains unclear.

Comparison of Plume Traces with Magnetic Polar Wander Curves

If we assume that plume traces are due to lithospheric plate motion only, that plumes are fixed in the mantle, and that the geomagnetic field is that of an axial, centred dipole, then a magnetic polar wander curve, caused by the postulated continental drift can be predicted using the lithospheric plate rotation poles (Fig. 4).

Fig. 5 Polar projection comparing the polar wander curve due to continental drift predicted by the plume-traces (solid line) and the palaeomagnetic data obtained at the actual plume-related volcanic centres. Numbering of poles as in Table 3.

The bulk of the contemporaneous palaeomagnetic data from the European plate comes from plutonic igneous or volcanic centres associated with the plume traces (Table 3). Data from individual plume sites are plotted in Fig. 5, for comparison with the predicted polar wander curve.

The palaeomagnetic pole positions are in relatively good agreement, especially those of Lower Tertiary age; they are, however, all distinctly removed from the position predicted by the plume traces (Fig. 5). Reconciliation of this discrepancy requires true polar wandering: either a rotation of the whole Earth, or a displacement of the magnetic pole.

After correction of observed palaeomagnetic pole positions for the inferred continental drift, any remaining difference from the present geographic pole is theoretically a result of true polar wander. We have "corrected" the Thulean and Eifel palaeomagnetic data by rotating them according to the rotation poles and rotation angles appropriate to the original sampling site in the plume traces (Table 3, Fig. 6). The difference which remains is true polar wandering, applicable in principle to the whole Earth. It can be generally described as involving a displacement of the N pole, since Lower Tertiary time, of

approximately 23° from an original position located along the 123° E meridian. The Upper Tertiary data are more scattered, but have an approximate displacement of 7° along the 150° E meridian.

Based on palaeomagnetic data alone, true wandering of the magnetic pole of $13 \pm 5^\circ$ along the 170° E meridian since early Tertiary time has been suggested by McElhinny and Wellman³⁹. Grommé and McKee⁴⁸ have inferred a similar excursion for the north pole from palaeomagnetic measurements on Tertiary volcanic rocks in the western United States.

Validity of the Model

The most controversial elements of our model are: first, the fact that igneous activity associated with the Eifel plume seems to have occurred contemporaneously over distances of 600 km, and to have continued at individual centres for over 20 m.y. Jackson *et al.*⁴⁹ report evidence from the Hawaiian chain of simultaneous tholeiitic eruptions over distances of 300 km. If activity over even greater distances is indeed characteristic of plume activity on continental crust, then the detection of linear trends in time and space will be extremely difficult, if not impossible.

Fig. 6 Polar projection showing location of palaeomagnetic poles corrected for the plume-predicted continental drift. The discrepancy between the corrected palaeo-poles and the present day pole is attributed to true polar wander. (About 23° from a point along the 123° meridian since Lower Tertiary.)

Second, although activity at most of the Scottish Tertiary centres commenced with alkali olivine basalt magmas (Kennedy's Plateau magma type), saturated and oversaturated mafic and felsic rocks are more conspicuous. But in the Eifel plume province, undersaturated, alkalic rocks are most prominent (Tables 1 and 2).

Third, some cratonic igneous activity may result from fracturing or local tectonism or other unknown magma generating agencies, which are quite unrelated to plumes (even if these exist). For example, we have at present no satisfactory explanation for the Upper Tertiary basaltic province of Central France (Massif Central). If there are additional magma generating mechanisms, the pattern of igneous activity would further be confused.

We feel, however, that the weight of the evidence is favour-

able to the model that plumes are relatively stationary in the mantle, and, in addition to producing aseismic ridges in oceans, are responsible for alkalic igneous intrusive-extrusive centres on continents. The hypothesis remains a working one, with many complex and uncertain features. If continental manifestations of plumes can be identified, and conform to the model, then in conjunction with palaeomagnetism, it will be possible to unravel plate motions and distinguish true polar wander in times prior to the oldest existing ocean floor. For this reason, if no other, the concept merits continued consideration and evaluation.

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