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PRIMITIVE MAGMAS IN ARC-TYPE VOLCANIC ASSOCIATIONS: EXAMPLES FROM THE SOUTHWEST PACIFIC

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ABSTRACT

In widely accepted models for the origin of volcanic arcs at convergent plate margins, magmas originate in subduction-modified peridotitic mantle. Primary arc magmas are predictably unique in different arcs because of the variety of potential components and processes involved in their generation, but they should share the common characteristics of high concentrations of Mg, Ni and Cr, reflecting their mantle origin. An examination of suites of samples taken from three southwest Pacific volcanic arcs (the Kermadec, New Zealand, and Papuan arcs) shows contrasting geochemical patterns that correlate with different tectonic settings. Magmas with primitive chemical characteristics are comparatively rare, and appear to occur where an extensional tectonic setting has allowed paths of relatively rapid ascent. In typical arc settings, magma ponds at one or more positions above its source and is modified by a combination of fractionation, eruption, assimilation and recharge processes, so that the most primitive magma erupted is a chemically evolved high-Al basalt. The factors that dictate the geochemical nature of magmas in volcanic arcs are tectonic setting together with the environment (depth) at which primary mantle-derived magmas are modified.

Keywords: arcs, primary magma, high-Mg basalt, Eu anomaly, southwest Pacific.

SOMMAIRE

Dans les modèles les plus répandus qui servent à expliquer l'origine des arcs volcaniques, aux sites de convergence de plaques tectoniques, les magmas proviendraient d'un manteau péridotitique modifié par processus liés à la subduction. Les magmas primaires des arcs semblent par définition uniques dans chaque arc à cause d'une variété de composants potentiels et de processus impliqués dans leur formation, mais ils devraient avoir, comme caractéristiques communes, des teneurs élevées en Mg, Ni et Cr, qui témoignent de leur origine dans le manteau. Nous examinons des suites d'échantillons prélevés de trois arcs volcaniques, Kermadec, Nouvelle Zélande et Papouasie, dans le secteur sud-ouest de l'océan Pacifique, afin de mettre en évidence des comportements géochimiques distincts qui pourraient dépendre des contextes tectoniques variés. Les laves possédant des caractéristiques primitives sont relativement rares, et semblent caractériser les milieux en extension, qui ont facilité la montée relativement rapide du magma. Dans le cas d'arcs typiques, les venues de magma s'arrêtent une ou plusieurs fois au cours de leur montée, et deviennent modifiés par une combinaison de processus associés au fractionnement, à l'éruption, à l'assimilation, et au reflux périodique de magma primitif, de telle sorte que le magma le plus primitif qui apparaisse à la surface correspond à la composition d'un basalte évolué, riche en Al. Les facteurs qui régissent la nature géochimique d'un magma dans les arcs volcaniques sont le contexte tectonique de mise en place et la profondeur à laquelle le magma primaire dérivé du manteau devient modifié.

(Traduit par la Rédaction)

Mots-clés: arcs, magma primaire, basalte magnésien, anomalie en Eu, secteur sud-ouest de l'océan Pacifique.

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INTRODUCTION

The spectrum of rock types that make up the volcanic systems of convergent plate boundaries are arguably the most complex of the igneous associations, from both petrographic and geochemical standpoints. Despite intensive study during the past three decades, fundamental questions of their petrogenesis remain to be answered. In particular, these are the linked questions of the origin of arc-type magmas and the processes that modify these magmas during their ascent to the Earth's surface.

At the source, the variety of potential components (mantle wedge, subducted sediments, subducted oceanic crust) and processes (slab dehydration, mantle fluxing, degree of melting, extent of equilibration) leads us to expect that the compositions of primary arc-type magmas will in some way be unique for each arc. However, these primary magmas should share some common characteristics. If it is assumed that arc-type magmas originate in the mantle overlying subduction zones, then the primary magmas will have equilibrated with a peridotitic residue; we follow Tatsumi & Eggins (1995) who, on this basis, expect these primary magmas to have high magnesium numbers (≈ 70 or higher) and high abundances of mantle-compatible trace elements, such as Ni (>200 ppm) and Cr (>400 ppm).

The identification of primary magmas in volcanic arc associations is dependent on seeing through the effects of processes that modify them during their transit to the surface. These processes are aptly represented by the acronym FEAR (fractionation, eruption, assimilation, recharge; Defant & Nielsen 1990). It is to be expected that they will modify the primary magmas in ways that vary with different crustal settings. If the characteristics of primary magmas can be identified either directly or by stripping away the effects of FEAR, both the processes of magma generation in arcs and the fundamental processes that generate continental crust will be clarified.

Lavas that fulfill the criteria for equilibration with peridotitic mantle residue are relatively uncommon in most volcanic arc associations. In our opinion, this is because the relatively dense primary magmas have insufficient buoyancy to ascend through the crust; instead, they spend prolonged intervals trapped in crustal reservoirs, where FEAR processes operate. Similar arguments have been advanced to explain the fractionated but remarkably uniform chemical characteristics of mid-ocean ridge basalts (*e.g.*, Stolper & Walker 1980). In consequence, we follow Woodhead (1988) in applying the term "parental magma" to describe the magma having the most primitive composition that we can recognize in an arc suite, and we will show later that this composition is usually that of a magma that has undergone neither removal nor addition of plagioclase; therefore, the compositional gap between primary and parental magmas is due to the fractionation of phases other than plagioclase.

In this paper, we use our data sets from three southwest Pacific volcanic arcs as a basis for an investigation of the occurrence and nature of primary and parental arc-type magmas. Our goal is to identify lavas that represent primary or parental magmas in these arcs and to examine how FEAR processes operate upon these

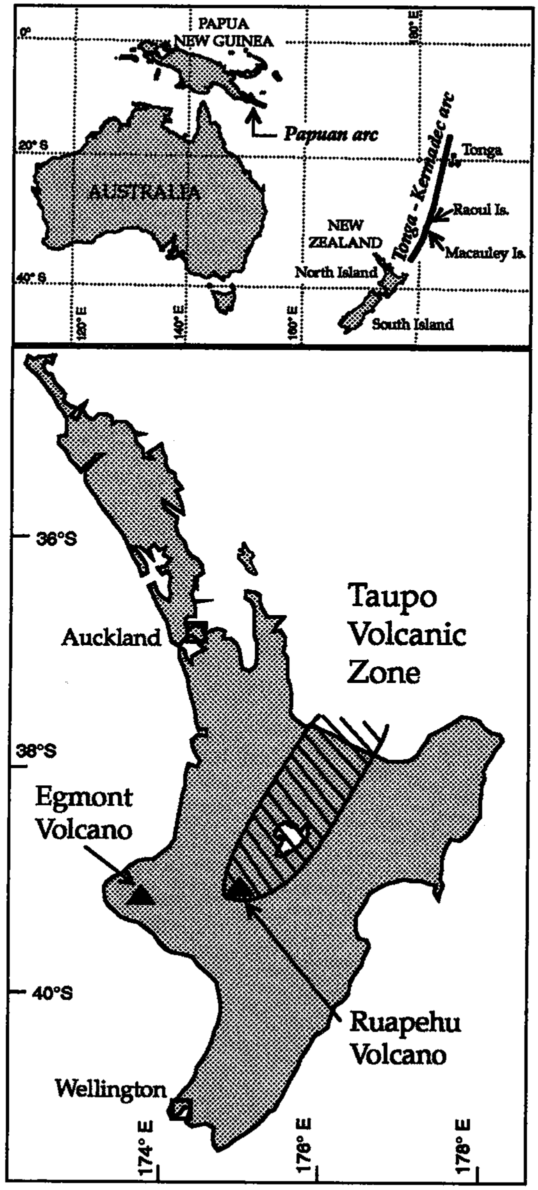

FIG. 1. Location map of the Kermadec and Papuan arcs, and Ruapehu and Egmont volcanoes, in the southwest Pacific.

magmas to generate the wide spectrum of lava compositions. The three arcs studied are the northern Kermadec segment of the Tonga–Kermadec arc, the southern end of the Taupo Volcanic Zone (TVZ) in New Zealand, and the Papuan arc (Fig. 1); each of these arcs occupies a distinct geological and tectonic environment.

Southwest Pacific Arcs

The southwest Pacific has been one of the Earth's principal loci of subduction since the Eocene; consequently, volcanic arcs have been developed on oceanic crust, continental crust, and crust whose nature is transitional between these extremes. Episodes of normal subduction and arc volcanism have been punctuated by periods when oceanic and continental microplates collided with subduction zones, resulting in temporary reversals of arc polarity or the jumping of subduction to new arc–trench systems (*e.g.*, Honza 1991). Another common feature of southwest Pacific arcs has been repeated intervals of rifting and development of back-arc basins, usually accompanied by significant variation in the intensity and composition of arc volcanism (Clift 1995). This plethora of tectonic environments is matched by the chemical variety of the volcanic rocks, which range from arc tholeiite through the low-, medium-, and high-K calc-alkaline suites to shoshonite and, locally, the more exotic boninite and adakite. The arcs we examine in this paper are representative of this tectonic and geochemical smorgasbord.

The simplest environment for a volcanic arc is represented by Raoul and Macauley islands at the northern end of the Kermadec arc (Fig. 1). Both are the emergent summits of large active stratovolcanoes that are located 90–100 km above a well-defined Wadati–Benioff Zone. A combination of geophysical and geochemical data, together with dredgings from the Tonga–Kermadec trench, suggests that the present arc is built upon an 18-km-thick crust of supra-subduction-zone character, consisting of serpentinite, gabbro and peridotite, overlain by an Eocene arc basalt and andesite association (Bloomer & Fisher 1987, Planck & Langmuir 1988, Bloomer *et al.* 1994). Both Raoul and Macauley volcanoes are composed predominantly of highly porphyritic basalt and basaltic andesite containing abundant phenocrysts of plagioclase and lesser clinopyroxene, although recent eruptions (<5 ka) from both have been dominated by dacitic to rhyolitic pyroclastic rocks (Lloyd & Nathan 1981, Nathan *et al.* 1996).

A more complex environment is represented by Ruapehu and Egmont volcanoes, at the southern end of the TVZ in northern New Zealand (Fig. 1). Both of these volcanoes are also large active stratovolcanoes; Ruapehu is located 90–100 km above the Wadati–Benioff Zone, whereas Egmont is located further west and behind the volcanic front, approximately 180 km above the Wadati–Benioff Zone. Geophysical studies indicate that Ruapehu is built upon a 15-km-thick attenuated continental crust, whereas Egmont is built upon a 25-km-thick continental crust (Stern 1985). Lavas from Ruapehu are typical arc-type plagioclase-phyric two-pyroxene basaltic andesite and andesite, with rare basalt and dacite also represented (Graham & Hackett 1987). Egmont lavas are similar to those of Ruapehu, but hornblende is a common phenocryst phase (Price *et al.* 1992).

The most complex environment is represented by the Papuan arc (Fig. 1). The tectonic evolution in southeastern Papua New Guinea involves subduction (possibly with a reversal in polarity), obduction and rifting, the complexities of which are not totally resolved (Davies *et al.* 1984, Smith & Milsom 1984). The Papuan volcanic arc is a curvilinear feature that extends approximately 300 km from the northern coast of the Papuan Peninsula through the D'Entrecasteaux Islands into the Louisiade Archipelago (Smith 1982a). The age of volcanism varies systematically from late Miocene in the east to Recent at the western end of the arc. The lavas range in composition from basalt to dacite and, locally, rhyolite (Smith & Johnson 1981); they are typically porphyritic, with sparse to abundant phenocrysts of plagioclase, olivine, orthopyroxene and clinopyroxene. Hornblende-bearing andesite and dacite are an important feature of this association (Smith 1982a). A group of more sparsely porphyritic lavas has been distinguished as a high-Mg suite by Smith & Mitchell (1989).

Both the Kermadec and the TVZ volcanoes have developed along the same active convergent plate margin, but in different geological settings. The subducting Pacific Plate is early to mid-Cretaceous in age throughout this plate margin, although the rate of subduction progressively decreases from 9 cm per year in the northern Kermadec arc to 4 cm per year at the southern end of the TVZ (Walcott 1978, Pelletier & Louat 1989). Thus, the principal difference in setting between the Kermadec and TVZ volcanoes is that the former overlie mafic supra-subduction-zone crust and are typical of oceanic arcs, whereas the latter are developed on the continental lithosphere of New Zealand and are representative of continental margin arcs. Ruapehu and Egmont volcanoes differ primarily in that Ruapehu is situated on the volcanic front, whereas Egmont is located behind the arc.

The Papuan arc has developed in a convergent margin setting, but is an atypical arc in two respects. Firstly, although the western end of the arc is still active (Smith 1982b), there is no definition of a Wadati–Benioff Zone in the underlying array of seismic foci. This absence has led to the suggestion that either magmatism is related to recently extinct subduction, or that magma generation is a multi-stage process that has involved a delay between subduction-related processes of source enrichment and later partial melting to produce the magmas (Johnson *et al.* 1978). Secondly, the development of the

TABLE 1. REPRESENTATIVE WHOLE-ROCK COMPOSITIONS OF PRIMITIVE LAVAS FROM THE KERMADEC AND PAPUAN ARCS, AND RUAPEHU AND EGMONT VOLCANOES*

Sample	A7114	A7125	A46316	V14765	V14854	V16721	89/77	90/4A	C33614	C33658	C33629	C33652
Source	Kerm.	Kerm.	Kerm.	Ruap.	Ruap.	Ruap.	Egm.	Egm.	Papua	Papua	Papua	Papua
SiO_2	48.84	49.52	68.44	53.81	52.23	59.97	49.54	54.22	51.91	55.22	53.42	53.53
TiO_2	0.65	0.72	0.60	0.70	0.66	0.77	1.17	0.96	1.93	1.46	1.23	1.43
Al_2O_3	20.83	17.69	14.30	17.49	15.57	15.08	16.39	17.63	15.53	17.66	14.53	14.17
Fe_2O_3	10.39	10.59	5.34	3.45	2.70	1.14	4.82	3.08	2.51	3.45	2.83	2.32
FeO				5.12	6.34	4.82	5.51	4.87	6.71	3.83	4.97	3.91
MnO	0.20	0.19	0.18	0.14	0.16	0.07	0.17	0.15	0.15	0.13	0.13	0.11
MgO	4.61	6.43	1.19	5.38	8.73	5.28	6.40	4.81	5.88	3.08	8.60	6.87
CaO	12.82	12.97	4.65	8.51	9.63	6.09	11.09	8.91	8.33	6.92	7.46	9.61
Na_2O	1.47	1.31	3.52	2.94	2.59	3.19	2.99	3.35	3.59	4.36	3.31	2.87
K_2O	0.14	0.18	0.62	0.74	0.58	1.98	1.53	1.77	1.57	2.29	1.79	2.33
P_2O_5	0.06	0.06	0.15	0.10	0.09	0.16	0.26	0.29	0.62	0.56	0.37	0.62
H_2O^-	0.03	0.06	0.02				0.21	0.03	0.39	0.17	0.48	0.33
LOI	-0.28	-0.20	0.69	1.51	0.75	0.59	0.25	0.09	0.42	0.41	0.55	1.22
Total	99.76	99.52	99.70	99.89	100.03	99.14	100.33	100.16	99.54	99.54	99.67	99.32
Sc	35	43	19	35	40	25	30	15	23	16	20	20
V	335	309	29	251	254	163	308	190	271	194	162	206
Cr	13	116	5	86	389	209	92	100	119	0	457	245
Ni	12	44	1	36	135	64	25	41	87	5	291	110
Cu	86	92	0	79	81	43	197	98	30	28	27	141
Zn	71	71	94	88	83	62	78	69	117	74	85	70
Ga	14	13	13	18	17	17	21	21	18	20	18	17
Rb	5	5	9	23	14	89	34	43	30	41	33	53
Sr	203	158	164	216	200	240	552	648	669	996	670	1346
Y	12	15	37	21	18	22	23	20	28	25	21	20
Zr	22	26	67	61	56	173	73	92	182	238	214	428
Nb	0.4	0.5	1.5	3.4	2.4	7.2	2.8	6.7	7	6	7	6
Ba	59	87	202	218	193	464	665	705	460	1011	882	1088
Cs	0.2	0.2	0.6	1.2	0.8	3.6	1.3	0.9	1.3	1.1	0.4	0.6
Hf	0.8	1.0	2.0	1.5	1.4	4.5	2.9	2.3	4.1	4.6	4.0	7.9
Pb	1.9	1.9	2.8	4.9	3.5	13.1	6.6	10.9	9.8	10.8	5.9	8.8
Th	0.2	0.3	0.2	1.8	1.1	8.2	4.2	3.8	3.8	5.9	4.5	10.5
U	0.1	0.1	0.2	0.5	0.4	2.3	6.1	1.1	1.0	1.8	0.9	2.4
La	1.5	2.0	2.7	6.0	4.7	14.0	11.6	13.0	22	39	29	207
Ce	4.5	5.7	8.2	13.4	10.8	31.1	24.7	29.0	54	69	53	311
Pr	0.8	1.0	1.4	1.8	1.5	3.8	3.1		6.1	7.0	5.5	19.7
Nd	4.1	5.1	8.3	8.5	7.3	14.9	14.6	16.8	27.6	30.9		65.1
Sm	1.4	1.8	2.9	2.2	1.9	3.5	4.2	3.8	5.7	6.1	4.9	7.8
Eu	0.6	0.7	1.0	0.8	0.7	1.0	1.2	1.2	1.7	2.0	1.5	2.0
Gd	1.8	2.3	3.7	2.5	2.3	3.4	4.0		4.9	4.8	3.8	5.3
Tb	0.3	0.4	0.8	0.5	0.4	0.5	0.6	0.7	0.9	0.8	0.6	0.6
Dy	2.3	2.8	5.4	3.0	2.7	3.5	4.0		4.8	4.3	3.7	3.5
Ho	0.5	0.6	1.2	0.6	0.6	0.7	0.8	0.7	1.0	0.8	0.7	0.6
Er	1.5	1.6	3.2	1.7	1.5	1.8	2.2		2.6	2.1	1.7	1.4
Tm	0.2	0.3	0.6	0.3	0.2	0.3			0.4	0.4	0.3	0.2
Yb	1.6	1.9	3.7	1.9	1.6	2.0	2.1	1.9	2.3	2.2	1.6	1.2
Lu	0.2	0.2	0.6	0.3	0.3	0.3		0.3	0.4	0.3	0.3	0.2

* Data on two samples of more evolved lavas, A46316 from the Kermadec arc and V16721 from Ruapehu, also are presented.

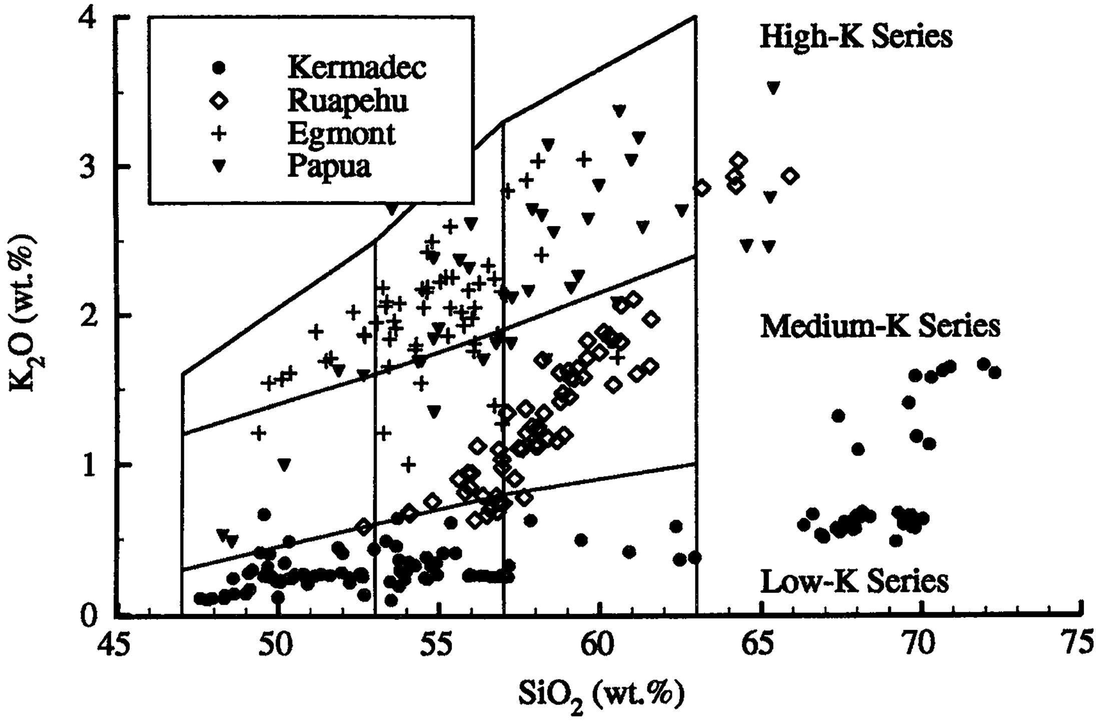

FIG. 2. Lavas from the northern Kermadec and Papuan arcs, and Ruapehu and Egmont volcanoes, plotted on a K_2O *versus* SiO_2 diagram. The classification fields are those of Tatsumi & Eggins (1995).

arc has been associated with the formation of metamorphic core complexes, dramatic uplift and crustal extension linked to the westward migration of sea-floor spreading from the Woodlark Basin. Thus, although the rocks of the Papuan arc are typical of those in many subduction settings, and their origin is considered to have been closely linked to subduction, their present tectonic environment is very different from that of a normal volcanic arc. Nevertheless, we have included the Papuan lavas in our study because they provide an example of a volcanic arc where an unusual tectonic setting may have permitted primitive arc-related magmas to erupt.

The database we utilise to examine lavas from the northern Kermadec, Ruapehu, and Egmont volcanoes and the Papuan arc consists of results of 580 major- and trace-element analyses. Representative compositions of primitive lavas from each of these are listed in Table 1; other representative compositions from this database have been published by Gamble *et al.* (1993), Price *et al.* (1992), Graham & Hackett (1987), and Smith (1982a).

GEOCHEMISTRY

Tatsumi & Eggins (1995) recommended classifying arc lavas according to their position on a plot of K_2O *versus* SiO_2; they used suite boundaries slightly modified from those of Gill (1981). We also adopt this scheme for the southwest Pacific arcs (Fig. 2); we recognize that the boundaries between these geochemical suites are of descriptive significance only. The majority of northern Kermadec lavas belong to the low-K suite, and their array is distinctly bimodal in SiO_2 content, with basalt and basaltic andesite forming the main mode at 47–58 wt.% SiO_2, and the younger dacite and rhyolite, a subordinate mode at 66–72 wt.% SiO_2. Kermadec lavas with intermediate compositions are rare and exhibit disequilibrium features that lead us to believe they were generated by magma mingling. Lavas from Ruapehu form a continuous array within the medium-K field, but the steep slope of this array takes it from the boundary with the low-K field at 56 wt.% SiO_2 into the high-K field at 63 wt.% SiO_2. Most of the Egmont and the Papuan lavas plot within the high-K

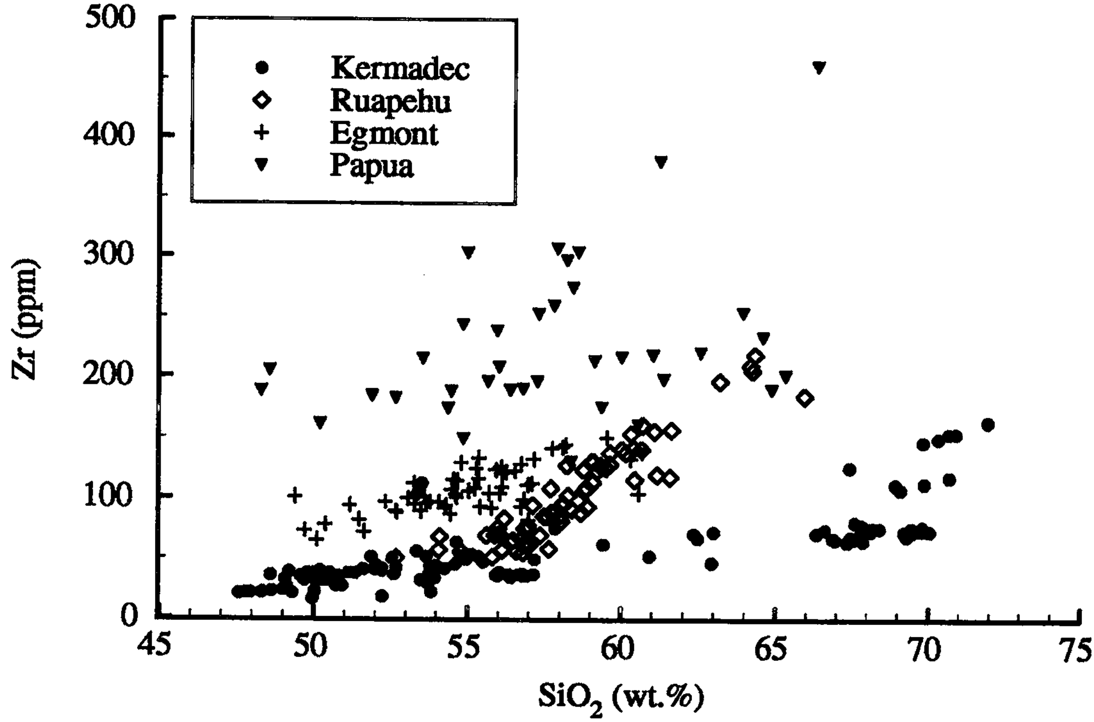

FIG. 3. Lavas from the four arc segments plotted on a Zr *versus* SiO_2 diagram. Note the parallel, but progressively higher, arrays of the northern Kermadec, Egmont and Papuan lavas, and the different slope of the Ruapehu array.

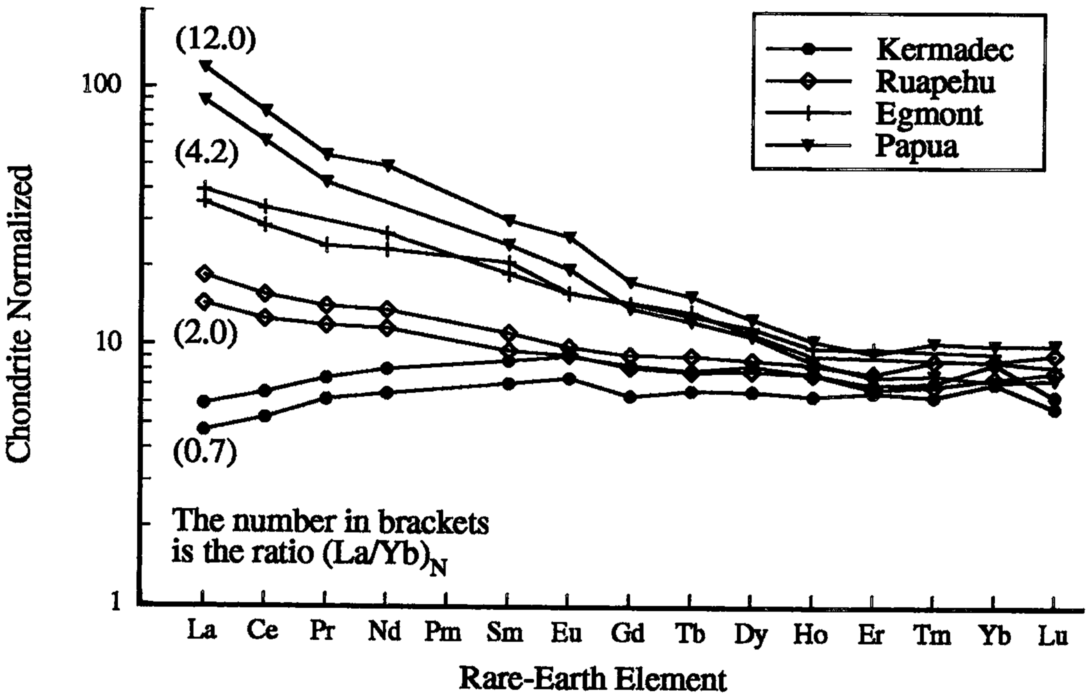

FIG. 4. Plot of chondrite-normalized *REE* abundances for representative primitive basalt and basaltic andesite from the four arc segments. Normalization factors after Nakamura (1974).

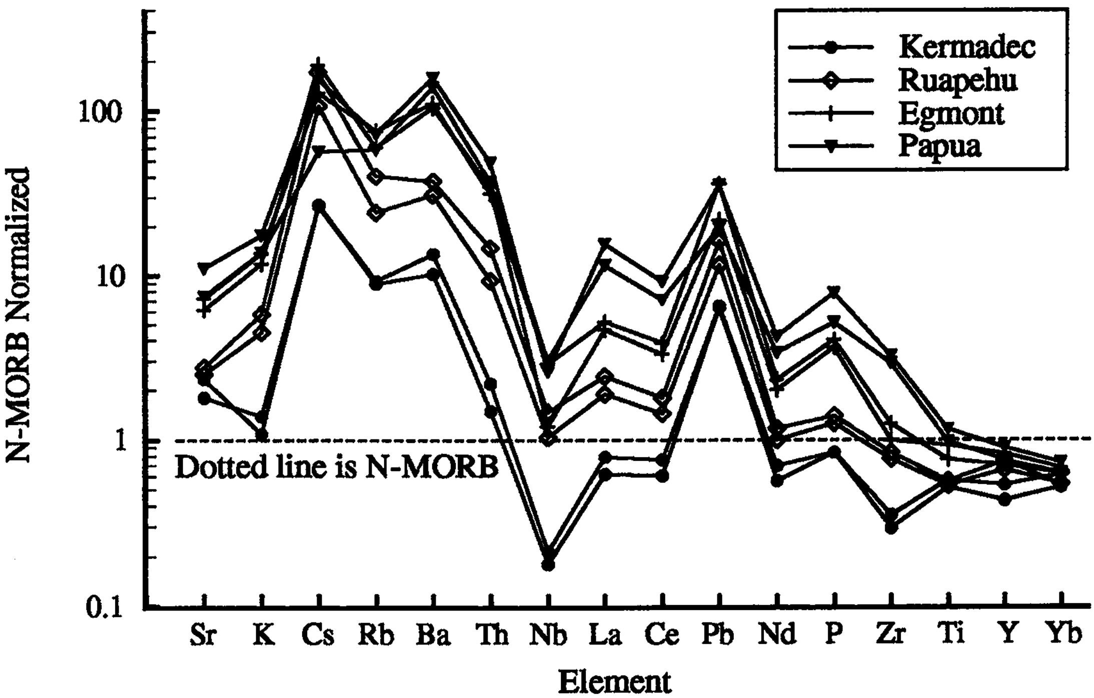

FIG. 5. Plot of N–MORB normalized *LILE* and *HFSE* abundances in representative primitive basalt and basaltic andesite from the four arc segments. Normalization factors after Pearce & Parkinson (1993).

field, but their arrays are less steep than that of the Ruapehu suite. Therefore, in terms of K_2O–SiO_2 variation, our chosen arcs encompass the entire range of typical arc-type volcanic suites.

The covariation of other incompatible elements, both the large-ion lithophile elements (*LILE*) and the high field-strength elements (*HFSE*), with indices of fractionation (*e.g.*, SiO_2, Zr) generates similar trends (Fig. 3). The common characteristic of these types of plots is that the slope of the northern Kermadec array is relatively flat, the Ruapehu array steep, whereas the Egmont and Papuan arrays are of intermediate slope, but usually at higher absolute levels of the incompatible elements (this is especially so for the Papuan array).

Chondrite-normalized abundances of the rare-earth elements (*REE*) in primitive basalts from the chosen arcs range from 4–8 times chondrite and *LREE*-depleted, with $(La/Yb)_N \approx 0.7$, at the northern Kermadec arc, to 60–120 times chondrite and strongly *LREE*-enriched, with $(La/Yb)_N > 8$, for the Papuan suite (Fig. 4). For the Kermadec, Egmont and Papuan arc lavas, the absolute level of *REE* abundances progressively increases with increasing degrees of fractionation from basalt through to dacite and rhyolite, but their chondrite-normalized slopes remain nearly constant. However, Ruapehu is distinct in that $(La/Yb)_N$ also increases markedly with fractionation from $(La/Yb)_N \approx 2$ in basalt to $(La/Yb)_N \approx 6.5$ in dacite.

If normalized to the composition of a normal mid-ocean ridge basalt (N–MORB), the primitive basalts from each of the arcs show the classic features of arc magmatism, namely *LILE* enrichment and either flat or depleted *HFSE* patterns (Fig. 5). These plots again serve to highlight differences among the less evolved lavas of the arcs; although their N–MORB-normalized patterns are similar, Kermadec basalt is mildly enriched in the *LILE*, whereas both Egmont basalt and Papuan basalt are strongly enriched.

THE IDENTIFICATION OF PRIMITIVE MAGMAS

A widely quoted chemical characteristic of arc-type volcanic rocks is their low content of certain transition-group elements, in particular Mg, Ni and Cr, relative to primitive compositions in lava suites known to be derived from mantle sources (*e.g.*, MORB and ocean-island basalt, OIB). This view has been challenged by the recognition of high-Mg lavas in the arc-type associations of the Aleutian Islands (Kay 1978, Nye & Reid 1986, Gust & Perfit 1987), Japan (Tatsumi & Ishizaka 1981, 1982a, b, Aoki & Fujimaki 1982), Vanuatu (Barsdell & Berry 1990, Eggins 1993), the Solomon

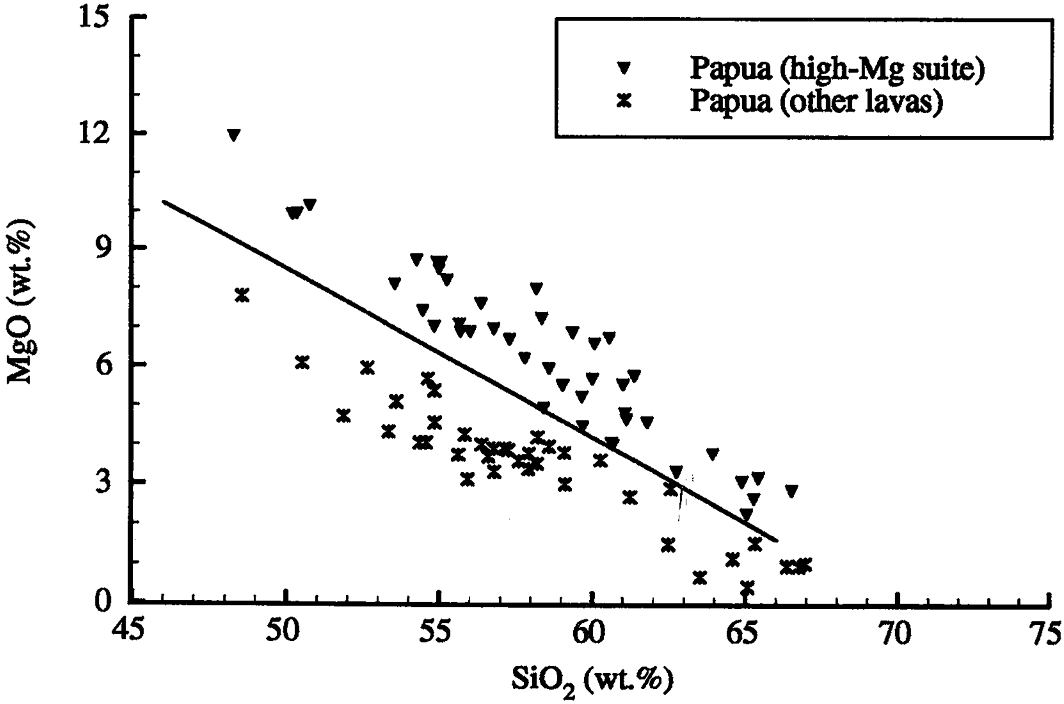

FIG. 6. Plot of MgO *versus* SiO_2 contents of Papuan arc lavas. The high-Mg and normal (other lavas) suites are separated by the line MgO = 30 – (0.43*SiO_2), after Smith & Mitchell (1989).

Islands (Ramsay *et al.* 1984) and as part of comparable associations in continental settings [Mexico: Saunders *et al.* (1987), USA: Meen & Eggler (1987)]. Magnesium-rich lavas also have been described, but not explicitly recognized, from a number of areas around the circum-Pacific rim, for example in northern California (Smith & Carmichael 1968). A review of the literature suggests that high-Mg rocks may be widespread among circum-Pacific arc-type volcanic associations, but because they are a relatively rare component, their importance has been overlooked (Smith & Mitchell 1989).

As noted in the introduction, if primary arc magmas originate in the peridotitic mantle wedge above the subducting slab as partial melts complementary to an olivine- and orthopyroxene-rich residue, then they should have relatively high Mg-numbers, and high Ni and Cr contents. Smith & Mitchell (1989) recognized a group of high-Mg lavas in the Papuan arc, with Mg-numbers of 64 to 76, which approximately fulfil these criteria (Fig. 6). These lavas have a relatively low content of phenocrysts; their chemical characteristics are not, therefore, attributable to crystal accumulation. Smith & Mitchell (1989) defined an empirical reference-line, here modified to the formula MgO = 30 – (0.43*SiO_2) based on further results of analyses and a different technique of normalization, to separate the petrographically distinct high-Mg lavas from those with more typical arc-type compositional characteristics. They estimated that the high-Mg suite constituted 15% of the total volume of outcropping lavas in the Papuan arc.

Using the Smith & Mitchell (1989) criterion for distinguishing high-Mg from "normal" arc-type compositions, several interesting patterns emerge from our data set (Fig. 7). The northern Kermadec suite includes a small proportion of basalts that are relatively rich in Mg and are candidates for relatively primitive magmas, but these are strongly porphyritic and contain disaggregated olivine – clinopyroxene – plagioclase clots. The majority of northern Kermadec basalts and all northern Kermadec lavas of intermediate composition are poor in Mg. In contrast, a significant proportion of Ruapehu compositions are high-Mg basaltic andesites and andesites comparable to the high-Mg Papuan suite. Egmont lavas are entirely poor in Mg. Although the high-Mg basaltic compositions can be interpreted as relatively primitive compositions, and the low-Mg basalts, basaltic andesites and andesites as their fractionated derivatives, a more difficult observation to explain is the relatively high-Mg andesites and dacites of both the Ruapehu and Papuan suites.

A similar picture emerges from a plot of Mg-number against SiO_2 (Fig. 8). Most northern Kermadec and all Egmont lavas have relatively low Mg-numbers for a range of SiO_2 contents from basaltic to dacitic. In con-

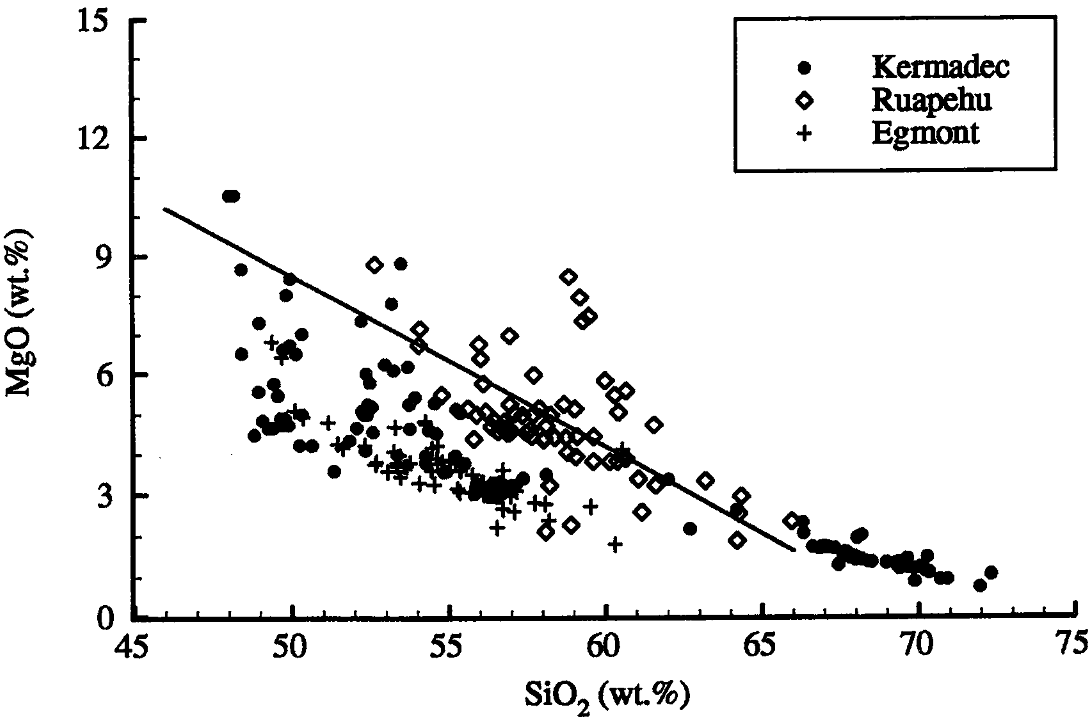

FIG. 7. Plot of MgO *versus* SiO_2 contents of the northern Kermadec, Ruapehu and Egmont lavas. High-Mg field as for Figure 6.

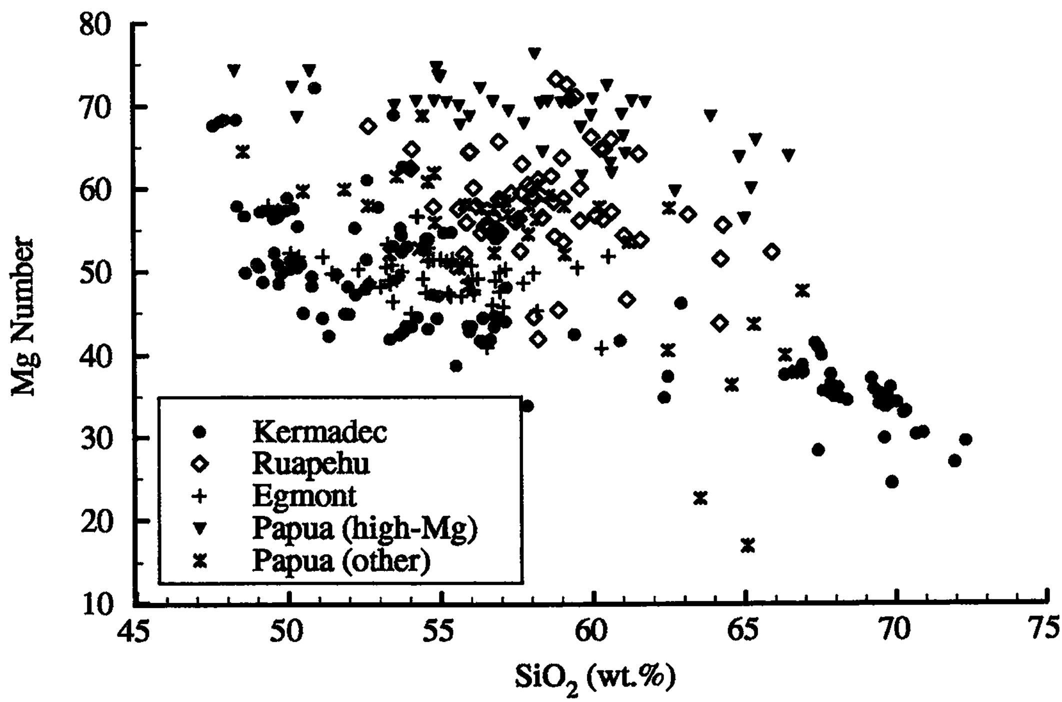

FIG. 8. Plot of Mg-number *versus* SiO_2 content of lavas from the four arc segments. Mg-number = 100*Mg/(Mg+Fe) at Fe_2O_3/FeO = 0.2.

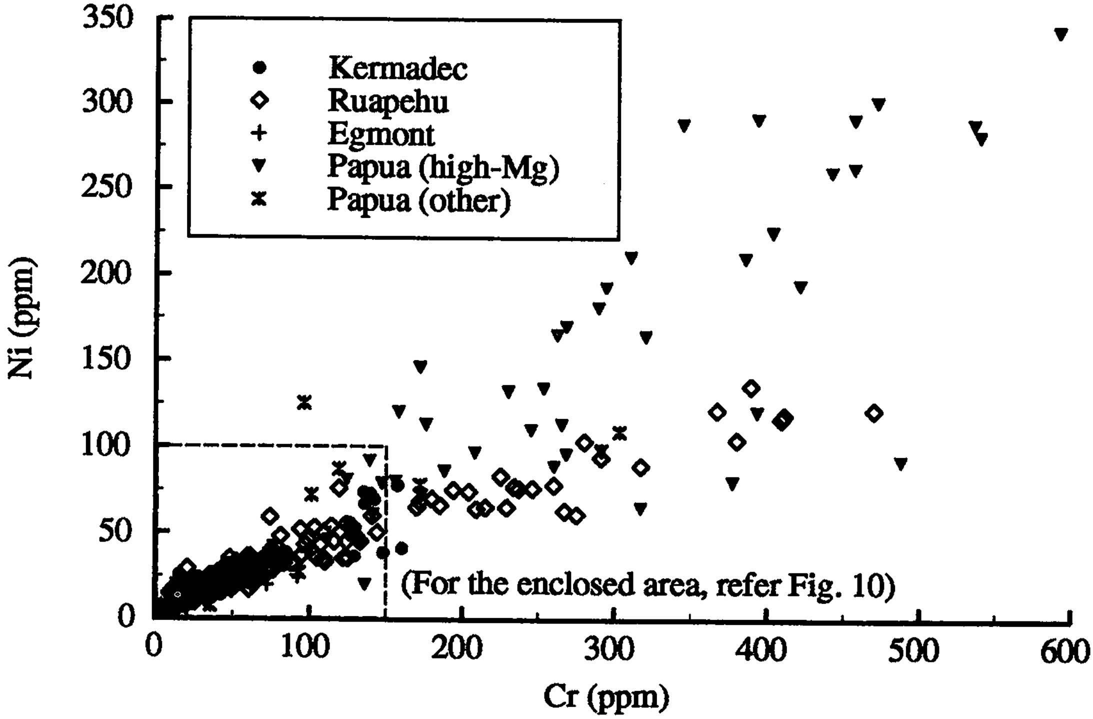

FIG. 9. Plot of Ni *versus* Cr concentrations in lavas from the four arc segments. Note the different slope of the Ruapehu and high-Mg Papuan arrays.

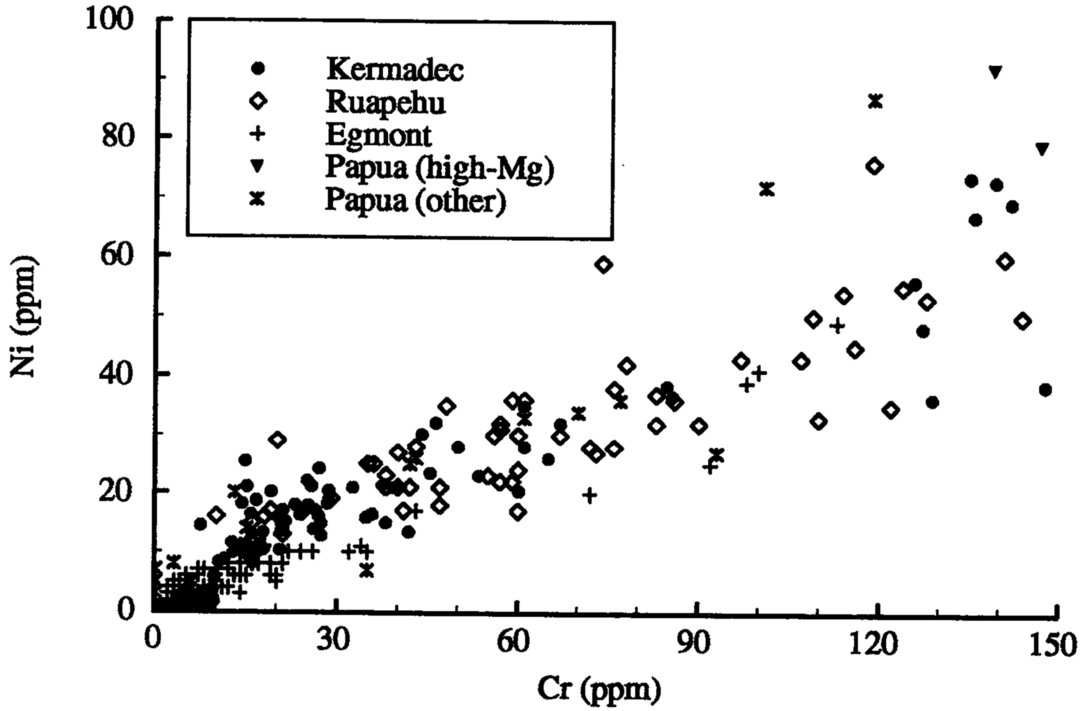

FIG. 10. Expanded view of Figure 9 for low values of Ni and Cr.

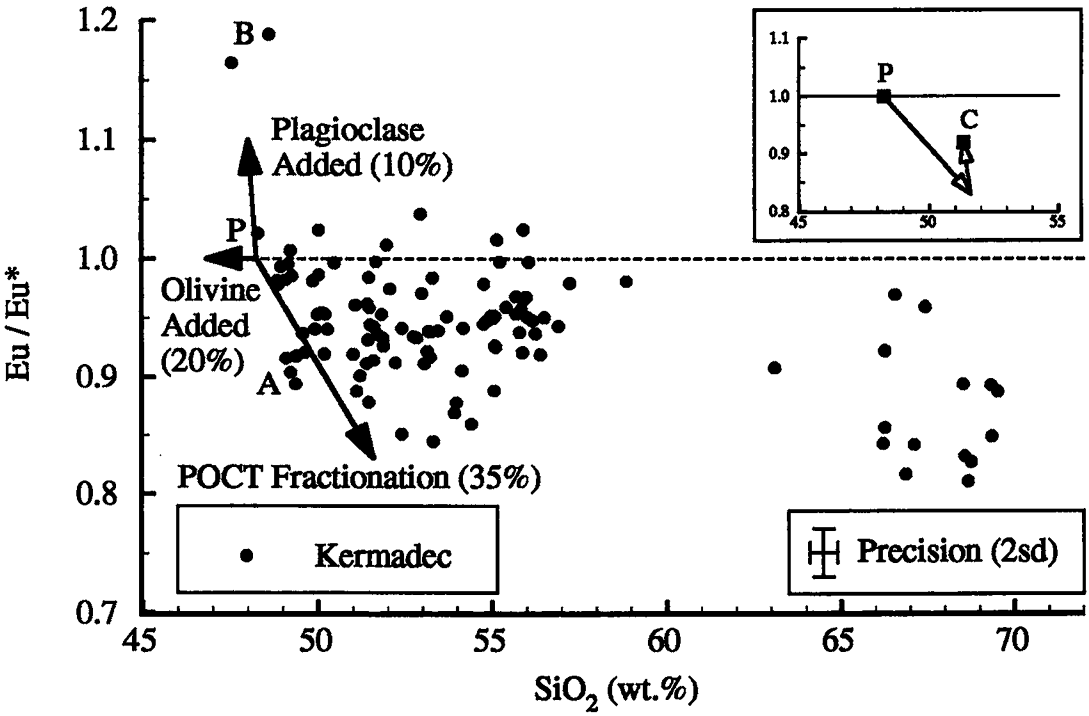

FIG. 11. Plot of Eu/Eu* *versus* SiO_2 content of lavas from Raoul Island, northern Kermadec arc. Fractionation vectors are calculated assuming that the plagioclase is An_{87}, the olivine is Fo_{85}, and the POCT assemblage consists 60% plagioclase + 5% olivine + 25% clinopyroxene + 10% titaniferous magnetite, where K_d for Eu^{2+} for plagioclase/melt = 0.8 [after compilation by Rollinson (1993)]. The inset depicts how lava C can be generated from parental magma P by 35% POCT fractionation followed by addition of 10% plagioclase; note that lava C still has a negative Eu anomaly. The precision bar is appropriate for a basaltic andesite with Eu/Eu* = 1.

trast, the Ruapehu suite contains a significant proportion of compositions with high Mg-numbers, and these are mainly in the andesite range. The Papuan lavas on this diagram are clearly separated into a "normal" group that overlaps the Kermadec, Egmont and low-Mg Ruapehu arrays, and the high-Mg group, in which high Mg-numbers persist through the range basalt to dacite.

Variations in Ni and Cr abundances (Figs. 9, 10) show the patterns and groupings established in Figures 6, 7 and 8. The high-Mg suites from Ruapehu and Papua typically contain 80 to 300 ppm Ni and 120 to 500 ppm Cr, though they form arrays with different slopes.

Simple application of the principles outlined earlier for the identification of primitive magmas in volcanic arcs leads to the conclusion that the Papuan arc contains compositions that are close to primary arc-type magma; in the northern Kermadec and New Zealand settings, such primitive magmas are either volumetrically subordinate to more fractionated lavas, or absent from the exposed sequences of lavas. Further, the existence of two distinct groupings in the Papuan arc suggests two lineages created by a different blend of FEAR processes.

THE IDENTIFICATION OF PARENTAL MAGMAS

It is more difficult to deduce the characteristics of primary magmas in the majority of arcs where high-Mg basalt has either not been sampled or not erupted. In such arcs, high-Al basalt predominates; considerable debate has raged over whether this lava-type represents a primary magma derived by partial melting of a quartz eclogite source (*e.g.*, Brophy & Marsh 1986), or the fractionated product of a primary high-Mg magma that has accumulated plagioclase (*e.g.*, Crawford *et al.* 1987). Where sparsely phyric lavas have been erupted, one technique is to add olivine and clinopyroxene in cotectic proportions until the magma has a composition that fulfills the criteria for a primary magma discussed above and is in equilibrium with its most magnesian olivine phenocrysts (*e.g.*, Eggins 1993). However, most arc lavas are strongly plagioclase-phyric, and thus the effects of plagioclase contamination must be removed before such calculations.

A valuable tool in deciphering the role of plagioclase is the magnitude of any Eu anomaly, defined as Eu/Eu*, where Eu is the chondrite-normalized Eu content of the lava and Eu* is the geometric mean of the chondrite-normalized Sm and Gd contents (Taylor & McLennan 1985). Under the redox conditions that prevail in most magmas and their source regions, Eu exists predominantly as Eu^{2+} and substitutes readily for Ca^{2+} in plagioclase (Sun *et al.* 1974, Drake & Weill 1975). Conversely, Eu^{3+} and the other trivalent *REE* behave incompatibly in the main silicate phases. Therefore, a primary magma generated by partial melting of mantle peridotite at a depth greater than 30 km should lack a Eu anomaly; progressive fractionation of plagioclase from this magma would produce an increasingly negative Eu anomaly, and addition of plagioclase would produce a positive Eu anomaly.

Northern Kermadec lavas form an array of increasingly negative and scattered Eu anomalies when plotted against indices of fractionation such as SiO_2 or Zr (Fig. 11). We have modeled the effects of removal and addition of plagioclase, olivine and clinopyroxene on the ratio Eu/Eu* for a Kermadec basalt; the more significant of these results are shown as vectors on Figure 11. Successful mass-balance calculations for the production of an andesite or a dacite by fractional crystallization of an arc basaltic magma typically require a bulk extracted assemblage composed of approximately 60% plagioclase, 25% clinopyroxene, 10% titaniferous magnetite and 5% olivine (*e.g.*, Woodhead 1988); a vector showing the effect of this fractionation of plagioclase – olivine – clinopyroxene – titaniferous magnetite (POCT) upon Eu/Eu* of a Kermadec basalt also is shown on Figure 11. An important result of our modeling is that the POCT fractionation vector is subparallel to a line drawn through the lowest Eu/Eu* of the Kermadec basalts, and can be superimposed on this line if the starting point of the vector is placed at 48 wt.% SiO_2 (point P on Fig. 11) and if the occasional presence of cumulus olivine is permitted (*i.e.*, in the lavas labeled A on Fig. 11).

We recognize that our depiction of POCT fractionation is simplistic; in reality, the composition of each mineral phase will vary as fractionation proceeds, as will the proportion of each phase in the extracted assemblage, and different batches of magma may follow different P–T–$f(O_2)$ paths. It is apparent that most Kermadec basalts and andesites plot between the POCT fractionation vector and the line Eu/Eu* = 1; the complications outlined above are probably responsible for part of this spread, because either a decrease in the proportion of plagioclase in the extract assemblage or a change to more oxidizing conditions will cause the POCT vector to be less steep than shown. However, the highly porphyritic plagioclase-rich nature of most Kermadec lavas, in which plagioclase characteristically occurs as phenocrysts of An_{85-90} composition, except for a strongly zoned thin (15 μm wide) rim, leads us to suspect that many of the plagioclase phenocrysts crystallized in different parts of the magmatic system and were incorporated into the magma only shortly before its eruption. Contamination of the magma by these plagioclase crystals would also increase the Eu/Eu* value of the magma, as indicated by the inset to Figure 11. An unequivocal example of plagioclase accumulation is provided by the two basalts (labeled B on Fig. 11) that have Eu/Eu* ≈ 1.17; these contain approximately 5% modal olivine, 10% clinopyroxene and 25% plagioclase as disaggregated clots of crystals. Their Eu anomaly and composition can be modeled as 5% fractionation of a POCT assemblage from a parental magma of composition P followed by the incorporation of the clots of crystals.

Previously, we have noted that no Kermadec lavas have the geochemical characteristics of primary magmas. Here, we emphasize that once plagioclase is removed from the magma, the Eu/Eu* value of the magma will be less than unity unless the magma is contaminated by a phase bearing significant amounts of Eu and with Eu/Eu* > 1, for which the only obvious candidate is plagioclase. The application of this simple principle has led us to infer that the parental Kermadec magma is represented by point P on Figure 11 and cannot be represented by points A or B, the former having undergone significant POCT fractionation, and the latter, plagioclase accumulation.

A similar trend is shown by Ruapehu lavas (Fig. 12). Again, the POCT vector can be superimposed upon the lowest Eu/Eu* of the Ruapehu lavas, and most lavas plot between the POCT vector and the line Eu/Eu* = 1. Following the principles outlined in our interpretation of the Kermadec lavas, we infer that the Ruapehu parental magma has a composition of 52 wt.% SiO_2. Our *REE* data set for Egmont lavas is too small for a valid comparison, but we note that most Egmont basalts contain a large negative Eu anomaly (Fig. 12). This suggests that the SiO_2 content of their parental magma is more similar to that of the Kermadec arc than to that of Ruapehu.

The correlation of Eu/Eu* with fractionation indices enables us to deduce the composition of parental magmas by removing the effects of fractionation and accumulation of plagioclase that are prevalent in most suites of arc-type lava. However, any comparison between parental magmas requires care, because plagioclase crystallization is sensitive to $P(H_2O)$ (Yoder & Tilley 1962); for example, two batches of the same primary magma may begin to fractionate plagioclase after different amounts of olivine – clinopyroxene – spinel fractionation if each batch encounters different $P(H_2O)$ conditions in the crust. The effect of this variable would be to produce two different compositions of parental magma. To us, the important point is that the parental magma calculated by our technique is that derived by the least amount of olivine – clinopyroxene – spinel fractionation from the primary magma. If we now desire to calculate the composition of the primary magma, we can add cotectic proportions of olivine –

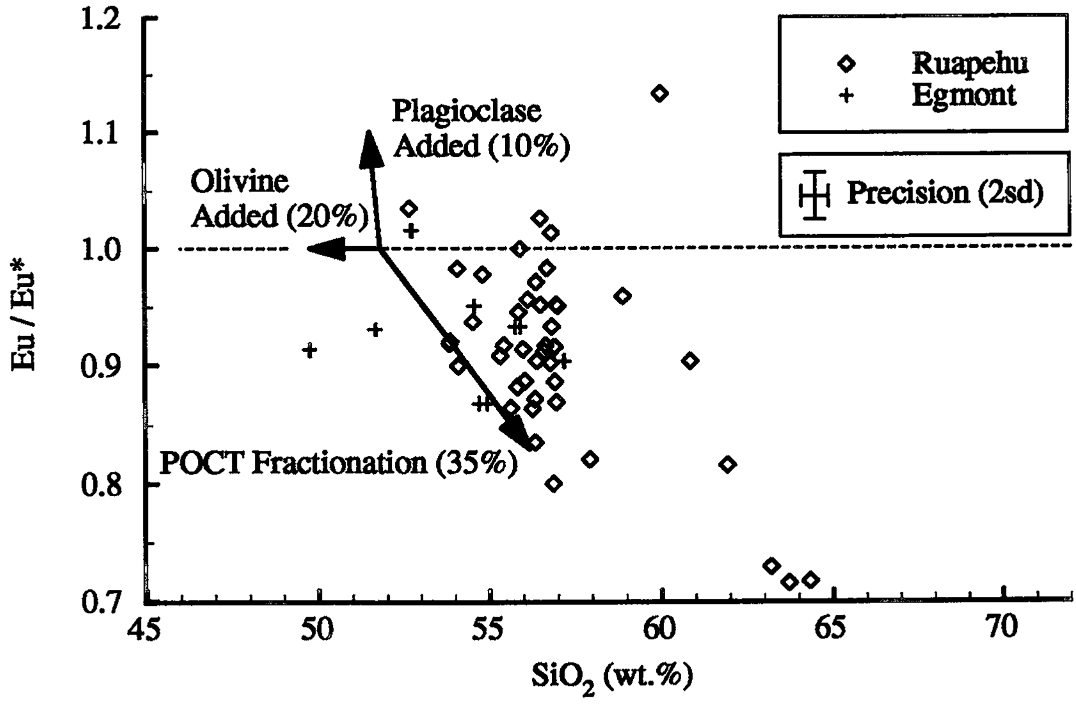

FIG. 12. Plot of Eu/Eu* *versus* SiO_2 content of lavas from Ruapehu and Egmont. Vectors as for Figure 11, except that the composition of plagioclase has been changed to An_{75} to reflect the mineral assemblage at Ruapehu (Graham & Hackett 1987).

clinopyroxene – spinel to the parental magma until it fulfills the criteria for a primary magma discussed in the preceding section (*e.g.*, Eggins 1993).

DISCUSSION

In the general model proposed for the genesis of arc magmas, primary magmas are produced by partial melting of the mantle-wedge peridotite in response to fluxing by fluids derived from the subducting slab of oceanic lithosphere. The chemical character of these primary magmas is influenced by the slab component, the wedge component and the melting process, all of which are potentially independent variables. Thereafter, the ascending primary magma may interact with the lithosphere in a variety of ways, and the extent of this interaction will depend upon the composition of the lithosphere, its thermal state, and the duration of the interaction. A consequence of such multi-stage, multi-component and multi-process models is that certain groups of elements can decouple from other elements during the series of processes that affect the magma. The compositional trends and contrasts shown by the southwest Pacific volcanic arcs provide a means of assessing the relative importance of some of these processes.

The most primitive magmas that we can recognize in our data set are represented by the comparatively abundant high-Mg basalts of the Papuan arc and the rarer high-Mg basalts of Ruapehu. In both instances, the high-Mg signature persists to more evolved compositions; for example, lavas with 61 wt.% SiO_2 still have Mg-numbers of 65 to 70 (Fig. 8). This leads us to infer that the high-Mg andesites were generated by fractionation of a low-Si phase that is less magnesian than olivine from the primary magma; this requirement, together with the higher $(La/Yb)_N$ values of the high-Mg lavas, causes us to suspect amphibole. We further suggest that the presence of a high-Mg and a high-Al lava series in the Papuan arc is due to two different paths of fractionation from the same primary magma; amphibole fractionation at relatively deep levels in the system generates the high-Mg series, whereas POCT fractionation at shallow levels produces the high-Al series.

The paucity of high-Mg basalt in volcanic arcs has been discussed above, together with the notion that this is due to the relatively buoyant crust acting as a barrier to the ascent of relatively dense high-Mg magma. The question then arises; what is so different about the Papuan arc that enables large volumes of primitive high-Mg basalt to erupt? We argue that high-Mg magma can breach the crust only in unusual tectonic

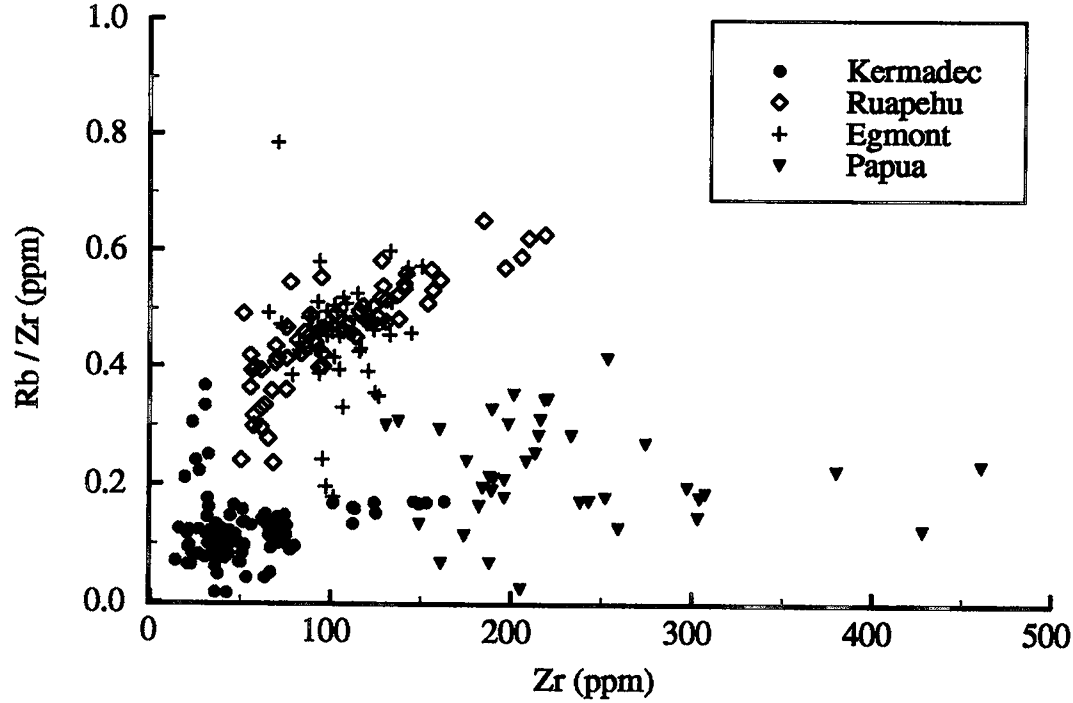

FIG. 13. Plot of Rb/Zr *versus* Zr concentration in lavas from the four arc segments. Note the trend of increasing Rb/Zr with Zr concentration in the Ruapehu suite.

circumstances, and draw attention to the impingement of sea-floor spreading in the Woodlark Basin upon the Papuan arc, generating an extensional tectonic regime unlike that of most volcanic arcs (Smith *et al.* 1977). A similar, but less extreme, situation occurs at Ruapehu, where the continental crust beneath the TVZ is anomalously thin compared to that on either side of the TVZ; indeed, geodetic measurements demonstrate that crustal extension is continuing there (Darby & Williams 1991).

Lavas from the relatively normal oceanic setting of the northern Kermadec arc are high-Al basalt and basaltic andesite, except for a few high-Mg lavas that are clearly contaminated by disaggregated clots of olivine – clinopyroxene – plagioclase crystals. Here, we argue that the high-Mg primary magma has been unable to penetrate the uppermost crust, and has instead undergone extensive olivine – clinopyroxene – spinel fractionation to produce a parental magma with 48 wt.% SiO_2; fractionation of these phases has removed most vestiges of the original high Mg, Ni and Cr signature. The composition of the parental magma is that at which plagioclase joins olivine and clinopyroxene as a liquidus phase; thereafter, POCT fractionation of this magma generates an increasingly negative Eu anomaly. As discussed in the previous section, the negative Eu anomaly is generally smaller than might be expected from simple mass-balance calculations that produce andesite from basalt; both the porphyritic plagioclase-rich nature of the lavas and the zoning patterns of the plagioclase lead us to infer that late-stage contamination of the magma by plagioclase crystals contributes more to this effect than variations in the proportion of plagioclase in the POCT assemblage.

Most Ruapehu lavas also are high-Al lavas. We interpret their origin in similar fashion to that of the northern Kermadec lavas, but with two important differences. Firstly, the parental magma at Ruapehu has a composition of approximately 52 wt.% SiO_2 as opposed to 48 wt.% SiO_2 for the Kermadec arc. We attribute this to plagioclase being later to arrive on the liquidus at Ruapehu, because the primary magma is trapped near the base of the 15-km-thick continental crust, as opposed to within the oceanic crust; that is, the primary magma fractionates under conditions that are more hydrous and at greater pressure, both factors tending to reduce the plagioclase field while enhancing the amphibole field (Yoder & Tilley 1962, Foden & Green 1992). Secondly, some high-Mg lavas have been erupted by Ruapehu, and these we relate to fractionation of the primary magma within the stability field of amphibole, in the same manner as we propose for the Papuan high-Mg suite.

Other noteworthy features of the Ruapehu lavas are their steeper arrays on plots of incompatible elements against indices of fractionation, compared to those of the northern Kermadec lavas, and their progressive increase in *LILE/HFSE* values with fractionation. We have investigated this decoupling among the incompatible elements by plotting Rb/Zr *versus* Zr (Fig. 13). Closed-system POCT fractionation of any magma should generate a nearly horizontal line on this plot, yet Ruapehu is alone among our volcanic suites in showing a marked increase in Rb/Zr with fractionation. We attribute this trend to progressively greater degrees of assimilation of continental crust with high *LILE/HFSE* values; this argument is further supported by the higher $^{87}Sr/^{86}Sr$ values of the more evolved Ruapehu lavas (Graham & Hackett 1987). The inferred assimilation of significant quantities of continental crust by Ruapehu magmas is probably facilitated by unusually high temperatures throughout the crust, as the heat flow of the TVZ has been calculated to exceed that of other comparable volcanic arcs by a factor of four (Hochstein 1995).

No Egmont lavas have primitive characteristics. We consider that their evolution is comparable to that of the northern Kermadec lavas, with considerable fractionation of a primary magma and parental magma occurring at the base of, or within, the crust. Some intriguing clues to the production of arc lavas in this backarc environment are revealed by Figure 13. On this plot, older Egmont lavas have both high and low Rb/Zr values, whereas recent Egmont lavas consistently have Rb/Zr $\approx$ 0.5. The variation in Rb/Zr is unlikely to be an artefact of crustal assimilation, as there is no correlation of the Rb/Zr value with fractionation indices, nor is there a trend to more extreme Rb/Zr values with time. Instead, we suspect that magma generation in a backarc environment is more complex than that beneath frontal arc volcanoes, and that we are seeing the effects of mantle metasomatic processes in the older Egmont lavas.

In conclusion, we infer that the occurrence of rocks representing primitive magmas in arc-type associations depends on the processes that operate on the magma in transit to the Earth's surface. These seem primarily linked to tectonic setting. Primitive geochemical characteristics in arc-type magmas are determined by melting processes in the mantle-wedge source, and such primary magmas have high Mg-numbers and high Ni and Cr contents. However, these magmas are only erupted in unusual, strongly extensional tectonic settings. More commonly, ascending dense high-Mg magmas are trapped beneath the relatively low-density crust, where they fractionate and evolve until they become sufficiently buoyant to break through. The depth of these trapped primitive magmas is critical to the nature of their fractionating assemblage. Deep fractionation of hydrous magma is likely to occur within the stability field of amphibole and, if erupted, the evolved magmas may preserve evidence of their ancestry by forming a fractionated but still high-Mg suite (*e.g.*, the high-Mg Papuan lavas). More commonly, shallow fractionation of an olivine – clinopyroxene – spinel assemblage produces parental magmas that contain few traces of their mantle origin, and that may further fractionate POCT assemblages, assimilate wallrock, or be contaminated with crystals of plagioclase before erupting to form high-Al convergent-margin lavas typical of arcs.

References

Aoki, K. & Fujimaki, H. (1982): Petrology and geochemistry of calc-alkaline andesite of presumed upper mantle origin from Itinome-gata, Japan. *Am. Mineral.* **67**, 1-13.

Barsdell, M. & Berry, R. (1990): Origin and evolution of primitive island arc ankaramites from western Epi, Vanuatu. *J. Petrol.* **31**, 747-777.

Bloomer, S.H., Ewart, A., Hergt, J.M. & Bryan, W.B. (1994): Geochemistry and origin of igneous rocks from the outer Tonga forearc (site 841). *In* Proc. of the Ocean Drilling Program, Scientific Results **135** (J.W. Hawkins, L.M. Parson, J.F. Allan, *et al.*, eds.). Ocean Drilling Program, College Station, Texas (625-646).

________ & Fisher, R.L. (1987): Petrology and geochemistry of igneous rocks from the Tonga Trench – a non-accreting plate boundary. *J. Geol.* **95**, 469-495.

Brophy, J.G. & Marsh, B.D. (1986): On the origin of high-alumina arc basalt and the mechanics of melt extraction. *J. Petrol.* **27**, 763-789.

Clift, P.D. (1995): Volcaniclastic sedimentation and volcanism during the rifting of western Pacific backarc basins. *In* Active Margins and Marginal Basins of the Western Pacific (B. Taylor & J.P. Natland, eds.). *Am. Geophys. Union, Geophys. Monogr. Ser.* **88**, 67-96.

Crawford, A.J., Falloon, T.J. & Eggins, S. (1987): The origin of island arc high-alumina basalts. *Contrib. Mineral. Petrol.* **97**, 417-430.

Darby, D. & Williams, R.O. (1991): A new geodetic estimate of deformation in the Central Volcanic Region of the North Island, New Zealand. *N.Z. J. Geol. Geophys.* **34**, 127-136.

Davies, H.L., Symonds, P.A. & Ripper, I.A. (1984): Structure and evolution of the southern Solomon Sea region. *Bureau Mineral. Res. J. Aust. Geol. Geophys.* **9**, 49-68.

Defant, M.J. & Neilsen, R.L. (1990): Interpretation of open system petrogenetic processes: phase equilibria constraints on magma evolution. *Geochim. Cosmochim. Acta* **54**, 87-102.

Drake, M.J. & Weill, D.F. (1975): Partition of Sr, Ba, Ca, Y, Eu^{2+}, Eu^{3+}, and other *REE* between plagioclase feldspar and magmatic liquid: an experimental study. *Geochim. Cosmochim. Acta* **39**, 689-712.

Eggins, S.M. (1993): Origin and differentiation of picritic arc magmas, Ambae (Aoba), Vanuatu. *Contrib. Mineral. Petrol.* **114**, 79-100.

FODEN, J.D. & GREEN, D.H. (1992): Possible role of amphibole in the origin of andesite: some experimental and natural evidence. *Contrib. Mineral. Petrol.* **109**, 479-493.

GAMBLE, J.A., SMITH, I.E.M, MCCULLOCH, M.T., GRAHAM, I.J. & KOKELAAR, B.P. (1993): The geochemistry and petrogenesis of basalts from the Taupo Volcanic Zone and Kermadec island arc, S.W. Pacific. *J. Volcanol. Geotherm. Res.* **54**, 265-290.

GILL, J.B. (1981): *Orogenic Andesites and Plate Tectonics.* Springer-Verlag, New York, N.Y.

GRAHAM, I.J. & HACKETT, W.R. (1987): Petrology of calc-alkaline lavas from Ruapehu Volcano and related vents, Taupo Volcanic Zone, New Zealand. *J. Petrol.* **28**, 531-567.

GUST, D.A. & PERFIT, M.R. (1987): Phase relations of a high-Mg basalt from the Aleutian island arc: implications for primary island arc basalts and high-Al basalts. *Contrib. Mineral. Petrol.* **97**, 7-18.

HOCHSTEIN, M.P. (1995): Crustal heat transfer in the Taupo Volcanic Zone (New Zealand): comparison with other volcanic arcs and explanatory heat source models. *J. Volcanol. Geotherm. Res.* **68**, 117-151.

HONZA, E. (1991): The Tertiary arc chain in the western Pacific. *Tectonophys.* **187**, 285-303.

JOHNSON, R.W., MACKENZIE, D.E. & SMITH, I.E.M. (1978): Delayed partial melting of subduction-modified mantle in Papua New Guinea. *Tectonophys.* **46**, 197-216.

KAY, R.W. (1978): Aleutian magnesian andesites: melts from subducted Pacific Ocean crust. *J. Volcanol. Geotherm. Res.* **4**, 117-132.

LLOYD, E.F. & NATHAN, S. (1981): Geology and tephrochronology of Raoul Island, Kermadec Group, New Zealand. *N.Z. Geol. Surv., Bull.* **95**.

MEEN, J.K. & EGGLER, D.H. (1987): Petrology and geochemistry of the Cretaceous Independence volcanic suite, Absaroka Mountains, Montana; clues to the composition of the Archean sub-Montana mantle. *Geol. Soc. Am., Bull.* **98**, 238-247.

NAKAMURA, N. (1974): Determination of *REE*, Ba, Fe, Mg, Na and K in carbonaceous and ordinary chondrites. *Geochim. Cosmochim. Acta* **38**, 757-775.

NATHAN, S., LLOYD, E.F., SMITH, I.E.M. & STEWART, R.B. (1996): Recent lavas from Macauley Island, Kermadec Group, New Zealand. *N.Z. J. Geol. Geophys.* **39**, 295-308.

NYE, C.J. & REID, M.R. (1986): Geochemistry of primary and least fractionated lavas from Okmok Volcano, central Aleutians: implications for arc magma genesis. *J. Geophys. Res.* **91**, 10271-10287.

PEARCE, J.A. & PARKINSON, I.J. (1993): Trace element models for mantle melting: application to volcanic arc petrogenesis. *In* Magmatic Processes and Plate Tectonics (H.M. Prichard, T. Alabaster, N.B.W. Harris & C.R. Neary, eds.). *Geol. Soc., Spec. Publ.* **76**, 373-403.

PELLETIER, B. & LOUAT, R. (1989): Seismotectonics and present-day relative plate motions in the Tonga–Lau and Kermadec–Havre region. *Tectonophys.* **165**, 237-250.

PLANK, T. & LANGMUIR, C.H. (1988): An evaluation of the global variations in the major element chemistry of arc basalts. *Earth Planet. Sci. Lett.* **90**, 349-370.

PRICE, R.C., MCCULLOCH, M.T., SMITH, I.E.M. & STEWART, R.B. (1992): Pb–Nd–Sr isotopic compositions and trace element characteristics of young volcanic rocks from Egmont Volcano and comparisons with basalts and andesites from the Taupo Volcanic Zone, New Zealand. *Geochim. Cosmochim. Acta* **56**, 941-953.

RAMSAY, W.R.H., CRAWFORD, A.J. & FODEN, J.D. (1984): Field setting, mineralogy, chemistry, and genesis of arc picrites, New Georgia, Solomon Islands. *Contrib. Mineral. Petrol.* **88**, 386-402.

ROLLINSON, H.R. (1993): *Using Geochemical Data: Evaluation, Presentation, Interpretation.* Longman Group, Singapore.

SAUNDERS, A.D., ROGERS, G., MARRINER, G.F., TERRELL, D.J. & VERMA, S.P. (1987): Geochemistry of Cenozoic volcanic rocks, Baja California, Mexico: implications for the petrogenesis of post-subduction magmas. *J. Volcanol. Geotherm. Res.* **32**, 223-245.

SMITH, A.L. & CARMICHAEL, I.S.E. (1968): Quaternary lavas from the southern Cascades, western U.S.A. *Contrib. Mineral. Petrol.* **19**, 212-238.

SMITH, I.E.M. (1982a): Volcanic evolution in eastern Papua. *Tectonophys.* **87**, 315-333.

________ (1982b): Young volcanoes in eastern Papua. *In* Cooke–Ravian Volume of Volcanological Papers (R.W. Johnson, ed.). *Geol. Surv. Papua New Guinea, Mem.* **10**, 257-265.

________, CHAPPELL, B.W., WARD, G.K. & FREEMAN, R.S. (1977): Peralkaline rhyolites associated with andesitic arcs of the southwest Pacific. *Earth Planet. Sci. Lett.* **37**, 230-236.

________ & JOHNSON, R.W. (1981): Contrasting rhyolite suites in the late Cenozoic of Papua New Guinea. *J. Geophys. Res.* **86**, 10257-10272.

________ & MILSOM, J.S. (1984): Late Cenozoic volcanism and extension in eastern Papua. *In* Marginal Basin Geology: Volcanic and Associated Sedimentary and Tectonic Processes in Modern and Ancient Marginal Basins (B.P. Kokelaar & M.F. Howells, eds.). *Geol. Soc., Spec. Publ.* **16**, 163-171.

________ & MITCHELL, P.A. (1989): High-magnesium lavas in the late Cenozoic volcanic arc associations of southeastern

Papua New Guinea. *Neues Jahrb. Mineral., Monatsh.*, 481-497.

STERN, T.A. (1985): A back-arc basin formed within continental lithosphere: the Central Volcanic Region of New Zealand. *Tectonophys.* **112**, 385-409.

STOLPER, E. & WALKER, D. (1980): Melt density and the average composition of basalt. *Contrib. Mineral. Petrol.* **74**, 7-12.

SUN, CHING-OH, WILLIAMS, R.J. & SUN, SHINE-SOON (1974): Distribution coefficients of Eu and Sr for plagioclase–liquid and clinopyroxene–liquid equilibria in oceanic ridge basalt: an experimental study. *Geochim. Cosmochim. Acta* **38**, 1415-1433.

TATSUMI, Y. & EGGINS, S. (1995): *Subduction Zone Magmatism.* Blackwell Science Inc., Cambridge, Massachusetts.

________ & ISHIZAKA, K. (1981): Existence of andesitic primary magma: an example from southwest Japan. *Earth Planet. Sci. Lett.* **53**, 124-130.

________ & ________ (1982a): Magnesian andesite and basalt from Shodo-shima Island, southwest Japan, and their bearing on the genesis of calc-alkaline andesites. *Lithos* **15**, 161-172.

________ & ________ (1982b): Origin of high-magnesian andesites in the Seteouchi volcanic belt, southwest Japan. I. Petrographical and chemical characteristics. *Earth Planet. Sci. Lett.* **60**, 293-304.

TAYLOR, S.R. & MCLENNAN, S.M. (1985): *The Continental Crust: its Composition and Evolution.* Blackwell, Oxford, U.K.

WALCOTT, R.I. (1978): Present tectonics and late Cenozoic evolution of New Zealand. *Geophys. J.R. Astr. Soc.* **52**, 137-164.

WOODHEAD, J.D. (1988): The origin of geochemical variations in Mariana lavas: a general model for petrogenesis in intra-oceanic island arcs? *J. Petrol.* **29**, 805-830.

YODER, H.S., JR. & TILLEY, C.E. (1962): Origin of basalt magmas: an experimental study of natural and synthetic rock systems. *J. Petrol.* **3**, 342-532.

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