# Andesites in Island Arcs and Continental Margins: Relationship to Crustal Evolution

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## ABSTRACT

Andesites of both island arc and continental margin environments contain petrologic evidence of mixing of mantle and crustal melts. Andesitic volcanism appears to involve addition of mantle-derived basaltic magma to the crust and fractionation of preexisting crustal material. Changes in andesitic volcanism with increasingly continental character of the crust reflect changes in a rhyolitic component derived from increasingly aged and fractionated crust. The initial stage in development of continental crust is partial melting of oceanic crust.

## INTRODUCTION

A number of recent papers have called attention to the observation that andesites commonly contain phases derived both from basaltic and silicic magmas and have suggested that mixing of basaltic and silicic magma plays an important role in petrogenesis and volcanism (EICHELBERGER, 1975; ANDERSON, 1976; SPARKS *et al.*, 1977; JOHNSTON, 1977; EICHELBERGER and GOOLEY, 1977). EICHELBERGER *et al.* (1976) argued that such a process, combined with crustal melting, can be the basis for a general model for intermediate and silicic igneous activity which is consistent with other geochemical and geophysical data. The purpose of this paper is to discuss the application of this model to island arc volcanism and the problem of crustal evolution.

## SUMMARY OF MIXING MODEL

The primary basis for proposing that andesites and dacites are hybrids is the presence within them of disequilibrium phenocryst assemblages. In addition to phenocrysts appropriate to magma systems of bulk intermediate composition, andesites and dacites commonly contain phenocrysts of a type and composition appropriate to basaltic and silicic magmas. Forsteritic olivine, highly calcic plagioclase, and sometimes magnesium- and aluminum-rich augite which occur as equilibrium assemblages in basalt also occur in andesite and dacite as individual grains and within cognate xenoliths of basaltic bulk composition. Inclusions of basaltic glass have been reported in olivine and calcic plagioclase in andesites (ANDERSON, 1976), indicating that these phases grew from a basaltic liquid. Moderately sodic plagioclase and sometimes quartz, sanidine, and biotite (roughly in order of frequency of occurrence) which occur as equilibrium assemblages in rhyolite also occur as resorbed or reacted phenocrysts in andesite and dacite. Sodic plagioclase is invariably surrounded by a dark resorbed zone consisting of melt channels in more calcic plagioclase, which is in turn surrounded by normally zoned calcic overgrowth. Quartz contains either embayments indicating resorption or is surrounded by an augite reaction rim. Sanidine is resorbed. Biotite usually reacts to form hornblende, plagioclase, and titanomagnetite. Inclusions of rhyolitic glass can often be found within the unresorbed or unreacted cores of these phases, indicating that these phases grew from a rhyolitic liquid. The cognate xenoliths consist of euhedral plagioclase and hornblende or pyroxene with interstitial glass. Plagioclase and hornblende are often highly elongate and hollow suggesting rapid growth. In composition the hornblende resembles hornblende coexisting with melt in basaltic systems from 1000°C to 700°C (HELZ, 1973). In some cases grain size decreases markedly toward the edge of the xenolith. Some large xenoliths are concentrically zoned with basaltic cores surrounded by andesitic rinds which contain resorbed plagioclase and quartz. The texture and bulk and phase compositions indicate that the xenoliths are essentially pillows formed by rapid crystallization of basaltic magma in cooler silicic magma. Mixing of basaltic and silicic magmas results in rapid crystallization of basaltic magma and resorption or reaction of any crystals present in the silicic magma. Crystals which originally grew in basaltic or silicic systems survive the mixing event if they become coated with overgrowths or reaction products. The system is homogenized by rapid convection, driven by the large temperature differences accompanying the initial heterogeneity in composition. This process accounts for the elemental and isotopic composition of andesites and dacites, since the composition of these rocks can be closely approximated as a linear combination of basaltic and rhyolitic compositions. The origins of the parental basaltic and rhyolitic liquids are attributed to partial melting in the upper mantle and lower crust, respectively, since these regions contain appropriate materials and conditions for magma genesis. The association of basaltic and rhyolitic liquids in time and space is attributed to heating and consequent melting of crust by basaltic volcanism. The rise of basaltic magma from the mantle and emplacement in the lower crust causes crustal melting and development of rhyolitic magma bodies. Once formed, these rhyolitic magma bodies trap subsequent intrusions of denser basaltic magma and mix by convection, forming andesitic and dacitic magmas, depending on the proportions. Thus the basalt-andesite-dacite-rhyolite association and its plutonic equivalent, the gabbro-diorite-granodiorite-granite association involve both crustal and mantle melts. Comparison of silicic volcanism in different crustal and tectonic settings indicates that continental crust favors generation of rhyolitic magma and compressional tectonic activity favors mixing.

## ISLAND ARCS AND ANDESITES

The debate over mantle versus crustal origin of andesitic magma is often viewed in terms of mantle melting and fractionation versus contamination. Andesitic magma might come from the mantle directly by melting or by fractionation of mantle-derived basaltic magma or by melting in a subduction zone of mantle-derived oceanic crust. The alternative is often viewed as assimilation of granitic crustal rocks by basaltic magma. The recognition that island arcs erupt large volumes of andesite where continental crust is absent or poorly developed (GORSHKOV, 1969), as well as the relatively low $^{87}$Sr/$^{86}$Sr of most andesites (DICKINSON, 1970), have ruled out a major role for assimilation of granitic rocks in andesite genesis. Continental crust is not required for andesite genesis, but this does not necessarily rule out crustal involvement altogether. Island arcs have thus become a focus for the debate over andesites, both because they place important constraints on how andesites can form and because their voluminous andesites and related intrusives appear to represent an early stage in development of continental crust.

## EVIDENCE OF MIXING IN SOME ISLAND ARC LAVAS

The Lesser Antilles are an active island arc in the Caribbean, apparently repre-

TABLE 1 - Electron microprobe analyses of phases in Pitons du Carbet and La Soufrière andesite.

*Plagioclase*

| | *Resorbed* | | | | | | *Normal Xeno* | | | | | | *Microphenocrysts* | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | *Core* | | *In Rim* | | *Out Rim* | | *Calcic* | | *Core* | | *Rim* | | *Core* | | *Rim* | |
| $SiO_2$ | 55.4 | | 46.7 | | 54.2 | | 45.2 | | 45.7 | | 55.0 | | 48.2 | | 53.1 | |
| $TiO_2$ | — | | — | | — | | — | | — | | — | | — | | — | |
| $Al_2O_3$ | 28.1 | | 33.2 | | 29.2 | | 35.9 | | 34.0 | | 28.3 | | 32.6 | | 28.7 | |
| $Cr_2O_3$ | — | | — | | — | | — | | — | | — | | — | | — | |
| FeO | 0.2 | | 0.8 | | 0.5 | | 0.6 | | 0.6 | | 0.4 | | 0.9 | | 0.9 | |
| MnO | — | | — | | — | | — | | — | | — | | — | | — | |
| MgO | — | | — | | — | | — | | — | | — | | — | | — | |
| CaO | 10.9 | | 17.2 | | 11.6 | | 18.5 | | 17.6 | | 10.9 | | 16.0 | | 11.4 | |
| $NaO_2$ | 5.4 | | 1.6 | | 5.1 | | 0.9 | | 1.1 | | 5.1 | | 2.2 | | 4.9 | |
| $K_2O$ | 0.2 | | 0.1 | | 0.4 | | 0.1 | | 0.1 | | 0.3 | | 0.2 | | 0.4 | |
| | 100.2 | | 99.6 | | 101.0 | | 101.1 | | 99.1 | | 100.0 | | 100.1 | | 99.4 | |
| Cations Per N Oxygens | 8 | | | | | | | | | | | | | | | |
| Si | 2.49 | 3.98 | 2.16 | 3.97 | 2.43 | 3.98 | 2.07 | 4.00 | 2.13 | 3.99 | 2.48 | 3.98 | 2.21 | 3.98 | 2.43 | 3.98 |
| Al IV | 1.49 | | 1.81 | | 1.55 | | 1.93 | | 1.87 | | 1.51 | | 1.77 | | 1.55 | |
| Al VI | — | | — | | — | | — | | — | | — | | — | | — | |
| Ti | — | | — | | — | | — | | — | | — | | — | | — | |
| Cr | — | | — | | — | | — | | — | | — | | — | | — | |
| Fe | 0.01 | 1.02 | 0.03 | 1.03 | 0.02 | 1.05 | 0.02 | 1.01 | 0.02 | 1.01 | 0.02 | 1.01 | 0.04 | 1.03 | 0.03 | 1.05 |
| Mn | — | | — | | — | | — | | — | | — | | — | | — | |
| Mg | — | | — | | — | | — | | — | | — | | — | | — | |
| Ca | 0.53 | | 0.85 | | 0.56 | | 0.90 | | 0.88 | | 0.53 | | 0.79 | | 0.56 | |
| Na | 0.47 | | 0.14 | | 0.45 | | 0.08 | | 0.10 | | 0.44 | | 0.19 | | 0.44 | |
| K | 0.01 | | 0.01 | | 0.02 | | 0.01 | | 0.01 | | 0.02 | | 0.01 | | 0.02 | |
| | 5.00 | | 5.00 | | 5.03 | | 5.01 | | 5.01 | | 5.00 | | 5.01 | | 5.03 | |
| An | 52.2 | | 85.1 | | 54.4 | | 91.9 | | 89.4 | | 53.3 | | 79.3 | | 54.9 | |
| Ab | 46.4 | | 14.3 | | 43.6 | | 7.7 | | 10.1 | | 44.8 | | 19.6 | | 42.9 | |

| | | | | | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| $TiO_2$ | 0.1 | | 0.1 | | — | | — | | — | | — | | — | | — | |
| $Al_2O_3$ | 28.9 | | 33.8 | | 29.2 | | 35.2 | | 32.9 | | 28.3 | | 33.4 | | 28.4 | |
| $Cr_2O_3$ | 0.0 | | — | | — | | — | | — | | — | | — | | — | |
| FeO | 0.4 | | 0.6 | | 0.8 | | 0.5 | | 0.4 | | 0.4 | | 0.8 | | 1.1 | |
| MnO | 0.0 | | — | | — | | — | | — | | — | | — | | — | |
| MgO | 0.1 | | 0.1 | | — | | — | | — | | — | | — | | — | |
| CaO | 11.4 | | 16.6 | | 12.1 | | 19.3 | | 16.9 | | 10.9 | | 16.1 | | 12.4 | |
| $NaO_2$ | 5.0 | | 1.5 | | 4.8 | | 0.7 | | 2.0 | | 5.1 | | 1.7 | | 4.1 | |
| $K_2O$ | 0.2 | | 0.1 | | 0.4 | | 0.0 | | 0.1 | | 0.3 | | 0.1 | | 0.2 | |
| | 99.7 | | 98.8 | | 100.0 | | 100.2 | | 99.9 | | 100.0 | | 98.5 | | 98.9 | |
| Cations Per N Oxygens | 8 | | | | | | | | | | | | | | | |
| Si | 2.43 | 3.99 | 2.14 | 3.99 | 2.40 | 3.97 | 2.06 | 3.98 | 2.19 | 3.98 | 2.48 | 3.98 | 2.18 | | 2.42 | |
| Al IV | 1.55 | | 1.85 | | 1.57 | | 1.92 | | 1.79 | | 1.50 | | 1.81 | | 1.54 | |
| Al VI | — | | — | | — | | — | | — | | — | | — | | — | |
| Ti | 0.00 | | — | | — | | — | | — | | — | | — | | — | |
| Cr | 0.00 | | — | | — | | — | | — | | — | | — | | — | |
| Fe | 0.01 | 1.02 | 0.02 | 1.00 | 0.03 | 1.06 | 0.02 | 1.04 | 0.02 | 1.04 | 0.02 | 1.01 | 0.03 | 1.00 | 0.04 | |
| Mn | 0.00 | | — | | — | | — | | — | | — | | — | | — | |
| Mg | 0.00 | | 0.01 | | — | | — | | — | | — | | — | | — | |
| Ca | 0.56 | | 0.83 | | 0.59 | | 0.96 | | 0.83 | | 0.53 | | 0.80 | | 0.61 | |
| Na | 0.44 | | 0.14 | | 0.42 | | 0.06 | | 0.18 | | 0.44 | | 0.16 | | 0.37 | |
| K | 0.01 | | 0.00 | | 0.02 | | 0.00 | | 0.01 | | 0.02 | | 0.01 | | 0.04 | |
| | 5.00 | | 4.99 | | 5.03 | | 5.02 | | 5.02 | | 4.99 | | 4.99 | | 5.02 | |
| An | 55.2 | | 85.2 | | 57.1 | | 93.5 | | 82.0 | | 53.3 | | 83.2 | | 62.0 | |
| Ab | 43.8 | | 14.3 | | 40.8 | | 6.3 | | 17.4 | | 44.8 | | 16.2 | | 36.9 | |
| Or | 1.0 | | 0.5 | | 2.1 | | 0.2 | | 0.6 | | 1.9 | | 0.6 | | 1.1 | |

*Segue:* TABLE 1 - Electron microprobe analyses of phases in Pitons du Carbet and La Soufrière andesite.

| | *Hornblende* | | | | *Orthopyroxene* | | *Clinopyroxene* | | | | | | | | *Olivine* | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | *Xeno* | | *Phen* | | *Xeno* | *Phen* | *Aluminous* | | | | *Normtl Xeno* | *Phen* | *Rim On Quartz* | | | |
| | | | | | | | *Core* | | *Rim* | | | | | | | |
| SiO$_2$ | 42.7 | | 42.4 | | | | 45.6 | | 52.4 | | | | 52.6 | | 38.9 | |
| TiO$_2$ | 1.6 | | 1.4 | | | | 1.1 | | 0.2 | | | | 0.2 | | — | |
| Al$_2$O$_3$ | 13.5 | | 13.3 | | | | 10.2 | | 1.5 | | | | 1.0 | | — | |
| Cr$_2$O$_3$ | 0.0 | | 0.0 | | | | 0.1 | | 0.0 | | | | 0.0 | | — | |
| FeO | 11.7 | | 11.3 | | | | 7.7 | | 10.6 | | | | 9.9 | | 15.6 | |
| MnO | 0.2 | | 0.2 | | | | 0.1 | | 0.6 | | | | 0.3 | | — | |
| MgO | 14.8 | | 15.1 | | | | 12.1 | | 15.5 | | | | 14.4 | | 45.3 | |
| CaO | 11.0 | | 11.2 | | | | 22.2 | | 18.7 | | | | 20.8 | | 0.0 | |
| NaO$_2$ | 2.9 | | 2.8 | | | | 0.2 | | 0.2 | | | | 0.2 | | — | |
| K$_2$O | 0.4 | | 0.5 | | | | 0.1 | | 0.0 | | | | 0.1 | | — | |
| | 98.9 | | 98.2 | | | | 99.4 | | 99.7 | | | | 99.5 | | 99.8 | |
| Cations Per N Oxygens | 23 | | | | 6 | | | | | | | | | | 4 | |
| Si | 6.19 | 8.00 | 6.19 | 8.00 | | | 1.71 | 2.00 | 1.96 | 2.00 | | | 1.97 | 2.00 | 0.98 | |
| Al IV | 1.81 | | 1.81 | | | | 0.29 | | 0.04 | | | | 0.03 | | — | |
| Al VI | 0.50 | 5.33 | 0.48 | 5.33 | | | 0.16 | 2.04 | 0.03 | 20.4 | | | 0.02 | 2.01 | — | |
| Ti | 0.18 | | 0.15 | | | | 0.03 | | 0.01 | | | | 0.01 | | — | |
| Cr | 0.00 | | 0.00 | | | | 0.00 | | 0.00 | | | | 0.00 | | — | |
| Fe | 1.42 | | 1.38 | | | | 0.24 | | 0.33 | | | | 0.31 | | 0.33 | 2.03 |
| Mn | 0.03 | | 0.03 | | | | 0.01 | | 0.02 | | | | 0.01 | | 1.70 | |
| Mg | 3.20 | | 3.29 | | | | 0.68 | | 0.86 | | | | 0.81 | | — | |
| Ca | 1.17 | 2.60 | 1.76 | 2.63 | | | 0.90 | | 0.75 | | | | 0.84 | | 0.00 | |
| Na | 0.81 | | 0.78 | | | | 0.02 | | 0.01 | | | | 0.01 | | — | |
| K | 0.08 | | 0.09 | | | | 0.00 | | 0.00 | | | | 0.00 | | — | |
| | 15.93 | | 15.96 | | | | 4.04 | | 4.01 | | | | 4.01 | | 3.01 | |
| An | | | | | | Wo | 49.2 | | 38.6 | | | | 42.8 | | | |
| Ab | | | | | | En | 37.4 | | 44.4 | | | | 41.2 | Fo | 83.9 | |

| | | | | | | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| $TiO_2$ | | 0.2 | | 0.2 | | 1.4 | | 0.5 | | 0.3 | | 0.4 | | 0.3 | | — | |
| $Al_2O_3$ | | 0.9 | | 0.9 | | 9.7 | | 2.5 | | 1.8 | | 2.0 | | 2.1 | | — | |
| $Cr_2O_3$ | | 0.0 | | 0.0 | | — | | — | | 0.0 | | 0.0 | | — | | — | |
| FeO | | 25.7 | | 25.7 | | 8.8 | | 9.9 | | 12.6 | | 13.1 | | 8.3 | | 17.5 | |
| MnO | | 1.0 | | 0.9 | | — | | — | | 0.6 | | 0.5 | | — | | — | |
| MgO | | 19.7 | | 19.6 | | 12.7 | | 15.8 | | 13.2 | | 13.1 | | 16.3 | | 43.6 | |
| CaO | | 1.4 | | 1.4 | | 20.6 | | 19.1 | | 19.8 | | 19.3 | | 20.5 | | 0.2 | |
| $NaO_2$ | | 0.0 | | 0.0 | | 0.3 | | 0.2 | | 0.3 | | 0.3 | | 0.3 | | — | |
| $K_2O$ | | 0.1 | | 0.1 | | — | | — | | 0.1 | | 0.0 | | — | | — | |
| | | 101.1 | | 100.8 | | 99.4 | | 99.9 | | 100.5 | | 100.3 | | 100.6 | | 100.2 | |
| Cations Per N Oxygens | | 6 | | | | | | | | | | | | | | 4 | |
| Si | | 1.96 | 2.00 | 1.96 | 2.00 | 1.73 | 2.00 | 1.93 | 2.00 | 1.95 | 2.00 | 1.94 | 2.00 | 1.94 | 2.00 | 0.99 | |
| Al IV | | 0.04 | | 0.04 | | 0.27 | | 0.07 | | 0.05 | | 0.06 | | 0.06 | | — | |
| Al VI | | 0.00 | 2.01 | 0.00 | 2.01 | 0.16 | 2.04 | 0.04 | 2.01 | 0.03 | 2.02 | 0.03 | 2.01 | 0.03 | 2.02 | — | |
| Ti | | 0.00 | | 0.01 | | 0.04 | | 0.01 | | 0.01 | | 0.01 | | 0.01 | | | |
| Cr | | 0.00 | | 0.00 | | — | | — | | 0.00 | | 0.00 | | — | | — | |
| Fe | | 0.81 | | 0.81 | | 0.28 | | 0.31 | | 0.40 | | 0.41 | | 0.26 | | 0.37 | 2.01 |
| Mn | | 0.03 | | 0.03 | | — | | — | | 0.02 | | 0.02 | | — | | — | |
| Mg | | 1.11 | | 1.10 | | 0.71 | | 0.87 | | 0.74 | | 0.74 | | 0.89 | | 1.65 | |
| Ca | | 0.06 | | 0.06 | | 0.83 | | 0.76 | | 0.80 | | 0.78 | | 0.81 | | 0.01 | |
| Na | | 0.00 | | 0.00 | | 0.02 | | 0.02 | | 0.02 | | 0.02 | | 0.02 | | — | |
| K | | 0.00 | | 0.00 | | — | | — | | 0.00 | | 0.00 | | — | | — | |
| | | 4.01 | | 4.01 | | 4.04 | | 4.01 | | 4.02 | | 4.01 | | 4.02 | | 3.02 | |
| An | Wo | 2.8 | | 2.9 | | 45.7 | | 39.1 | | 41.2 | | 40.4 | | 41.4 | | | |
| Ab | En | 56.1 | | 56.0 | | 39.1 | | 45.1 | | 38.3 | | 38.2 | | 45.6 | Fo | 81.7 | |
| Or | Fs | 41.1 | | 41.1 | | 15.2 | | 15.9 | | 20.4 | | 21.4 | | 13.1 | | | |

senting an intermediate stage of island arc development. The Lesser Antilles are smaller and younger than the Japan island arc but larger and older than the Mariana island arc. In this paper I present new data on rocks from two major andesitic centers of the Lesser Antilles, pyroxene andesite from La Soufriere, Guadeloupe, and hornblende andesite from Pitons du Carbet, Martinique (Fig. 1). These lavas appear to be fairly typical of young andesitic rocks of the Lesser Antilles.

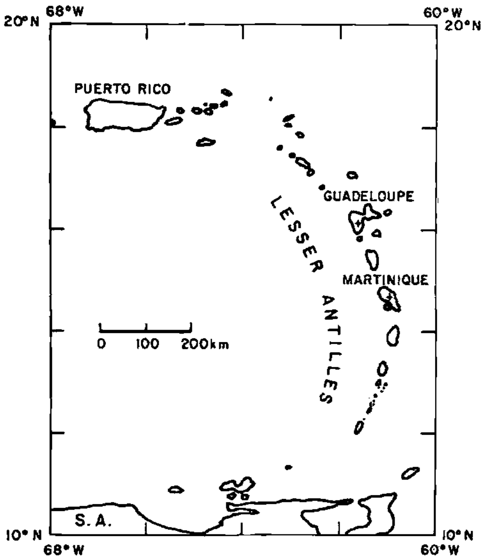

FIG. 1 - The Lesser Antilles island arc. Crosses indicate the locations of La Soufrière Volcano on Guadeloupe and Pitons du Carbet Volcano on Martinique.

## PETROLOGY OF PITONS DU CARBET AND LA SOUFRIERE ANDESITE

In terms of petrologic features, lavas of Pitons du Carbet and La Soufriere have much in common with each other and with other orogenic andesites, but there are some differences. The dominant mafic phase in La Soufriere andesite is pyroxene, while in Pitons du Carbet andesite it is hornblende. Unlike La Soufriere andesite, Pitons du Carbet andesite contains fine-grained mafic aggregates pseudomorphic after biotite. A major petrologic difference between these andesites and andesites of continental margins is that cores of resorbed plagioclase in continental andesites are generally much more sodic. Representative electron microprobe analyses of phases in Pitons du Carbet and La Soufriere andesite are given in Table 1. For phases which are relatively unzoned, a single analysis is reported. Core and rim analyses are reported for strongly zoned phases. Examples of some of these phases are shown in Fig. 2.

### *Plagioclase*

The total range of plagioclase compositions observed in both lavas is about $An_{50}$ to $An_{95}$. Except for the outermost rims, the most sodic plagioclase occurs as the cores of resorbed phenocrysts. Resorbed plagioclase phenocrysts are present in abundance in the lavas and also occur in cognate xenoliths of intermediate composition. The cores are only weakly zoned, usually from about $An_{55-50}$. In contrast, cores of similar resorbed plagioclase in Cascade andesites range from $An_{30}$ in the Lassen Volcanic Field to $An_{45}$ at Mt. Hood (EICHELBERGER, 1978). Similar plagioclase cores in the Andes are commonly $An_{25-35}$ (PICHLER and ZEIL, 1972; EICHELBERGER, 1978). Composition of rhyolite glass inclusions in cores from La Soufriere is reported in Table 2. Similar inclusions appear to be present in cores from Pitons du Carbet, but were too small for satisfactory analysis. The cores are surrounded by a « cloudy » or « dusty » zone of fine glass channels, skeletal calcic plagioclase, and fine pyroxene and opaque grains. This dark zone is in turn surrounded by a clear, normally and strongly zoned overgrowth. In both lavas, compositions range from about $An_{85}$ at the inside of this rim (« in rim » in Table 1) to about

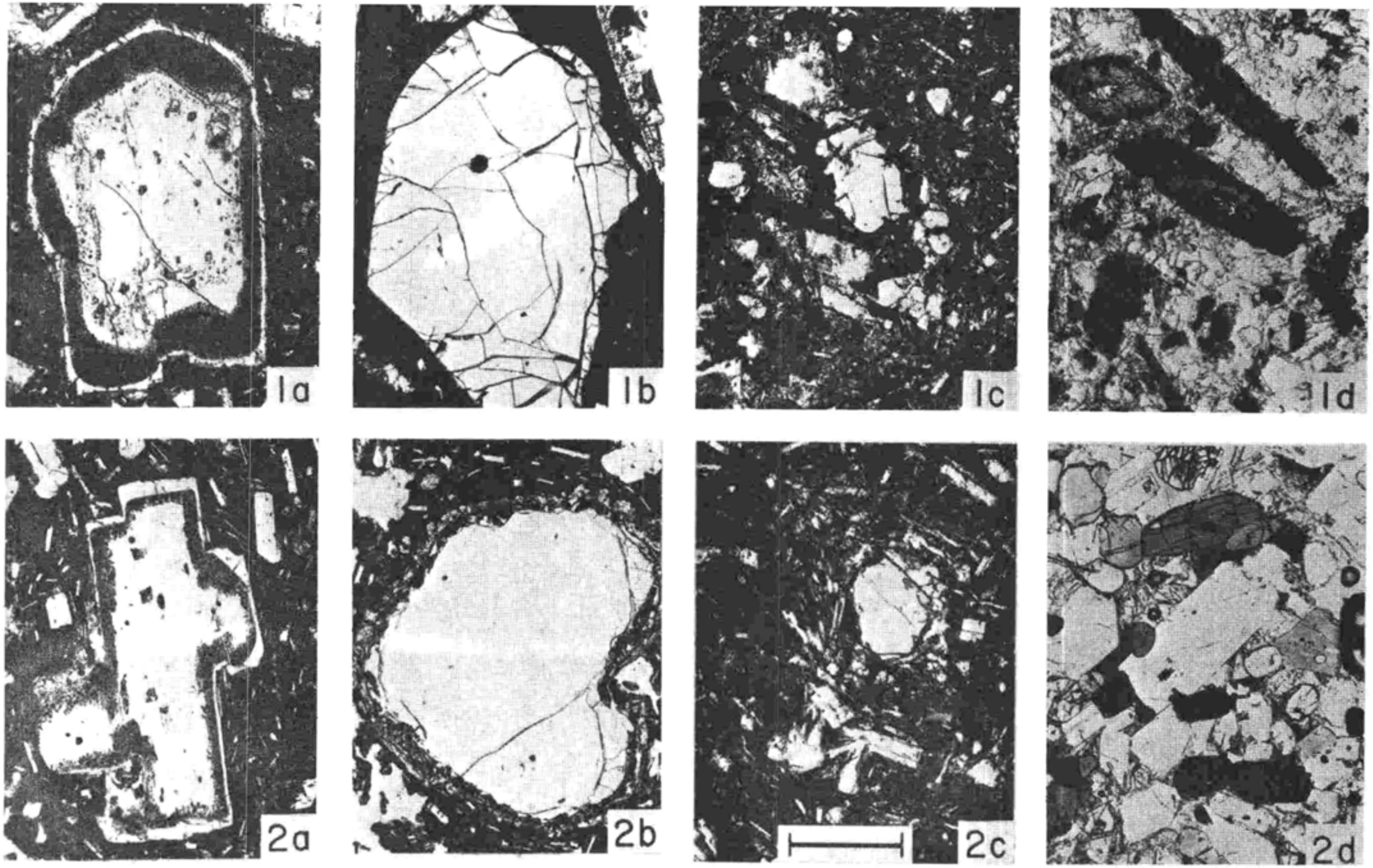

FIG. 2 - Photomicrographs of features in andesite from Pitons du Carbet (line 1) and La Soufrière (line 2). Columns are as follows: (a) resorbed plagioclase with calcic overgrowth (bar length = 0.5 mm), (b) quartz (bar length = 0.5 mm), (c) olivine (bar length = 0.2 mm), and (d) cognate xenolith (bar length = 0.5 mm). Note euhedral inclusion of rhyolite glass in quartz phenocryst (1b) and xenolithic material attached to olivine grains (1c and 2c). Xenolith in 1d is plagioclase, acicular hornblende, and interstitial glass. Xenolith in 2d is plagioclase, augite, orthopyroxene, and interstitial glass. Polars crossed for 1a through 2c, partly crossed in 2d, and not crossed in 1d. Compositions of crystalline phases are reported in Table 1. Composition of glass inclusions in resorbed plagioclase and quartz are reported in Table 2.

$An_{55}$ at the outside of the rim (« out rim » in Table 1).

Highly calcic plagioclase ($An_{90-95}$) occurs both as phenocrysts and within cognate xenoliths. It occurs as clear, weakly zoned, equant laths or as cores of more sodic grains.

The dominant plagioclase of the xenoliths (« normal xeno » in Table 1) is intermediate in composition between the resorbed sodic and highly calcic plagioclase. It occurs most often as elongate laths with normal zoning. In the Piton du Carbet xenoliths this phase is smaller, more elongate, and more strongly zoned than in La Soufrière xenoliths. Some grains in xenoliths in La Soufrière andesite lack calcic cores.

Plagioclase of similar morphology and composition («microphenocrysts» in Table 1) are dispersed throughout the host lava. Because of this similarity and because other xenolith phases are sometimes attached to them, they appear to be derived from dissaggregation of the xenoliths (or vice versa if the xenoliths are to be interpreted as cumulates).

### *Hornblende*

The dominant mafic phase of Pitons du Carbet andesites is hornblende. It invariably occurs as accicular grains rimmed with, or sometimes completely converted to, fine-grained opaque aggregates. This latter feature is probably due to oxidation during or after eruption. Some grains appear to be hollow (Fig. 2). As with the dominant plagioclase phase of the xenoliths, hornblende of the xenoliths (« xeno » in Table 1) and lava (« phen » in Table 1) are very similar in morphology and composition. Composition of the hornblende is also similar to compositions reported by HELZ (1973) for hornblende coexisting with melt in basaltic systems. Distinctive compositional features are low Si, high Al, high Na + K, and moderate Ti concentrations. Acicular hornblendes in many other orogenic andesites are strongly zoned from compositions similar to Pitons du Carbet hornblende to high Si, and low Al, Na + K, and Ti (EICHELBERGER and BARTHOLOMEW, 1978). Since the rims of Pitons du Carbet hornblende were destroyed by oxidation, some of the range in composition may have been lost.

### *Pyroxene*

Only trace amounts of orthopyroxene are present in Pitons du Carbet andesite, but it is a major phase in La Soufrière andesite. Morphologically and compositionally similar orthopyroxenes occur as major phases in both cognate xenoliths (« xeno » in Table 1) and lava (« phen » in Table 1) of La Soufrière. They are subhedral to euhedral and relatively unzoned.

Three kinds of augite occur in La Soufrière andesite and two kinds occur in Piton du Carbet andesite. One type of augite which is present in lavas of both volcanoes (« aluminous » in Table 1) occurs as small laths with hourglass extinction in both xenoliths and lava. It is strongly zoned, with very aluminous cores and low Al, low Ti rims. The dominant type of augite in La Soufrière lava is absent in Pitons du Carbet lava. Its mode of occurrence is the same as orthopyroxene, appearing in both xenoliths and lava (« xeno » and « phen » in Table 1). It is also relatively unzoned, resembling in composition the rims of the aluminous augite. The third type of augite comprises the reaction rims on quartz phenocrysts present in both lavas. In its low Ti and Al contents it resembles the dominant augite of La Soufrière and differs greatly from the aluminous augites of both lavas.

### *Quartz*

Quartz phenocrysts in various stages of resorption and reaction are present in both lavas. Remnants of well-formed

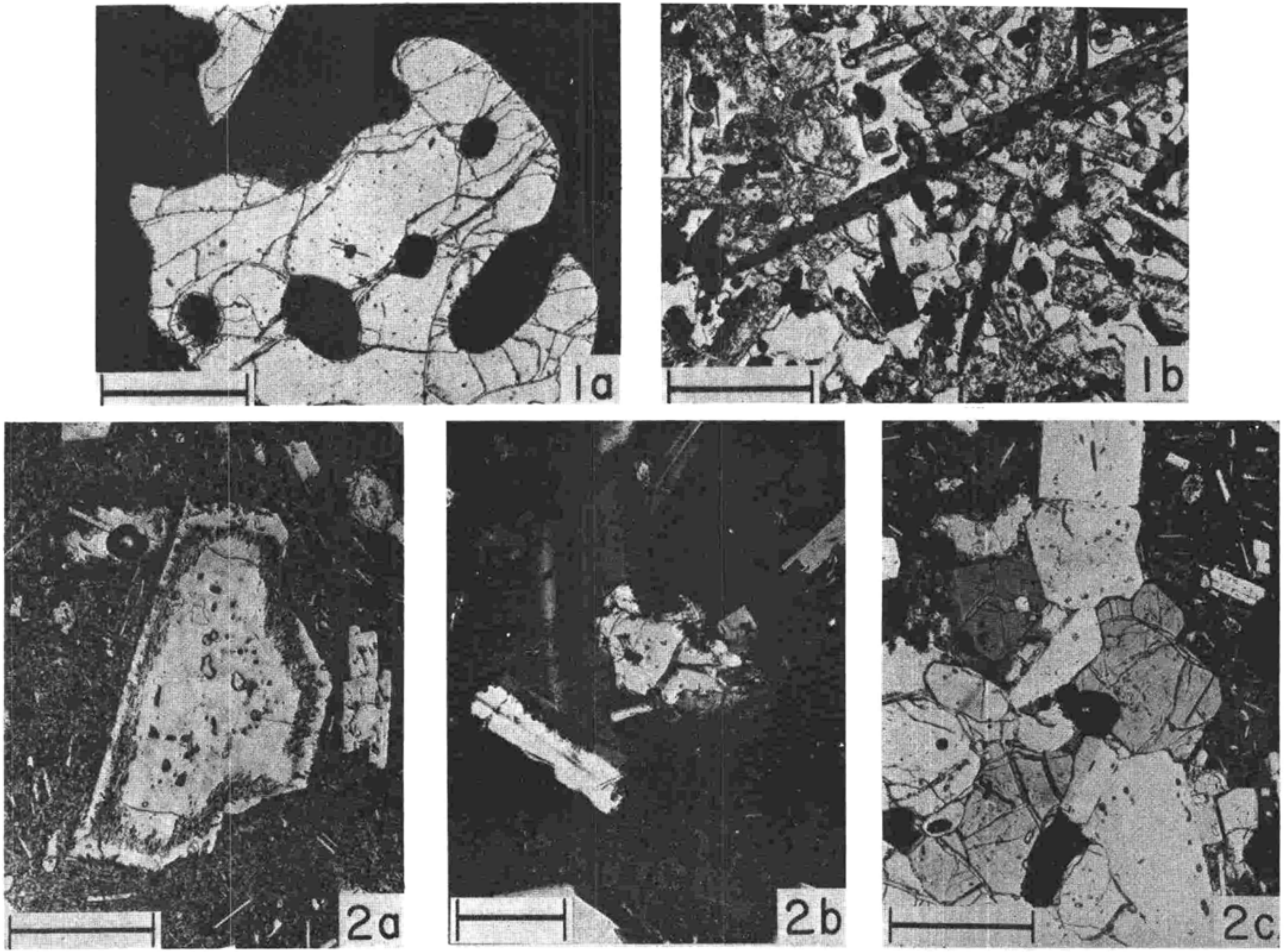

FIG. 3 - Features in a dacite and an andesite from Japan, analogous to features in Fig. 2. Line 1 (dacite from Onikobe): (a) quartz (bar length = 0.5 mm), (b) cognate xenolith (bar length = 0.5 mm). Line 2 (andesite from Kirishima): (a) resorbed plagioclase (bar length = 0.4 mm), (b) olivine (bar length = 0.1 mm), (c) cognate microxenolith (bar length = 0.5 mm). Note inclusions of rhyolite glass in quartz phenocryst (1a) and xenolithic material attached to olivine grain (2b). Xenolith in 1b is plagioclase, accicular hornblende, and interstitial glass. Xenolith in 2c is plagioclase, augite, orthopyroxene, and interstitial glass.

faces are present on some of the grains, suggesting that the phenocrysts were originally euhedral. Euhedral holes or « negative crystals » are present within quartz in Pitons du Carbet andesite (Fig. 2) and contain rhyolitic glass (Table 2). Formation of augite rims appears to begin with growth of fine-grained aggregates at the edge of the crystal. As the melt fraction at this boundary increases with increasing $SiO_2$ content of the melt, larger, elongate augites grow within the early formed augite ring and normal to the retreating crystal. SATO (1975) has carefully described these features in some andesitic lavas from Shikoku, Japan.

### *Olivine*

Olivine occurs as resorbed, relatively unzoned grains within both lava and cognate xenoliths of Pitons du Carbet and La Soufrière andesites. Where it occurs in the lava, it often has some xenolithic material attached (Fig. 2). It is seldom found in direct contact with either the interstitial glass of the xenoliths or the groundmass of the lava.

TABLE 2 - Observed and inferred compositions of magmas related to Pitons du Carbet and La Soufrière andesite.

| | *1* | *2* | *3* | *4* | *5* | *6* | *7* | *8* | *9* | *10* | *11* |
|---|---|---|---|---|---|---|---|---|---|---|---|
| $SiO_2$ | 50.1 | 51.4 | 50.0 | 50.0 | 61.2 | 60.3 | 73.8 | 73.9 | 73.8 | 74.4 | 74.0 |
| $TiO_2$ | 0.6 | 0.5 | 0.9 | 1.0 | 0.5 | 0.6 | 0.1 | 0.3 | 0.2 | 0.1 | 0.5 |
| $Al_2O_3$ | 18.1 | 18.6 | 19.3 | 19.7 | 17.5 | 17.7 | 12.7 | 12.5 | 13.3 | 16.0 | 16.6 |
| FeO | 11.6 | 10.9 | 10.3 | 10.8 | 5.9 | 6.4 | 1.2 | 2.3 | 2.0 | 1.4 | 0.2 |
| MnO | — | — | — | — | 0.1 | 0.2 | — | — | — | — | — |
| MgO | 6.0 | 5.3 | 4.6 | 6.8 | 2.4 | 3.7 | 0.2 | 0.4 | 0.4 | 0.0 | (1.1) |
| CaO | 10.7 | 10.3 | 11.5 | 9.2 | 7.1 | 6.0 | 1.4 | 2.3 | 2.7 | 2.5 | 3.5 |
| $Na_2O$ | 2.4 | 2.5 | 2.7 | 1.7 | 3.2 | 2.6 | 3.9 | 4.0 | 3.9 | 3.7 | 4.0 |
| $K_2O$ | 0.6 | 0.5 | 0.8 | 0.9 | 1.3 | 1.3 | 4.7 | 3.2 | 3.3 | 1.8 | 2.3 |
| | 100.0 | 100.0 | 100.1 | 100.1 | 99.2 | 98.8 | 98.0 | 98.9 | 99.6 | 99.9 | 100.0 |

1. Average of three high-Al basalts from Martinique (WESTERCAMP, 1975), recalculated water-free.
2. Average of two cognate basaltic xenoliths in Pitons du Carbet andesite, Martinique (WESTERCAMP, 1975), recalculated water-free.
3. Composition of basaltic parent of Pintons du Carbet andesite calculated from andesitic composition 5, recalculated water-free. and rhyolitic composition 10.
5. Average of seven andesites from Pitons du Carbet (GUNN *el al.*, 1974).
6. Composition of andesite from summit dome of La Soufrière. Guadeloupe. Electron microprobe analysis of fused bead. R. Gooley, analyst.
7. Average of three analyses of glass inclusions in quartz phenocrysts, Pitons du Carbet andesite.
8. Average of thirteen analyses of glass inclusions in cores of resorbed plagioclase, La Soufrière andesite.
9. Average of five analyses of rhyolite lava, South Aegean Arc (NICHOLL, 1971).
10. Melt composition, Picture Gorge Tholeiite, T = 750°C, $P = 5$ kb, $H_2O$ saturated (HELZ, 1976).
11. Composition of rhyolitic parent of Pitons du Carbet andesite calculated from andesitic composition 5, recalculated water-free, and basaltic composition 2.

### *Biotite Pseudomorphs*

Pitons du Carbet andesite contains large (≤ 2 cm) phenocrysts of biotite which are almost completely converted to masses of oxides, pyroxene and plagioclase.

### *Other Features*

Two other, relatively rare, features deserve comment. Some ortho-pyroxenes have rims of moderately aluminous augite. These grains appear to be derived from the cognate xenoliths. In addition to cognate xenoliths described earlier, Pitons du Carbet andesites contain small fragments with corroded grains of plagioclase and hornblende, possibly representing lower crustal material.

## DISCUSSION OF DATA

For the reasons outlined above, the data indicate that both andesites formed by mixing of high-Al basaltic magma crystallizing calcic plagioclase, forsteritic olivine, and aluminous augite with rhyolitic (or rhyodacitic) magma crystallizing moderately sodic plagioclase and quartz. The rhyolitic parent of Piton du Carbet andesite may have, in addition, contained biotite. In the case of Pitons du Carbet, mixing caused growth of the rims on resorbed sodic plagioclase, the similarly zoned « normal » plagioclase of the cognate xenoliths, hornblende of the cognate xenoliths, and phenocrysts derived by disaggregation of the xenoliths. In addition, augite reaction rims formed on quartz. The situation is similar for La Soufrière, except that orthopyroxene and clinopyroxenes formed in place of hornblende. Mixing explains the occurrence of the non-equilibrium phases, the rhyolitic glass inclusions in the resorbed plagioclase and quartz, the cognate basaltic xenoliths, and the textures suggestive of rapid growth in the xenoliths.

The occurrence of hornblende in place of pyroxene as a mixing product in Pitons du Carbet has a number of possible explanations: (1) lower temperature, (2) effect of bulk composition especially water content and, (3) mixing at higher pressure. Lower temperature can be due to either lower initial temperature of one or both of the parental magmas or a larger proportion of the cooler rhyolitic melt in the mixture. In the Lassen volcanic suite (EICHELBERGER and GOOLEY, 1977) the cooler, more silicic mixtures contain plagioclase-hornblende xenoliths while the hotter, more mafic mixtures contain plagioclase-pyroxene xenoliths. However, the compositions of Pitons du Carbet and La Soufrière andesite are so similar that proportions of parental magmas in the mixtures are unlikely to be very different. The only other obvious difference between the two rocks is the presence of mafic aggregates pseudomorphic after biotite in Pitons du Carbet. This may indicate a higher water content and/or lower temperature of Pitons du Carbet parental rhyolite, which would account for the stability of hornblende in the hybrid.

## COMPOSITION
## OF PARENTAL MAGMA

The bulk composition of the parental magmas can be inferred from: (1) composition of associated, end member, unmixed magmas or where such data is lacking, by analogy to other regions or appropriate experimental data, (2) extrapolated composition of end members assuming that the observed variation in composition is due entirely to mixing, (3) for the basaltic parent, composition of the cognate xenoliths, (4) for the rhyolitic parent, composition of glass inclusions in resorbed plagioclase and quartz. Results of those approaches for which data are available are given in Table 2. The average composition of high Al basalts from Martinique (WESTERCAMP, 1975) is very close to the composition of cognate basaltic xenoliths in Pitons du Carbet andesite (WESTERCAMP, 1975) and probably closely approximates the composition of the basaltic parent. Calculated compositions for the Pitons du Carbet and La Soufrière basaltic parents using a rhyolitic melt composition from HELZ (1976) are also shown. The appropriateness of this melt is discussed below.

Since the rhyolitic glass inclusions in the two andesites were clearly trapped after growth of plagioclase and quartz phenocrysts began in the rhyolitic magma, they represent the composition of the liquid fraction but may differ somewhat from the bulk composition of the magma. Growth of plagioclase would deplete the liquid in Al and Ca. Growth of quartz and plagioclase have opposite effects on Si in the liquid. The exact composition of the parental rhyolite cannot be determined, but the glass inclusion compositions, an island arc rhyolite, a rhyolitic melt composition (HELZ, 1976), and a composition calculated from the basaltic parent composition are listed for comparison in Table 2.

## SOURCE OF RHYOLITIC MAGMA

Evidence that these andesites formed by mixing of basaltic and rhyolitic magmas implies that the rhyolitic magma is not related to basaltic magma by fractional crystallization, since the intermediate members of the series do not represent a « liquid line of descent ». Rather, it appears probable that the rhyolitic magma, like basaltic magma, is a product of melting.

The occurrence of rhyolitic magma in a relatively young island arc places severe constraints on possible sources of the magma. Geological and geophysical data indicate that the Lesser Antilles arc is built on oceanic crust of Jurassic age (WESTBROOK, 1975). Further, those features which make the arc more continental in character than surrounding oceanic crust, greater crustal thickness and presence of silicic rocks, appear to be due to the calc-alkaline arc magmatism itself. To attribute the rhyolitic magma to remelting of silicic intrusives begs the question of how silicic magmas form in an initially oceanic environment. It seems unlikely that these volcanoes are simply recycling previous magmatic products. If this were true, the arc has ceased to grow, since no new silicic material is being produced. There are no other data to support such a conclusion. If on the other hand, the parental rhyolitic magma represents the generation of new silicic material in the arc, as the history of the arc would seem to suggest, then it must form by large scale melting subducted sediments or partial melting of ultramafic mantle or basaltic oceanic crust. The thick ridge of highly deformed sediments east of the Lesser Antilles suggests that sediments are being scraped off the down-going plate rather than subducted (WESTBROOK, 1975). Even if enough sediments were subducted to produce the rhyolitic magma, it seems unlikely that such a magma would survive its journey through 100 km of ultramafic mantle. Yet such a journey would be required, since the rhyolitic magma began precipitating plagioclase before mixing with basalt. An ultramafic source for rhyolitic magma appears to be ruled out by experimental data (STERN and WYLLIE, 1973). Two possible oceanic crustal sources lie beneath the volcanoes: the crust directly beneath them and the subducted slab. The slab lies as 100 km to 200 km depth (WESTBROOK, 1975) and experimental data indicate that partial melting at these mantle pressure would not produce highly silicic liquids (STERN and WYLLIE, 1973; GREEN and RINGWOOD, 1968). However, the lower Lesser Antilles crust represents a large volume of material under conditions appropriate to produce rhyolitic magma. Data on melting of basaltic material under lower crustal conditions indicate that melts of appropriate composition will be produced up to about 20 % melt fraction (HELZ, 1976). Such a melt is consistent with the La Soufrière and Pitons du Carbet compositional data (Table 2) and coexists with $An_{50}$ plagioclase, as does the parental rhyolitic magma.

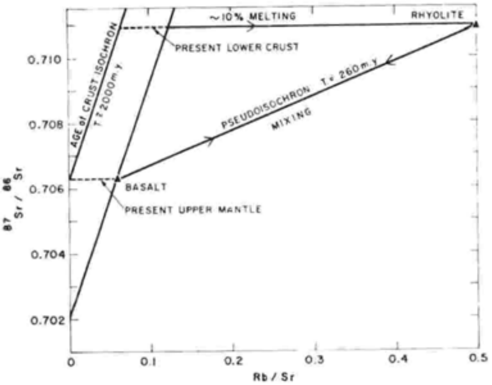

FIG. 4 - $^{87}Sr/^{86}Sr$ vs. Rb/Sr diagram illustrating interpretation of Sr isotopic data in mixing model. Separate crust and mantle systems develop during some previous crust-forming event (T = ∾ 2000 m.y.). During a recent magmatic event ($T = 0$ m.y.) melting occurs in both the mantle and lower crust. A small degree of melting in the lower crust produces a rhyolitic magma with Rb/Sr of 5 to 10 times that of its source. Mixing of this magma with basaltic magma from the mantle produces andesitic magma with $^{87}Sr/^{86}Sr$ only slightly higher than the basaltic parent magma. The mixtures fall on a pseudoisochron giving an age much less than the age of the crust/mantle system. Note that for young, low Rb/Sr crust such as is under island arcs, the Sr isotopic difference between hybrid andesite and basaltic parent is negligible. The pseudoisochron shown is based on data from Barroso, Peru (JAMES *et al.*, 1976), projected to basaltic and rhyolitic compositions.

Mixing of a crustal partial melt with a mantle partial melt does not conflict with the two main constraints which have been used to argue against crustal involvement in andesite genesis. Because the crustal component in andesite is derived from a mafic source, continental crust which is lacking or poorly developed in island arcs is not required. Because the crustal component is derived from a mafic source and diluted with a mantle component, the andesite has low $^{87}Sr/^{86}Sr$.

This interpretation of the isotopic data is shown in Fig. 4.

## DEPTH OF MIXING

The presence of large volumes of relatively homogeneous hybrids implies mixing of large batches of parental magmas within plutons. The composition and phase assemblage of the parental magmas before mixing, and the phases which grow within the pluton during and after mixing, place constraints on the depth of the pluton. Possibly significant features in these andesites bearing on depth of mixing are Al content of basalt-derived and mixing-generated pyroxenes, coexistence of forsteritic olivine and calcic plagioclase in the xenoliths, early precipitation of plagioclase within basaltic magma before and during mixing, and involvement of rhyolitic magma.

The Al content of pyroxene in basaltic magma may increase with increasing depth of crystallization (GREEN and RINGWOOD, 1967). Thus the high Al content of basalt-derived augite in the andesites would indicate that crystallization of the basaltic parent began at high pressure. The low Al content of pyroxenes produced during mixing by crystallization of basaltic magma and by reaction of quartz would indicate that mixing occurred at low pressure. There is however, little correlation between Al content of pyroxene in basaltic systems and pressure in available experimental data and there are very few data for high Al basalt.

KUSHIRO and YODER (1966) have shown that pure forsterite and anorthite react to form clinopyroxene, orthopyroxene, and spinel above 8 kb. Absence of this reaction within cognate xenoliths containing forsteritic olivine and calcic plagioclase implies a relatively low pressure of formation and hence of mixing. Sodium content of plagioclase raises the pressure of this reaction over the pure end members, but iron content of olivine lowers it. Although offering a different interpretation of the xenoliths,

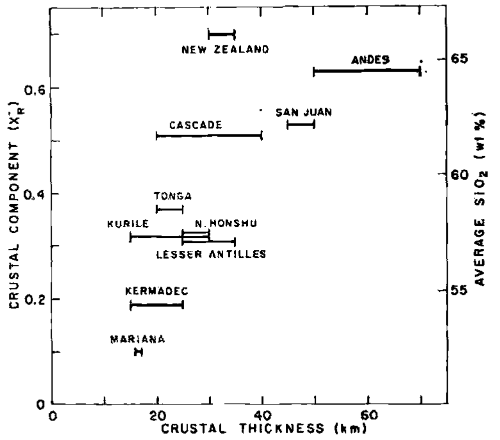

FIG. 5 - Crustal melt component of surface lavas ($X_R$) versus crustal thickness for andesitic volcanism in several island arcs and continental margins, where $X_R = R/R + B$, R = rhyolitic parent magma and B = basaltic parent magma, by volume. $X_R$ is calculated from the average silica content of lavas (average of equally weighted analyses), assuming that the basaltic parent is 50 wt. % $SiO_2$ and the rhyolitic parent is 73 wt. % $SiO_2$. There is a general increase in crustal component with increasing thickness of crust. Data sources are as follows: Mariana: MIYASHIRO (1974), MARAUCHI *et al.* (1968); Kermadec: EWART *et al.* (1977); KARIN (1970), BRODIE (1964); Lesser Antilles: BAKER (1968), WESTBROOK (1957); Kurile: MIYASHIRO (1974), GORSHKOV (1970); Northern Honshu: YOSHII and ASANO (1972), MIYASHIRO (1974); Tonga: EWART *et al.* (1977), BRODIE (1964); Cascade Range: MIYASHIRO (1974), WARREN and HEALY (1973); San Juan: LIPMAN *et al.* (1978); WARREN and HEALY (1973); Andes (central): MIYASHIRO (1974), JAMES (1971); New Zealand: EWART *et al.* (1977), BRODIE (1964).

YODER (1969) suggested that they must form at less than 7 kb.

Early precipitation of plagioclase in the basaltic magma, beginning before mixing, indicates that the basaltic magma reached shallow depth before mixing with rhyolitic magma. Experimental work (GREEN and RINGWOOD, 1967; YODER and TILLEY, 1962) shows that plagioclase is a late appearing phase at 10 kb, and is an early phase at 1-2 kb. The pressure effect on the order of crystallization is difficult to quantify however, because water content of magma in addition to pressure strongly influences stability of plagioclase.

The involvement of rhyolitic magma places a further constraint on the depth of mixing. Clearly mixing must occur at or above the level where both magmas are present. Since the lower Lesser Antilles crust seems the most likely rhyolite source region, the plutons in which mixing occurs and from which the volcanoes tap their lavas must lie within the Lesser Antilles crust. In accordance with the model outlined earlier, the relationship of these volcanoes to active subduction is attributed to the rise of basaltic magma from the Benioff zone. This magma supplies both the heat for crustal melting and the mantle component of the andesite.

## OCCURRENCE OF EVIDENCE OF MIXING IN ANDESITES

Examples of features in lavas of Japan, similar to those just described in the Lesser Antilles, are shown in Fig. 3. KUNO (1950) estimated that about half the rocks of his « hypersthenic series » and a quarter of the rocks of his « pigeonitic series » contain resorbed sodic plagioclase. At least some features discussed here as indicative of mixing are present in nearly all the andesites I have studied. Both cognate xenoliths and resorbed sodic plagioclase are present in a large portion of andesites and dacites of the Cascade Range. Because evidence of mixing appears to be present in the majority of andesites and because it can account for the bulk composition of these rocks, I will proceed with the assumption that it is a dominant process by which these rocks form.

## EVOLUTION OF THE CRUST

Continental crust differs from oceanic crust in that it is thicker and contains

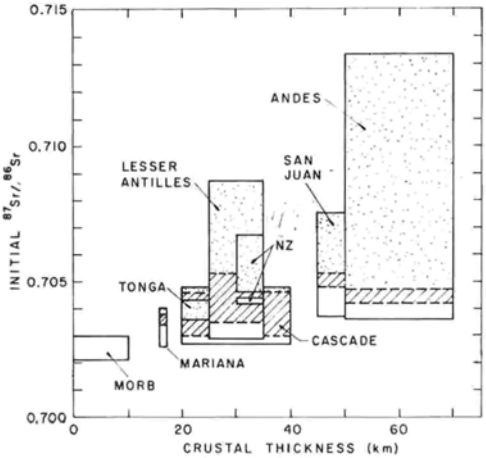

FIG. 6 - Initial $^{87}$Sr/$^{86}$Sr versus crustal thickness for some of the volcanic arcs and fields shown in Fig. 7. Cross-ruled areas indicate region of overlap of isotopic compositions of basaltic and more silicic rocks. Stippled areas and cross-ruled areas indicate range of intermediate and silicic volcanic rocks. White areas within boxes and cross-ruled areas are range of basaltic rocks. Rocks with $SiO_2$ < 53 wt. % were defined as basalt except where $SiO_2$ content was not reported. In the latter case, the rock name employed in the data source was used. Mid-ocean ridge basalt (MORB) is shown for comparison. There is a general increase in $^{87}$Sr/$^{86}$Sr of silicic rocks with increasing crustal thickness. References, in addition to those used in Fig. 5, are as follows: MORB: compiled by MEIJER (1976); Mariana: MEIJER (1976); Tonga: GILL and COMPSTON (1973); Lesser Antilles: HEDGE and LEWIS (1971), DONNELLY *et al.* (1971), PUSHKAR (1968); New Zealand: EWART and STIPP (1968); Cascade: PETERMAN *et al.* (1970), HEDGE *et al.* (1970), CHURCH and TILTON (1973; San Juan: LIPMAN *et al.* (1978); Andes (central): KLERKX *et al.* (1977), JAMES *et al.* (1976), MCNUTT *et al.* (1975).

low density, silica-rich igneous rocks, especially in its upper portion. Because island arcs appear to be built on oceanic crust and represent thickened crust enriched in silica-rich igneous rocks, they seem to represent an early stage in development of continental crust. Therefore any model of andesite genesis which applies to island arcs has implications for crustal evolution. For a model to apply to both island arc and continental margin andesites, it must account for the more silicic and radiogenic character of andesitic volcanism in a continental environment (Fig. 5 and 6). The model discussed here can account for the crustal evolution which island arcs represent, because it involves both thickening of the crust and fractionation of the lower crust with upward migration and concentration of the low melting component. It also appears to account for changes in volcanism accompanying crustal evolution (Fig. 7).

The most primitive crust, oceanic crust derived from the mantle, is the starting point. Where this crust is trapped over a mantle heat source it is heated and thickened by emplacement of basaltic sills. Eventually the solidus is exceeded at its base, and the low melting silicic component migrates upward and becomes concentrated at high levels. Intermediate magmas are generated by mixing of the mantle and crustal melts simply as a consequence of basalt-induced crustal melting. Thickening of the crustal and concentration of crustal melt products at high levels results in a crust more continental in character. Subsequent crustal melting events may involve both the basaltic sills of previous events and any previous products of crustal fractionation which remain at depth sufficient for remelting. Thus with increasingly continental character of the crust, more fractionated and aged material becomes available for melting. This provides an explanation for correlation of volume and isotopic composition of silicic lavas with crustal setting (Fig. 5 and 6). The increasing volume and radiogenic composition of these lavas with increasing crustal thickness reflects the increasing volume and radiogenic composition of the rhyolitic parent magma, which in turn reflects an increasingly aged, fractionated and larger source. The crustal melts

Table 3 - Some models for andesitic volcanism and their consequences for crustal evolution.

| | *Model* | *Primary basis* | *Crustal evolution if dominant process* | *Comments* | *References (Partial List)* |
|---|---|---|---|---|---|
| Mantle process | Andesitic liquid from partial melting of ultramafic mantle or subducted basaltic crust or deep fractionation of basaltic melt from mantle. Melt source in or near Benioff zone. | Voluminous nature of andesitic volcanism in subduction zones. Feasibility indicated by experimental data. | Addition of andesitic material to crust. Subduction of water-rich crust required for development of continental crust. | Does not by itself account for genesis of andesite outside subduction zones, more voluminous character and radiogenic composition of continental silicic magmas, or granitic character of upper continental crust. | Yoder (1969), Green and Ringwood (1968), Kushiro (1972, 1973, 1974), Mysen and Boettcher (1975). |
| | Melting of subducted sediments. | Transport of lowmelting material to mantle depth. | No new continental material generated. | As above, but limited by isotopic data. | Coats (1962), Donnelly *et al.* (1971). |
| Crustal process | Andesitic liquid from large degree of melting of lower crust. | Voluminous nature on andesitic volcanism and intermediate batholiths in areas of thickened crust. | Mild fractionation of crust but no growth. | Does not account for thickening of crust accompanying volcanism. Requires excessively high temperature for lower crust. | Forbes *et al.* (1969), Bateman and Wahrhaftig (1966). |
| | Andesitic mush from lower crust consisting of rhyolitic liquid and refractory residuum. | As above. | As above | As above, but avoids problem of high temperature. | Piwinskii (1973), Presnall and Bateman (1973). |
| Mantle and crustal processes | Assimilation of granitic rocks by basaltic magma from mantle. | Xenocrysts in andesite. | Gradual oceanization of continents by dilution with basalt. | Does not account for andesites in island arcs. Limited by isotopic data. | Turner and Verhoogen (1960). |
| | Low pressure fractionation of basaltic magma from mantle to produce andesite. Partial melting of lower crust to produce dacitic liquid after thickening of crust. | Association of basalt with andesite. Development of batholiths in mature island arcs and continental margins. | Growth and fractionation of crust with time. | Yields basaltic bulk composition of crust unless refractory material returns to mantle. Requires very high temperature for lower crust. Does not account for higher $^{87}Sr/^{86}Sr$ of continental andesite. | Kuno (1968). |
| | Mixing of basaltic magma from mantle with rhyolitic magma from lower crust. | Phase assemblages and textures in andesite. | As above. | As above, but avoids problem of high temperature in lower crust while accounting for correlation of isotopic composition of andesite with crustal setting. | Piwinskii and Wyllie (1968), Younker and Vogel (1976).<br>this paper. |
| | Andesitic melt from subducted slab and overlying mantle plus rhyolitic melt from lower crust. | Voluminous nature of andesitic volcanism in subduction zones.<br>Silicic character of continental volcanism. Sr isotopic data. | As above. | Similar to above except that mafic parent magma is andesitic. Yields continental crust of andesitic composition. Does not account for higher $^{87}Sr/^{86}Sr$ of continental andesite or genesis of andesite outside subduction zones. | Brown (1977), Younker and Vogel (1976). |

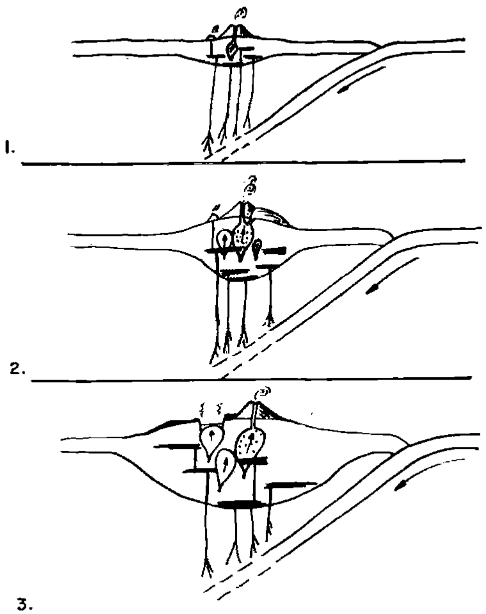

FIG. 7 - Model for crustal evolution based on evidence of mixing in andesite. (1) Young island arc: thickening and heating of oceanic crust with emplacement of basaltic sills. Melting of amphibolite begins at base of crust and andesitic magma form by mixing of mantle and crustal melts. Surface lavas are basaltic to andesitic, *e.g.* Mariana. (2) Mature island arc: Zone of melting expands as crust approaches continental thickness. Material undergoing melting is still generally amphibolite of basaltic composition. Surface lavas are basalt to rhyolite, andesitic on average. *e.g.* Lesser Antilles. (3) Continental margin or old island arc: Large volume of lower crust undergoes partial melting, including earlier emplaced basaltic material and silicic material from earlier episodes of crustal melting. Some refractory material may sink back into mantle. Surface lavas are basalt to rhyolite, dacitic on average. Granitic batholiths accumulate in upper crust and average more silicic than volcanic piles because mixing favors eruption, *e.g.* Andes.

become less calcic as well, and this is reflected in the less calcic composition of the resorbed plagioclase of continental andesites.

In this view of andesite genesis and crustal evolution, plate interactions at subduction zones merely provide a heat source, in the form of foci of basaltic volcanism, for the crustal melting which produces silicic magmas and hence continental crust. Special conditions within the mantle of subduction zones are not required for development of continental crust. Rather, the critical factor in evolution of the crust is that the solidus is exceeded within the crust. This depends on the presence of water, thickness of the crust, and heat transfer either by conduction or movement of magma. Early heat flow may have been sufficient to cause crustal melting without the rise of basaltic magma from the mantle. If so, the process of generation of continental crust would have been more efficient early in the earth's history than in modern island arcs. Development of granitic crust can therefore be viewed as the next step in evolution of a planet possessing basaltic crust, and occurs by fractionation of the basaltic crust itself.

## COMPARISON OF MODELS

Some of the major features and implications for crustal evolution of models of andesite genesis are summarized in Table 3. In general, models emphasizing mantle derivation fail to account for the more silicic character of continental magmatism, while those emphasizing crustal origin fail to account for crustal thickening and encounter difficulties with the high temperatures required. In this paper I have presented the view that petrologic data support a common origin of island arc and continental andesites, but equally strongly indicate crustal involvement. The mixing model presented here is really a middle approach between the extremes of mantle or crustal derivation. It is of course possible that some or all of the possible processes apply to different andesites, or in different degrees to the same andesite. Yet andesites from widely dispersed volcanic fields contain distinctive and remarkably similar features suggestive of a common and perhaps simple origin. Most of the models can be tested, particularly with isotopic data. Although many workers have used the gross isotopic similarity of magmas in the andesitic association to support fractionation models involving mantle-derived magmas, many well sampled suites show significant correlation between elemental and isotopic composition. What appears to be needed is data on volcanic units more closely related in time and space, so that the variables induced by dealing with different magma batches are minimized. A good focus for more detailed work are the voluminous, compositionally zoned ash flow sheets from andesitic centers such as the Aso Caldera of Japan and Crater Lake of the United States. If andesites are related to other magmas by fractionation, then these are plutons quenched in the process of fractionation (HEDGE and NOBLE, 1976). If andesites are mixtures of basaltic and silicic magmas, then these are plutons quenched in the process of mixing (EICHELBERGER and CROWE, 1978).

NOTE ADDED IN PROOF: Further work on basaltic xenoliths indicates that mixing is driven by vapor exsolution in wet basaltic magma as it enters the cooler silicic magma chamber. This process operates at pressures no greater than a few kilobars.

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