

Studies of volcanic series related to the origin of some marginal sea floors

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Abstract—The evolution of basalt series reflects the thermal regime and dynamics of mantle diapirism as well as interaction between ultramafic magma and crustal rocks. The pressure–temperature regime of the origin and fractionation for two types of basalt series has been determined with the following semi-empirical equations:

$$T(^{\circ}\text{C}) = 1047 + 44X - 13.77X^2 + Y(314.97 + 1941.55X - 630.69X^2) + 4P(\text{kbar}), \quad (1)$$

where X is the mol fraction of forsterite in olivine and Y , the abundance of nonbridging oxygen associated with Fe and Mg in the melt. The pressure may be estimated from;

$$P = 40.57 + 21.33 \ln(a_{\text{K}}^{\text{K}}/a_{\text{Na}}^{\text{K}}) + 3.794 \ln(a_{\text{K}}^{\text{K}}/a_{\text{Na}}^{\text{K}})^2, \quad (2)$$

which was derived by least-squares regression from the bulk compositions of 23 nondifferentiated magmatic liquids, from experimental data, and partial molar volumes of the components involved. The results are compared with pressure estimates for ultramafic inclusions in alkali basalts.

The first group of basalts includes tholeiite and alkali basalt that are typical of growing igneous crust at both convergent and divergent plate margins. The evolution of these rocks show typical fractional crystallization behavior with gradual enrichment of alkalis and silica in the magma. This series evolves rapidly with the tholeiitic members equilibrating at 35–40 km depth and the alkali basalt members at depths between 60 and 100 km in the upper mantle.

The second volcanic series shows an evolution from acid to mafic members over a long time period (~60 m.y.). The pressures and temperatures required to produce the initial magmas for this group are 4–10 kbar and $\leq 1470^{\circ}\text{C}$, respectively. This rock series is typical of marginal sea floors, and may reflect interaction (thermal erosion and magma mixing) between molten diapirs of ultramafic composition with crustal rocks. This process may include magmatic replacement of continental crust by newly-formed oceanic crust at an average rate of ~ 0.6 cm/year. It is proposed that this process is typical for active continental margins as exemplified by the western Pacific sea floor (*e.g.*, the Philippine, Japan and Okhotsk sea floors). A possible mechanism for simultaneous origin of deep-sea depressions and island arcs is also discussed.

INTRODUCTION AND FORMULATION OF THE PROBLEM

BOTH NONDIFFERENTIATED and well-differentiated basalts series are widespread in the crust of the Earth. Figure 1 illustrates the relationship between alkalis and silica for several basalt series that are typical of continents, island arcs and oceanic islands. These series do not represent the total diversity of the volcanic series known, but are the most typical. In particular, the most acidic series and high-alkali groups will not be discussed here. The former shows a maximum alkali content, whereas the latter exhibits a silica maximum on the $(\text{Na}_2\text{O} + \text{K}_2\text{O}) - \text{SiO}_2$ differentiation curves (*e.g.*, COX *et al.*, 1979; PERCHUK and FROLOVA, 1980).

The paths in Figure 1 may be viewed as projections of cotectic minima in basalt systems onto the plane silica + alkalis. The position of each projection depends principally on bulk composition and pressure. These parameters are not well known. The bulk composition of a series (*i.e.*, composition of the initial basalt magma) is difficult to calculate be-

cause of lack of knowledge about the proportions of the products differentiated. Pressure estimates are not always available due to lack of calibrated petrological barometers. Nevertheless, the values of some of these parameters may be estimated.

Nondifferentiated basalts have been suggested as analogues of "primary" magmas (PERCHUK and FROLOVA, 1979; PERCHUK, 1984a). The principal nondifferentiated basalts are alkali basalt and tholeiite (see Figure 2). These magmas may undergo fractionation and subsequent formation of volcanic series. At low pressures (shallow depths), tholeiitic magmas are produced. As the depth increases, the silica content and degree of partial melting of the mantle material decrease, whereas the amount of alkalis may increase (*e.g.*, KUSHIRO, 1983). Bulk compositions of 23 nondifferentiated and slightly differentiated volcanic series are listed in Table 1 with additional data on Al–(Fe, Mg)–Ca variation shown in Figure 2.

The dots in Figure 2C are distributed within a path limited by clear lines of a negative slope. These

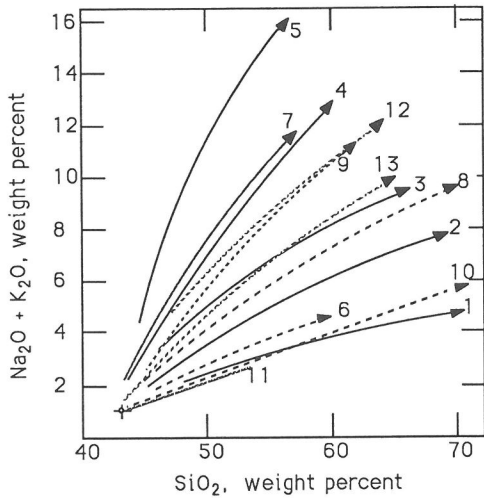


FIG. 1. Curves fitted by least squares regression of alkalis and silica for basalt series developed in the crust of continents (1–5), island arcs (6–10) and oceanic islands (11–13). 1, 2 = Karoo, South Africa (WALKER and POLDEVART, 1949); 3 = volcanic suite of Aden and Little Aden (COX and MALICK, 1970); 4 = Ethiopia trap formation (MOHR, 1960; SCHEINMANN, 1968); 5 = high alkali basalts from Central Africa (SCHAIMANN, 1968; PERCHUK and VAGANOV, 1977); 6, 7 = Little Kuriles (FROLOVA *et al.*, 1985; PERCHUK and FROLOVA, 1982); 8 = Shidra suite, Japan (KUNO, 1968); 9 = Morotu subvolcanics, Sakhalin (YAGI, 1953); 10 = South Sandwich Islands (FROLOVA and RUDNIK, 1974); 11, 12 = Hawaii (MACDONALD and KATSURA, 1964; MURATA, 1970; WINCHELL, 1947; THOMPSON and TILLEY, 1969); 13 = Galapagos Islands (MCBIRNEY and WILLIAMS, 1969; PERCHUK and FROLOVA, 1979). The symbol denotes coordinates of intersection of the curves (an average of the Earth's mantle after PERCHUK and VAGANOV, 1980). Numbers indicate the series discussed in the text.

lines can be described with the following two equations:

$$\text{Alkalis (weight percent)} = 44.35 - 0.84\text{SiO}_2; \quad r^2 = 0.954, \quad (1)$$

and

$$\text{Alkalis (weight percent)} = 39.06 - 0.68\text{SiO}_2; \quad r^2 = 0.910. \quad (2)$$

In order to understand the regularities of the magmatism in marginal sea floors, three basins (Okhotsk, Japan and Philippine sea floors) of different geodynamic history have been compared. The northernmost of these, the Okhotsk sea floor, is composed (~80 percent) of typical continental crust (Precambrian gneisses, Jurassic granites, etc.). Its thickness is about 50 percent of that of the adjacent continents. The crustal thickness in the South Okhotsk depression is 11–15 km, including a 4 km thick sediment layer (Neogene basalts and terrigenous material have been dredged from the floor). The basement of the Okhotsk sea floor (the Okhotsk dome) is covered by Upper Cretaceous (113–133 m.y.) and Paleogene (45–55 m.y.) rhyolites, andesites and basalts. An inverse volcanic sequence appears typical for this region (GNIBIDENKO and KHVEDCHUK, 1984). A regular change in volcanism from acidic to basic over the last 50–60 m.y. (Figures 3 and 4; see also Table 2) has also been suggested for the Japan sea floor (KONOVALOV, 1984; PERCHUK, 1985; PERCHUK *et al.*, 1985).

Basalts (10–15 m.y.; see also Table 2) with high Na, K, Ti and P contents and variable Al/Mg occur

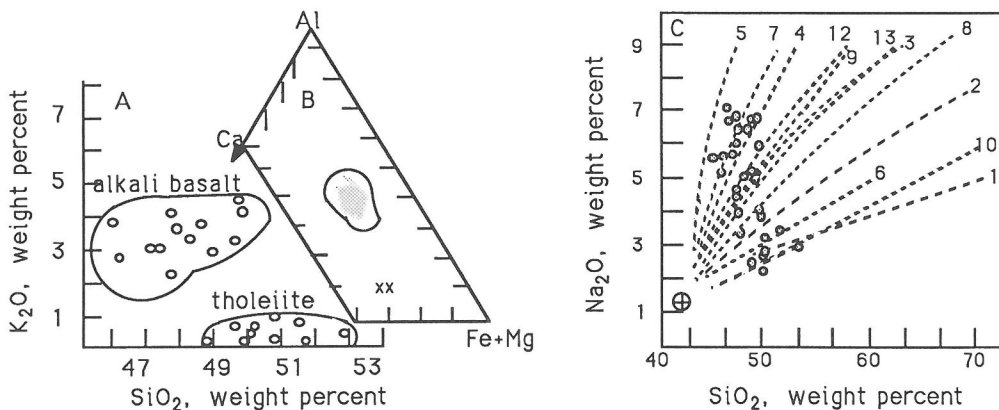


FIG. 2. Average values of petrochemical parameters for 23 nondifferentiated and slightly differentiated basalt complexes formed in continents, oceans and island arcs (see also Table 1). Symbols and trend numbers in Figure 2C as in Figure 1. Asterisks in Figure 2B indicate the average Yakutian kimberlite and composition of the upper mantle (PERCHUK and VAGANOV, 1980). Dashed lines from Figure 1.

Table 1. Some petrochemical parameters for nondifferentiated and slightly differentiated basalt complexes (after PERCHUK and FROLOVA, 1979; KEPHEZHINSKAS, 1979)

| No. | Province | Weight percent | | Mol ratio | | $a_{\text{SiO}_2}^{**}$ |
|---------------------------------|---|------------------|--------------------------------------|-----------|----------|-------------------------|
| | | SiO ₂ | Na ₂ O + K ₂ O | K/Na | ψ^* | |
| Mid-Atlantic Ridge | | | | | | |
| 1 | Rift valley, 30°N | 50.11 | 2.30 | 0.057 | 0.086 | 0.463 |
| 2 | Rift valley, 35°N | 50.35 | 3.35 | 0.122 | 0.122 | 0.469 |
| 3 | Red Sea Rift | 50.09 | 2.67 | 0.036 | 0.099 | 0.464 |
| 4 | Hess depression and the Galapagos Rift, Pacific | 48.96 | 2.55 | 0.045 | 0.099 | 0.453 |
| Varton Depression, Indian Ocean | | | | | | |
| 5 | Tholeiites | 51.06 | 2.58 | 0.067 | 0.067 | 0.472 |
| 6 | Alkali basalt | 51.10 | 4.75 | 0.194 | 0.166 | 0.475 |
| 7 | Jan Mayen Island, North Atlantic | 46.69 | 5.59 | 0.539 | 0.196 | 0.429 |
| Primorije, Far East, USSR | | | | | | |
| 8 | Shufan Plateau | 53.24 | 2.90 | 0.099 | 0.101 | 0.494 |
| 9 | Shkotov Plateau | 51.61 | 3.49 | 0.230 | 0.199 | 0.477 |
| 10 | Berezovskaya Formation: Ural | 49.72 | 4.93 | 0.143 | 0.181 | 0.458 |
| 11 | South Australia and Victoria | 47.75 | 5.75 | 0.319 | 0.207 | 0.437 |
| Eastern Siberia, USSR | | | | | | |
| 15 | Irkut River | 47.75 | 4.47 | 0.351 | 0.161 | 0.440 |
| 13 | East Sayan | 48.04 | 3.47 | 0.309 | 0.147 | 0.442 |
| 16 | Tuva | 49.21 | 5.24 | 0.517 | 0.175 | 0.455 |
| Mongolia | | | | | | |
| 12 | Jida | 48.44 | 6.58 | 0.369 | 0.231 | 0.446 |
| 14 | SW Mongolia | 46.02 | 5.18 | 0.382 | 0.231 | 0.458 |
| 17 | Potassium-rich formation | 47.96 | 6.79 | 0.434 | 0.237 | 0.434 |
| 18 | Sodium-rich formation | 45.99 | 5.60 | 0.314 | 0.210 | 0.419 |
| 21 | Hangay | 48.90 | 6.82 | 0.516 | 0.230 | 0.447 |
| 23 | Dariganga | 48.21 | 5.92 | 0.315 | 0.202 | 0.458 |
| 19 | Eastern Pacific Elevation | 49.74 | 3.91 | 0.166 | 0.142 | 0.462 |
| 20 | Kunlun and NW Tibet | 47.60 | 7.19 | 0.955 | 0.233 | 0.437 |
| 22 | Northern China | 48.21 | 5.92 | 0.432 | 0.206 | 0.443 |

* $\psi = (\text{Na} + \text{K})/\text{Si}$ ** Ideal mixing: $a_{\text{SiO}_2} \sim X_{\text{SiO}_2}$

in the Tsushima basin and on the submarine islands of Ulyndo and Chukto, whereas younger tholeiites (3–10 m.y.) are typical of the main Japan sea floor depressions (Figure 4). These tholeiites differ, however, from Mid-Ocean Ridge Basalts (MORB) by their high MgO content and higher K/Na (PERCHUK *et al.*, 1985). Almost all Japan Sea floor depressions are characterized (RODNIKOV *et al.*, 1982) by high heat flow ($2.75\text{--}2.92 \times 10^{-6}$ cal/cm² sec), seismic inhomogeneity, and the absence of a granitic layer.

Tholeiitic basalts are widely distributed in the Philippine sea floor (SHARASKIN, 1984; PEIVE *et al.*, 1980). Its western portion, *i.e.*, the Philippine depression, shows an east–west orientation of paleomagnetic and tectonic lineaments, whereas in the eastern portion of this sea floor these lineaments

are oriented in north–south directions. The eastern portion of the Philippine sea floor is also younger and has a complex morphology and tectonic setting; three submarine ridges may be relict island arcs. The geological structure of all three ridges is similar to submarine ridges on the Japan sea floor. They comprise calc-alkaline volcanic rocks as well as ophiolites, greenschists and amphibolites of Jurassic to Miocene age. In contrast to the Japan sea floor, boninites occur regularly in the volcanic sequences from ophiolite formations of the Philippine sea floor.

All marginal sea floors in southeast Asia show an inverse correlation between crustal thickness and the square root of their age (HAYES, 1984). An inter-arc spreading mechanism has been suggested to ex-

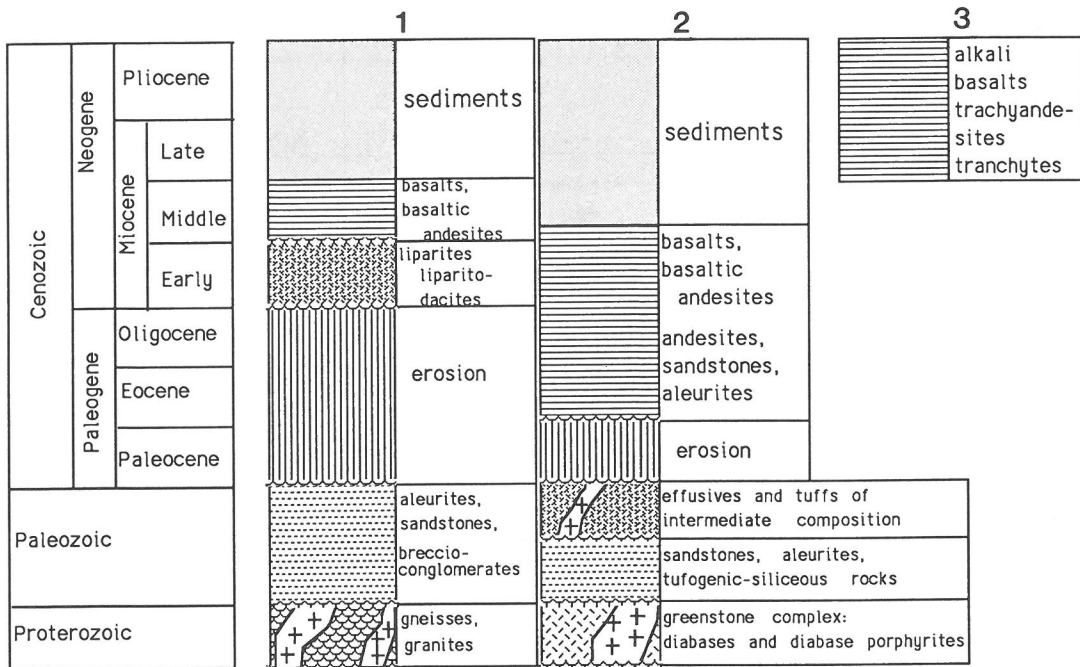


FIG. 3. Typical stratigraphic columns reflecting inverse volcanic sequences in morphostructures in the Japan sea floor (after KONOVALOV, 1984; PERCHUK *et al.*, 1985). 1 = sea floor margin; 2 = submarine ridges and islands (submarine arcs); 3 = depressions.

plain the origin of all marginal sea floors in this region (KARIG, 1974, 1975; HAYES, 1984; SHARASKIN, 1984).

Typical continental crust (gneisses, amphibolites, granites, etc.) is exposed in the basement of some island arcs. For example, the crustal thickness beneath the Ryukyu arc is about 20 km including a 10 km thick granitic layer. Its crust resembles the crust of the Okhotsk sea floor. The basement of the eastern Philippine sea floor, however, is composed principally of ophiolites, greenschists and amphibolites (MEIJER, 1976; PEIVE *et al.*, 1980).

Basalts resembling MORBs are the main constituents of the Philippine sea floor. However, some petrochemical variation is observed with differences between younger basalts and older typical oceanic tholeiites. For example, in the western Mariana depression, the basalts exhibit a "boninite tendency", with relatively high concentrations of SiO₂, MgO, Al₂O₃, Ba, Sr, Th and Ni (PEIVE *et al.*, 1980; SHARASKIN, 1984; KARPENKO *et al.*, 1984).

In order to explain the magmatic activity in marginal sea floors, three interrelated problems must be addressed. These are (1) the temperature and pressure of formation of primary (nondifferentiated) mantle-derived magmas, (2) the origin of boninites and related rocks, and (3) the origin of back-arc

basins and island arcs adjacent to oceanic trenches. For example, in order to support the idea of magmatic replacement of the Earth's crust by high-temperature, ultramafic magmas, it is necessary (a) to estimate the pressure-temperature constraints by the mantle on diapirism and magmatism in the mantle and the crust, (b) to find volcanic equivalents of the proposed mixed magmas, and (c) to interpret how back-arc depressions are related to the growth of igneous crust in island arcs. In view of the limited field documentation, available experimental data and the absence of adequate numerical models of the proposed processes, I will discuss only a petrological hypothesis for the origin of some marginal sea depressions.

BASALT SERIES: PRESSURE-TEMPERATURE CONTROL BASED ON GEOTHERMOMETRY AND GEOBAROMETRY

Xenoliths in alkali basalts

Mafic and ultramafic inclusions are found in many oceanic and continental alkali basalts. Because these inclusions are xenoliths, geothermometers and geobarometers provide only estimates of the minimum pressure-temperature conditions of the basalt magma generation. Mineral compositions

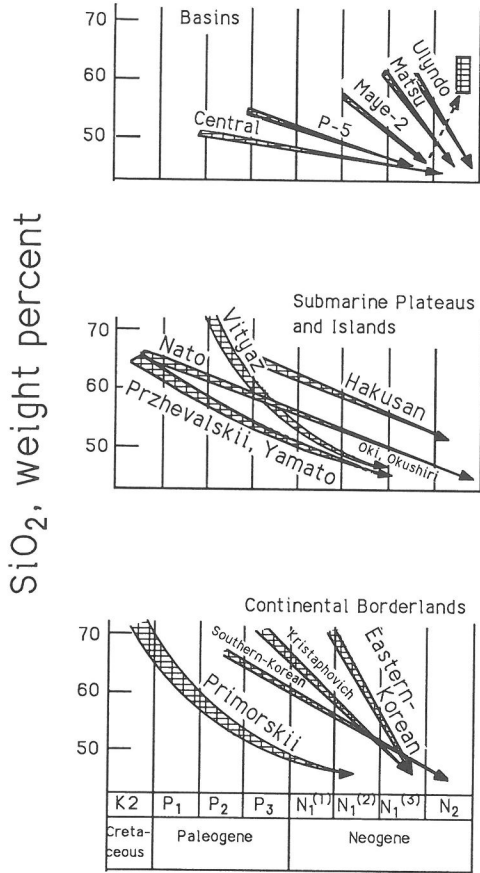


FIG. 4. Time-composition paths illustrating inverse sequence of volcanism in the Japan sea floor. Arrows summarize petrochemical data for each area. Data sources as in Figure 3; see also text for additional information.

Table 2. Age of the Japan Sea submarine plateaus (K-Ar method) (after KONOVALOV, 1984).

| Plateau | Rocks | Age (m.y.) |
|-------------------------|--------------------------|--------------------|
| Vityaz | Basaltic andesite | 16 |
| | Andesitic dacite | 32 |
| | Rhyodacite | 24 |
| | Rhyodacite | 53.2 |
| Yamato | Basalt | 35, 34, 32, 23, 22 |
| | Basaltic andesite | 46, 34.5, 34, 32 |
| | Andesite | 49, 54, 61.1 |
| | Dacite | 46 |
| Korean Sea floor margin | Basalt | 15.5 |
| | Trachyrhyolite | 17 |
| | Rhyodacite tuff | 23.4 |
| | Tuff of acid composition | 53.2 |
| Bogorov | Basalt | 52.7, 17.9, 6.5 |
| | Rhyolite | 49.5 |

of xenoliths in alkali basalts are in places notably different from those in kimberlites, and the two-pyroxene thermometer calibrated originally by DAVIS and BOYD (1966) at 30 kbar cannot be used. Instead, for the high-temperature range I have used experimental data (AKELLA and BOYD, 1973, 1974; GREEN and RINGWOOD, 1966; HENSEN, 1973), and for the low-temperature range, estimates for the same equilibria from PERCHUK (1967, 1968, 1969, 1977c) have been employed in order to improve on the two-pyroxene thermometer (PERCHUK, 1977a). The empirical form of a modified version of this thermometer is (PERCHUK, 1977a) (all symbols in this and subsequent equations are defined in Table 3):

$$T(K) = 10^3 / (0.4305 - 0.1651 \ln K_{FeO} + 1.1071 X_{Mg} - 1.0525 X_{Mg}^2 - 0.0304 X_{Mg}^2 \ln K_{FeO}), \quad (3)$$

Table 3. Symbols used in formulae and equations

| | |
|---------------------|---|
| $N_{Al_2O_3}^{Opx}$ | Weight percent Al_2O_3 in orthopyroxene |
| X_i^θ | mol fraction of component, i , in phase θ |
| a_i^θ | activity of component, i , in phase θ |
| ΔG | Gibbs free energy change in a reaction |
| K_D | distribution coefficient |
| K_i | the partition coefficient of component i between two phases α and β ; $K_i = X_i^\alpha / X_i^\beta$ |
| P | pressure, bar (kbar) |
| T, K | temperature, K |
| $T, ^\circ C$ | temperature, $^\circ C$ |
| R | gas constant, 1.987 cal/K mol |
| ΔV | volume change in a reaction |
| V_m^i | partial molar volume of component i |
| V_m^e | excess molar volume of component i |

Abbreviations used in the text and figures

| | |
|------|-----------------|
| Ab | Albite |
| Ak | Akermanite |
| Am | Amphibole |
| An | Anorthite |
| Bi | Biotite |
| CaTs | Ca-tschermakite |
| Cpx | Clinopyroxene |
| Di | Diopside |
| En | Enstatite |
| Fo | Forsterite |
| Gr | Garnet |
| Ks | Kalsilite |
| Ne | Nepheline |
| Ol | Olivine |
| Opx | Orthopyroxene |
| Py | Pyrope |
| Qz | Quartz |
| Usp | Ulvospinel |

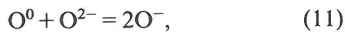
Systems

| | |
|------|------------------------------|
| KFQ | Ks-Fo-Qz |
| NFQ | Ne-Fo-Qz |
| CMAS | CaO-MgO- Al_2O_3 - SiO_2 |

The liquidii of basalts generated in the mantle and the crust

Since ROEDER and EMSLIE (1970), the liquidus temperature of basalt saturated with olivine at 1 bar pressure could be estimated. The influence of alkalis on the liquid–olivine equilibria has not been studied experimentally. The activity of an olivine component in liquids as a function of the activity coefficient for Na has been derived from systems involving fayalite, albite and nepheline (PERCHUK and VAGANOV, 1977). However, those equations are valid only for a relatively narrow range of liquidus temperatures and bulk compositions. Recently, GHIORSO and KELEMEN (1987) summarized data and completed their version of the super-liquidus phase relations of magmatic rocks.

Because of the high liquidus temperatures of komatiite, basaltic komatiite and boninite, thermodynamic treatment of olivine–liquid equilibria for temperatures in the range 1350–1650°C is necessary. Figure 5 has been constructed from a thermodynamic treatment of all available phase diagrams involving olivine on the liquidus at 1 bar (see APPENDIX for sources, and also references in PERCHUK and VAGANOV, 1977; PERCHUK, 1984a; LEEMAN, 1978). The diagram makes use of the oxygen model for silicate melts (see review by BOTTINGA *et al.*, 1970):



where O^0 is bridging oxygen (*i.e.*, Si–O–Si bridge), O^{2-} is free oxygen (*i.e.*, Me–O–Me, where Me is a metal cation not in tetrahedral coordination) and O^- is nonbridging oxygen (*i.e.*, Si–O–Me). The abscissa in Figure 5 indicates the fraction, Y , of oxygen (nonbridging) associated with Fe^{2+} and Mg^{2+} in the melt. Least squares regression of Y versus temperature (153 data points) yields the expression (see APPENDIX; Table A1, for data sources)

$$T(^{\circ}C) = 1089 + 1634Y; \quad r^2 = 0.988. \quad (12)$$

Equation (12) is valid for all systems in the composition range $Y = 0-0.5$.

For volatile-free systems with potassium contents ≥ 5 weight percent, the liquidus temperatures have been estimated ($\pm 25^{\circ}C$) with the equation:

$$T(^{\circ}C) = 1047 + 44X - 13.77X^2 + Y(314.97 + 1941.55X - 630.69X^2), \quad (13)$$

where,

$$X = X_{Mg}^{Ol} = 3.33X_{Mg}^{rock} / (1 + 2.33X_{Mg}^{rock}),$$

$$X_{Mg}^{rock} = Mg / (Mg + Fe + Mn),$$

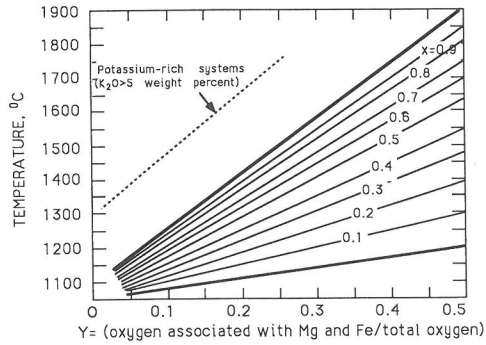


FIG. 5. Temperature-dependence of the structure-chemical parameter, Y , for Fe–Mg systems with olivine on the liquidus at $P = 1$ bar. Dashed line is calculated from the data of SCHAIRER (1954) and WENDLANDT and EGGLEER (1980) for the system Fo–Ks–Qz. Other data sources are listed in the APPENDIX.

and

$$Y = \frac{\text{oxygen associated with Mg and Fe}}{\text{total oxygen}}$$

At $K_2O \geq 5$ weight percent, the activity coefficient for MgO increases rapidly, and olivine appears on the liquidus at a temperature about $210^{\circ}C$ higher than that predicted from Equation (13). This effect is so pronounced that the activity coefficient of MgO becomes practically independent of the potassium content in the system up to $K_2O = 16-17$ weight percent. For iron-free, potassium-rich systems, the liquidus temperature may be expressed as a linear function of Y (see Figure 5 and APPENDIX; Table A2):

$$T(^{\circ}C) = 1294 + 1690.84Y; \quad r^2 = 0.955; \quad \text{accuracy} = \pm 30^{\circ}C. \quad (14)$$

This inference is of great importance when using Equations (13) and (14) as petrological thermometers. The effect of Na_2O is not yet known.

Equation (13) was deduced for $P = 1$ bar. From the experimental data of DAVIS and ENGLAND (1963), the pressure-dependence of the forsterite liquidus temperature may be approximated with the equation;

$$T(^{\circ}C) = 1898 + 4.77P(\text{kbar}), \quad (15)$$

whereas for fayalite (HSU, 1967), the expression is:

$$T(^{\circ}C) = 1205 + 4.85P(\text{kbar}). \quad (16)$$

The temperatures of the melting curves for basalt compositions with olivine on the liquidus (typically up to pressures near 10 kbar) show a relatively constant pressure derivative, $(\partial T / \partial P) = 4.08^{\circ}C / \text{kbar}$

(PERCHUK, 1983; see also APPENDIX; Table A3). It is suggested, therefore, that in the pressure range where olivine is on the liquidus, the pressure dependence of the liquidus temperature can be approximated with the expression:

$$T_P(^{\circ}\text{C}) = T_{1\text{ bar}}(^{\circ}\text{C}) + 4P(\text{kbar}). \quad (17)$$

There are two unknown parameters in Equation (17)—pressure and temperature. Methods are required to calculate the values of these variables. One method was proposed by PERCHUK (1984b) and PERCHUK *et al.* (1982). The approach is based on the observation (KUSHIRO, 1975; PERCHUK, 1973; p. 269), also postulated by KORZHINSKII (1959), that for volatile-free silicate systems the shift in liquidus boundaries with pressure is correlated with the radius of additional cations in the system. The data in Figure 2C may indicate a decrease of silica activity in the parent magma as the alkali activity increases. This observation is supported by thermodynamic calculations (NICHOLLS and CARMICHAEL, 1972; NICHOLLS *et al.*, 1971).

Silicate melts are not ideal mixtures (*e.g.*, CARMICHAEL *et al.*, 1977; PERCHUK and VAGANOV, 1977; NAVROTSKY, 1980; GHIORSO *et al.*, 1983). NICHOLLS (1977) and DEPAOLO (1979) have shown, however, that for simple silicate systems such as MgO-SiO₂, the following inference is valid:

$$\left(\frac{\partial \ln a_{\text{SiO}_2}}{\partial T}\right)_P \approx 0. \quad (18)$$

Some, but not all, silicate melts may, therefore, be approximated as ideal solutions. The validity of this inference is also supported by constant a_{SiO_2} along isobaric liquidus boundaries between forsterite and enstatite in the systems forsterite-kalsilite-silica (Fo-Ks-Qz) and forsterite-nepheline-silica (Fo-Ne-Qz) at high pressure (KUSHIRO, 1968, 1980) and at atmospheric pressure (SCHAIRER, 1954). From the data reported by KUSHIRO (1968, 1980), it is inferred that:

$$\text{Fo-Ks-Qz at 30 kbar } a_{\text{SiO}_2}^L \approx 0.47; \text{ and}$$

$$\text{Fo-Ne-Qz at 30 kbar } a_{\text{SiO}_2}^L \approx 0.49.$$

At other pressures, the SiO₂ activity varies, and the higher the alkali content, the greater the effect of pressure on the shift of the enstatite-forsterite (En-Fo) liquidus boundaries. The change in SiO₂ mol fraction with pressure at constant Na₂O and K₂O ($X_{\text{alkalies}}^L = 0.12$) can be described with the linear equations for the systems

(1) Fo-Ks-Qz:

$$-\ln K_{\text{SiO}_2}^L = 0.39 - 0.0123P(\text{kbar}), \quad (19)$$

(2) Fo-Ne-Qz:

$$-\ln K_{\text{SiO}_2}^L = 0.49 - 0.0167P(\text{kbar}). \quad (20)$$

In such a case the activity coefficient of silica is considered independent of pressure and the partial molar volume of SiO₂, $(V_{\text{SiO}_2}^m)^L$ can be estimated. By taking into account Equation (18), integration of

$$RT \ln a_{\text{SiO}_2}^L = \int_0^P (V_{\text{SiO}_2}^m)^L dP, \quad (21)$$

at $X_{\text{alkalies}}^L = 0.12$ and $T = 1500^{\circ}\text{C}$ gives the following average value for both the system Fo-Ne-Qz and Fo-Ks-Qz:

$$(V_{\text{SiO}_2}^m)^L = V_{\text{SiO}_2} - V_{\text{SiO}_2}^0 = -0.051 \text{ cal/bar}. \quad (22)$$

This value is much larger than that calculated with Equation (5) of DEPAOLO (1979) for the alkali-free system MgO-SiO₂ under the assumption that $\ln a_{\text{SiO}_2}^L$ is a linear function of pressure:

$$(V_{\text{SiO}_2}^m)^L = V_{\text{SiO}_2} - V_{\text{SiO}_2}^0 = -0.012 \text{ cal/bar}, \quad (23)$$

where in Equations (22) and (23), V_{SiO_2} is the partial molar volume of SiO₂ in the melt (BOTTINGA and WEILL, 1970) and $V_{\text{SiO}_2}^0$ is the molar volume of molten SiO₂ at the same temperature. Thus, Maxwell's relation for the alkali-bearing systems,

$$\left(\frac{\partial \ln a_{\text{SiO}_2}^L}{\partial P}\right) = -1.23 \times 10^{-5}, \quad (24)$$

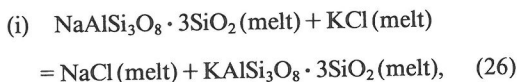
leads to the conclusion that with increasing pressure, addition of alkalis to volatile-free systems results in a large decrease of silica activity and forces the En-Fo boundary to shift toward alkali-rich compositions. There are also significant deviations from ideality in alkali-bearing silicate melts. Nonideal behavior of these silicate solutions has been demonstrated at 1 bar pressure both in terms of the random network model (PERCHUK and VAGANOV, 1977, 1980; LEEMAN, 1978) and the Temkin model of silicate melt structure (HERZBERG, 1979). The use of ESIN's (1946) polymer theory-based model allows us to avoid excess enthalpy of mixing.

In order to calibrate the pressure-dependence of the silica activity, isobaric cross-sections from the system MgO-SiO₂-Al₂O₃-(Na, K)₂O should be studied in detail. Only a limited number of experimental data is, however, available for thermodynamic treatment of the systems Fo-Ks-Qz and Fo-Ne-Qz (KUSHIRO, 1965, 1980; WENDLANDT and EGGLE, 1980; MODRESKI and BOETTCHER, 1973; SCHAIRER, 1954; IRVINE, 1976). As was shown above, the alkali/silica is a good indicator of the pressure effect on the peritectic in Fe-Mg silicate systems. This effect can be roughly calibrated under

the assumption of statistically equal association of free oxygen with Na and K in the silicate melt, which, in turn, can be described with the internal distribution coefficient:

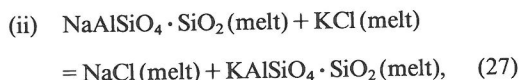
$$K_D = \left(\frac{a_{K_2O}^L}{a_{SiO_2}^L} \right)_{FoKsQz} \left(\frac{a_{SiO_2}^L}{a_{Na_2O}^L} \right)_{FoNeQz} \approx 1.0. \quad (25)$$

This assumption is based on experimental data on sodium and potassium distribution in haplogranite melt–chloride melt, haplophonolite melt–chloride melt and haplophonolite melt–chloride + water fluid (ZYRIANOV, 1987; ZYRIANOV and PERCHUK, 1978; PERCHUK and LINDSLEY, 1982). The following equilibria have been studied under the pressure–temperature conditions of interest:



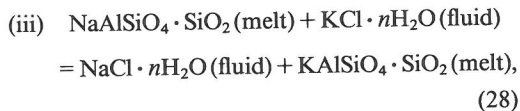
$$1100^\circ\text{C}, 6 \text{ bar}, K_D = 0.78$$

$$(n = 10 \text{ data points}; \sigma = \pm 0.102)$$



$$1300^\circ\text{C}, 6 \text{ bar}, K_D = 1.24$$

$$(n = 7 \text{ data points}; \sigma = \pm 0.177)$$



$$1100^\circ\text{C}, 5 \text{ kbar}, K_D = 1.146$$

$$(n = 10 \text{ data points}; \sigma = \pm 0.207)$$

where;

$$K_D = \left(\frac{X_K}{1 - X_K} \right)_{\text{silicate}} \left(\frac{1 - X_K}{X_K} \right)_{\text{chloride}}, \quad (29)$$

where X_K is the mol fraction of potassium.

Equilibria (i) and (ii) were studied in sealed quartz tubes at high temperature with a calculated internal pressure of approximately 6 bar. Equilibrium (iii) was studied by using sealed platinum containers at 5 kbar in an internally–heated, gas–medium apparatus. Silicate glasses in the system Ne–Ks–Qz were prepared from oxides, melted at 1200°C and examined with the electron microprobe for chemical homogeneity. Experimental charges were similarly studied. The K/Na of coexisting glasses, salts and fluids was determined by flame photometry.

The K_D equals 0.78 for the silica–rich composition, (i), whereas for the silica–poor compositions,

(ii) and (iii), $K_D \approx 1.2$. To a first approximation, these data permit the approximation of ideal mixing of Na and K in silicate melt systems.

By assuming that K_D in Equation (25) is independent of pressure and temperature, the pressure–dependence of $\ln(a_{Na_2O}^L/a_{SiO_2}^L)$ for the systems Fo–Ne–Qz and Fo–Ks–Qz, respectively, has been calculated for pressures corresponding to constant silica activity. From Equation (24) the position of the Fo–En boundary in the system Na_2O – K_2O – MgO – SiO_2 – Al_2O_3 has been calculated (Figure 6). The values for the calculated parameters are similar to those given in Table 1 and Figure 2C for natural magmatic liquids. By taking into account Equation (18), the lines in Figure 6 can be described with the following expressions:

$$P(\text{kbar}) = 51.3(\Psi/a_{SiO_2}^L - 0.017), \quad (30)$$

where,

$$\Psi = \frac{a_{(Na,K)_2O}^L}{a_{SiO_2}^L}, \quad (31)$$

to yield the expression:

$$a_{SiO_2}^L = \sqrt{\frac{a_{(Na,K)_2O}^L}{0.0195P - 0.017}}. \quad (32)$$

Equations (30) and (32) show the dependence of the liquidus pressure on the alkali content of a melt in equilibrium with olivine and orthopyroxene.

The equations above are analytical forms of the regularities in liquidus phase relations predicted by KUNO (1960). Despite the fairly large uncertainty ($\pm 30\%$ compared with the input data), Equations (30) and (32) could be used to calculate the last

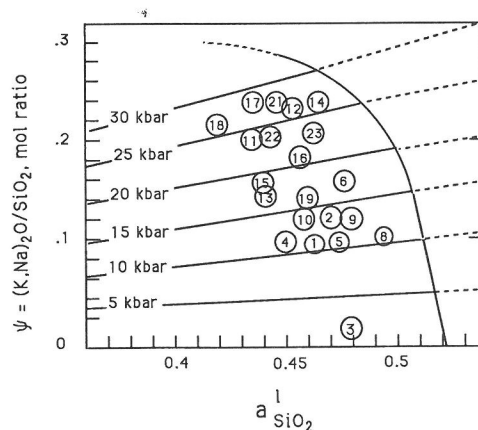
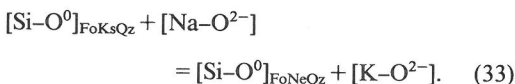


FIG. 6. Pressure effect on interrelation between composition parameters in the volatile–free system MgO – $(Na, K)_2O$ – SiO_2 – Al_2O_3 – SiO_2 . Solid line: Theoretically calculated shift of the incongruent melting point as a result of pressure. Numbered circles correspond to the numbers in Table 1. For other details, see text.

term in Equation (17), deduced for volatile-free conditions.

The value of K_D in Equation (25) is based on data collected for simple systems. In order to approximate a mixing model for natural magmatic compositions, the following internal oxygen exchange equilibrium in the melts (equilibrated with a solid residue in the mantle) should be considered:



In this expression, $[\text{Si-O}^0]_{\text{FoKsQz}}$ and $[\text{Si-O}^0]_{\text{FoNeQz}}$ are activities of bridging oxygen in the systems Fo-Ks-Qz and Fo-Ne-Qz, respectively, whereas $[\text{K-O}^{2-}]$ and $[\text{Na-O}^{2-}]$ are activities of free oxygen associated with potassium and sodium in the melts. Because it is assumed that the activities of bridging oxygen in Equation (33) are the same, the equilibrium constant is:

$$K_{(33)} = \frac{[\text{Na-O}^{2-}]}{[\text{K-O}^{2-}]} = \exp\left(\frac{\Delta G_{(33)}}{RT}\right)_P. \quad (34)$$

The volume change of reaction (33) can be calculated from partial molar volumes of K_2O and Na_2O in silicate melts (BOTTINGA and WEILL, 1970; STEBBINS *et al.*, 1984). The value of $\Delta V_{(33)}$, calculated per atom of K and Na, is 0.204 cal/bar at 1400°C, which results in:

$$\left(\frac{\partial \ln K_{(17)}}{\partial P}\right)_{T=1673} = -6.13 \times 10^{-5}. \quad (35)$$

This value is quite similar to the value that can be calculated, to a first approximation, by comparing K/Na of volcanic rocks. For example, in mid-ocean ridge basalts, $(\text{K}_2\text{O}/\text{Na}_2\text{O}) = 0.056$; whereas in kimberlite this ratio is near 5.54 (see Table 1). If it assumed that the mid-ocean ridge basalts were formed at pressures near 10 kbar, kimberlite would be formed at $P = 75\text{--}80$ kbar (PERCHUK, 1977b). By assuming linear relations between $\ln K_D$ and pressure and that basalt and kimberlite were formed at the same temperature, the following value of the derivative has been calculated:

$$\left(\frac{\partial \ln K_{(17)}}{\partial P}\right)_T = -6.23 \times 10^{-5}. \quad (36)$$

The value from Equation (36) is surprisingly similar to that of Equation (35). Consequently, in addition to Equation (30), we can use this large effect of pressure on initial magma composition as a petrological barometer. By using 25 data points from Table 1, the following expression has been derived:

$$P(\text{kbar}) = 40.57 + 21.33\psi + 3.794\psi^2, \quad (37)$$

where $\psi = \ln(\text{K}/\text{Na}) \simeq 1/K_{(33)}$ in basalt (molar ratio). The pressures obtained with Equations (30) and (37) are in accord to within ± 4.3 kbar.

The least-squares fitted lines in Figures 1 and 2C intersect at $\text{SiO}_2 = 42.5 \pm 1.2$ weight percent. In the concentration range between 42.5 and 54 weight percent SiO_2 , each trend could be approximated by a straight line. From Equations (30) and (37), the following empirical expression has been obtained:

$$P(\text{kbar}) = 9.13 + 20.76\alpha, \quad (38)$$

where

$$\alpha = (\text{Na}_2\text{O} + \text{K}_2\text{O} - 1.25)/(\text{SiO}_2 - 42.5). \quad (39)$$

Equation (38) is useful for pressure estimates of the most mafic members of differentiated basalt series as discussed further below.

The deduced equations imply an empirical correlation between the alkalinity of a cotectic basalt and degree of melting of the mantle material. This effect is several orders of magnitude greater than any other variation. Equations (30), (36) and (37) were deduced on the basis of phase relations in portions of the system $\text{Na}_2\text{O}\text{--}\text{K}_2\text{O}\text{--}\text{MgO}\text{--}\text{Al}_2\text{O}_3\text{--}\text{SiO}_2$. It is, therefore, possible that their application to calculate the pressure of formation of magmatic liquids in the mantle may be restricted. Any significant variation in bulk compositions of the mantle-derived basalt will affect the value of the calculated pressure. It is likely, though, that these equations will provide useful estimates of relative pressures of formation of basaltic magmas with different alkalinity.

Pressure estimates from Equations (30) and (37) can be compared with results from geobarometry from xenoliths from alkali basalts and kimberlites (Tables 4 and 5). Because the xenoliths are not genetically related to the host magmas, the pressures estimated from their mineral assemblages are minimal values insofar as the magma is concerned. Average compositions from Kilauea and Mauna Kea are also included in Table 5, where the average depth of their magma chambers is known from seismic data. Two volcanic samples from Mongolia are also included (KEPENZHINSKAS, 1979), known as sodium-rich and potassium-rich basalts. The first group is characterized by spinel-bearing xenoliths (Solhitin Volcano), whereas the second group is characterized by magma-types found in the Shavarin Volcano.

As can be seen from the data in Table 5, the pressure estimates obtained with the three different methods are in fair agreement. As expected, the

Table 5. Chemical analyses (recalculated to 100 weight percent) and some compositional parameters of volcanic rocks containing deep-seated ultramafic xenoliths

| Country | USA, Hawaii | | | Mongolia | | Tanzania |
|--------------------------------------|----------------|---------------------|----------------------------|-----------------|---------------|----------|
| | Kilauea (1) | Mauna Kea (2) | Salt Lake, Oahu, (3) | Solhitin (4) | Tsaram (5) | (6) |
| SiO ₂ | 49.82 | 47.48 | 40.82 | 49.61 | 50.86 | 41.39 |
| TiO ₂ | 2.57 | 2.68 | 2.54 | 3.12 | 2.47 | 2.49 |
| Al ₂ O ₃ | 12.87 | 14.17 | 11.66 | 9.98 | 15.99 | 6.13 |
| Fe ₂ O ₃ | 2.06 | 4.05 | 2.52 | 5.75 | 2.37 | 9.67 |
| FeO | 9.51 | 8.38 | 12.41 | 7.43 | 7.57 | 4.82 |
| MgO | 10.09 | 9.59 | 13.02 | 9.62 | 5.86 | 18.52 |
| MnO | 0.17 | 0.18 | 0.20 | 0.20 | 0.14 | 0.18 |
| CaO | 10.22 | 9.83 | 13.50 | 9.21 | 6.05 | 12.85 |
| Na ₂ O | 2.20 | 2.71 | 2.62 | 3.91 | 4.84 | 2.07 |
| K ₂ O | 0.50 | 0.92 | 0.71 | 1.15 | 3.84 | 1.04 |
| P ₂ O ₅ | — | — | — | — | — | 0.80 |
| Na ₂ O + K ₂ O | 2.70 | 3.64 | 3.33 | 5.06 | 8.68 | 3.11 |
| α | 0.198 | 0.479 | -1.24 | 0.535 | 0.89 | -1.69 |
| a_{SiO_2} | 0.462 | 0.439 | 0.374 | 0.458 | 0.463 | 0.377 |
| $\chi_{\text{FeO}}^{\text{ol}}$ | 0.838 | 0.823 | 0.839 | 0.817 | 0.779 | 0.889 |
| T °C at 1 atm | 1265 | 1258 | 1406 | 1265 | 1193 | 1405 |
| P , kbar* | 12 | 17 | 20 | 15 | 20 | 17-20 |
| (Na, K)/Si | 0.098 | 0.136 | 0.146 | 0.82 | 0.280 | 0.129 |
| P , kbar† | 12 | 17 | 21 | 21 | 32 | 18 |
| $\ln(K/Na)$ | -1.48 | -1.07 | -1.224 | -1.22 | -0.232 | -0.688 |
| P , kbar‡ | 14 | 17 | 15 | 16 | 28 | 19 |
| T , °C at P | 1316 | 1326 | 1407 | 1340 | 1313 | 1485 |

* Pressure estimated with geobarometers or seismic data.

† Pressure estimated with Equation (30).

‡ Pressure estimated with Equation (37).

pressures from the ultramafic xenoliths are slightly lower than the pressure values obtained with either Equation (30) or (37).

APPROXIMATE PRESSURE-TEMPERATURE CONDITIONS OF INITIAL MAGMA FRACTIONATION

The equations derived above may be used to estimate the approximate depth of magma chambers and their thermal regime during fractionation. A few examples of such estimates will be provided.

Two volcanic series (alkali basalt and tholeiite) are known in the Little Kurile arc north of Hokkaido. The island arc is bounded by the South Okhotsk depression and Kurile-Kamchatka trench to the northwest and southeast, respectively. The arc comprises two subparallel island chains divided by a basin. The frontal arc consists of three islands—Zeleny, Polonsky and Panfilov composed of volcanics with common trachybasaltic sills. Radiometric dating of the rocks yields ages in the range 50–68 m.y. The volcanic rocks are younger than the host flysch (FROLOVA *et al.*, 1985) and occur

as thick layered sills (olivine-bearing gabbro-norites, teshenites, monzonites and syenites) and as small bodies and dikes composed of trachybasalt and trachydolerite. These rocks make up one alkali basalt series denoted 7 in Figure 1. The depth of the initial magma chamber is estimated [by Equation (37)] to be near 70 km (Figure 7). The calculated temperature interval of fractionation is about 150°C [from Equation (17)].

Tholeiite series from the Tyatya volcano (Kunashir Island in the inner portion of the Little Kurile arc) was generated at a depth of about 40 km (see Figure 7). In spite of the fact that the calculated temperature interval of fractionation was only about 80°C [from Equation (17)], the series is well differentiated, possibly in a magma chamber located directly beneath the island arc crust. This conclusion is consistent with the reconstruction of the lithosphere beneath Kunashir Island based on recent seismic data (ZLOBIN *et al.*, 1982; FROLOVA *et al.*, 1985) suggesting a magma chamber near the crust-mantle boundary at 40–60 km depth. Similar volcanism is typical of oceanic islands such as Hawaii

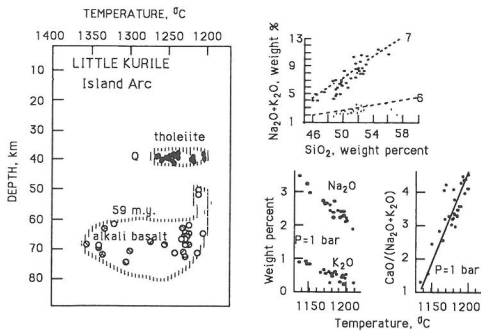


FIG. 7. Approximate depth and temperature regime of differentiation of the Paleogene (59 m.y.) alkali basalts and Holocene (Q) tholeiites from the Little Kuriles. Left-hand diagram shows temperatures of magma generation calculated with Equation (17) for appropriate pressures from Equation (37). The temperature regime of differentiation of tholeiite basalts is calculated with Equation (17).

(WINCHELL, 1947; MACDONALD and KATSURA, 1964) and continents.

The origin of boninite, characteristic of some ancient and present-day island arcs, is considered next. It has been suggested that boninite and komatiite series exhibit kindred petrology, geochemistry and very high temperature melt inclusions. Furthermore, in some greenstone belts, both series are located within ophiolite formations (CAMERON *et al.*, 1979; NESBITT *et al.*, 1979). According to existing views (PEIVE *et al.*, 1980; SHARASKIN, 1980; MEIJER, 1980, 1983; CAMERON *et al.*, 1979, 1983; SOBOLEV *et al.*, 1986) and also suggested from experimental data (KUSHIRO, 1972; GREEN, 1976; TATSUMI, 1987), boninite could be a product of partial melting of ultramafic upper mantle material at high water pressures. However, recent studies on melt inclusions in olivine and orthopyroxene show low contents (1–3 weight percent) of primary water (SOBOLEV *et al.*, 1986; DANUSHEVSKII *et al.*, 1986; SOBOLEV and DANUSHEVSKII, 1986) with homogenization temperatures in the range 1090–1430°C. Earlier, WALKER and CAMERON (1983) suggested that parent boninite magma in equilibrium with Fo_{94} most likely would have been generated at 1250–1300°C at $P < 10$ kbar and with relatively low water contents (2–3 weight percent).

Compositions of boninites from some western Pacific and Australian island arcs are plotted in Figure 8. The stratigraphic position of boninites in ophiolite formations suggests fractionation of the initial magma to produce a differentiated series beginning with olivine-rich, enstatite-bearing products and ending with olivine-free, high-silica and low-alkali melts. Figure 8 shows, however, a regular change in boninite composition with respect to Mg,

Al and Si, whereas there is no correlation between silica and alkalis. Boninite occupies the field between ultramafics and dacite-andesite in Figure 8. Similar relationships have been demonstrated by SAKUYAMA (1979, 1981) for some calc-alkaline series in the Japanese island arcs. This relationship suggests that perhaps boninite might be a product of interaction between ultramafic (harzburgitic or komatiitic?) magma and crustal rocks, where the latter were melted before mixing with the postulated, mantle-derived ultramafic magma. The lack of correlation of alkalis with silica might be the result of such a process. In other words, this possible interaction should lead to thermal erosion of the Earth's crust and magma mixing rather than forming these rock series through fractional crystallization. In order to support this possibility, we have to show (1) that boninite and related rocks have a shallow origin, (2) very high temperature of magma mixing, (3) absence of correlation between alkalis and silica, and (4) mixed sources of compatible and incompatible elements. Evidence relating to the first three requirements is shown in Figure 9.

Let us consider the abundance of compatible and incompatible trace elements in boninites and related rocks. Boninite series rocks are characterized by the following geochemical features (HICKEY and FREY, 1982; KARPENKO *et al.*, 1984; SHARASKIN, 1984; ARMSTRONG and NIXON, 1981; DEPAOLO and WASSERBURG, 1977):

- (1) $54 < SiO_2 < 59$ weight percent at $X_{Mg} > 0.5$ with Mg, Ni, Co and Cr contents typical of primary mantle-derived magmas.

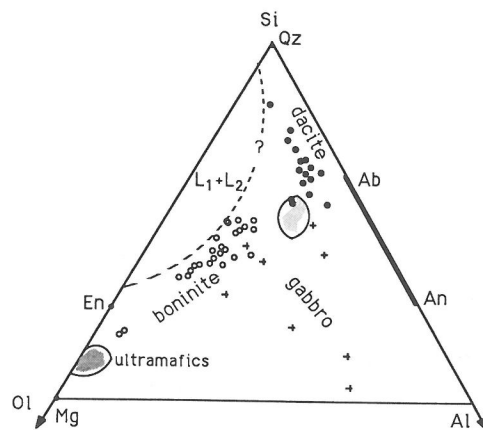


FIG. 8. Compositional variations in boninite series (open circles) ophiolites and calc-alkaline rocks from Bonin Island, the Mariana Trench (Philippine Sea) and Cape Vogel, Papua, New Guinea (WALKER and CAMERON, 1983; PEIVE *et al.*, 1980). Dashed lines = likely immiscibility gap. Small dots = composition of glasses from boninites.

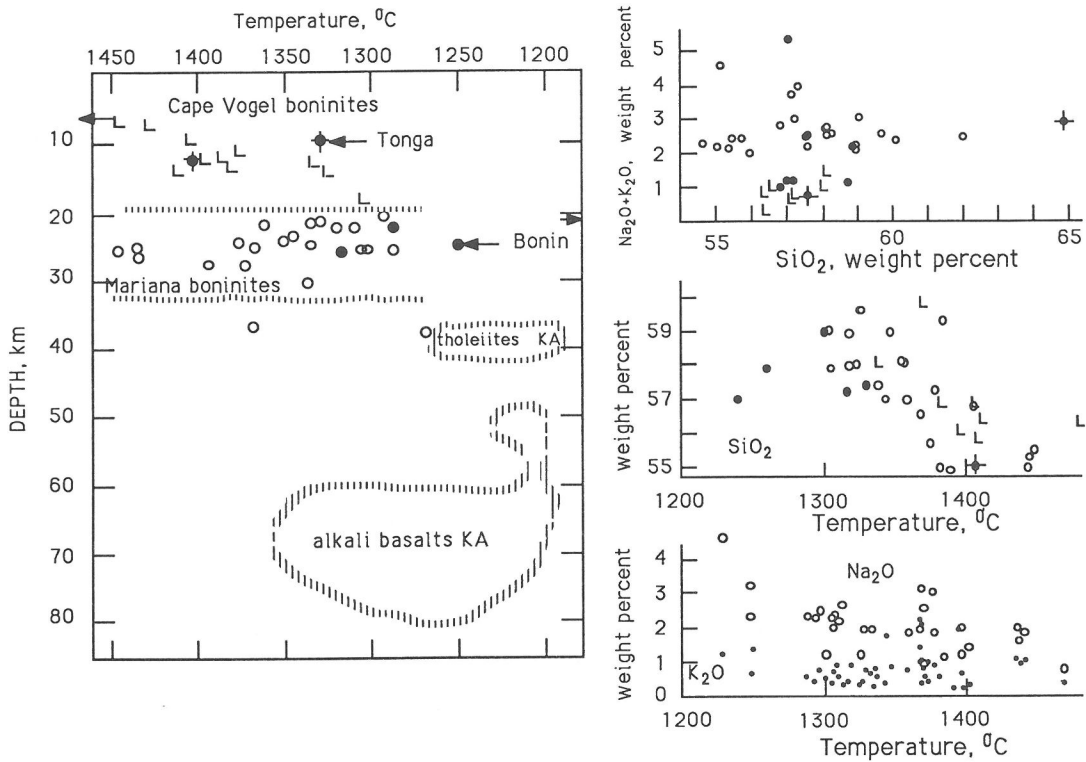


FIG. 9. Temperature regime of boninite series formation in the Earth's lower crust from contemporary island arcs calculated under the assumption of volatile-free conditions with Equations (17) and (30). Note the absence of silica versus alkali correlation in the initial magma in the temperature range 1220–1460°C. Symbols as in Figure 8. KA = Kurile arc (from Figure 7).

(2) Abundance of alkalis, Ba and Sr typical of island arc rocks.

(3) Variations in $^{143}\text{Nd}/^{144}\text{Nd}$ between those of chondrite and MORBs. The $^{143}\text{Nd}/^{144}\text{Nd}$ shows a good inverse correlation with Zr/Ti and $^{87}\text{Sr}/^{86}\text{Sr}$ and positive correlation with Sm/Nd and La/Sm .

(4) High and variable Zr/Ti and variable HREE/LREE.

(5) Correlation between Ti and Th as well as between $^{143}\text{Nd}/^{144}\text{Nd}$ and the abundance ratios, Sm/Nd and Ti/Zr .

(6) A wide variation of Ti/Zr and La/Yb .

Many models have been suggested to explain this variable geochemistry of boninites (e.g., metasomatic alteration of the mantle before its partial melting, partial melting of a subducted, lithospheric slab). The geochemical data from ophiolite complexes (involving boninites) suggest, however, that mixing of primary magma with crustal material might be possible in the course of their formation. On the basis of REE data (Figure 10), drawn mainly from data for rocks in the western Pacific island arcs, heterogeneous sources of boninites might be

suggested. KARPENKO *et al.* (1984) also suggested two sources (at the minimum) to explain the isotopic and rare earth element data from boninites, although ultramafic magma–crustal rock mixing was not postulated in their model.

The pressure–temperature range for generation and fractionation of basaltic liquids in the Earth's crust and mantle is summarized in Figure 11. Alkali basalts are produced by ~5 percent melting of ultramafic rocks in the mantle at depths down to about 100 km, whereas boninites and related rocks (including some tholeiite complexes; e.g., Bushveld and Karoo) were generated at shallower depth.

HYPOTHESIS FOR THE ORIGIN OF THEIR TERNARY GEOSTRUCTURES: MARGINAL SEA FLOORS–ISLAND ARCS–TRENCHES

From the evidence discussed in the preceding section, it is possible to conclude that the eastern and southeastern margins of Asia have been gradually destroyed over the last 50–60 m.y. and converted into oceanic crust. The appearance of typical sea floor instead of continental crust as well as

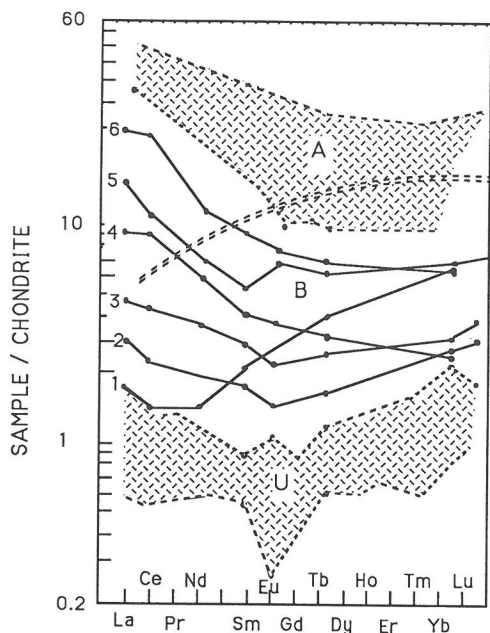


FIG. 10. Distribution of rare earth elements from rocks from western Pacific island arcs (after KARPENKO *et al.*, 1984; SHARASKIN *et al.*, 1983; ISHIZAKA and CARLSON, 1983; FROLOVA *et al.*, 1985). A = andesites and basalts in the northern Kuriles; high-magnesian andesites from southwest Japan fall in the same region. U = ultrabasites from the Mariana fore-arc. B = boninites from Bonin Islands (2), the Mariana Trench (3), Cape Vogel, Papua, New Guinea (4), Tonga Trench (5), and New Caledonia (6). Data from Cyprus boninites are given for comparison. See text for data sources.

thickening of the island arc crust have been observed in many areas. Possible mechanisms to explain these observations have been widely discussed, and two groups of possible mechanisms have been proposed. Back-arc spreading was suggested by, for example, KARIG (1974, 1975), HAYES (1984), SHARASKIN (1984) and others. A second mechanism has been proposed by ARTYUSHKOV (1979), BELOUSOV (1981, 1982) and PERCHUK (1984). This group of mechanisms is of a different nature and includes rock density inversions (BELOUSOV, 1981, 1982), eclogitization of the lower crust (ARTYUSHKOV, 1979) and magmatic replacement as a result of interaction of mantle-derived, high-temperature magma with crustal rocks (PERCHUK, 1985). Supporting evidence (see also discussion above) for the latter hypothesis includes:

(1) Gradual change in bulk composition of volcanic rocks from rhyolite and alkali basalt to trachybasalt and tholeiite during the last 60 m.y. in the Japan Sea. The volume of basic magma erupted increased significantly in the late Neogene.

(2) Increase in intensity of transformation of continental into oceanic crust, *i.e.* from the Okhotsk sea floor through the Japan Sea toward the Philippine sea floor. The process is accompanied either by crustal thinning beneath the sea floors or by thickening of the island arc igneous crust (KUSHIRO, 1985; TATSUMI, 1987; PEIVE *et al.*, 1980; SHARASKIN, 1984). This may reflect different stages of the same process.

(3) Alkali basalts that are diagnostic of the early (rift volcanism) and later stages (post-tholeiite volcanism) of continental destruction.

(4) Lateral correlation of alkalis and incompatible trace element contents with degree of partial melting of the mantle [across Japan and the Kuriles (KUSHIRO, 1983; TATSUMI, 1987)].

(5) Magma mixing beneath island arcs (SAKUYAMA, 1979, 1981).

(6) Fluctuations in differentiation paths in terms of alkalis and silica with simultaneous strong correlation between Mg, Al and Si in boninite series; high temperature (up to 1450°C) and low pressure (less than 10 kbar) for the generation of boninites

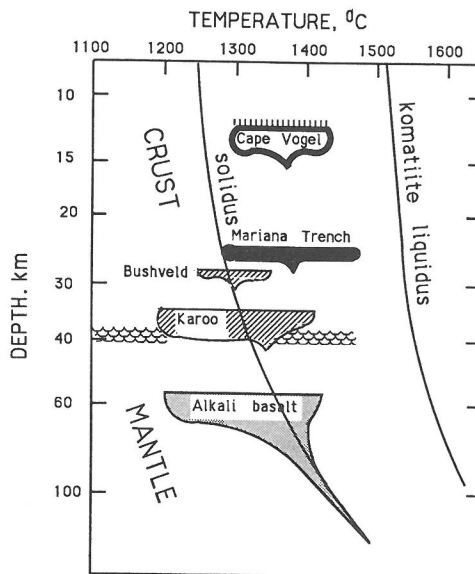


FIG. 11. Depth—temperature diagram illustrating possible regimes of basalt magma generation in the crust and the mantle. The contoured arcs show depth of each magma chamber and the temperature range of differentiation (not necessarily at a given pressure). Bushveld Complex—analytical data from CAWTHORN and DAVIES (1983). Cape Vogel and Mariana Trench are represented by boninite series. Karoo Province data are from WALKER and POLDEVAART (1949) to represent tholeiite and high-alumina basalt series. Alkali basalt series from numerous regions in the world. Wavy line represents the crust—mantle boundary beneath continental shields.

and fractionation of magma with 1–3 weight percent H₂O (WALKER and CAMERON, 1983; SOBOLEV *et al.*, 1986).

(7) Predominance of alkali basalt and tholeiite among differentiated and nondifferentiated volcanic complexes.

(8) Relatively high heat flow at low seismic and volcanic activity in the depressions of marginal sea floors and inverse relationships in island arcs (HAYES, 1984; SHARASKIN, 1984; RODNIKOV *et al.*, 1980).

(9) Distribution of major and trace elements and their isotopes in boninites and related rocks.

By taking into account these observations together with the discussion above, the following hypothesis is being proposed. Mantle magmatism is caused by decompression melting of a plume, enriched in incompatible trace elements, that ascends from the core–mantle boundary (*e.g.*, ARTUSHKOV, 1979; RAMBERG, 1972). Because the rate of heat dissipation by conduction is several orders of magnitude lower than by thermal convection, the plume mechanism may be the only mechanism responsible for mantle diapirism (*e.g.*, RAMBERG, 1972). The temperature of the plume (~500 km in diameter at the boundary between the lower and upper mantle) could have reached 1900–2000°C where, however, only a small amount of melting would occur ($P = 130\text{--}150$ kbar). At a depth of about 300 km, the first komatiite composition melts could be produced (NESBITT *et al.*, 1979). The amount of komatiitic magma depends on the thermal environment, *i.e.* heat distribution in the upper part of the diapir. The relationships between the temperature and crystallization of the magma and the thermal conditions in a given volume of the mantle determine the composition of the diapiric front and the extent of its interaction with the surrounding rocks. The appearance of komatiitic liquid might lead to accumulation of volatiles in the melt and, as a result, to an increase in the ascent rate. Some of this melt may segregate and rise as an intrusion in the mantle. As the pressure and temperature decrease, differentiation of the komatiitic magma might lead to crystallization in the form of layered intrusions. If tectonic conditions allow, a small amount of the komatiitic liquid might appear on the Earth's surface along deep-seated faults. This feature is illustrated with the well-known ophiolitic complex in Cyprus.

The next event in the diapiric ascent is illustrated by the appearance of late Cretaceous alkali basaltic volcanism in rift zones in island arcs. For example, pressure–temperature and age determinations of

alkali basalt magma generation in the Little Kurile island arcs (Figure 7) reflect the partial melting of the upper mantle 50–60 m.y. ago in the front of the diapir at depths of 60–100 km at a temperature near 1400°C. Fluid and melt inclusions in glasses and olivine phenocrysts in alkali basalts reflect enrichments in the initial magma of CO₂, H₂, N₂, CH₄, Ar, CO and H₂O (LETNIKOV *et al.*, 1977). These data show some differentiation processes near the solidus of a diapiric material with increasing abundance of volatiles in the frontal portions of the body (at 100–200 km or shallower). This enrichment might result in a lower temperature of partial melting at the depth corresponding to alkali basalt liquid formation as compared with a volatile-free environment. Deep-seated faults favor fast ascent of the low-density magma thus produced, and alkali basaltic volcanism is likely to be the first stage of the magmatic process observed on the surface.

The stratigraphy of the Japan sea floor shows erosion of sedimentary and igneous rocks during the Paleogene (VASILKOVSKII *et al.*, 1978; see also Figure 3). This period corresponds to the time of formation of an anticline structure observed in the modern Japan sea floor crust. Penetration of mantle-derived fluids into this structure might lead to partial melting of the crustal rocks and, as a result, to the development of acid magmatism in this area. The change in bulk composition of crustal-derived magmas (see Figure 4) correlates with the ascent of the diapir and its partial melt followed by complete melting resulting from the decompression (RAMBERG, 1972).

Thus, the ascent of the diapir results in an increase in heat flow upward and partial melting of both the crust and the mantle. This process results in the development of liparite–dacitic magmas followed by the basaltic andesite series shown in Figures 3 and 4. Komatiite could also be a product of lherzolitic or harzburgitic magma assimilated by crustal material when the completely molten diapir ascended toward the continental crust. Melting and mixing processes in the frontal part of the diapir would lead to cooling and subsequent fractionation. When the diapir reaches the crust–mantle interface, a completely molten diapir could react with the crust to produce mixed magmas whose compositions vary from komatiite through boninite to dacite. A peak in volcanic activity occurred in the Japan Sea area 3–5 m.y. ago when deep depressions formed in the sea floor (RODNIKOV *et al.*, 1980; PERCHUK *et al.*, 1985; see also Figure 3). In the Okhotsk sea floor, this stage was essentially limited to tholeiitic volcanism (accompanied by the formation of the South Okhotsk depression). In the

Philippine sea floor, the maximum intensity of this late stage was manifested by tholeiitic volcanism and tectonic activity.

From isotopic age determinations, an ascent rate from 350–400 km to the surface over the last ~60 m.y. can be estimated (0.4–0.6 cm/year). This rate is consistent with RAMBERG's (1972) prediction. Thermal convection cells must exist adjacent to the diapir, with a relatively high rate of horizontal displacement of the mixed magma toward cooler zones. This process is accompanied by fractionation. As the temperature decreases, isolated magma chambers may be formed. The largest extent of fractionation of the mixed magma might lead to magmatic activity and growth of igneous crust in island arcs and continents (PERCHUK *et al.*, 1985).

Deep-sea depressions are caused by magmatic replacement of the crust (melting, magma mixing and transport of material toward relatively low-temperature zones beneath the island arcs) and increasing density of the material beneath the sea floors. A decrease in crustal thickness beneath the sea floor is accompanied by increased thickness in the island arc provided by the thermal gradient discussed above. This gradient may be a function of the geometry of the subducted slab. With a very steep descent angle, the thermal gradient is low and horizontal displacement is slow, resulting in only slow growth of the igneous crust in a volcanic arc. This situation could have existed beneath the Philippine sea floor, where island arc growth at early stages of diapir–crust interaction was replaced by mantle-derived ultramafic magma during the latest stages. A sharp decrease in heat flow, and displacement of isotherms to greater depth, lead to the disappearance of the magma flow beneath the sea floor. The asthenospheric surface beneath the Philippine sea floor, at present, varies between 20 km beneath the Mariana trench and fore-arc and 80 km beneath the Philippine basin, between the Palau–Kyushu ridge and Ryukyu (A. G. RODNIKOV, personal communication, 1986).

According to the suggested hypothesis, fractionation and crystallization of diapiric material in separate convection cells produce numerous layered mafic–ultramafic bodies in the upper mantle. The body size depends on the relationships between the thermal gradient in a given volume of mantle and the solidus temperature of komatiite.

This hypothesis suggests a simultaneous origin of marginal sea floors and island arcs and also suggests replacement of early island arc volcanics by mantle-derived diapiric material. This hypothesis can explain, for example, the evolution of volcanism from acid to basic accompanied by a decrease of

sea floor crustal thickness. The hypothesis is not at variance with well-known models for calc–alkaline magma generation in subduction zones, and can also shed some light on some unresolved problems related to this mechanism. In contrast to models developed by Japanese petrologists, for example (*e.g.*, SAKUYAMA, 1983; KUSHIRO, 1985; TATSUMI, 1987), this hypothesis emphasizes interaction of high-temperature ultramafic magma with crustal rocks and the appearance of hybrid melts whose subsequent fractionation could lead to development of the basalt series in frontal arcs.

It is suggested that similar processes were inherent in many periods in the Earth's evolution. For example, the most suitable conditions for interaction processes existed in the Precambrian because of the higher geothermal gradients in the crust and upper mantle.

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APPENDIX

Table A1. Sources of data used for evaluation of Equation (13)

| A. System $CaO-MgO-Al_2O_3-SiO_2$ at atmospheric pressure | | B. Alkali-bearing systems at atmospheric pressure | |
|---|---|---|---|
| Fo-Di-SiO ₂ | BOWEN (1914); KUSHIRO (1972a) | Fo-Ne-Di | SCHAIRER and YODER (1960a,b); YODER and KUSHIRO (1972) |
| MgO-SiO ₂ | BOWEN and ANDERSEN (1914) | En-Ab-Di | SCHAIRER and MORIMOTO (1959) |
| Fo-An-SiO ₂ | ANDERSEN (1915); BIRD (1971) | Fo-Ab-Di | WATSON (1977); HART <i>et al.</i> (1976) |
| MgO-Al ₂ O ₃ -SiO ₂ | RANKIN and MERWIN (1918) | Di-Ak-Ne-CaTs | YAGI and ONUMA (1969) |
| Fo-Di-An | OSBORN and TAIT (1952) | Di-Ab-An-NiO | LINDSTROM (1976) |
| En-Di-An | HYTONEN and SCHAIRER (1961) | K ₂ O-Al ₂ O ₃ -MgO-SiO ₂ | SCHAIRER (1954); IRVINE (1976); WENDLANDT and EGGLER (1980); KUSHIRO (1980) |
| CaO-MgO-Al ₂ O ₃ -SiO ₂ | MACGREGOR (1965); PRESNALL <i>et al.</i> (1973) | | |
| FeO-MgO-SiO ₂ | BOWEN and SCHAIRER (1935) | | |

Table A1. (Continued)

| C. Synthetic systems at high pressure | |
|---|-------------------------------|
| Fo-Di-Py | DAVIS and SCHAIRER (1965) |
| CaO-MgO-Al ₂ O ₃ -SiO ₂ | PRESNALL <i>et al.</i> (1973) |
| CaO-MgO-Al ₂ O ₃ -SiO ₂ -Na ₂ O | KUSHIRO (1972b) |
| Fo-Di-SiO ₂ | KUSHIRO (1969) |
| MgO-TiO ₂ -SiO ₂ | MACGREGOR (1969) |

Table A2. Data used for linear regression of *Y* versus temperature [Equation (14)] for the system Fo-Ks-Qz (weight percent) for K₂O > 5 weight percent (SCHAIRER, 1954; WENDLANDT and EGGLEER, 1980)

| Fo | Ks | Qz | <i>Y</i> | <i>T</i> _{experimental} | <i>T</i> _{theoretical} |
|----|------|------|----------|----------------------------------|---------------------------------|
| 15 | 31.8 | 53.2 | 0.071 | 1450 | 1413 |
| 7 | 36.9 | 56.1 | 0.031 | 1400 | 1349 |
| 2 | 42.6 | 55.4 | 0.010 | 1200 | 1311 |
| 8 | 48.3 | 43.7 | 0.039 | 1400 | 1359 |
| 2 | 45.5 | 52.2 | 0.010 | 1300 | 1311 |
| 21 | 57 | 22 | 0.107 | 1490 | 1474 |
| 30 | 51 | 19 | 0.179 | 1560 | 1553 |
| 35 | 47 | 18 | 0.204 | 1640 | 1637 |
| 51 | 36 | 13 | 0.260 | 1700 | 1732 |

Table A3. Chemical compositions of basalts studied at dry conditions between 1 and 10 kbar

| Sample no. | (1) A-90 | (2) TF-38 | (3) KRB 2 | (4) NT-233 | (5) NM-51 | (6) 527-1-1 |
|--------------------------------|-------------|--------------|--------------|---------------|--------------|----------------|
| SiO ₂ | 49.73 | 44.31 | 47.47 | 49.25 | 47.93 | 48.60 |
| TiO ₂ | 2.10 | 1.58 | 2.43 | 0.79 | 1.34 | 0.61 |
| Cr ₂ O ₃ | n.d. | n.d. | n.d. | 0.20 | — | 0.06 |
| Al ₂ O ₃ | 15.82 | 13.89 | 17.15 | 13.64 | 16.75 | 16.30 |
| Fe ₂ O ₃ | 9.57* | 10.47* | 2.26 | — | — | 0.30 |
| FeO | — | — | 7.22 | 9.77† | 11.40† | 8.40 |
| MgO | 6.41 | 13.78 | 8.76 | 17.61 | 7.59 | 10.20 |
| CaO | 9.30 | 10.52 | 8.03 | 9.58 | 9.33 | 12.30 |
| Na ₂ O | 3.45 | 2.49 | 3.67 | 0.89 | 2.95 | 1.90 |
| K ₂ O | 1.34 | 0.73 | 1.89 | 0.06 | 0.19 | 0.07 |
| H ₂ O† | n.d. | 0.98 | 1.69 | n.d. | — | 0.08 |
| H ₂ O ⁻ | n.d. | — | 0.22 | n.d. | — | 0.13 |
| P ₂ O ₅ | n.d. | 0.17 | 0.66 | n.d. | — | 0.06 |
| Total | 99.05 | 99.09 | 100.38 | 99.93 | 100.00 | 99.16 |
| <i>dT/dP</i> °C/kbar | 3.8 | 4.0 | 4.5 | 4.7 | 4.0 | 3.5 |

(1)–(2): data from PERCHUK (1983); (3): TAKAHASHI (1979); (4): ELTHON and SCARFE (1979); (5): COHEN *et al.* (1967); (6): BENDER *et al.* (1978).

* All iron as Fe₂O₃

† All iron as FeO

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