Evolution of the mantle:
Geochemical evidence from alkali basalt

ABSTRACT

A careful consideration of the heterogeneities in the trace-element concentrations and lead and strontium isotope ratios for primary alkali basalt and nephelinite shows that the isotopic heterogeneities reflect real geochemical heterogeneities of their mantle source that have existed variously from about 1,000 to 3,000 m.y. ago. It is suggested that at present there is not mantle-wide convection and that the mantle source for alkali basalt is deeper than the presently convecting mantle. During Archean time, mantle convection was much deeper than at the present time, and this deeper convection led to hot-line tectonics that may have determined the tectonic style found in Archean greenstone belts.

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INTRODUCTION

A systematic comparison of the major-element compositions of alkali basalt by Schwarzer and Rogers (1974) shows that alkali basalt in ocean islands, on continents, and back of island arcs is a primary type of magma with a similar parentage regardless of its tectonic environment. This does not include alkali basalt and high-potassic basalt typically found in the east African rift valleys or the Rhine graben, as these are geochemically distinct (S. Kesson, 1974, oral commun.). Compared to the sources for ocean-ridge basalt, the sources for alkali basalt are more enriched in the incompatible elements, that is, K, Ba, Rb, and light rare-earth elements, and the alkali basalt appears to have been derived from a less depleted mantle than the ocean-ridge basalt (Gast, 1968). Kay and Gast (1973) studied the concentrations of the rare-earth elements, Ba, Rb, and Sr in alkali basalt and nephelinite and concluded that they result from several percent or less of partial melting of a garnet-bearing mantle with a chondritic rare-earth element (REE) pattern and REE concentrations of 2 to 5 times chondrite at depths greater than 60 km and that the chemical differences between nephelinite and alkali basalt can be explained by nephelinite having formed from the same mantle source but from lower extents of partial melting. We (S. Sun and G. Hanson, in prep.) have re-evaluated the evidence for the REE composition of the mantle source for alkali basalt and nephelinite and concluded that alkali basalt results from 7 to 15 percent melting, and nephelinite results from 3 to 7 percent melting, of a garnet-bearing mantle with a light REE-enriched pattern relative to chondrite, in which La is about 14 times chondrite and Yb is about 3 times chondrite. Lead and strontium isotopes from alkali basalt and tholeiite on ocean islands are generally more variable and more radiogenic than those from ocean-ridge basalt (Gast and others, 1964; Gast, 1967; Tatsumoto, 1966a, 1966b; Cooper and Richards, 1966; Doe, 1968; Oversby and Gast, 1970; Hedge and Peterman, 1970). The isotopic variations in the alkali basalt on ocean islands have been explained to be a result of contamination by pelagic sediment, frequent addition of conti-

nental crust material to the mantle, mixing of alkali-basalt source with ocean-ridge basalt source, isotopic disequilibrium during partial melting, or real heterogeneities within the mantle source that have existed for at least many hundreds of millions of years (Gast and others, 1964; Cooper and Richards, 1966; Tatsumoto, 1966b; Gast, 1967; Armstrong, 1968; Oversby and Gast, 1970; Armstrong and Hein, 1973; O'Nions and Pankhurst, 1973). The sources of alkali basalt and nephelinite may be attributed to melting of mantle in the low-velocity zone, deep-mantle plumes (Morgan, 1972), or less depleted portions of present-day convection cells.

This paper is an outgrowth of our study of the alkali basalt at Ross Island, Antarctica, as part of the Dry Valley Drilling Project (S. Sun and G. Hanson, in prep.) and the study of lead isotopes from oceanic volcanic rocks (S. Sun, in prep.). New ways of looking at the geochemical data resulted when we asked the question, "How similar in trace elements and Sr and Pb isotopes is the source of the volcanic rocks that occur on a continental crust to similar volcanic rocks on ocean islands?" With recent high-quality data from the literature (S. Sun and G. Hanson, in prep.; S. Sun, in prep.), it is possible to evaluate the proposed models for the geochemical heterogeneities found in alkali basalt and nephelinite. We find the most viable model to be one in which the isotopic heterogeneities are in the source; this has implications for present-day convection in the mantle as well as for convection during Precambrian time. The depth of convection may have had an effect on controlling the style of Archean and later tectonics.

TRACE-ELEMENT AND ISOTOPE DATA

We will assume that during melting the major and trace elements in the melt are in at least local equilibrium with the residual minerals. This seems reasonable, as Kay and Gast (1973) have explained the enrichment in light REE and incompatible elements and the relative depletion of the heavy REE by using

mineral-melt distribution coefficients for an experimentally determined garnet peridotite mineralogy necessary to produce alkali basalt and nephelinite (Green, 1973). Also, if there were large and varying extents of disequilibrium, it would be difficult to explain why the major and trace elements in alkali basalt and nephelinite are so similar.

Figure 1 compares REE for primary[1] alkali basalt from Ross Island, Antarctica, with highest precision REE data from other primary alkali basalt from ocean islands and continents. The total variation for an REE for primary alkali basalt is less than a factor of 2.5 and may reflect variations in (1) REE abundances of mantle source, (2) mineralogic composition of mantle source, or (3) extents of partial melting.

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[1] To reduce the effect of fractional crystallization, primary alkali basalt and nephelinite are considered to be those with $Mg/(Mg+Fe^{+2})$ (atomic) >0.64 and nickel contents >100 ppm. Irving (1971) showed that primary basalt equilibrated with mantle olivine (Fo_{88-92}) should have $Mg/(Mg+Fe^{+2})$ ratios of 0.68 to 0.77. Since olivine has a mineral-melt distribution coefficient for Ni of about 10 (Gast, 1968), primary melts of a mantle with about 2,000 ppm Ni (a common value for peridotite nodules) should have about 300 ppm Ni. For 10-percent fractional crystallization of equal amount of olivine and clinopyroxene, the liquidus minerals for alkali basalt will reduce the $Mg/(Mg+Fe^{+2})$ ratio in the melt by about 5 percent of the original ratio and the Ni content by about 40 percent of the original Ni content. Thus the Ni concentration in particular is very sensitive for indicating small percentages of fractional crystallization of olivine and clinopyroxene.

Figure 2 presents the lead data for ocean-island volcanic rocks and the continental Ross Island volcanic rocks and ocean-ridge basalt. The ocean-ridge basalt occupies a restricted field on the diagram, whereas the ocean-island volcanic rocks vary linearly, and the lines for each island or island group are sub-parallel. The linear relation for the ocean-island volcanic rocks may be interpreted in two ways: (1) a result of mixing, such as an alkali-basalt component mixing with an ocean-ridge basalt or a sediment component either before or after melting, or (2) the alkali basalt sources having a history of isolation, with long-term mantle heterogeneity in the $^{238}U/^{204}Pb$ ratio, the heterogeneity having existed for approximately the last 2,000 m.y. Data for a precise $^{238}U/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ isochron age are sparse; however, the available data plot with very large scatter about an isochron of approximately 1,000 m.y.

Figure 3 is a $^{87}Sr/^{86}Sr$ versus $^{87}Rb/^{86}Sr$ plot of primary alkali basalt and nephelinite. For comparison purposes, a 4,600-m.y.-old isochron with an initial ratio of 0.699 and a 2,000-m.y.-old isochron with an initial ratio of 0.7015 are plotted. The primary alkali basalt and nephelinite can be seen to plot about the 2,000-m.y. isochron. For both the lead and strontium data (Figs. 2, 3), individual volcanic groups have slopes suggesting a heterogeneity of ages from about 1,000 to 3,000 m.y. but with a mode of about 2,000 m.y.

The approximate 2,000-m.y. age for the Pb and Sr isotope data would suggest that inhomogeneities in the $^{238}U/^{204}Pb$ and Rb/Sr ratios were imposed about 2,000 m.y. ago if they are not a result of recent processes. In order to determine if mixing may be a viable model, $^{206}Pb/^{204}Pb$ is plotted against $^{87}Sr/^{86}Sr$ in Figure 4. If there is any correlation between $^{206}Pb/^{204}Pb$ and $^{87}Sr/^{86}Sr$, it is an inverse relationship, and the data do not lie on a mixing line between ocean-ridge basalt and a primitive alkali basalt. For Iceland, however, where there is mixing with ocean-ridge basalt magma (Schilling, 1973; O'Nions and Pankhurst, 1973; Sun and others, in prep.), the data for a suite of volcanic rocks plot along a line between the ocean-ridge basalt and alkali basalt. Gast and others (1964) have shown that sedimentary contamination sufficient to
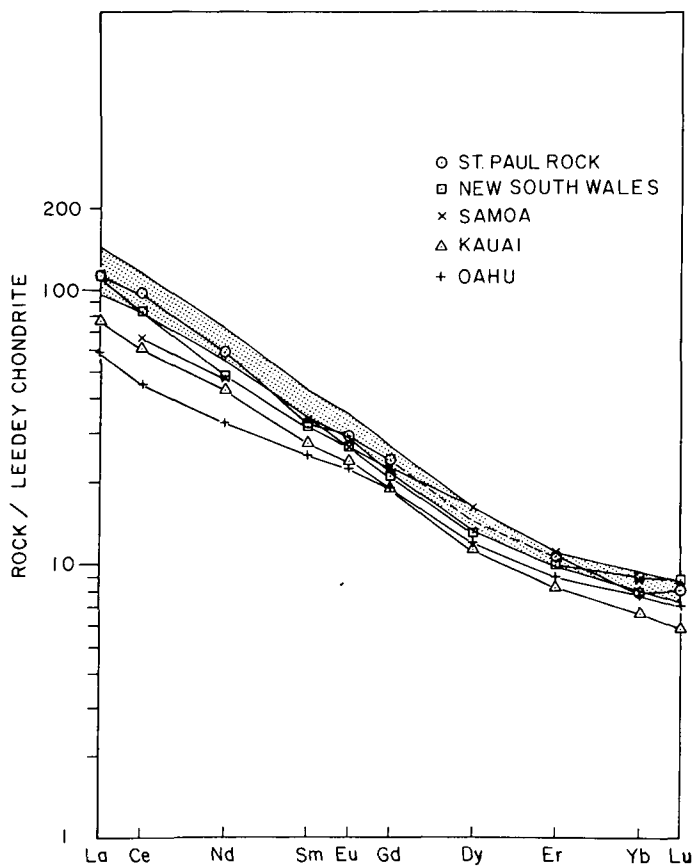

Figure 1. Comparison of REE for primary alkali basalt covering the total range in REE. The shaded field is for Ross Island, Antarctica. Data are normalized to Leedey chondrite (Masuda and others, 1973). REE data are from Frey (1970), Kay and Gast (1973), Hubbard (1971), and S. Sun and G. Hanson (in prep.).
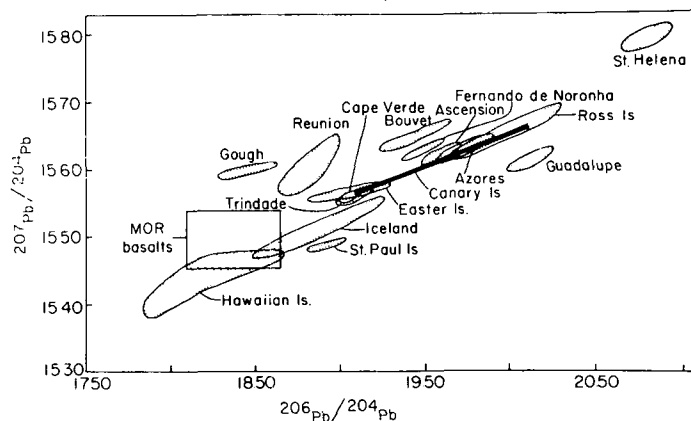

Figure 2. Plot of $^{207}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ fields for ocean-island volcanic rocks not restricted to primary alkali basalt or nephelinite and for ocean-ridge basalt. Data from S. Sun (in prep.), S. Sun and G. Hanson (in prep.), Oversby (1972), and Tatsumoto (1966a).

change significantly the $^{87}Sr/^{86}Sr$ ratio for alkali basalt requires so much sediment that it would greatly change the major-element chemistry and mineralogy and would completely dominate the lead-isotope composition.

O'Nions and Pankhurst (1973) have suggested that the variations in the strontium isotope ratios might be explained by isotopic disequilibrium during partial melting of a homogeneous mantle. They suggest that during melting the minerals that melt at low temperature, presumably amphibole, phlogopite, and apatite, give their isotopic character to the melt; that is, the melt is not in isotopic equilibrium with the total residue. This leads to melts with a more radiogenic isotopic composition than the total parent. Assuming that amphibole, phlogopite, and apatite in the mantle can retain a radiogenic isotopic composition for hundreds of millions of years at mantle temperatures, the lead data in Figure 2 suggest that the mantle source has maintained at least mineral-isotopic heterogeneities for about 2,000 m.y. In the melting process presumably more radiogenic lead would first be removed from a mineral such as apatite, and as melting proceeded the less radiogenic lead from phlogopite or amphibole would be added, or vice versa. In either case, the lead-isotopic data would be along an isochron defining the time when the minerals were last isotopically homogenized. If disequilibrium were an important process, the strontium-isotope ratios in melts from one region should show an inverse relation to the extent of partial melting. In the Hawaiian Islands, however, tholeiite as well as alkali basalt, and nephelinite formed by very different extents of partial melting (Kay and Gast, 1973; Hubbard, 1969), give essentially the same $^{87}Sr/^{86}Sr$ ratio. Thus we cannot seriously consider that isotopic disequilibrium during partial melting is the main source of the isotopic variations found in alkali basalt and nephelinite. Rather, the variations in the $^{238}U/^{204}Pb$ and Rb/Sr ratios seem to be in the mantle source and have existed for 1,000 to 3,000 m.y.

Trace-element data have been used by Kay and Gast (1973) to confirm that alkali basalt and nephelinite are derived by low percentages of partial melting of garnet peridotite. It

should also be possible to test for the presence of the isotopically important minerals apatite and phlogopite in the residue at the time the melt left the residue. The cerium content can be used to determine the extent of partial melting if the mantle source for alkali basalt is relatively homogeneous, if there is no apatite in the residue, and if the numerical value for the fraction of melting is greater than the bulk distribution coefficient for the mantle.[2] If apatite is a minor phase in the mantle during melting, it will control the Ce content of the melt, as apatite has a mineral-melt distribution coefficient for Ce of about 5 (inferred from Frey and Green, 1974), whereas the other presumed mantle minerals have a bulk distribution coefficient of less than 0.01 (Kay and Gast, 1973). If there is apatite in the residue, the P_2O_5 concentrations of the melts would be independent of the extent of melting, because P is a stoichiometric element in apatite, whereas the Ce concentration would be inversely proportional to the extent of melting and percent of apatite present. However, if there is no phosphate mineral in the residue, P_2O_5 and Ce should be covariant because

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[2]The concentration of trace elements in the magma relative to the original mantle (C_1/C_0) as a result of simple partial melting is given by Shaw (1970) as $C_1/C_0 = \dfrac{1}{D(1-F) + F}$ where D is the bulk distribution coefficient of the residue at the moment of removal of liquid from the residue. $D = \sum\limits_{a}^{n} X_a K_d^i$ where X_a is the weight fraction of each phase and K_d^i is the mineral-melt distribution coefficient. If D is very large for small percentages of melting, $C_1/C_0 = 1/D$. If D approaches zero, $C_1/C_0 = 1/F$. If zone refining is an important process, it will increase the concentration of the trace elements for alkali basalt to the limit of $C_1/C_0 = 1/D$ (Harris, 1974). In most cases the effect of zone refining cannot be distinguished from the effects of small extents of partial melting. Whether zone refining occurs or not has little effect on the conclusions derived regarding the mantle source for alkali basalt or nephelinite.
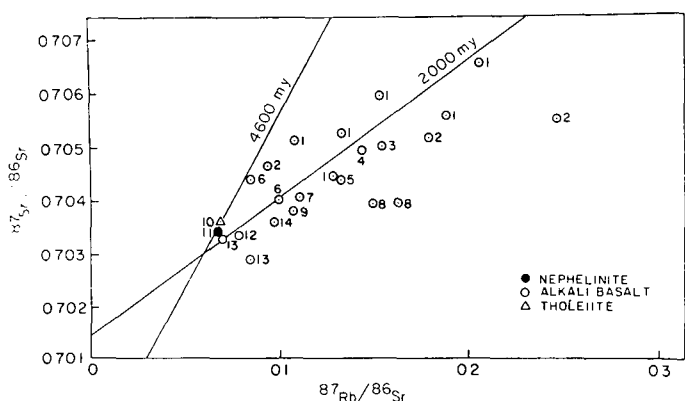

Figure 3. Plot of $^{87}Sr/^{86}Sr$ versus $^{87}Rb/^{86}Sr$ for primary alkali basalt and nephelinite for ocean islands. 1, Samoa; 2, Kerguelen; 3, Gough; 4, Tristan de Cunha; 5, Tahiti; 6, Reunion; 7, Amsterdam; 8, Crozet; 9, Eniwitok; 10, Kilauea, Hawaii (tholeiite); 11, Oahu, Hawaii; 12, St. Helena; 13, Galapagos; 14, Guadalupe. Data from Stuckless (1974, oral commun.), Hedge and Peterman (1970), Hedge and others (1973), Hedge (1974, written commun.), McBirney and Williams (1969), Peterman and Hedge (1971), Hedge and others (1972), Hart (1973), S. Sun and G. Hanson (in prep.), and O'Nions and Pankhurst (1974).
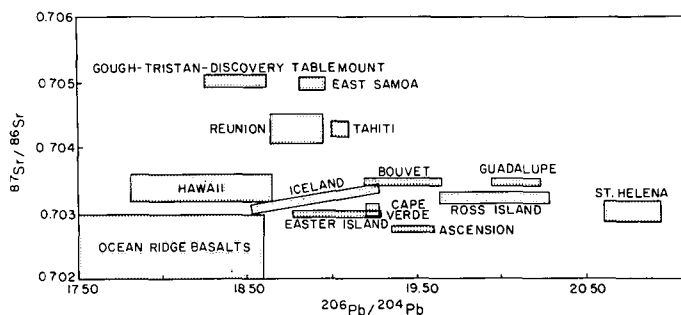

Figure 4. Plot of $^{87}Sr/^{86}Sr$ versus $^{206}Pb/^{204}Pb$ for volcanic rocks (not restricted to primary volcanics) from ocean islands and Ross Island. Data from references in Figure 2 and Figure 3 plus S. Sun (in prep.), Swainbank (1967), O'Nions and Pankhurst (1974), Sun and others (in prep.), McDougall and Compston (1965), Stuckless (1974, oral commun.), Grant and others (1973), and Klerkx and others (1974).
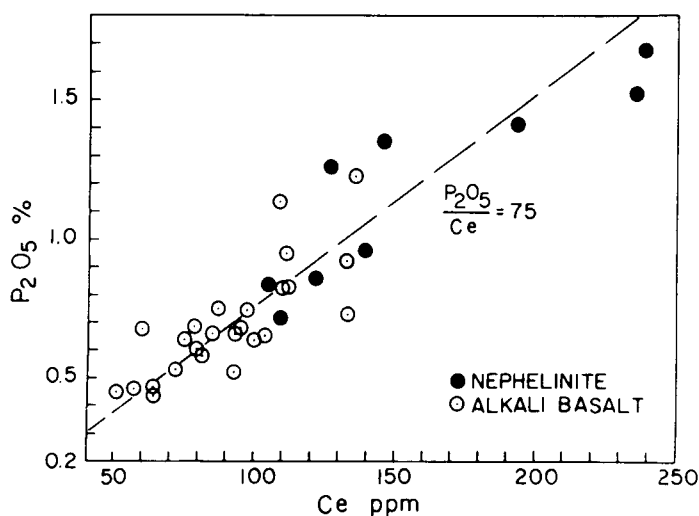
Figure 5. Plot of P_2O_5 versus cerium for primary alkali basalt and nephelinite. Data from Kay and Gast (1973), Stice (1968), Hubbard (1971), Price and Taylor (1973), Frey and Green (1974), Kesson (1973), and Goldich (1975, written commun.).
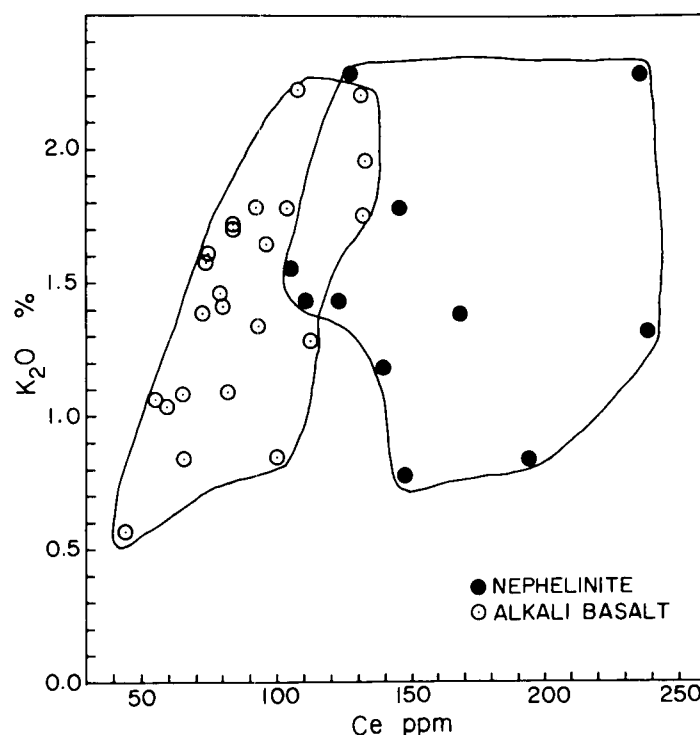

Figure 6. Plot of K_2O versus cerium for primary alkali basalt and nephelinite. Data from Philpotts and others (1972) and from references in Figure 5.

the concentration of both will be inversely proportional to the extent of partial melting as a result of dilution. Figure 5 is a plot of P_2O_5 versus Ce showing that both are roughly covariant for both alkali basalt and nephelinite. Much of the variation about the line in Figure 5 may be a result of relatively large analytical uncertainties, particularly for P_2O_5. However, from these data it can be suggested that the P_2O_5/Ce ratio in the mantle source is 75 ± 15. We conclude that apatite is probably not a mineral phase in the residue at the time of magma separation and that it should be possible to use Ce concentration as an indicator of the extent of partial melting, assuming alkali basalt and nephelinite were produced from relatively homogeneous mantle sources.

Figure 6 is a plot of K_2O versus Ce for primary alkali basalt and nephelinite. The mineral-melt distribution coefficients for K are low for all possible mantle minerals except phlogopite, for which K is stoichiometric, and perhaps amphibole. Thus if phlogopite is absent and amphibole is minor, there should be an inverse relationship between K_2O and the extent of melting, leading to a covariance of K_2O and Ce. This is the case for the alkali basalt. The nephelinite data, however, are more scattered, suggesting that K is either much more variable in the mantle source for the nephelinite or that a phase controlling K content, for example, phlogopite (Yoder and Kushiro, 1969), is a residual mineral in the mantle for nephelinite melts. The suggestion that nephelinite is a result of lower degrees of partial melting than is the alkali basalt (Green, 1973) is in agreement with the higher Ce concentration for nephelinite. For the nephelinite we interpret the K_2O-versus-Ce plot to indicate that a K-rich mineral such as phlogopite is in the residue and that the variation in the K content in nephelinite may be a function of mantle and melt composition (particularly volatiles such as F, Cl, and H_2O) as well as temperature and total pressure, which would control the stability of phlogopite and thus the concentration of potassium in the melt.

MANTLE EVOLUTION

We suggest that magma mixing, sediment contamination, or isotopic disequilibrium during partial melting are not controlling factors in determining the isotopic composition of Pb and Sr for alkali basalt and nephelinite. Rather they are controlled by heterogeneities in the mineralogy and chemistry of the mantle and history of the mantle. It is also suggested that alkali basalt and nephelinite have a mantle source which is similar to and independent of the environments in which they are found. The ocean-ridge basalt, derived presumably from the convecting mantle, has Rb/Sr ratios too low to support its [87]Sr/[86]Sr ratios, lower Ba/Sr ratios relative to alkali basalt, depleted light REE concentrations relative to heavy REE, and high K/Rb ratios relative to alkali basalt (Gast, 1968). These geochemical properties suggest that the mantle sources for ocean-ridge basalt have undergone more extensive melting than have the sources for alkali basalt and nephelinite. The sources of alkali basalt and ocean-ridge basalt must differ, and mantle convection must have a geometry that isolates these sources. Figure 7 is a graphical representation showing possible mantle activity and its surficial volcanic expression.

One possible source for the alkali basalt is near the middle of a convection cell shown by plume 1, which may not have as complicated a history as the outer parts of the cell. This source would restrict alkali basalt to regions away from plate margins and therefore is not a good source for all alkali basalt and nephelinite. Another possible source is the low-velocity zone. Leeds and others (1974) suggested that the lid of the low-velocity zone comes essentially to the base of the oceanic crust near the ridge and becomes deeper away from the ridge. Many ocean islands with alkali basalt occur near ridges (source 2 in Fig. 7), where the low-velocity zone would be too shallow to have garnet present as a stable phase (Ringwood, 1969) and where a depleted mantle would be involved. The low-velocity
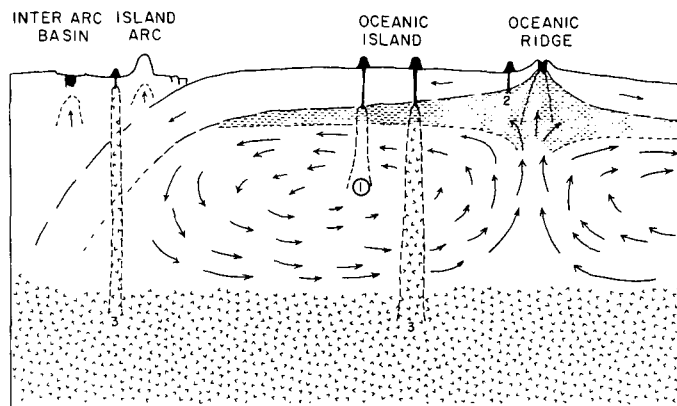
Figure 7. Diagrammatic representation of present-day mantle. 1 is possible source of alkali basalt and nephelinite in center of convection cell, 2 is possible source in low-velocity zone, and 3 is most probable source in mantle below present-day convection.

zone, therefore, is probably not the source. Unless there is some way to maintain an isolated closed system for long periods of time within a large part of the convecting mantle, we suggest that a mantle plume (Morgan, 1972; Wilson, 1973; 3 in Fig. 7) originating below the presently convecting mantle is the only viable alternative for a relatively similar source available under all tectonic environments.

If the alkali basalt is a result of mantle plumes, then this deeper mantle was involved in magma generation prior to some 1,000 to 3,000 m.y. ago, suggesting deeper convection and steeper geothermal gradients. Presumably, then, in early Precambrian time the convection was much deeper, and high percentages of melted materials from this less depleted mantle were the source of basalt in the Archean greenstone belts. This suggestion is supported by the trace-element geochemistry of tholeiite in Archean greenstone belts, whose mantle sources have strong affinities to the mantle sources of present-day alkali basalt (Hart and others, 1970). Richter (1973) suggested that with steeper geothermal gradients a second-order (or maybe even a higher order) convection may occur, forming horizontal rolls normal to the major convection. This results in hot lines, which may be capable of breaking up a continent, rather than hot spots. Tectonically this might be similar in some ways to present-day rift valleys or interarc basins on continents.

The hot lines would have associated with them massive amounts of tholeiitic magma resulting from high percentages of melting of the mantle. Basalt would be forming along hot lines on the oceanic crust as well as on the continental crust. If at the same time there was an oceanic crust acting somewhat like that of today, with ridges, island arcs, and subduction zones, the volcanic rocks associated with the hot lines on the oceanic crust would be returned to the mantle, whereas those volcanic rocks formed adjacent to or within the pre-existing continental crust would be preserved.

This model might explain why the Archean terrane is made up of low-grade greenstone belts and higher grade and sometimes older gneiss belts (Hepworth, 1971; Windley, 1973). The gneiss belts would represent the pre-existing continental crust derived by processes probably unrelated to greenstone belts (Bridgewater and others, 1974). The greenstone belts

would represent the uplifted and rifted parts that are filled with volcanics and mainly volcanogenic sediment. The later granitic rocks intrusive into the greenstones are derived by partial melting of the deeper parts of the volcanic and sedimentary pile or are derivatives of partial melting of material at mantle depths (Arth and Hanson, 1975). The lowered continental crust would be the basement for sedimentary basins in which the pre-existing continental crust and sediment are metamorphosed to granulite grade by the high heat flux. Due to the high heat flux, there may be melting of the lower continental crust, resulting in synkinematic granitic intrusions at higher levels. In the later stages of the cycle, with reduction of the heat flux, lower extents of melting would be represented in the greenstone belts by alkali basalt (Cooke and Moorhouse, 1969; Ridler, 1970) and by late alkalic stocks (Anhauesser and others, 1969; Arth and Hanson, 1975). At the end of the tectonic cycle, with isostatic readjustment, the rifted zones are lowered and only slightly eroded, leaving the relatively low-grade metavolcanic and metasedimentary rocks in the greenstone belts. The continental crust would rebound and be more deeply eroded, exposing the high-grade metamorphic rocks of the gneiss belts.

This model explains many of the major features found in an Archean terrane (Anhauesser and others, 1969) and also explains the lack of some features necessary for island-arc analogies, for example, the lack of blueschist-facies rocks, Alpine-type peridotite, and major compressional tectonic features in greenstone belts.

Perhaps because of loss of heat due to extensive volcanism during Archean time, the mantle cooled, convection became shallower, and there was separation of the present-day mantle source of ocean-ridge basalt and the mantle source of alkali basalt. The Rb-Sr and Pb-Pb ages would suggest that some parts of the mantle were isolated only for the last 1,000 m.y. Perhaps the depth of convection grew shallower until at least that time. The 1,000- to 3,000-m.y. ages for the present-day mantle source for alkali basalt and nephelinite may thus represent the last time of convection or beginning of isolation of different parts of the Earth's mantle, which may or may not be related to a time of special tectonic activity. In any case, it would appear that after Archean time, the style of convection changed and became more similar to that of the present.

REFERENCES CITED

Anhaeusser, C. R., Mason, R., Viljoen, M. J., and Viljoen, R. P., 1969, Reappraisal of some aspects of Precambrian shield geology: Geol. Soc. America Bull., v. 80, p. 2175-2200.

Armstrong, R. L., 1968, A model for the evolution of strontium and lead isotopes in a dynamic earth: Rev. Geophysics, v. 6, p. 175-199.

Armstrong, R. L., and Hein, S. M., 1973, Computer simulation of Pb and Sr isotope evolution of the Earth's crust and upper mantle: Geochim. et Cosmochim. Acta, v. 37, p. 1-18.

Arth, J. G., and Hanson, G. N., 1975, Geochemistry and origin of the early Precambrian crust of northeastern Minnesota: Geochim. et Cosmochim. Acta, v. 39, p. 325-362.

Bridgewater, D., McGregor, V. R., and Myers, J. S., 1974, A horizontal tectonic regime in the Archean of Greenland and its implications for early crustal thickening: Precambrian Research, v. 1, p. 179-197.

Cooke, D. L., and Moorhouse, W. W., 1969, Timiskaming volcanism in the Kirkland Lake area, Ontario, Canada: Canadian Jour. Earth Sci., v. 6, p. 117-132.

Cooper, J. A., and Richards, J. R., 1966, Lead isotopes and volcanic magmas: Earth and Planetary Sci. Letters, v. 1, p. 259-269.

Doe, B. R., 1968, Lead and strontium isotopic studies of Cenozoic volcanic rocks in the Rocky Mountain region: A summary: Colorado School Mines Quart., p. 149-174.

Frey, F. A., 1970, Rare earth and potassium abundances in St. Paul's rocks: Earth and Planetary Sci. Letters, v. 7, p. 351-360.

Frey, F. A., and Green, D. H., 1974, The mineralogy, geochemistry, and origin of lherzolite inclusions in Victorian basanites: Geochim. et Cosmochim. Acta, v. 38, p. 1023-1059.

Gast, P. W., 1967, Isotope geochemistry of volcanic rocks, in Hess, H. H., and Poldervaart, A., eds., Basalts: New York, Interscience Pubs., p. 325-358.

———— 1968, Trace element fractionation and the origin of tholeiitic and alkaline magma types: Geochim. et Cosmochim. Acta, v. 32, p. 1057-1086.

Gast, P. W., Tilton, G. R., and Hedge, C. E., 1964, Isotopic composition of lead and strontium from Ascension and Gough Islands: Science, v. 145, p. 1181-1185.

Grant, N. K., Powell, J. L., Walther, J. V., and Burkholder, F. R., 1973, Isotopic composition of strontium in lavas from the island of St. Helena, South Atlantic: EOS (Am. Geophys. Union Trans.), v. 54, p. 501-502.

Green, D. H., 1973, Experimental melting studies on a model upper mantle composition at high pressure under water-saturated and water-undersaturated conditions: Earth and Planetary Sci. Letters, v. 19, p. 37-53.

Harris, P. G., 1974, Origin of alkaline magmas as a result of anatexis, in Sorensen, H., ed., The alkaline rocks: New York, John Wiley & Sons, Inc., p. 427-436.

Hart, S. R., 1973, Submarine basalts from Kilauea rift, Hawaii: Non-dependence of trace element composition on extrusion depth: Earth and Planetary Sci. Letters, v. 20, p. 201-203.

Hart, S. R., Brooks, C., Krogh, T. E., Davis, G. L., and Nava, D., 1970, Ancient and modern volcanic rocks: A trace element model: Earth and Planetary Sci. Letters, v. 10, p. 17-28.

Hedge, C. E., and Peterman, Z. E., 1970, The strontium isotopic composition of basalt from the Gordo and Juan de Fuca Rises, northeastern Pacific Ocean: Contr. Mineralogy and Petrology, v. 27, p. 114-120.

Hedge, C. E., Peterman, Z. E., and Dickinson, W. R., 1972, Petrogenesis of lavas from western Samoa: Geol. Soc. America Bull., v. 83, p. 2709-2714.

Hedge, C. E., Watkins, N. D., Hildreth, R. A., and Doering, W. P., 1973, $^{87}Sr/^{86}Sr$ ratios in basalts from islands in the Indian Ocean: Earth and Planetary Sci. Letters, v. 21, p. 29-34.

Hepworth, J. V., 1971, Reappraisal of some aspects of Precambrian shield geology: Discussion: Geol. Soc. America Bull., v. 82, p. 803-804.

Hubbard, N. J., 1969, A chemical comparison of oceanic ridge, Hawaiian tholeiitic and Hawaiian alkalic basalts: Earth and Planetary Sci. Letters, v. 5, p. 346-352.

———— 1971, Some chemical features of lavas from the Manu'a Islands, Samoa: Pacific Sci., v. 25, p. 178-187.

Irving, A. J., 1971, Geochemical and high pressure experimental studies of xenoliths, megacrysts, and basalts from southeastern Australia [Ph.D. thesis]: Canberra, Australian National Univ.

Kay, R. W., and Gast, P. W., 1973, The rare earth content and origin of alkali-rich basalts: Jour. Geology, v. 81, p. 653-682.

Kesson, S. E., 1973, The primary geochemistry of the Monaro alkaline volcanics, southeastern Australia—Evidence for upper mantle heterogeneity: Contr. Mineralogy and Petrology, v. 42, p. 93-108.

Klerkx, J., Deutsch, S., and DePaepe, P., 1974, Rubidium, strontium content and strontium isotopic composition of strongly alkalic basaltic rocks from the Cape Verde Islands: Contr. Mineralogy and Petrology, v. 45, p. 107-118.

Leeds, A. R., Knopoff, L., and Kausel, E. G., 1974, Variations of upper mantle structure under the Pacific Ocean: Science, v. 186, p. 141-143.

Masuda, A., Nakamura, N., and Tanaka, T., 1973, Fine structures of mutually normalized rare-earth patterns of chondrites: Geochim. et Cosmochim. Acta, v. 37, p. 239-248.

McBirney, A. R., and Williams, H., 1969, Geology and petrology of the Galapagos Islands: Geol. Soc. America Mem. 118, 197 p.

McDougall, I., and Compston, W., 1965, Strontium isotope composition and potassium-rubidium ratios in some rocks from Reunion and Rodriguez, Indian Ocean: Nature, v. 207, p. 252-253.

Morgan, W. J., 1972, Deep mantle convection plumes and plate motions: Am. Assoc. Petroleum Geologists Bull., v. 56, p. 203-213.

O'Nions, R. K., and Pankhurst, R. J., 1973, Secular variation in the Sr-isotope composition of Icelandic volcanic rocks: Earth and Planetary Sci. Letters, v. 21, p. 13-21.

———— 1974, Petrogenetic significance of strontium isotope variations in volcanic rocks from the mid-Atlantic [abs.]: Paris, International meeting for geochronology, cosmochronology, and isotope geology.

Oversby, V. M., 1972, Genetic relations among the volcanic rocks of Reunion: Chemical and lead isotopic evidence: Geochim. et Cosmochim. Acta, v. 36, p. 1167-1179.

Oversby, V. M., and Gast, P. W., 1970, Isotopic composition of lead from oceanic islands: Jour. Geophys. Research, v. 75, p. 2097-2114.

Peterman, Z. E., and Hedge, C. E., 1971, Related strontium isotopic and chemical variations in oceanic basalts: Geol. Soc. America Bull., v. 82, p. 493-499.

Philpotts, J. A., Schnetzler, C. C., and Thomas, H. H., 1972, Petrogenetic implications of some new geochemical data on eclogitic and ultrabasic inclusions: Geochim. et Cosmochim. Acta, v. 36, p. 1131-1166.

Price, R. C., and Taylor, S. R., 1973, The geochemistry of the Dunedin volcano, East Otago, New Zealand: Rare earth elements: Contr. Mineralogy and Petrology, v. 40, p. 195-205.

Richter, F. M., 1973, Dynamical models for sea floor spreading: Rev. Geophysics and Space Physics, v. 11, p. 223-287.

Ridler, R. H., 1970, Relationships of mineralization to volcanic stratigraphy in Kirkland-Larder Lakes area, Ontario: Geol. Assoc. Canada Proc., v. 21, p. 33-42.

Ringwood, A. E., 1969, Composition and evolution of the upper mantle, in Hart, P. J., ed., The earth's crust and upper mantle: Am. Geophys. Union Mon. 13, p. 1-17.

Schilling, J-G., 1973, Iceland mantle plume: Geochemical study of Reykjanes Ridge: Nature, v. 242, p. 565-571.

Schwarzer, R. R., and Rogers, J.J.W., 1974, A world-wide comparison of alkali olivine basalts and their differentiation trends: Earth and Planetary Sci. Letters, v. 23, p. 286-296.

Shaw, D. M., 1970, Trace element fractionation during anatexis: Geochim. et Cosmochim. Acta, v. 34, p. 237-243.

Stice, G. D., 1968, Petrography of the Manu'a Islands, Samoa: Contr. Mineralogy and Petrology, v. 19, p. 343-357.

Swainbank, J. G., 1967, The isotopic composition of lead and strontium from the volcanic rocks of the islands of the South Pacific [Ph.D. thesis]: New York, Columbia Univ., 114 p.

Tatsumoto, M., 1966a, Isotopic composition of lead in volcanic rocks from Hawaii, Iwo Jima, and Japan: Jour. Geophys. Research, v. 71, p. 1721-1733.

———— 1966b, Genetic relations of oceanic basalts as indicated by lead isotopes: Science, v. 153, p. 1094-1101.

Wilson, J. T., 1973, Mantle plumes and plate motions: Tectonophysics, v. 19, p. 149-164.

Windley, B. F., 1973, Crustal development in the Precambrian: Royal Soc. London Philos. Trans., v. A273, p. 321-341.

Yoder, H. S., and Kushiro, I., 1969, Melting of a hydrous phase; phlogopite: Am. Jour. Sci., v. 267-A, p. 558-582.