

Theory of the Earth

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Chapter 8. Chemical Composition of the Mantle

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Abstract:

Considerations from cosmochemistry and the study of meteorites permit us to place only very broad bounds on the chemistry of the Earth's interior. These tell us little about the distribution of elements in the planet. Seismic data tell us a little more about the distribution of the major elements. General considerations suggest that the denser major elements will be toward the center of the planet and the lighter major elements, or those that readily enter melts or form light minerals, will be concentrated toward the surface. To proceed further we need detailed chemical information about crustal and mantle rocks. The bulk of the material emerging from the mantle is in the form of melts, or magmas. It is therefore important to understand the chemistry and tectonic setting of the various kinds of magmatic rocks and the kinds of sources they may have come from.

Chemical Composition of the Mantle

*These rocks . . .
Shall yet be touched with beauty,
and reveal
the secrets of the book of earth to man.*

—ALFRED NOYES

Considerations from cosmochemistry and the study of meteorites permit us to place only very broad bounds on the chemistry of the Earth's interior. These tell us little about the distribution of elements in the planet. Seismic data tell us a little more about the distribution of the major elements. General considerations suggest that the denser major elements will be toward the center of the planet and the lighter major elements, or those that readily enter melts or form light minerals, will be concentrated toward the surface. To proceed further we need detailed chemical information about crustal and mantle rocks. The bulk of the material emerging from the mantle is in the form of melts, or magmas. It is therefore important to understand the chemistry and tectonic setting of the various kinds of magmatic rocks and the kinds of sources they may have come from.

METHODS OF ESTIMATING MANTLE CHEMISTRY

The chemical composition of the mantle is one of the most important yet elusive properties of our planet. Attempts to estimate mantle composition fall into two broad categories.

Cosmochemical approaches take meteorites or mixtures of meteoritic material as the basic building blocks, and mixing ratios are adjusted to satisfy such constraints as core size, heat flow and crustal ratios of certain elements. An example is the six-component model of Morgan and Anders (1980). Cosmochemical models constrain the bulk chemistry of the Earth rather than that of the mantle alone; nevertheless they provide important input into models of the bulk chemistry of the Earth.

Petrological models begin with the reasoning that since basalts represent melts, and peridotites are thought to

be residues, some mixture of these should approximate the composition of the upper mantle. With only two components this approach does not yield chondritic ratios for many key elements, most notably Si/Mg. An alternate approach is to search for the most "primitive" ultramafic rock (that is, the one with most nearly chondritic ratios of the refractory elements) and attribute its composition to the whole mantle. Unfortunately, even the most "primitive" ultramafic nodules are depleted in many of the trace elements and have nonchondritic rare-earth ratios, but theorizing has proceeded beyond the first crude models.

Both approaches utilize terrestrial and meteoritic data. The common theme is that the Earth should have an unfractionated chondritic pattern of the refractory elements. This assumption, justified by the observation that these elements occur in roughly constant proportions in the various meteorite classes, has led to the generally accepted hypothesis that the refractory elements do not suffer any preaccretionary fractionation. This can be used as a formal a priori constraint in geochemical modeling of the composition of the Earth. Some recent estimates of the composition of the mantle are given in Table 8-1. Once they are in a planet, the refractory elements become fractionated by a variety of processes. The refractory siderophiles enter the core, the compatible refractories are retained in mantle silicates and the incompatible refractories preferentially enter melts and the crust along with the more volatile elements.

The volatile elements are fractionated by preaccretionary processes, and they exhibit a wide range in meteorites. It is therefore difficult to estimate the volatile content of the Earth or to obtain estimates of such key volatile-to-refractory ratios as K/U, Rb/Sr, Pb/U and others. It is often assumed that ratios of this type are the same in the Earth or in primitive mantle as they are in the continental crust. The

TABLE 8-1
Estimates of Average Composition of Mantle

Oxide	(1)	(2)	(3)	(4)	(5)
SiO ₂	45.23	47.9	44.58	47.3	45.1
Al ₂ O ₃	4.19	3.9	2.43	4.1	3.9
MgO	38.39	34.1	41.18	37.9	38.1
CaO	3.36	3.2	2.08	2.8	3.1
FeO	7.82	8.9	8.27	6.8	7.9
TiO ₂	—	0.20	0.15	0.2	0.2
Cr ₂ O ₃	—	0.9	0.41	0.2	0.3
Na ₂ O	—	0.25	0.34	0.5	0.4
K ₂ O	—	—	0.11	0.2	(0.13)

(1) Jacobsen and others (1984): extrapolation of ultramafic and chondritic trends.

(2) Morgan and Anders (1980): cosmochemical model.

(3) Maaløe and Steel (1980): extrapolation of lherzolite trend.

(4) 20 percent eclogite, 80 percent garnet lherzolite (Anderson, 1980).

(5) Ringwood and Kesson (1976, Table 7): pyrolite adjusted to have chondritic CaO/Al₂O₃ ratio and Ringwood (1966) for K₂O.

crust, of course, is just one repository of the incompatible elements and is less than 0.6 percent of the mass of the mantle. The validity of this assumption is therefore not obvious and needs to be tested by an independent approach.

"Primitive mantle" as used here is the silicate fraction of the Earth, prior to differentiation and removal of the crust and any other parts of the present mantle that are the result of differentiation, or separation, processes. In some geochemical models it is assumed that large parts of the Earth escaped partial melting, or melt removal, and are therefore still "primitive." Some petrological models assume that melts being delivered to the Earth's surface are samples from previously unprocessed material. I find it difficult to believe that any part of the Earth could have escaped processing during the high-temperature accretional process. "Primitive mantle," as used here, is a hypothetical material that is the sum of the present crust and mantle. "Primitive magma" is a hypothetical magma, the parent of other magmas, which formed by a single-stage melting process of a parent rock and has not been affected by loss of material (crystal fractionation) prior to sampling.

The view that there is a single primitive mantle magma type that leaves behind a single depleted peridotite, the essence of the pyrolite model, is clearly oversimplified. There is increasing evidence that ophiolitic peridotites, for example, are not simply related to the overlying basalts. Isotopic data on basalts and nodules show that there are at least two major mantle reservoirs. The identification of a component in ocean-island tholeiites and alkali olivine basalts that is "enriched," both chemically (in LILs) and isotopically, also is not adequately accounted for in single mantle reservoir models. I will refer to this enriched component as Q.

The use of three components of primitive mantle is conventional: basalts, ultramafic rocks, and continental crust. The assumption that the crust and depleted mantle are strictly complementary and are together equivalent to the bulk Earth, however, is not consistent with isotopic results. Basalts cover a broad compositional range, from LIL-poor to LIL-rich. "Large-ion-lithophile" (LIL) is commonly, although loosely, used to refer to elements (including small high-charge elements!) that do not substitute readily for magnesium or iron and are therefore excluded from olivine and orthopyroxene. Some LILs are relatively compatible in garnet and clinopyroxene. One recent proposal is that most mantle magmas are composed of a depleted MORB component and an enriched component (Q) with high potassium, LIL, ⁸⁷Sr/⁸⁶Sr, ¹⁴⁴Nd/¹⁴³Nd and ²⁰⁶Pb/²⁰⁴Pb. Mid-ocean-ridge basalt (MORB) represents the most uniform and voluminous magma type and is an end member for LIL concentrations and isotopic ratios. This is logically taken as one of the components of the mantle. The MORB source has been depleted by removal of a component—Q—that must be rich in LIL but relatively poor in Na and the garnet-clinopyroxene-compatible elements (such as Al, Ca, Yb, Lu and Sc). Kimberlitic magmas have the required complementary relationship to MORB, and I adopt them here as the Q component. Peridotites are the main reservoirs for elements such as magnesium, chromium, cobalt, nickel, osmium and iridium. The continental crust is an important reservoir of potassium, rubidium, barium, lanthanum, uranium and thorium. Thus, each of these components plays an essential role in determining the overall chemistry of the primitive mantle. It is conventional to adopt a single lherzolite or harzburgite as the dominant silicate portion of the mantle. An orthopyroxene-rich component is also present in the mantle and is required if the Mg/Si and Ca/Al ratios of the Earth are to be chondritic. Some peridotites appear to have been enriched (metasomatized) by a kimberlite-like component.

Figure 8-1 shows representative compositions of kimberlite, crust, MORB, and ultramafic rock. For many refractory elements kimberlite and crust have a similar enrichment pattern. However, the volatile/refractory ratios are quite different, as are ratios involving strontium, hafnium, titanium, lithium, yttrium, ytterbium and lutetium. Kimberlite and MORB patterns are nearly mirror images for the refractory elements, but this is only approximately true for MORB and crust, especially for the HREE, and the small ion-high charge elements. MORB and kimberlite also represent extremes in their strontium and neodymium isotopic compositions.

An important development in recent years has been the recognition of an LIL-enriched "metasomatic" component in the mantle. The most extreme magmas from the mantle (high LIL, high LREE/HREE, high ⁸⁷Sr/⁸⁶Sr, low ¹⁴³Nd/¹⁴⁴Nd) are kimberlites and lamproites (McCulloch and others, 1982, 1983). When these are mixed with a depleted

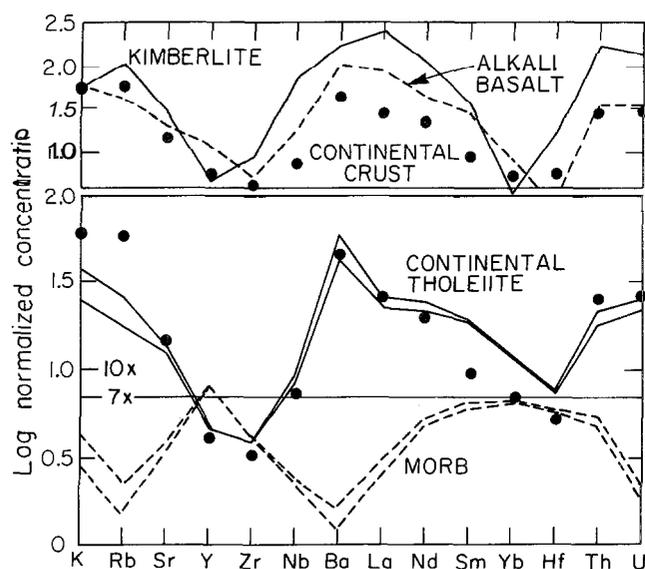


FIGURE 8-1

Trace-element concentrations in the continental crust (dots), continental basalts and mid-ocean ridge basalts (MORB), normalized to average mantle compositions derived from a chondritic model. Note the complementary relationship between depleted basalts (MORB) and the other materials. MORB and continental tholeiites are approximately symmetric about a composition of $7 \times C_1$. This suggests that about 14 percent of the Earth may be basalt. For other estimates, see text.

magma (MORB), the resulting blend can have apparently paradoxical geochemical properties. For example, the hybrid magma can have such high La/Yb, Rb/Sr and $^{206}\text{Pb}/^{204}\text{Pb}$ ratios that derivation from an enriched source is indicated, but the $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios imply derivation from an ancient depleted reservoir. Many ocean-island, island-arc and continental basalts have these characteristics. These apparently paradoxical results simply mean that ratios do not average as do concentrations.

In order to estimate the volatile and siderophile content of the mantle, we seek a linear combination of components that gives chondritic ratios for the refractory elements. We can then estimate such key ratios as Rb/Sr, K/U, and U/Pb. In essence, we replace the five basic building blocks of mantle chemistry (ol, opx, cpx, ga, and Q, or their high- and low-pressure equivalents) with four composites; peridotite (ol \pm opx), orthopyroxenite (opx \pm ol), basalt (cpx \pm ga), and Q. In practice, we can use two different ultramafic rocks (UMR and OPX) with different ol/opx ratios, to decouple the ol + opx contributions. The chemistry of the components (MORB, UMR, KIMB, OPX, and crust) are given in Table 8-2.

Having measurements of m elements in n components where m far exceeds n , we can find the weight fraction x_j of each component, given the concentration C_j of the j th element in the i th component, that yields chondritic ratios

of the refractory oxyphile elements. In matrix form,

$$C_{ij}x_j = kC_j \quad (1)$$

where C_j is the chondritic abundance of element j and k is a dilution or enrichment factor, which is also to be determined. The least-squares solution is

$$x_j/k = (C^T C)^{-1} C^T C_j \quad (2)$$

where C^T is the matrix transpose of C , with the constraints

$$\sum x_j = 1 \quad \text{and} \quad k = \text{constant} \quad (3)$$

When x_j and k are found, equation 1 gives the mantle concentrations of the volatile and siderophile elements, elements not used in the inversion.

The mixing ratios found from equations 2 and 3 are UMR, 32.6 percent, OPX, 59.8 percent, MORB, 6.7 percent, crust, 0.555 percent, and Q, 0.11 percent. This is a model composition for primitive mantle, that is, mantle plus crust. The Q component is equivalent to a global layer 3.6 km thick. The MORB component represents about 25 percent of the upper mantle. This solution is based on 18 refractory elements and, relative to carbonaceous chondrite (C1) abundances, $k = 1.46 \pm 0.09$. The result is given in Table 8-3 under "Mantle plus crust." Concentrations normalized to C1 are also given. Note that the high-charge, small ionic radius elements (Sc, Ti, Zr, Nb, and Hf) have the highest C1-normalized ratios. Taking all the refractory oxyphile elements into account (23 elements), the C1-normalized enrichment of the refractory elements in the mantle plus crust is 1.59 ± 0.26 . If six elements (Fe, S, Ni, Co, P, and O) are removed from C1 to a core of appropriate size and density, the remaining silicate fraction will be enriched in the oxyphile elements by a factor of 1.48 relative to the starting C1 composition. This factor matches the value for k determined from the inversions. Thus it appears that the Earth's mantle can be chondritic in major- and refractory-element chemistry if an appreciable amount of oxygen has entered the core (see Chapter 4).

Table 8-3 also compares these results with Morgan and Anders's cosmochemically based model. This model can be viewed as providing a first-order correction for volatile-refractory fractionation and inhomogeneous accretion. Rather than treating each element separately, Morgan and Anders estimated abundances of groups of elements: refractories, volatiles, and so on. There is strong support for unfractionated behavior of refractories prior to accretion, but the volatile elements are likely to be fractionated. Both the volatile elements and the siderophile elements are strongly depleted in the crust-mantle system relative to cosmic abundances.

In the pyrolite models, it is assumed that primitive mantle is a mix of basalt and peridotite and that one knows the average compositions of the basaltic and ultramafic components of the mantle. These are mixed in somewhat arbitrary proportions and the results are comparable with

TABLE 8-2
Chemical Composition of Mantle Components (ppm)

	MORB	Ultramafic Rocks	KIMB	Crust	Picrite	OPX	Morgan and Anders
Li	9	1.5	25	10	7	1.5	1.85
F	289	7	1900	625	219	—	14
Na	15900	2420	2030	26000	12530	—	1250
Mg*	5.89	23.10	16.00	2.11	10.19	21.00	13.90
Al*	8.48	2.00	1.89	9.50	6.86	1.21	1.41
Si*	23.20	21.80	14.70	27.10	22.85	22.60	15.12
P	390	61	3880	1050	308	—	1920
S	600	8	2000	260	452	—	14600
Cl	23	1	300	1074	17	—	20
K	660	35	17600	12500	504	10	135
Ca*	8.48	2.40	7.04	5.36	6.96	1.32	1.54
Sc	37.3	17	15	30	32	12	9.6
Ti	5500	1000	11800	4800	4375	800	820
V	210	77	120	175	177	60	82
Cr	441	2500	1100	55	956	2500	4120
Mn	1080	1010	1160	1100	1063	1010	750
Fe*	6.52	6.08	7.16	5.83	6.41	6.08	32.07
Co	53	105	77	25	66	105	840
Ni	152	2110	1050	20	642	2110	18200
Cu	77	15	80	60	62	30	31
Zn	74	60	80	52	71	20	74
Ga	18	4	10	18	14	3	3
Ge	1.5	1.1	0.5	1.5	1.4	1.1	7.6
Se	0.181	0.02	0.15	0.05	0.14	0.006	9.6
Rb	0.36	0.12	65	42	0.30	0.04	0.458
Sr	110	8.9	707	400	85	4.5	11.6
Y	23	4.6	22	22	18	0.023	2.62
Zr	70	11	250	100	55	5.5	7.2
Nb	3.3	0.9	110	11	2.7	0.45	0.5
Ag	0.019	0.0025	—	0.07	0.015	0.001	0.044
Cd	0.129	0.0255	0.07	0.2	0.103	0.01	0.016
In	0.072	0.002	0.1	0.1	0.055	—	0.002
Sn	1.36	0.52	15	2	1.15	0.52	0.39
Cs	0.007	0.006	2.3	1.7	0.01	0.002	0.0153
Ba	5	3	1000	350	4.50	2	4
La	1.38	0.57	150	19	1.18	0.044	0.379
Ce	5.2	1.71	200	38	4.33	0.096	1.01
Nd	5.61	1.31	85	16	4.54	0.036	0.69
Sm	2.08	0.43	13	3.7	1.67	0.009	0.2275
Eu	0.81	0.19	3	1.1	0.66	0.0017	0.079
Tb	0.52	0.14	1	0.64	0.43	0.0017	0.054
Yb	2.11	0.46	1.2	2.2	1.70	0.0198	0.229
Lu	0.34	0.079	0.16	0.03	0.27	0.0072	0.039
Hf	1.4	0.34	7	3	1.14	0.17	0.23
Ta	0.1	0.03	9	2	0.08	—	0.023
Re [†]	1.1	0.23	1	1	0.88	0.08	60
Os [†]	0.04	3.1	3.5	5	0.81	3.1	880
Ir [†]	0.0011	3.2	7	1	0.80	3.2	840
Au [†]	0.34	0.49	4	4	0.38	0.49	257
Tl [†]	0.01	0.02	0.22	0.25	0.01	—	0.00386
Pb	0.08	0.2	10	7	0.11	—	0.068
Bi	0.007	0.005	0.03	0.2	0.01	—	0.00294
Th	0.035	0.094	16	4.8	0.05	—	0.0541
U	0.014	0.026	3.1	1.25	0.02	—	0.0135

*Results in percent

†Results in ppb.

Anderson (1983a).

TABLE 8-3

Chemical Composition of Mantle (ppm)

	CI	Morgan and Anders	UDS	Upper Mantle	Lower Mantle	Mantle + Crust	Normalized to M&A	Normalized to CI
Li	2.4	1.85	7.48	3.49	1.50	2.09	0.76	0.87
F	90	14	259	91	1	28	1.39	0.31
Na	7900	1250	13172	6004	359	2040	1.10	0.26
Mg*	14.10	13.90	9.80	18.67	21.31	20.52	1.00	1.46
Al*	1.29	1.41	6.95	3.65	1.33	2.02	0.97	1.57
Si*	15.60	15.12	23.00	22.20	22.48	22.40	1.00	1.44
P	1100	1920	387	170	9.06	57	0.02	0.05
S	19300	14600	458	158	1.19	48	0.0022	0.0025
Cl	1000	20	79	27	0.07	8	0.27	0.008
K	890	135	1356	475	14	151	0.76	0.17
Ca*	1.39	1.54	6.87	3.89	1.48	2.20	0.96	1.58
Sc	7.8	9.6	31.92	22	13	15	1.09	1.99
Ti	660	820	4477	2159	830	1225	1.01	1.86
V	62	82	176	110	63	77	0.63	1.24
Cr	3500	4120	907	1969	2500	2342	0.38	.67
Mn	2700	750	1066	1029	1010	1016	0.91	0.38
Fe*	27.20	32.07	6.39	6.18	6.08	6.11	0.13	0.22
Co	765	840	64	91	105	101	0.08	0.13
Ni	15100	18200	611	1610	2110	1961	0.07	0.13
Cu	160	31	62	31	28	29	0.62	0.18
Zn	455	74	70	63	26	37	0.34	0.08
Ga	14	3	15	7	3	4	0.94	0.31
Ge	47	7.6	1.40	1.20	1.10	1.13	0.10	0.02
Se	29	9.6	0.14	0.06	0.01	0.02	0.0016	0.0008
Rb	3.45	0.458	3.32	1.19	0.05	0.39	0.57	0.11
Sr	11.4	11.6	109	42.3	5.2	16.2	0.94	1.42
Y	2.1	2.62	18.64	9.28	0.70	3.26	0.84	1.55
Zr	5.7	7.2	60	27	6	13	1.18	2.20
Nb	0.45	0.5	4.30	2.03	0.52	0.97	1.31	2.15
Ag	0.27	0.044	0.02	0.01	0.00	0.003	0.05	0.01
Cd	0.96	0.016	0.11	0.05	0.01	0.02	1.03	0.03
In	0.12	0.002	0.06	0.02	0.0003	0.006	2.13	0.05
Sn	2.46	0.39	1.34	0.79	0.52	0.60	1.04	0.24
Cs	0.29	0.0153	0.13	0.05	0.003	0.02	0.68	0.05
Ba	3.60	4	34	13	1.72	5.22	0.88	1.45
La	0.367	0.379	3.75	1.63	0.12	0.57	1.02	1.56
Ce	0.957	1.01	8.28	3.90	0.34	1.40	0.93	1.46
Nd	0.711	0.69	6.03	2.88	0.23	1.02	1.00	1.43
Sm	0.231	0.2275	1.90	0.92	0.07	0.32	0.96	1.40
Eu	0.087	0.079	0.70	0.362	0.03	0.13	1.10	1.48
Tb	0.058	0.054	0.44	0.241	0.02	0.09	1.09	1.51
Yb	0.248	0.229	1.72	0.88	0.09	0.32	0.95	1.30
Lu	0.038	0.039	0.27	0.144	0.02	0.06	0.96	1.46
Hf	0.17	0.23	1.30	0.66	0.20	0.33	0.98	1.96
Ta	0.03	0.023	0.28	0.11	0.004	0.04	1.10	1.24
Re†	60	60	0.89	0.45	0.10	0.21	0.0023	0.0034
Os†	945	880	1.08	2.43	3.10	2.90	0.0022	0.0031
Ir†	975	840	0.88	2.43	3.20	2.97	0.0024	0.0031
Au†	255	257	0.62	0.53	0.49	0.50	0.0013	0.002
Tl†	0.22	0.0039	0.03	0.02	0.00	0.01	1.28	0.033
Pb	3.6	0.068	0.60	0.33	0.03	0.12	1.19	0.033
Bi	0.17	0.0029	0.02	0.009	0.0007	0.0033	0.75	0.019
Th	0.051	0.0541	0.48	0.224	0.014	0.0765	0.96	1.50
U	0.014	0.0135	0.12	0.057	0.004	0.0196	0.98	1.40

*Results in percent.

†Results in ppb.

Anderson (1983a).

pleted in these elements, including the heat producers potassium, uranium and thorium.

In an alternative approach we can replace UMR and OPX by their primary constituent minerals, olivine, orthopyroxene, and clinopyroxene. The present mantle is therefore viewed as a five-component system involving olivine, orthopyroxene, clinopyroxene, MORB (cpx and ga), and Q. In this case the LIL inventory of the primitive mantle is largely contained in four components: MORB, Q, clinopyroxene, and CRUST. The results are: olivine, 33.0 percent, orthopyroxene, 48.7 percent, clinopyroxene, 3.7 percent, MORB, 14.0 percent, Q, .085 percent and CRUST, 0.555 percent. Concentrations of certain key elements are sodium, 2994 ppm, potassium, 205 ppm, rubidium, 0.53 ppm, strontium, 25 ppm, and cesium, 0.02 ppm. The alkalis are generally within 50 percent of the concentrations determined previously. Key ratios are Rb/Sr, 0.021, K/U, 4323, and K/Na, 0.07. The Rb/Sr and K/Na ratios are essentially the same as those determined previously; the K/U ratio is 44 percent lower.

In summary, a four-component (crust, basalt, peridotite and Q) model for the upper mantle can be derived that gives chondritic ratios for the refractory trace elements. The model gives predictions for volatile/refractory ratios such as K/U, Rb/Sr and Pb/U. An orthopyroxene-rich component is required in order to match chondritic ratios of the major elements. Such a component is found in the upper mantle and is implied by the seismic data for the lower mantle. The abundances in the mantle-plus-crust system are 151 ppm potassium, 0.0197 ppm uranium and 0.0766 ppm thorium, giving a steady-state heat flow of $0.9 \mu\text{cal}/\text{cm}^2 \text{ s}$. This implies that slightly more than half of the terrestrial heat flow is due to cooling of the Earth, consistent with convection calculations in a stratified Earth (McKenzie and Richter, 1981).

Primitive mantle can be viewed as a five-component system; crust, MORB, peridotite, pyroxenite and Q (quintessence, the fifth essence) or, alternatively, as olivine, orthopyroxene, garnet plus clinopyroxene (or basalt) and incompatible and alkali-rich material (crust and kimberlite or LIL-rich magmas).

THE UPPER MANTLE

The mass-balance method gives the average composition of the mantle but makes no statement about how the components are distributed between the upper and lower mantle. If we assume that the only *ultramafic* component of the upper mantle is UMR, we can estimate the composition of the upper and lower mantles and, as an intermediate step, the composition of the MORB source region prior to extraction of crust and Q. The lower mantle is UMR plus OPX. Orthopyroxenite, the most uncertain and to some extent arbitrary

of the components, plays a minor role in the mass-balance calculations for the trace refractories and is required mainly to obtain chondritic ratios of Ca/Al and Mg/Si.

My approach to the upper mantle is similar to the conventional approach in that I consider a basaltic and an ultramafic component. However, instead of making an *a priori* selection of basalt, I have decomposed it into a depleted (MORB) and an enriched (Q) component. These represent extremes in both LIL contents and isotopic ratios. For example, fresh MORB has $^{87}\text{Sr}/^{86}\text{Sr}$ as low as 0.7023 and kimberlite usually has $^{87}\text{Sr}/^{86}\text{Sr}$ well above 0.704; alkalic basalts are intermediate in both LIL contents and isotopic ratios. The procedure is as follows. The mixing ratios of MORB, crust and Q are known from the previous section, and these ingredients are assumed to be entirely contained in the primitive upper mantle. The absolute sizes of the crustal and upper-mantle reservoirs (above 650 km) are known, so we know both the relative and absolute amounts of each component. As an intermediate step, we estimate the composition of a possible picritic parent to MORB. The relation $\text{PICRITE} = 0.75 \text{ MORB} + 0.25 \text{ UMR}$ gives the results tabulated under "picrite" in Table 8-2 and 8-4.

The mixing ratios which were determined to give a chondritic pattern for the refractory elements yield

$$\text{UDS} = 0.9355\text{PICRITE} + 0.0106\text{Q} + 0.0559\text{CRUST}$$

The composition of this reconstructed Undepleted Source Region is tabulated under UDS. The fraction of the crustal component is about 10 times the crust/mantle ratio, so UDS accounts for 10 percent of the mantle. The remainder of the upper mantle is assumed to be ultramafic rocks UMR. This gives the composition tabulated under "Upper Mantle" in Tables 8-3 and 8-4. This region contains 23.4 percent basalt (MORB). The resulting upper mantle has refractory-element ratios (Table 8-5), which, in general, are in agreement with chondritic ratios. The La/Yb, Al/Ca and Si/Mg ratios, however, are too high. These are balanced by the lower mantle in the full calculation. The solution for the lower mantle is 0.145 UMR and 0.855 OPX. This gives chondritic ratios for Mg/Si and Ca/Al for the Earth as a whole. An orthopyroxene-rich lower mantle is expected for a chondritic model for the major elements, particularly if the upper mantle is olivine-rich. At low pressure olivine and orthopyroxene are refractory phases and are left behind as basalt is removed. However, at high pressure the orthopyroxene-rich phases, majorite and perovskite, are both refractory and dense. If melting during accretion extended to depths greater than about 350 km, then the melts would be olivine-rich and separation of olivine from orthopyroxene can be expected.

Figure 8-3 shows the concentrations of the lithophile elements in the various components, upper mantle and mantle-plus-crust, all normalized to the Morgan and Anders mantle equivalent concentrations. The refractory elements in the upper mantle have normalized concentrations of

TABLE 8-4
Elemental Ratios

	C1	Morgan Anders*	UDS	Upper Mantle	Lower Mantle	Mantle + Crust	Normalized M&A*	C1
Rb/Sr	0.3026	0.0395	0.0304	0.0281	0.0101	0.024	0.61	0.08
K/U	63571	10000	11429	8356	3552	7693	0.77	0.12
Sm/Nd	0.3249	0.3297	0.315	0.319	0.318	0.319	0.968	0.982
Th/U	3.64	4.01	4.08	3.94	3.62	3.98	0.97	1.07
U/Pb	0.0039	0.20	0.20	0.17	0.13	0.16	0.82	42
La/Yb	1.48	1.66	2.18	1.85	1.43	1.77	1.07	1.20
K/Na	0.11	0.11	0.10	0.08	0.04	0.07	0.69	0.66
Mg/Si	0.90	0.92	0.43	0.84	0.95	0.92	1.00	1.01
Ca/Al	1.08	1.09	0.99	1.07	1.12	1.09	1.00	1.01
Yb/Sc	0.032	0.024	0.054	0.04	0.007	0.021	0.87	0.65
Ce/Nd	1.35	1.46	1.37	1.35	1.49	1.37	0.94	1.02
Eu/Nd	0.12	0.11	0.12	0.13	0.13	0.13	1.10	1.03
Yb/Lu	6.53	5.87	6.26	6.10	4.77	5.80	0.99	0.89
Sr/Ba	3.17	2.90	3.17	3.14	2.99	3.11	1.07	0.98
U/La	0.038	0.036	0.03	0.035	0.032	0.034	0.97	0.90

*Model of Morgan and Anders (1980).

about 3; this includes the crustal contribution. Since the upper mantle is about one-third of the whole mantle, a strongly depleted lower mantle is implied. Note that the upper mantle is depleted in lithium and titanium. These elements may be in the lower mantle, or the Morgan and Anders estimates may be too high.

The upper mantle is not necessarily uniform. The basaltic fraction, as eclogite, is denser than peridotite and may form the major part of a separate layer. Seismic data are consistent with an eclogite-rich transition region and also suggest that the roots of midocean ridges extend to the tran-

sition region. The conventional view of basalt petrogenesis, however, is that the basaltic fraction of the mantle is uniformly dispersed, on a microscopic scale, in a mainly olivine matrix. The picritic fraction of the mantle corresponds to a layer about 200 km in thickness. At high pressure this would be a picritic eclogite, or piclogite. The transition region, it happens, is just over 200 km thick. The shallow mantle is probably harzburgite-rich. Basalts, or their parents, of course, must pass through the shallow mantle on their way to the surface, so parts of the shallow mantle are basalt-rich.

TABLE 8-5
Elemental Ratios in Upper-Mantle Components

	MORB	Ultramafic Rocks	KIMB	Crust	Picrite	OPX	Morgan and Anders
Rb/Sr	0.0033	0.0135	0.0919	0.105	0.0035	0.0089	0.039
K/U	47143	1346	5677	10000	29632	—	10000
Sm/Nd	0.371	0.328	0.153	0.231	0.368	0.25	0.3297
Th/U	2.50	3.62	5.16	3.84	2.94	—	4.01
U/Pb	0.18	0.13	0.31	0.18	0.15	—	0.20
La/Yb	0.65	1.24	125	8.64	0.69	2.22	1.66
K/Na	0.84	0.01	8.67	0.48	0.04	—	0.11
Mg/Si	0.25	1.06	1.09	0.08	0.45	0.93	0.92
Ca/Al	1.00	1.20	3.72	0.56	1.01	1.09	1.09
Yb/Sc	0.057	0.027	0.08	0.073	0.053	0.0017	.0239
Ce/Nd	0.93	1.31	2.35	2.38	0.95	2.67	1.46
Eu/Nd	0.14	0.15	0.04	0.07	0.14	0.05	0.11
Yb/Lu	6.21	5.82	7.50	7.33	6.18	2.75	5.87
Sr/Ba	22.00	2.97	1.18	1.14	18.83	3.00	2.90
U/La	0.01	0.46	0.021	0.066	0.014	—	0.036

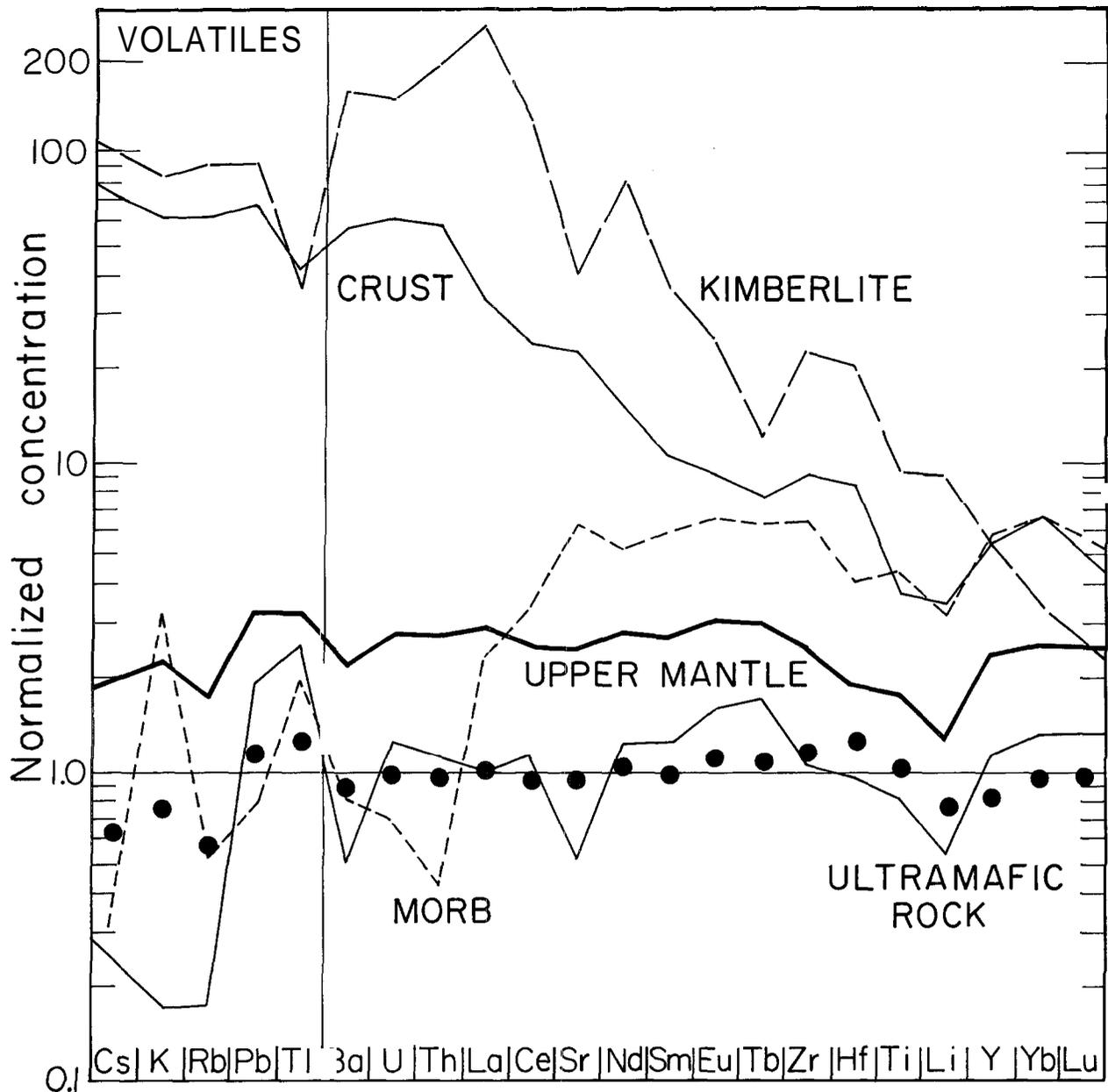


FIGURE 8-3
Lithophile elements in ultramafic rocks, MORB, continental crust and kimberlite normalized to 1.48 times the Morgan and Anders (1980) model. Derived values for upper mantle and primitive mantle (dots) are also shown (after Anderson, 1983a).

THE LOWER MANTLE

Whether the Earth is chondritic in major-element ratios, as assumed above, or is pyrolitic with a large silicon deficiency depends on the composition of the lower mantle. This section condenses the more detailed discussion in Chapter 4. Seismic data show that the lower mantle has a higher density and bulk modulus than olivine-rich peridotite or pyrolite whether in a mixed oxide or perovskite-bearing

assemblage. The lower mantle is separated from the upper-mantle by a sharp discontinuity at about 650 km. The sharpness of the discontinuity and the magnitude of the density and velocity jumps suggest that this is a chemical discontinuity as well as a phase change (Burdick and Anderson, 1975; Anderson, 1977).

There has been much controversy over the composition of the lower mantle. Ringwood (1975; Ringwood and Kesson, 1977) has advocated a chemically uniform pyrolite

mantle with Mg/Si of about 1.5. Others have shown, however, that the seismic data imply a chondritic or pyroxene-rich lower mantle with Mg/Si roughly 1 (Anderson, 1977; Butler and Anderson, 1978). The main seismic evidence for a high SiO₂ content for the lower mantle is the high bulk modulus, K_0 .

In an attempt to resolve the density discrepancy, Ringwood (1975) suggested that enstatite might transform to MgSiO₃ (perovskite) with a density 3 to 7 percent greater than the isochemical mixed oxides. This transformation has subsequently been found (Liu, 1974), but the density increase is only 2 percent relative to the mixed oxides, and the density difference is negligible at high pressure. The pyrolite density discrepancy therefore still remains.

When the newer, more accurate data are used, one concludes that the zero-pressure density (4.1 g/cm³) and bulk modulus (2.6 Mbar) of the lower mantle are coincident with a perovskite rock ($\rho = 4.1 \text{ g/cm}^3$, $K_0 = 2.6 \text{ Mbar}$). Substantial amounts of (Mg,Fe)O, as in the pyrolite models, cannot be tolerated because of the very low K_0 of (Mg,Fe)O. Thus, the methodology advocated by Ringwood, when implemented with modern data, gives support to the alternate hypothesis of a lower mantle rich in pyroxene (in the perovskite structure) with a near-chondritic Mg/Si ratio. These results, based on physical arguments, agree with the chemical model presented here. The issue is not closed since there are trade-offs between composition and temperature, and temperature corrections to physical properties are uncertain. There is certainly no reason, however, to invoke a grossly nonchondritic (major elements) lower mantle unless the recent estimates of solar composition are used for TiO₂ and CaO and the high solar Fe-content translates to a high FeO mantle content.

MAGMAS

Magmas are an important source of information about conditions in the Earth's interior. The bulk composition, trace-element chemistry, isotope geochemistry and volatile content of magmas all contain information about the source region and the processes that have affected the magmas before their eruption. Mantle fragments, or xenoliths, found in these magmas tell us about the material through which the magmas have passed on the way to the surface. Unfortunately, none of the above, even when combined with results from experimental petrology, can determine the depth of the source region, the degree of partial melting involved, or the bulk chemistry of the source region. It will prove useful to discuss the various magma types in the context of their tectonic environment. Representative compositions of various magmas are given in Table 8-6.

Three principal magma series are recognized: tholeiite, calc-alkaline and alkali. The various rock types in each series may be related by varying degrees of partial melting or

crystal separation. The dominant rock type in the voluminous tholeiitic series is tholeiite, a fine-grained dark basalt containing little or no olivine. Tholeiites are found in both oceanic and intraplate settings. Those formed at midocean ridges are low-potassium and LIL-depleted and relatively high in Al₂O₃, while those found on oceanic islands and continents are generally LIL-enriched. The calc-alkaline series is characterized by andesite in island-arc volcanic terranes and granodiorite in plutonic terranes. Tholeiites and calc-alkaline basalts may be derived from a common parent and differ because of the depth of fractionation and the nature and proportions of the minerals that crystallize prior to eruption. The alkali series is characterized by alkali basalt, an olivine-bearing rock that is relatively rich in alkali elements. Tholeiites and alkali basalts are common in midplate environments. Large marginal basins have basalts similar to MORB, while smaller back-arc and fore-arc basins are more commonly calc-alkaline.

Tholeiites often contain large plagioclase crystals but seldom contain mantle fragments or xenoliths. Ridge tholeiites differ from continental and island tholeiites by their higher contents of Al₂O₃ and chromium, low contents of large-ion lithophile (LIL) elements (such as K, Rb, Cs, Sr, Ba, Zr, U, Th and REE) and low ⁸⁷Sr/⁸⁶Sr and ²⁰⁶Pb/²⁰⁴Pb ratios. Tholeiitic basalts have low viscosity and flow for long distances, constructing volcanic forms of large area and small slope. The major- and trace-element differences between the tholeiitic and calc-alkaline suites can be explained by varying proportions of olivine, plagioclase and pyroxene crystallizing from a basaltic parent melt.

Calc-alkaline volcanoes are commonly large, steep-sided stratovolcanoes composed of lava and fragmented material. Their eruptions are typically explosive, in contrast to the eruptions of oceanic island and continental-rift volcanoes.

Explosive volcanism is the most dramatic evidence for high temperatures in the Earth's interior. Spectacular as they are, however, the best known active volcanoes in Italy, Japan, Hawaii, Indonesia, Iceland, Alaska and Washington state represent only a small fraction of the magma that is brought to the Earth's surface. The world-encircling mid-ocean-ridge system, mostly hidden from view beneath the oceans, accounts for more than 90 percent of the material flowing out of the Earth's interior. The whole ocean floor has been renewed by this activity in less than 200 Ma. Hot-spots represent less than 10 percent of the material and the heat flow.

The presence of a crust and core indicates that the Earth is a differentiated body. The age of the crust and the evidence for an ancient magnetic field indicate that differentiation was occurring prior to 3 Ga and may have been contemporaneous with the accretion of the Earth. The high concentrations of incompatible elements in the crust and the high argon-40 content of the atmosphere indicate that the differentiation has been relatively efficient. The presence of

TABLE 8-6
Average Compositions of Basalts and Andesites (Condie, 1982)

	Tholeiite					Oceanic	Cont. Rift	Andesite		
	Ridge	Arc	Continental Rift	Island	High-A1	Alkali Basalt	Alkali Basalt	Arc	Low-K	High-K
SiO ₂	49.8	51.1	50.3	49.4	51.7	47.4	47.8	57.3	59.5	60.8
TiO ₂	1.5	0.83	2.2	2.5	1.0	2.9	2.2	0.58	0.70	0.77
Al ₂ O ₃	16.0	16.1	14.3	13.9	16.9	18.0	15.3	17.4	17.2	16.8
Fe ₂ O ₃ *	10.0	11.8	13.5	12.4	11.6	10.6	12.4	8.1	6.8	5.7
MgO	7.5	5.1	5.9	8.4	6.5	4.8	7.0	3.5	3.4	2.2
CaO	11.2	10.8	9.7	10.3	11.0	8.7	9.0	8.7	7.0	5.6
Na ₂ O	2.75	1.96	2.50	2.13	3.10	3.99	2.85	2.63	3.68	4.10
K ₂ O	0.14	0.40	0.66	0.38	0.40	1.66	1.31	0.70	1.6	3.25
Cr	300	50	160	250	40	67	400	44	56	3
Ni	100	25	85	150	25	50	100	15	18	3
Co	32	20	38	30	50	25	60	20	24	13
Rb	1	5	31	5	10	33	200	10	30	90
Cs	0.02	0.05	0.2	0.1	0.3	2	>3	~0.1	0.7	1.5
Sr	135	225	350	350	330	800	1500	215	385	620
Ba	11	50	170	100	115	500	700	100	270	400
Zr	85	60	200	125	100	330	800	90	110	170
La	3.9	3.3	33	7.2	10	17	54	3.0	12	13
Ce	12	6.7	98	26	19	50	95	7.0	24	23
Sm	3.9	2.2	8.2	4.6	4.0	5.5	9.7	2.6	2.9	4.5
Eu	1.4	0.76	2.3	1.6	1.3	1.9	3.0	1.0	1.0	1.4
Gd	5.8	4.0	8.1	5.0	4.0	6.0	8.2	4.0	3.3	4.9
Tb	1.2	0.40	1.1	0.82	0.80	0.81	2.3	1.0	0.68	1.1
Yb	4.0	1.9	4.4	1.7	2.7	1.5	1.7	2.7	1.9	3.2
U	0.10	0.15	0.4	0.18	0.2	0.75	0.5	0.4	0.7	2.2
Th	0.18	0.5	1.5	0.67	1.1	4.5	4.0	1.3	2.2	5.5
Th/U	1.8	3.3	3.8	3.7	5.9	6.0	8.0	3.2	3.1	2.5
K/Ba	105	66	32	32	12	28	16	58	49	68
K/Rb	1160	660	176	630	344	420	55	580	440	300
Rb/Sr	0.007	0.022	0.089	0.014	0.029	0.045	0.13	0.046	0.078	0.145
La/Yb	1.0	1.7	10	4.2	3.7	11	32	1.1	6.3	4.0

*Total Fe as Fe₂O₃.

helium-3 in mantle magmas shows, however, that outgassing has not been 100 percent efficient. The evidence for differentiation and a hot early Earth suggest that much of the current magmatism is a result of recycling or the processing of already processed material. The presence of helium-3 in the mantle suggests that a fraction of the magma generated remains in the mantle; that is, magma removal is inefficient.

MIDOCEAN-RIDGE BASALTS (MORB)

The floor of the ocean, under the sediments, is veneered by a layer of tholeiitic basalt that was generated at the mid-ocean ridge systems. The pillow basalts that constitute the upper part of the oceanic crust extend to an average depth

of 1 to 2 km and are underlain by sheeted dikes, gabbros and olivine-rich cumulate layers. The average composition of the crust is much more MgO-rich and lower in CaO and Al₂O₃ than the surface MORB, if ophiolites can be used as a guide. The crust rests on a depleted harzburgite layer of unknown thickness. In certain models of crustal genesis, the harzburgite layer is complementary to the crust and is therefore about 24 km thick if an average crustal thickness of 6 km and 20 percent melting are assumed. This is just a model, and it is not clear that harzburgites are genetically related to the basalts. Seismic data are consistent with a harzburgite layer in the shallow mantle, but its thickness has not been determined. Oceanic plateaus and aseismic ridges have anomalously thick crust (>30 km), and the corresponding depleted layer would be more than 120 km thick if the simple model is taken at face value. Since depleted peridotites, including harzburgites, are less dense than

Al₂O₃-rich or fertile peridotite, or eclogite, several cycles of plate tectonics, crust generation and subduction, would fill up the shallow mantle with harzburgite. Oceanic crust may in fact be deposited on ancient, not contemporaneous, harzburgite.

Obviously, midocean-ridge basalts result from the partial or complete melting of something with a basaltic component. The composition of the source region and the degree of partial melting are unknown. The smaller the degree of melting, the larger is the problem with excess residual peridotite. Also unknown are the amount and nature of phases that have crystallized from the parent material prior to the eruption of MORB. Some olivine fractionation is commonly assumed, but garnet and pyroxene crystallization may be involved if the parent magma is trapped beneath the lithosphere. At shallow depths the crystallization of plagioclase can extensively modify the composition of the remaining melt. In general, the crystallization of refractory phases increases the content of potassium, rubidium and other incompatible elements in the melt. These questions are explored in more detail in Chapter 9. Note the similarity of tholeiites from various tectonic environments as far as the major elements are concerned but the depletion of ridge tholeiites in the LIL elements (Table 8-6).

Models of MORB genesis fall into two categories. In the first, the most primitive MORB samples are considered to represent "primary" magmas formed by partial melting of the mantle beneath the midocean ridges and little modified by crystal fractionation or other processes once they segregated from this source region. In the second category it is maintained that MORB is not primary; it evolved from a parent magma by substantial amounts of fractionation, principally of olivine and perhaps other phases as well, at pressures less than those at which the primary magma separated from the mantle source region. Primary magmas would be picritic, or rich in an olivine component, and may contain about 20 percent MgO. If the parent magma ascends from great depth and is allowed to fractionate at depths of the order of 50 km or greater, it may have precipitated garnet and clinopyroxene as well as olivine.

In most models of MORB genesis, it is assumed not only that olivine and orthopyroxene are contained in the source rock but that these are the dominant phases. Petrological data alone, however, cannot rule out a source that is mainly garnet and clinopyroxene. Primitive mantle and samples from the shallow mantle contain abundant olivine and orthopyroxene. However, MORB are extremely depleted in the incompatible elements, and isotopic ratios show that the depletion of the MORB source occurred early in Earth history. Therefore, the MORB source is not primitive, nor has it been mixed with much primitive or enriched material. It may represent a deep garnet-clinopyroxene-rich cumulate layer or a peridotite layer that experienced a small amount of partial melting and melt removal. The melting in the latter case could not have been too extensive since the

source is still fertile, that is, it contains clinopyroxene and garnet, the low-melting components.

The eclogite (garnet plus clinopyroxene) or olivine eclogite source hypothesis differs only in scale and melt extraction mechanism from the fertile peridotite hypothesis. In the fertile peridotite model the early melting components, garnet and clinopyroxene, are distributed as grains in a predominantly olivine-orthopyroxene rock. On a larger scale, eclogite might exist as pods or blobs in a peridotite mantle. Since eclogite is denser than peridotite, these blobs would tend to sink and coalesce in a hot, low-viscosity mantle. In the extreme case an isolated eclogite-rich layer might form below the lighter peridotite layer. Such a layer could also form by subduction or by crystal settling in an ancient magma ocean. The isotopic evidence for isolated reservoirs and the geophysical evidence for gross layering suggest that differentiation and chemical stratification may be more important in the long run than mixing and homogenization. This is not to deny that magma mixing occurs in the processes leading up to eruption.

The important geophysical questions regarding the MORB source region and MORB genesis are (1) depth of source, (2) amount of olivine and orthopyroxene in the source, (3) the extent of melting, (4) the depth of melt-crystal separation (or final equilibration), (5) the amount of crystal fractionation subsequent to melt separation from the source but prior to eruption, and (6) the amount of contamination. The answers to most of these questions are assumed in many models of basalt genesis. It is also generally assumed that the upper mantle is homogeneous and that non-MORB magmas must originate elsewhere. In fact, the upper mantle is neither radially nor laterally homogeneous.

HOTSPOTS

Most of the world's visible volcanoes occur at plate boundaries and are associated with plate creation and consumption. There are about 50 other volcanic centers, scattered about the globe, that are either remote from plate boundaries or that are anomalous in the volume or chemistry of eruptive material (Figure 8-4). The numerous linear island chains in the central Pacific are perhaps the most dramatic manifestations of long-lived hotspot activity. Hawaii and Tahiti are examples of hotspots in the Pacific; Iceland, the Azores, Tristan de Cunha and the Canaries are Atlantic examples. Réunion and Kerguelen are hotspots in the Indian Ocean, and tiny Bouvet Island is in the Southern Ocean.

J. Tuzo Wilson (1963) and W. Jason Morgan (1971) proposed an important additional element to plate tectonics. The concepts of seafloor spreading and continental drift leave unexplained many aspects of seafloor and continental geology and geochemistry; not the least of these are the driving forces for plate tectonics and the requirement for

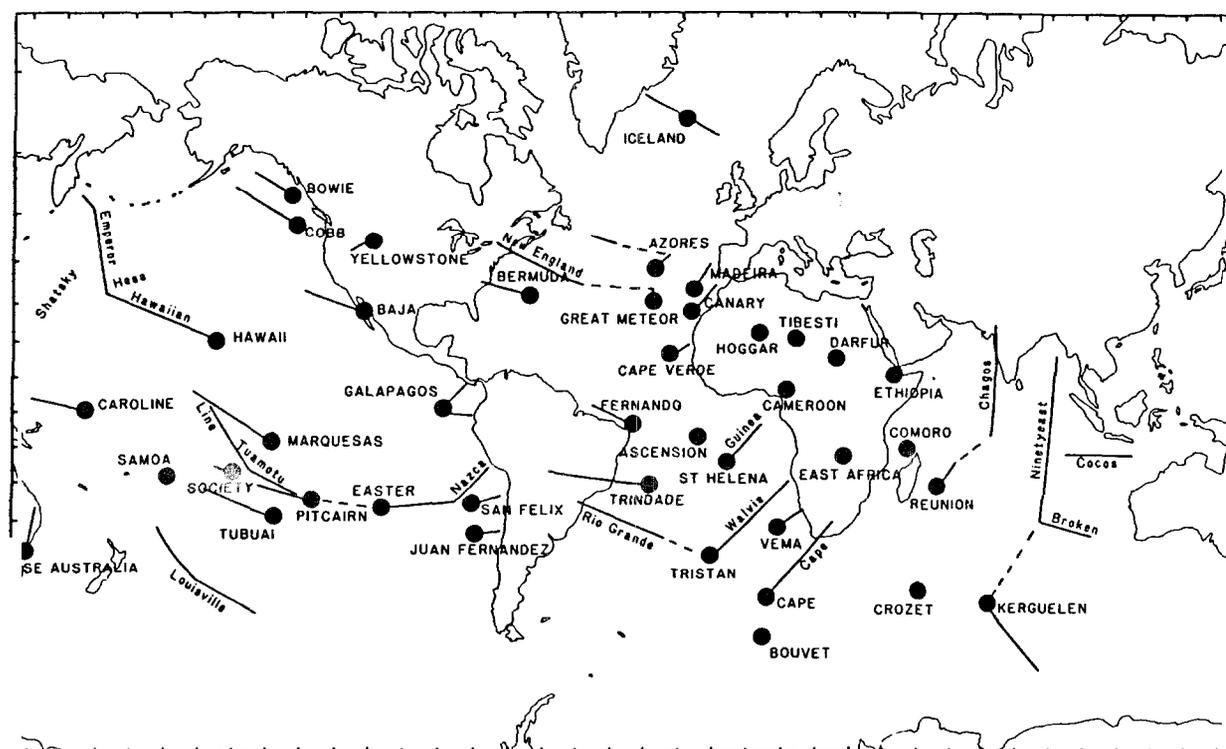


FIGURE 8-4
Global distribution of hotspots and hotspot tracks (after Crough, 1983)

multiple magma sources in the mantle. Wilson proposed that hotspots, or melting spots in the mantle, could be used as absolute reference points. Morgan (1982) used hotspots as fixed reference points in his continental reconstructions to about 200 million years ago. Morgan (1972) further proposed that hotspots had a deep-mantle, thermal, origin and that they drove plate motions. He interpreted hotspots in terms of thermal convection, a concentrated upwelling of hot material from the deep mantle.

The Wilson-Morgan hypothesis attributes linear volcanic chains to thermal plumes that originate deep in the mantle. A relatively fixed mantle source of heat seems to be required by the age progression data, but a deep mantle source for the material is not required by either the geochemical or geophysical data. For this reason others have proposed that localities such as Hawaii and Iceland be termed "hotspots" or "melting spots" rather than "plumes," a term that has deep mantle implications. Alternative mechanisms include propagating fractures, shear melting, bumps on the asthenosphere and gravitational anchors. These hypotheses do not directly address the problem of why melts, tears or bumps occur in the first place.

In 1975 I introduced the concept of "chemical plumes" and proposed that plumes differ in intrinsic chemistry from ordinary mantle, rather than being just thermal perturbations (Anderson, 1975). Chemical inhomogeneity of the

lower mantle, for example, could be caused by upwelling of material from a chemically distinct layer at the base of the mantle, D", which is also the location of the mantle-core thermal boundary layer. If this material is intrinsically denser than normal lower mantle it would rise only when hot, much in the manner of "lava lamps" where hot wax blobs rise through oil, cool and then sink. Even if lower-mantle material cannot cross into the upper mantle, the locations of hot regions may fix the locations of melting anomalies in the upper mantle. The composition of lower mantle plumes, thermal or chemical, may be unrelated to hotspot magmas. If the upper mantle is chemically stratified, upwellings from the deeper layer may contribute material different in chemistry from melts from the shallow mantle. Plumes in the upper mantle could originate near or above the 650-km discontinuity or where melting is most extensive.

There is now abundant evidence that hotspots provide magmas that are chemically distinct from MORB. The thermal plume hypothesis must therefore be abandoned. Hotspot magmas, however, are similar to MORB in major elements, and it may not be necessary to invoke a distinct reservoir (in bulk chemistry or location). The trace-element and isotopic characteristics of hotspot magmas may have been acquired during ascent, or during evolution in the shallow mantle.

The concept of relatively fixed deep mantle plumes was initially proposed to explain the age progression of linear volcanic chains in the Atlantic and Pacific. Before the introduction of the plume concept, many geophysicists believed that convection in the mantle should be restricted to the low-velocity layer or asthenosphere. This seemed reasonable since the combined effects of temperature and pressure on melting and viscosity would lead to lower viscosities and greater degrees of partial melting in the upper mantle. In this view, most of the material presently rising to the Earth's surface is recycled, and present centers of volcanic activity, such as midoceanic rises, are passive reflections of where cracks form. If "fundamental" volcanic centers, variously called plumes, hotspots or melting spots, are relatively motionless or slowly moving, then a more profound fixed source of heat or melting is implied.

Hotspots appear to define a relatively stable reference frame that is sometimes assumed to be fixed to the lower mantle. However, if the upper mantle and lower mantle are convecting, either as a single system or separately, hotspots should wander relative to one another. The return flow associated with plate tectonics would be expected to disrupt the pattern or, at least, move cells in one hemisphere relative to another. However, this motion is expected to be much smaller than plate velocities. For example, if the slabs are 50 km thick and the return flow occurs over the upper 500 km of the mantle, the relative motions of hotspots will be only one-tenth of the relative motions of the plates. This is consistent with observations. Return flow may also be below the upper-mantle convection cells, down in the transition zone or the lower mantle.

Although it is generally agreed that the lower mantle is convecting, it is likely to be a much more sluggish convection than occurs in the upper mantle. Convection rates depend on viscosity and temperature contrasts. Pressure serves to increase viscosity. High stress decreases viscosity. Near-surface temperatures range from near the melting point of rocks to below 0°C. Such extremes cannot occur in the deeper mantle. The transition zone or the lower mantle is therefore a better candidate than the lithosphere, the spin axis or the liquid core for a stable, or at least slowly varying reference system. A convecting lower mantle will have relatively stable uprising and descending limbs, which may show up in the geoid. Convection deforms the core-mantle and the upper mantle-lower mantle boundaries. This deformation will influence convection in the core and the upper mantle, even if no material transfer takes place. For example, hotspots may be preferentially located over the hot upcurrents in the lower mantle, and then hotspots would indeed appear fixed to the lower mantle even if hotspot magmas originate in the upper mantle. On the other side of the system, downwellings in the core occur preferentially under colder parts of the lower mantle, these being where heat is most effectively removed.

Both continental and oceanic geology give ample evidence of uplift and subsidence throughout the recognizable

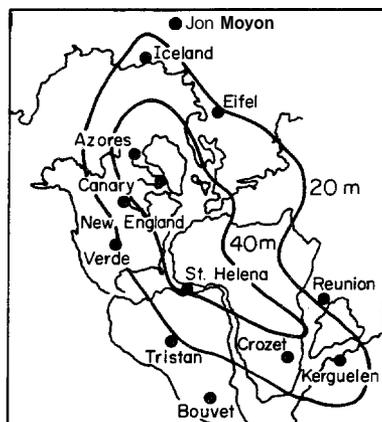


FIGURE 8-5 Distribution of hotspots relative to Pangaea and the Atlantic-Africa geoid high (contours) at 200 Ma (after Anderson, 1982c).

geologic record. The concepts of mobile plates combined with hotspots or mantle upwellings give a natural explanation for these phenomena. A plume, either thermal or chemical, will generate "bumps" at the base of the upper mantle, the asthenosphere or the lithosphere that can contribute to uplift and, ultimately, fracture and magmatism in the overlying lithosphere. Crustal uplift may be due to lithosphere overriding mantle upwellings. Thermal expansion, partial melting and upwelling all contribute to lithospheric uplift and extension.

The distribution of hotspots on the surface of the Earth is not random. Most hotspots occur in the long-wavelength geoid highs centered in the central Pacific and Africa (Figure 8-5); few occur in geoid lows. These geoid highs correlate well with low seismic velocity regions in the lower mantle (Chapter 15). In a general sense the broad-scale distribution of hotspots seems to be controlled by hotter than average regions in the lower mantle. This does not imply that material is actually transferred from the lower mantle to the surface. The presumed high viscosity of the lower mantle makes convection sluggish, and hot areas of the lower mantle appear more permanent than surface features. Hot areas in the lower mantle may be due to upwellings from the thermal boundary layer at the core-mantle boundary, region D". This, in turn, may be chemically distinct from most of the lower mantle.

The hotspot and plume concepts were developed from geometric and kinematic observations involving the relative fixity and large volume of certain melting spots. It later became evident that these hotspots differ in trace-element and isotopic chemistry from ocean-ridge volcanism. Hotspot magmas are richer in volatiles and LIL, and have high $^{87}\text{Sr}/^{86}\text{Sr}$, $^{144}\text{Nd}/^{143}\text{Nd}$, $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ ratios and, in some cases, high $^3\text{He}/^4\text{He}$ ratios. Their source is therefore chemically distinct, and has been for some time, from that providing the more voluminous MORB.

Hotspot magmas, including tholeiites and alkali basalts, are intermediate in LIL and isotope chemistry between MORB, the most depleted magmas, and kimberlites, in many ways the most enriched magmas. The bulk chemistry of ocean-island and continental flood basalts, two manifestations of hotspot magmatism, are similar to midocean-ridge basalts (Table 8-6), suggesting that the geochemical signatures of hotspot magmas may reflect contamination of MORB, or a magma mixture that may involve only a small fraction of an enriched component, such as Q. Although there are subtle chemical and isotopic differences between ocean-island (hotspot) and ocean-ridge basalts, most of the material coming out of both regions probably originates in the upper mantle, some or all of which has possibly been recycled. The dilution of plume material by "normal" upper-mantle melts (MORB) makes it difficult to unravel the chemistry of plumes.

Large oceanic islands, such as Iceland, Hawaii, Kerguelen and Tahiti, smaller islands such as Tristan da Cunha, St. Helena, Canaries and Ascension, and continental flood basalt and alkali basalt provinces are considered to be manifestations of hotspot volcanism. Carbonatites and kimberlites may be early continental manifestations of hotspot activity. Carbonatites and kimberlites are exotic assemblages that apparently result from anomalous chemical or thermal conditions of the upper mantle. The spatial, petrologic and geochemical relationship between kimberlites and carbonatites is well established. These suites commonly occur in stable continental areas and form trends from continent to continent. Carbonatites or related rock types occur on many of the proposed Atlantic and Indian Ocean hotspots or on their tracks. In several cases continental carbonatites, kimberlites or alkali-rich provinces abut proposed hotspots when the continents are reconstructed. The many geochemically similar seamounts on the ocean floor and evidence for simultaneity of magmatism over large areas, once thought to be hotspot tracks, makes the concept of a relatively small number of fixed deep mantle plumes less attractive than when originally proposed. Island-arc and back-arc basin basalts have many of the same geochemical properties as do hotspots, suggesting that the distinctive geochemical signatures of hotspot magmas may be acquired at shallow depth. In these regions a cold slab intervenes between the source region and the deep mantle.

The similarity in major-element chemistry of ocean floor, ocean island and continental rift basalts, or their inferred parents, suggests that all may share a common source, differing only in their trace elements, isotopes and cooling/fractionation history. For example, hotspot magmas may be contaminated MORB, intruded at depth and cooled prior to eruption. In fact, most basalts can be treated as mixtures of MORB and an enriched component.

An important and controversial problem in geochemistry and petrology is the source of the hotspot reservoir. This is actually three separate problems, usually treated as one:

1. The source of the thermal anomaly.
2. The source of the magma.
3. The source of the distinctive trace-element, volatile and isotopic signature.

That these are not necessarily the same question can be seen from the following possible scenario: Convection currents in the lower mantle provide nonuniform heating to the base of the upper mantle. If the mantle is chemically stratified, no mass transfer takes place from the lower to the upper mantle. This nonuniform heating localizes sites of upwelling and melting in the transition region, which provides depleted MORB-type magmas. These magmas rise through the LIL- and volatile-rich peridotitic shallow mantle and pick up the geochemical signature of the enriched low-velocity zone or lower lithosphere. Enriched alkali or plume-type magmas form by fractionation of this hybrid magma at sublithospheric depths. Ancient magmatic events probably involved a component of crustal and lithospheric underplating and intrusion as well as extrusion. Modern magmatism through these regions will acquire a hotspot geochemical signature even if the initial melt is MORB, particularly the early stages.

Ocean-Island Basalts

Oceanic islands are composed of both alkali and tholeiitic basalts. Alkali basalts are subordinate, but they appear to dominate the early and waning stages of volcanism. Ocean-island basalts, or OIB, are LIL-rich and have enriched isotopic ratios relative to MORB. Their source region is therefore enriched, or depleted parent magmas suffer contamination en route to the surface. The larger oceanic islands such as Iceland and Hawaii generally have less enriched magmas than smaller islands and seamounts. A notable exception is Kerguelen, which has very high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and low $^{143}\text{Nd}/^{144}\text{Nd}$ ratios (Dosso and Murthy, 1980).

The newest submarine mountain in the Hawaii chain, Loihi, is alkalic. Intermediate age islands are tholeiitic, and the latest stage of volcanism is again alkalic. The volcanism in the Canary Islands in the Atlantic also changes from alkalic to tholeiitic. The sequence suggests that the hotspot over which the islands are progressing is supplying voluminous tholeiitic magmas and the alkalis are being provided on the edges, perhaps from the wings of a magma chamber.

Trace-element and isotopic patterns of OIB overlap continental flood basalts (CFB), and a common source region and fractionation pattern can be inferred. Island-arc basalts (IAB) also share many of the same geochemical characteristics, suggesting that the enriched character of these basalts may be derived from the characteristics of the shallow mantle. This is not the usual interpretation. The geochemistry of CFB is usually attributed to continental

contamination, that of IAB to sediment and subducted crust involvement, and that of OIB to "primitive" lower mantle. In all three cases the basalts have evolved beneath thick crust or lithosphere, giving more opportunity for crystal fractionation and contamination by crust, lithosphere or shallow mantle (the low-velocity zone) prior to eruption.

Kimberlite

Kimberlite is a rare igneous rock that is volumetrically insignificant compared to other igneous rocks such as basalt. Kimberlite provinces themselves, however, cover very broad areas and occur in most of the world's stable craton, or shield, areas. Kimberlites are best known as the source rock for diamonds, which must have crystallized at pressures greater than about 50 kilobars. They carry other samples from the upper mantle that are the only direct samples of mantle material below about 100 km. Some kimberlites appear to have exploded from depths as great as 200 km or more, ripping off samples of the upper mantle and lower crust in transit. The fragments, or xenoliths, provide samples unavailable in any other way. Kimberlite itself is an important rock type that provides important clues as to the evolution of the mantle. It contains high concentrations of lithophile elements (Table 8-7) and higher concentrations of the incompatible trace elements than any other ultrabasic rock. Although rare, kimberlites or related rocks have been found on most continents including Africa, Asia, Europe, North America, South America, Australia, India and Greenland. Most kimberlites are found in ancient cratons and relatively few are found in circumcratonic fold-belts. The most important areas for diamond production are in South Africa, Siberia and northwest Australia, although diamonds are also found in North America and Brazil. Dia-

TABLE 8-7
Kimberlite Composition Compared with Ultrabasic and Ultramafic Rocks

Oxide	Average Kimberlite	Average Ultrabasic	Average Ultramafic
SiO ₂	35.2	40.6	43.4
TiO ₂	2.32	0.05	0.13
Al ₂ O ₃	4.4	0.85	2.70
	6.8	—	—
FeO	2.7	12.6	8.34
MnO	0.11	0.19	0.13
MgO	27.9	42.9	41.1
CaO	7.6	1.0	3.8
Na ₂ O	0.32	0.77	0.3
K ₂ O	0.98	0.04	0.06
H ₂ O	7.4	—	—
CO ₂	3.3	0.04	—
P ₂ O ₅	0.7	0.04	0.05

Wederpohl and Muramatsu (1979).

Dawson (1980).

mond-bearing kimberlites are usually close to the craton's core, where the lithosphere may be thickest. Barren kimberlites are usually on the edges of the tectonically stable areas. Kimberlites range in age from Precambrian to Cretaceous. Some areas have been subjected to kimberlite intrusion over long periods of geological time.

Compared to other ultrabasic rocks such as lherzolite, dunite or harzburgite, kimberlites contain unusually high amounts of K₂O, Al₂O₃, TiO₂, FeO, CaO, CO₂, P₂O₅ and H₂O. For most incompatible trace elements kimberlites are the most enriched rock type; important exceptions are elements that are retained by garnet and clinopyroxene.

Since kimberlites are extremely enriched in the incompatible elements, they are important in discussions of the trace-element inventory of the Earth's mantle. Such extreme enrichment implies that kimberlites represent a small degree of partial melting of a mantle silicate or a late-stage residual fluid of a crystallizing cumulate layer. The LIL elements in kimberlite show that it has been in equilibrium with a garnet-clinopyroxene-rich source region, possibly an eclogite cumulate. The LIL contents of kimberlite and MORB are complementary. Removal of a kimberlite-like fluid from an eclogite cumulate gives a crystalline residue that has the required geochemical characteristics of the depleted source region that provides MORB.

Carbonatites and other ultramafic alkaline rocks, closely related to kimberlites, are widespread. In spite of their rarity, kimberlites provide us with a sample of magma that probably originated below about 200 km and, as such, contain information about the chemistry and mineralogy of the mantle in and below the continental lithosphere. Kimberlites are anomalous with respect to other trace-element-enriched magmas, such as nephelinites and alkali basalts, in their trace-element chemistry. They are enriched in the very incompatible elements such as rubidium, thorium and LREE, consistent with their representing a small degree of partial melting or the final concentrate of a crystallizing liquid. However, they are relatively depleted in the elements (Sc, Ti, V, Mn, Zn, Y, Sr and the HREE) that are retained by garnet and clinopyroxene. They are also low in silicon and aluminum, as well as other elements (Na, Ga, Ge) that are geochemically coherent with silicon and aluminum. This suggests that kimberlite fluid has been in equilibrium with an eclogite residue. Kimberlites are also rich in cobalt and nickel.

Despite their comparative rarity, disproportionately high numbers of eclogite xenoliths have been found to contain diamonds. Diamond is extremely rare in peridotitic xenoliths. Eclogitic garnets inside diamonds imply a depth of origin of about 200–300 km if mantle temperatures in this depth range are of the order of 1400–1600°C (Cohen and Rosenfeld, 1979). Seismic velocities between 400 and 670 km depth in the mantle are consistent with picritic eclogite (Bass and Anderson, 1984). The bulk modulus in the transition zone for olivine in its spinel forms is higher than

is consistent with the seismic data. Eclogites may originate in the transition region or they may be subducted oceanic crust.

It is obvious that if a fluid similar to kimberlite is the enriching or metasomatic fluid that is responsible for the trace-element pattern of alkali basalts, and other enriched continental and ocean-island magmas, then these magmas will appear to have been in equilibrium with a garnet-rich residue, even if no garnet is present in their immediate parent reservoir. That is to say, a magma with a garnet-equilibrium signature in its trace-element pattern, for instance HREE depletion, does not necessarily imply that it is a result of small degrees of partial melting of a garnet-bearing parent, leaving behind a garnet-rich residue. The signature could be inherited from the parent reservoir of the enriching fluid, and probably has been if kimberlitic fluids are the enriching fluids. Similar comments apply if the immediate parent of continental basalts had experienced a prior episode of eclogite extraction, as in a deep magma chamber or a deep magma ocean.

Alkali basalts have LIL concentrations that are intermediate to MORB and kimberlite. Although kimberlite

pipes are rare, there may be a kimberlite-like component (Q) dispersed throughout the shallow mantle. Indeed, alkali basalts can be modeled as mixtures of a depleted magma (MORB) and an enriched magma (kimberlite), as shown in Table 8-8. Peridotites with evidence of secondary enrichment may also contain a kimberlite-like component.

Kimberlites commonly contain calcium zeolites and anorthite and contain higher concentrations of such refractory trace elements as barium, strontium, lanthanum, zirconium and niobium than *would* be expected *from racks* originating in normal mantle. There are many other minor elements in excess (including Li, B, Ca, Ga, Rb, Cs, Ta and Pb). Xenoliths in kimberlites commonly contain rutile, kyanite, garnet, corundum, spinel, ilmenite, perovskite, nepheline, anatase, strontianite and barite and are sometimes associated with flows of olivine basalt unusually rich in barium, titanium, zirconium, niobium, P₂O₅ and H₂O. The ⁸⁷Sr/⁸⁶Sr ratio for fresh kimberlites is often high. A rock often associated with kimberlite is melilite basalt. There is a complete chemical gradation between kimberlites and carbonatites; both are enriched in barium, phosphorus, lanthanum and niobium. In addition carbonatites are com-

TABLE 8-8
Trace-Element Chemistry of MORB, Kimberlite and Intermediate (Alkali) Basalts (ppm)

Element	Alkali Basalts						Kimberlite	
	MORB	EPR*	Australia	Hawaii	BCR [†]	Theoretical [‡]	Average	Max
Sc	37	27	25	33	34–37	35	30	—
V	210	297	260	170	399	197–214	120	250
Co	53	41	60	56	38	57–61	77	130
Ni	152	113	220	364	15	224–297	1050	1600
Rb	0.36	13	7	24	47	10–45	65	444
Sr	110	354	590	543	330	200–289	707	1900
Y	23	37	20	27	37	23–28	22	75
Zr	70	316	111	152	90	97–133	250	700
Nb	3.3	—	24	24	14	19–48	110	450
Ba	5	303	390	350	700	150–580	1000	5740
La	1.38	21	24	23	26	16–31	150	302
Ce	5.2	46	47	49	54	31–57	200	522
Nd	5.6	28	24	23	29	16–26	85	208
Sm	2.1	6.7	6.8	5.5	6.7	3.1–4.8	13	29
Eu	0.8	2.1	2.1	2.0	2.0	1.1–1.4	3	6.5
Tb	0.5	1.2	0.92	0.87	1.0	0.6–0.7	1	2.1
Yb	2.1	3.4	1.6	1.8	3.4	2.0–2.1	1.2	2.0
Lu	0.34	0.49	0.28	0.23	0.55	0.32–0.34	0.16	0.26
Hf	1.4	5.3	2.8	3.9	4.7	2.2–4.3	7	30
Ta	0.1	9	1.8	2.1	0.9	1.3–2.5	9	24
Pb	0.08	—	4.0	5	18	1.6–5.1	10	50
Th	0.35	2.7	2.0	2.9	6.8	2.4–5.4	16	54
K	660	8900	4700	8800	1200	3200–4770	17,600	41,800
U	0.014	0.51	0.4	0.6	1.8	0.48–1.84	3.1	18.3

* EPR = East Pacific Rise.

† BCR = Basalt, Columbia River (USGS Standard Rock).

‡ Mixture ranging from 85 percent MORB plus 15 percent average kimberlite to 90 percent MORB plus 20 percent maximum kimberlite.

monly enriched in uranium, thorium, molybdenum and tungsten, and kimberlites are enriched in boron, vanadium and tantalum.

Alkali ultrabasic rocks in ring complexes usually occur in stable or fractured continental regions; they contain nepheline, melilite-rich rocks, carbonatites, and kimberlite dikes. Kimberlites are mostly confined to interiors and margins of stable continental areas. The distribution appears to be related to deep-seated tectonics with linear trends. There appear to be genetic links between kimberlites, carbonatites and alkalic ultrabasic complexes.

The intrusion of kimberlites takes place, during uplift of platform areas, along deep-seated fractures either bounding or cutting across the uplifted areas. In South Africa, Brazil and the USSR, kimberlite eruptions closely followed widespread extrusions of flood basalt.

Extensive magmatic and metamorphic events in continental interiors may occur when sublithospheric heat sources are focused for long periods of time, many millions of years, on the same part of the lithosphere. The localization of this effect, rather than being a continent-wide phenomenon, also suggests a rather localized source. During the past 800 Ma Africa has been the site of three main thermal episodes. In some cases uplift and fracturing preceded volcanism.

The Kimberlite-MORB Connection

Kimberlites are enriched in the LIL elements, especially those that are most depleted in MORB. Figure 8-6 gives the composition of kimberlites, MORB and continental tholeiites. The complementary pattern of kimberlite and MORB is well illustrated as is the intermediate position of continental tholeiites (CFB). Note that kimberlite is not enriched in the elements that are retained by the eclogite minerals, garnet and clinopyroxene. This is consistent with kimberlite having been a fluid in equilibrium with subducted oceanic crust or an eclogite cumulate. If a residual cumulate is the MORB reservoir, the ratio of an incompatible element in kimberlite relative to the same element in MORB should be the same as the liquid/solid partition coefficient for that element. This is illustrated in Figure 8-7. The solid line is a profile of the MORB/kimberlite ratio for a series of incompatible elements. The vertical lines bracket the solid/liquid partition coefficients for garnet and clinopyroxene. Although MORB is generally regarded as an LIL-depleted magma and kimberlite is ultra-enriched in most of the incompatible elements, MORB is enriched relative to KIMB in yttrium, ytterbium and scandium, elements that have a high solid/melt partition coefficient for an eclogite residue. The trend of the MORB/KIMB ratio is the same as the partition coefficients, giving credence to the idea that kimberlite and MORB are genetically related. MORB is now generally regarded as descended from a more picritic magma by olivine fractionation. The dashed line shows the LIL ra-

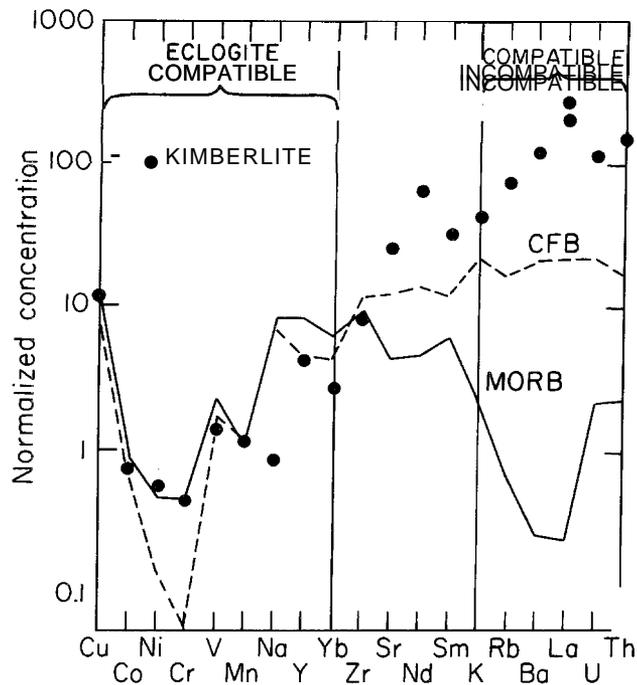


FIGURE 8-6 Trace-element concentrations in MORB, continental flood basalts (CFB) and kimberlites. The elements to the right are incompatible in all major mantle phases (olivine, pyroxene and garnet) while those to the left are retained in the eclogite minerals (clinopyroxene and garnet). Note the complementary pattern between MORB and kimberlite and the intermediate position of CFB. Concentrations are normalized to estimates of mantle composition.

tios of MORB plus 50 percent olivine, relative to KIMB. Its fit with the partition coefficients indicates that the MORB parent magma could have lost up to about 50 percent olivine, presumably in shallow upper-mantle magma chambers.

These results support the joint hypotheses that (1) the MORB reservoir is an eclogite-rich cumulate and (2) kimberlite was in equilibrium with eclogite and could be the late-stage melt of a deep crystallizing cumulate layer. A kimberlite-like fluid may be partially responsible for the LIL enrichment of the shallow mantle as well as the depletion of the MORB reservoir. An eclogite-rich cumulate could be the result of crystallization of a magma ocean. Such crystallization would tend to separate olivine and orthopyroxene, the shallow-mantle minerals, from garnet and clinopyroxene.

The LIL content of continental tholeiites and alkali-olivine basalts suggest that they are mixtures of MORB and a melt from a more enriched part of the mantle. If diapirs initiate in a deep depleted layer, they must traverse the shallow mantle during ascent, and cross-contamination seems unavoidable. The more usual model is that the whole upper

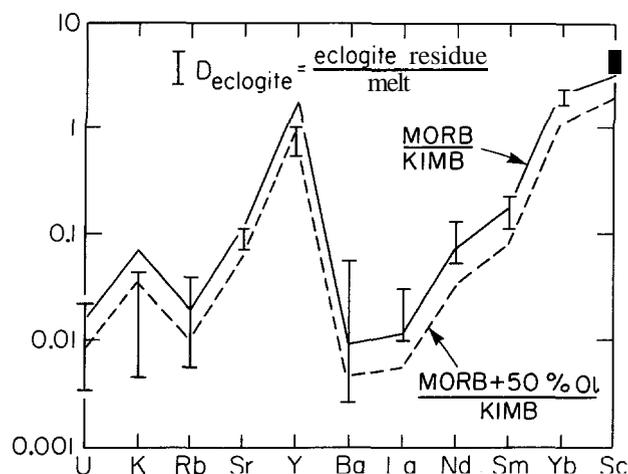


FIGURE 8-7

Solid line is ratio of concentrations in MORB and kimberlite (KIMB). Vertical bars are solid/liquid partition coefficients for garnet plus clinopyroxene. The dashed line gives the ratio for MORB plus 50 percent olivine, a possible picrite parent magma for MORB. If the MORB or picrite source region is the crystalline residue remaining after removal of a kimberlitic fluid, the ratio of concentrations, MORB/KIMB or picrite/KIMB, should equal the solid/liquid partition coefficient D_{sl} , where D_{sl} depends on the crystalline (xl) phases in the residue. D_{ec} is the solid/liquid partition coefficient for an eclogite residue (after Anderson, 1983b).

mantle is homogeneous and provides depleted (MORB) magmas. Enriched magmas, in this scenario, must come from the lower mantle or from the crust or continental lithosphere. The alternative is that the upper mantle is both radially and laterally inhomogeneous, an alternative adopted here. It should be stressed that there is no evidence that any magma or magma component originates in the lower mantle. This applies to both direct and indirect, or mass balance, evidence.

The Kimberlite-KREEP Relation

KREEP is a lunar material having very high concentrations of incompatible elements (K, REE, P, U, Th) and so on (see Chapter 2). It is thought to represent the residual liquid of a crystallizing magma ocean (Warren and Wasson, 1979). Given proposals of a similar origin for kimberlite (Anderson, 1982a, 1982b), it is of interest to compare the composition of these two materials. An element by element comparison gives the remarkable result that for many elements (K, Cs, P, S, Fe, Ca, Ti, Nb, Ta, Th, U, Ba and the LREE) kimberlite is almost identical (within 40 percent) to the composition of KREEP (Figure 8-8 and 8-9). This list includes compatible and incompatible elements, major, minor and trace elements, and volatiles as well as refractories. KREEP is relatively depleted in strontium and europium, elements that have been removed from the KREEP source

region by plagioclase fractionation. Kimberlite is depleted in sodium, HREE, hafnium, zirconium and yttrium, elements that are removed by garnet-plus-clinopyroxene fractionation. It appears that the differences between KIMB and KREEP are due to difference in pressure between the Earth and the Moon; garnet is not stable at pressures occurring in the lunar mantle. The plagioclase and garnet signatures show through such effects as the different volatile-refractory ratios of the two bodies and expected differences in degrees of partial melting, extent of fractional crystallization and other features. The similarity in composition extends to metals of varying geochemical properties and volatilities such as iron, chromium, manganese as well as phosphorus. KREEP is depleted in other metals (V, Co, Ni, Cu and Zn), the greater depletions occurring in the more volatile metals and the metals that are partitioned strongly into olivine. This suggests that olivine has been more important in the evolution of KREEP than in the evolution of kimberlite, or that cobalt, copper and nickel are more effectively partitioned into MgO-rich fluid such as kimberlite.

It appears from Figure 8-8 that KREEP and kimberlite differ from chondritic abundances in similar ways. Both are depleted in volatiles relative to refractories, presumably due to preaccretionary processes. Strontium is less enriched than the other refractories, although this is much more pronounced for KREEP. The pronounced europium and strontium anomalies for KREEP are consistent with extensive plagioclase removal. The HREE, yttrium, zirconium and hafnium are relatively depleted in kimberlite, suggesting eclogite fractionation or a garnet-rich source region for kimberlite. The depletion of scandium simply indicates that olivine, pyroxene or garnet have been in contact with both KREEP and KIMB. The depletion of sodium in KIMB is also consistent with the involvement of eclogite in its history. The depletion of sodium in KREEP, relative to the other alkalis, is presumably due to the removal of feldspar, and the greater relative depletion of sodium in kimberlite therefore requires another process, such as removal of a jadeite component by eclogite fractionation.

Chromium, manganese and iron are approximately one-third the chondritic level in both KREEP and kimberlite. This is about the level in the mantles of these bodies. Nickel and cobalt in kimberlite are about the same as estimated for the Earth's mantle. These elements are extremely depleted in KREEP, indicating that they have been removed by olivine or iron extraction from the source region.

The similarity of kimberlite and KREEP is shown in Figure 8-9. For many elements (such as Ca, Ba, Nd, Eu, Nb, Th, U, Ti, Li and P) the concentrations are identical within 50 percent. Kimberlite is enriched in the volatiles rubidium, potassium and sulfur, reflecting the higher volatile content of the Earth. Kimberlite is also relatively enriched in strontium and europium, consistent with a prior extraction of plagioclase from the KREEP source region. Kimberlite is depleted in the HREEs and sodium, consistent

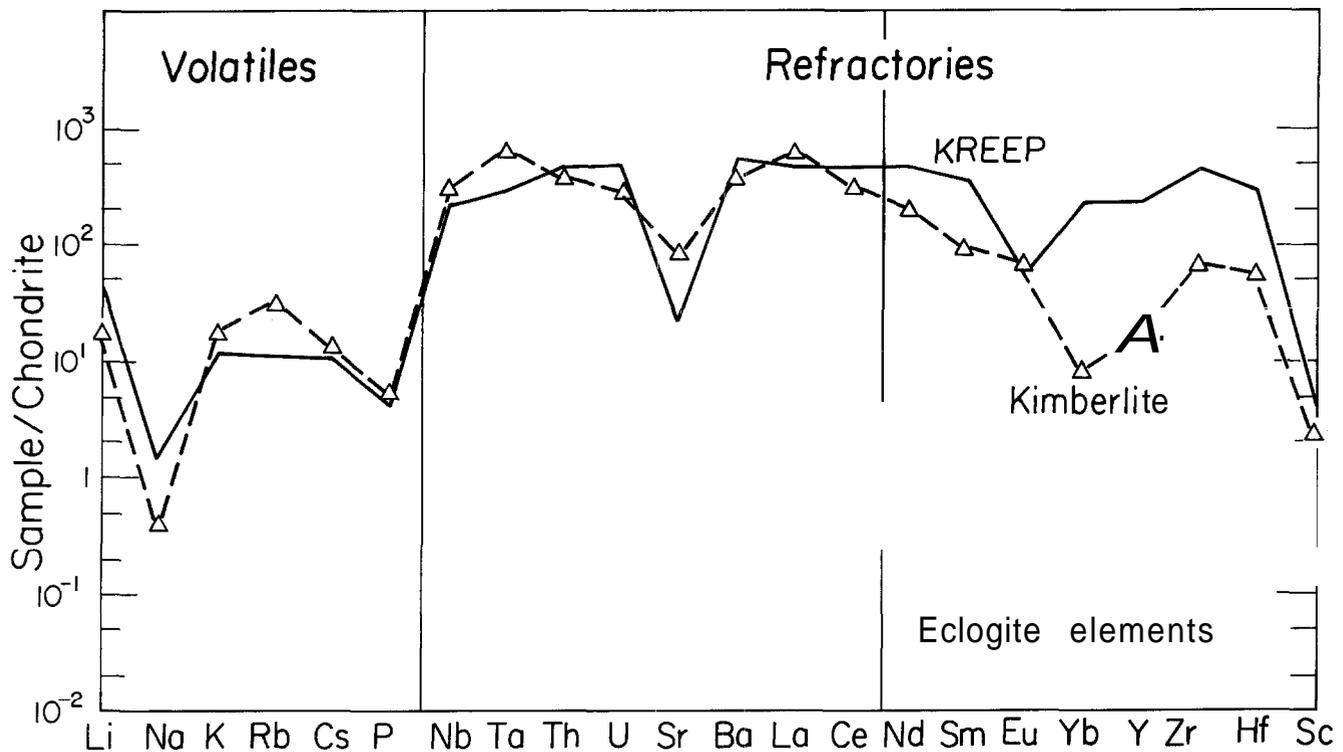


FIGURE 8-8
Chondrite-normalized trace-element compositions for kimberlite and lunar KREEP. The strontium and europium anomalies for KREEP are due to plagioclase extraction from the lunar source region. The relative depletion of kimberlite in the eclogite elements and sodium may reflect eclogite and jadeite extraction from the terrestrial source region. Lithium is plotted with the alkali metals although it is a refractory.

with an eclogite source region for kimberlite. This difference in controlling mineralogy is simply a reflection of the greater pressures in the Earth.

Kimberlite and lunar KREEP are remarkably similar in their minor- and trace-element chemistry. The main differences can be attributed to plagioclase fractionation in the case of KREEP and eclogite fractionation in the case of kimberlite. KREEP has been interpreted as the residual fluid of a crystallizing magma ocean. In a small body the Al_2O_3 content of a crystallizing melt is reduced by plagioclase crystallization and flotation. In a magma ocean on a large body, such as the Earth, the Al_2O_3 is removed by the sinking garnet. Kimberlite, in fact, is depleted in eclogite elements including the HREE and sodium. This suggests that kimberlite may represent the late-stage residual fluid of a crystallizing terrestrial magma ocean. A buried eclogite-rich cumulate layer is the terrestrial equivalent of the lunar anorthositic crust.

We have established that removal of a kimberlite-like fluid from a garnet-clinopyroxene-rich source region gives a crystalline residue that has the appropriate trace-element chemistry to be the reservoir for LIL-depleted magmas such as MORB. It is usually assumed that the continental crust

serves the role that we attribute to kimberlite, and in fact the crust may also be involved. I propose, however, that enriched fluid permeates the shallow mantle and is responsible for the LIL-enrichment of island-arc and oceanic island basalts. Accretion of arcs, islands and plateaus would be the mechanism for continental growth. The continental lithosphere would therefore also be enriched in incompatible elements even if depleted in a basaltic fraction. Clearly, the LIL-enrichment of the shallow mantle, including the continental lithosphere, can affect the chemistry of depleted magmas that traverse it.

ALKALI BASALT

Continental alkaline magmatism may persist over very long periods of time in the same region and may recur along lines of structural weakness after a long hiatus. The age and thickness of the lithosphere play an important role, presumably by controlling the depth and extent of crystal fractionation and the ease by which the magma can rise to the surface. Alkali magmatism occurs over broad areas and is not necessarily associated with rifts but is often accompanied

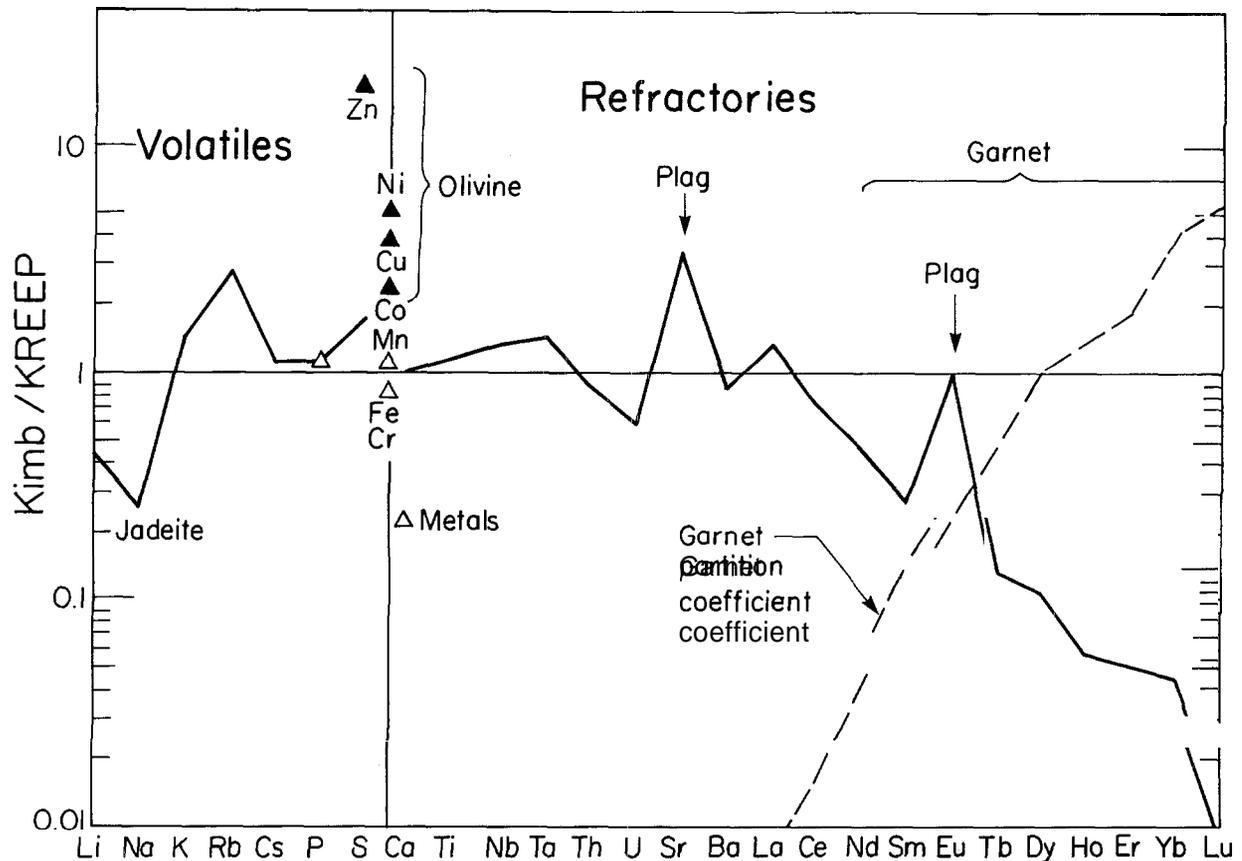


FIGURE 8-9

Trace-element concentrations in kimberlite relative to KREEP. Elements that are fractionated into olivine, plagioclase, garnet and jadeite are noted. The solid/liquid partition coefficient for garnet is shown. If kimberlite has been in equilibrium with garnet, or an eclogite cumulate, the concentrations in kimberlite will be proportional to the reciprocal of the garnet partition coefficient.

by crustal doming. Some of the more important alkaline provinces are in Africa, Siberia, the Baltic shield, Scotland, the Oslo graben, southwest Greenland, the Appalachians, the Basin and Range, the White Mountains (New England), and the St. Lawrence and Ottawa grabens. Many of these are associated with grabens and transcurrent faults and with the opening of ocean basins. Alkali magmatism precedes and follows the opening of the Atlantic. Alkali magmatism can be traced along a small circle, presumably representing a transform fault, from provinces in North America to the New England seamounts and the Canary Islands to the Moroccan alkaline province (Sykes, 1978). Likewise alkaline provinces in South America can be linked with those in west-central Africa through the Cape Verde Islands. Oceanic transform faults in general have long been recognized as the loci of alkali basalts.

The non-random distribution of alkaline provinces has been interpreted as hotspot tracks and structural weaknesses in the lithosphere. Sykes (1978) suggested that hotspots are

passive centers of volcanism whose location is determined by pre-existing zones of weakness, rather than manifestations of deep mantle plumes.

Examples are known of alkaline basalts preceding and following abundant tholeiitic volcanism. In general alkaline basalts occur beneath thicker lithosphere and are early when the lithosphere is thinning (rifting) and late when the lithosphere is thickening (flanks of ridges, downstream from midplate volcanic centers).

On Kerguelen, the third largest oceanic island, and on Ascension Island, the youngest basalts are alkalic and overlie a tholeiitic base. In Iceland, tholeiites dominate in the rift zone and grade to alkali basalts along the western and southern shores. Thus, there appears to be a relation between the nature of the magmatism and the stage of evolution. The early basalts are tholeiitic, change to transitional and mixed basalts and terminate with alkaline compositions. It was once thought that Hawaii also followed this sequence, but the youngest volcano, Loihi, is alkalic and

probably represents the earliest stage of Hawaiian volcanism. It could easily be covered up and lost from view by a major tholeiitic shield-building stage. Thus, oceanic islands may start and end with an alkalic stage.

In continental rifts, such as the Afar trough, the Red Sea, the Baikal rift and the Oslo graben, the magmas are alkalic until the rift is mature and they are then replaced by tholeiities, as in a widening oceanic rift. A similar sequence occurs in the Canary Islands and the early stages of Hawaii. The return of alkalis in the Hawaiian chain occurs when the islands drift off the hotspot. They, and seamounts on the flanks of midocean ridges, may be tapping the flanks of a mantle upwelling, where crystal fractionation and contamination by shallow mantle are most prevalent. In all cases lithospheric thickness appears to play a major role. This plus the long duration of alkalic volcanism, its simultaneity over large areas and its recurrence in the same parts of the crust argue against the simple plume concept.

There are two prevailing types of models to explain the diversity of basalts. In the primary magma models a parent magma evolves to a wide variety of derivative magmas depending on the depth of fractionation and, therefore, the nature of the fractionation crystals: olivine, plagioclase, orthopyroxene, clinopyroxene, garnet, ilmenite, magnetite and so on. Thus, alkalis can evolve from tholeiites by garnet and clinopyroxene fractionation. Isotopic data show that crystal fractionation is not enough; several isotopically distinct reservoirs must be involved. In the partial melt models the different basalt types are a result of varying degrees of partial melting at different depths and depend on the nature of the residual crystals in the solid rocks that are left behind. In this case alkalis represent small degrees of partial melting, small enough so that the low melting point crystals, clinopyroxene and garnet, are left behind in the residue. Tholeiites would represent a greater degree of melting, enough to eliminate the clinopyroxene and garnet. Another component is required to explain the isotopic ratios. Basalts probably have a complex history involving mixing of materials representing variable degrees of partial melting from different depths, followed by variable amounts of crystal fractionation and interaction with, or contamination by, shallower parts of the mantle and crust.

Highly alkaline products are characteristic of the later stages of many areas of extensive basaltic magmatism. These often occur as dikes, sills and caps associated with late tholeiitic layers. They constitute rock types such as basanites, melilitites, nephelinites, lamprophyres and trachybasalts. Similar lavas occur in the early stages of continental rifting and on some oceanic islands in the early stages of volcanism. The distinctive chemical character of these magmas is often attributed to low degrees of partial melting of a deep, previously untapped mantle source, or melting of mantle that has been metasomatized by a CO_2 , H_2O -rich fluid. An alternative mechanism involves crystal fractionation from a tholeiitic melt or melting of a shallow reservoir

caused by ascending tholeiitic magma. In these models the origin of tholeiitic and alkali magmas are closely related, but melts from several reservoirs may be involved. The HREE depletion of alkali magmas can be explained by eclogite fractionation at sublithospheric depths.

Figure 8-10 shows that alkali-rich basalts generally have trace-element concentrations intermediate between MORB and highly alkalic basalts such as nephelinite. In fact, they can be treated as mixtures of MORB, or a picritic MORB-parent, and nephelinite.

Alkaline rocks are the main transporters of mantle inclusions. They include basanites, nephelinites, hawaiites, alkali olivine basalts, phonolites, kimberlites and lamproites. All of these magmas or melts are known to transport spinel- or garnet-bearing peridotites or high-pressure crystals such as diamond, sapphire, garnet and so on to the surface. The presence of these inclusions demonstrates that these melts are from the mantle and that part of the mantle is enriched in LIL and isotopic ratios.

CONTINENTAL FLOOD BASALTS

Continental flood basalts, the most copious effusives on land, are mainly tholeiitic flows that can cover very large areas and are typically 3 to 9 km in thickness. These are also called plateau basalts and fissure basalts. Examples are the Deccan traps in India, the Columbia River province in the United States, the Parana basin in Brazil, the Karoo province in South Africa, the Siberian traps, and extensive flows in western Australia, Tasmania, Greenland and Antarctica. They have great similarity around the world. Related basalts are found in recent rifts such as East Africa and the Basin and Range province. Rift and flood basalts may both form during an early stage of continental rifting. In some cases alkali-rich basalts were erupted first, occurring on the edge of rifts while tholeiites occur on the floor of the rift. Continental alkali basalts and tholeiites are highly enriched in LIL elements and have high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. There are many similarities between continental and ocean-island basalts. Experiments indicate that fractional crystallization or varying degrees of partial melting can produce various members of the alkali series (Green and Ringwood, 1967; Frey and others, 1978). The large range of isotopic ratios in such basalts indicates that their source region is inhomogeneous or that mixing between at least two magma types is involved. The similarity between many continental basalts and ocean-island basalts indicates that involvement of the continental crust is not necessary to explain all continental flood basalts; high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, for example, are not necessarily indicative of contamination by the crust.

Continental flood basalts have been the subject of

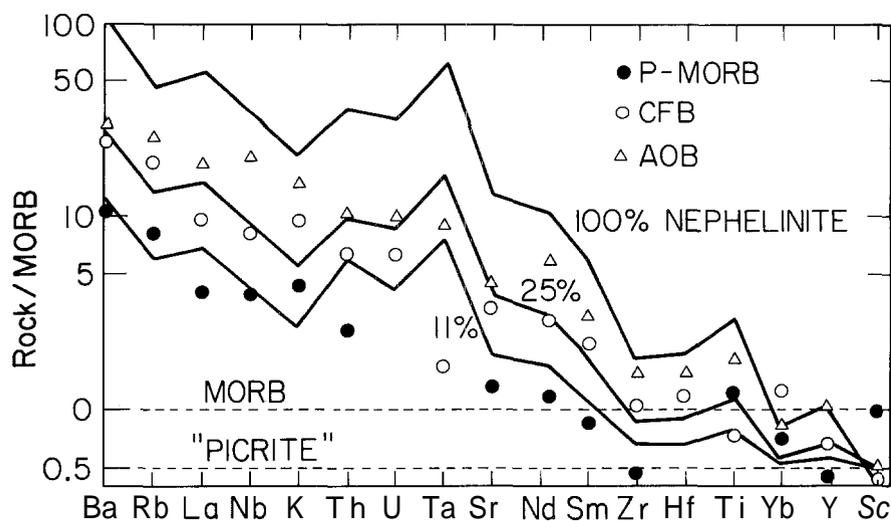


FIGURE 8-10

Trace-element concentrations of "enriched" magmas relative to MORB. P-MORB are "plume-type" MORB, found along midocean ridges; CFB are continental flood basalts; AOB are alkali olivine basalts, found in continents and on oceanic islands. Lines are mixing curves. This suggests that enriched magmas may be contaminated MORB, rather than melts from a different source region or varying degrees of melting from a common source region

much debate. Their trace-element and isotopic differences from depleted midocean-ridge basalts have been attributed to their derivation from enriched mantle, primitive mantle, magma mixtures or from a parent basalt that experienced contamination by the continental crust. In some areas these flood basalts appear to be related to the early stages of continental rifting or to the proximity to an oceanic spreading center. In other regions there appears to be a connection with subduction, and in these cases they may be analogous to back-arc basins. In most isotopic and trace-element characteristics they overlap basalts from oceanic islands and island arcs.

Chemical characteristics considered typical of continental tholeiites are high SiO_2 and incompatible-element contents, and low MgO/FeO and compatible-element concentrations. The very incompatible elements such as cesium and barium can be much higher than in typical oceanic island basalts.

The Columbia River flood basalts are probably the best studied, but they are not necessarily typical. They appear to have been influenced by the subduction of the Juan de Fuca plate beneath the North American plate and to also have involved interaction with old enriched subcontinental mantle (Carlson and others, 1980). The idea that continental flood basalts are derived from a primordial "chondritic" source (DePaolo and Wasserburg, 1979) does not hold up when tested against all available trace-element and isotopic data, even though some of these basalts have neodymium isotopic ratios expected from a primitive mantle.

Many of the characteristics of continental and ocean-

island basalts can be explained by sublithospheric fractionation of garnet and clinopyroxene from a MORB parent, combined with contamination by a high-LIL component. Superimposed on these effects are shallow-mantle or crustal fractionation and, possibly, crustal assimilation. It does not appear necessary to invoke two distinct, isolated fertile source regions such as the low-velocity zone of the upper mantle and some part of the lower mantle, for example, to explain MORB and hotspot magmas. Common basalts with different histories of cooling, fractionation and contamination can exhibit the observed chemical differences.

The relationships between some continental basalts and oceanic volcanism are shown in Figure 8-11. The basalts in the hatched areas are believed to result from hotspot volcanism. The dashed lines show boundaries that apparently represent sutures between fragments of continents. Most hotspots are inside the two noted geoid highs. Table 8-9 gives some flood basalt provinces and the possibly related hotspot. Also shown are some kimberlite provinces and their ages.

ANDESITE

Andesites are associated primarily with island arcs and other convergent plate boundaries and have bulk compositions similar to the continental crust. The accretion of island arcs onto the edges of continents is probably the primary mechanism by which continents grow or, at least, maintain their area. The "average" andesite contains about 60 per-

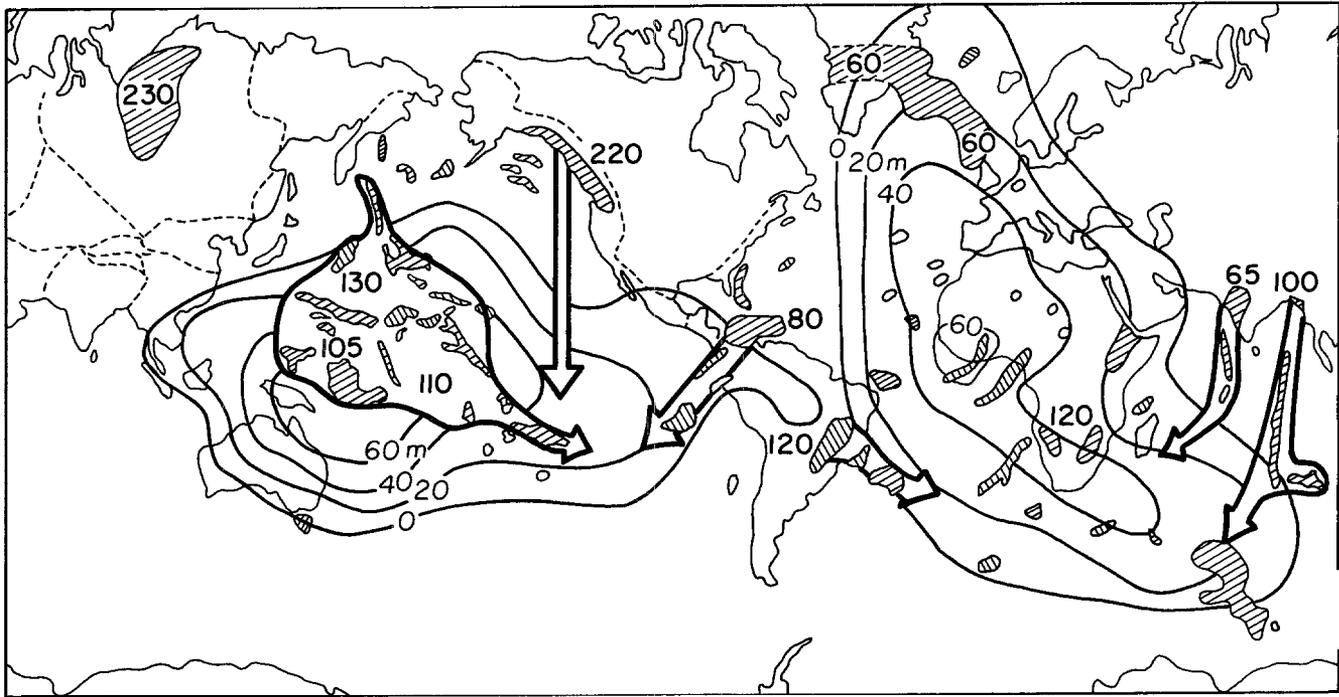


FIGURE 8-11
 Map of continental flood basalts, selected hotspot islands, and oceanic plateaus, with ages in millions of years. The period between 130 and 65 Ma was one of extensive volcanism and most of this occurred inside geoid highs (contours) that, in general, are areas of high elevation. Some of the basalt areas have drifted away from their point of origin (arrows). Geoid highs are probably due to hotter than average mantle.

cent SiO₂, 17 percent Al₂O₃, 3 percent MgO, 7 percent CaO, 3.3 percent Na₂O, 1.6 percent K₂O and 1.2 percent H₂O. However, the range in SiO₂, by convention, is from 53 to 63 percent. The K₂O content is variable and seems to correlate with the depth to the underlying slab. "Calc-alkaline" is often used as an adjective for andesite. Tholeiites are also found in island arcs.

Andesites can, in principle, originate from melting of continental crust, contamination of basalt by continental crust, partial fusion of hydrous peridotite, partial fusion of subducted oceanic crust, differentiation of basalt by crystal or vapor fractionation, or by magma mixing. Melting of hydrous peridotite or low-pressure differentiation of basalt by crystal fractionation seem to be the preferred mechanisms of most petrologists, although involvement of subducted material and crust, and magma mixing, are sometimes indicated. Andesites probably originate in or above the subducted slab and, therefore, at relatively shallow depths in the mantle. The thick crust in most andesitic environments means that shallow-level crystal fractionation and crustal contamination are likely. In their incompatible-element and isotopic chemistry, andesites are similar to ocean-island basalts and continental flood basalts. They differ in their bulk chemistry, a difference that can be explained by shallower crystal fractionation.

Volcanic or island arcs typically occur 100 to 250 km from convergent plate boundaries and are of the order of 50 to 200 km wide. The volcanoes are generally about 100 to 200 km above the earthquake foci of the dipping seismic zone. Crustal thicknesses at island arcs are generally 20 to

TABLE 8-9
 Continental Flood Basalts and Possibly Related Hotspots and Kimberlites

Region	Hotspot	Age (Ma)
Southeast Greenland	Iceland	60
India	Reunion	65
Columbia River	Yellowstone	20
Southern Brazil	Tristan	120
Southern Africa	Crozet	190
Siberia	Jan Mayen (?)	240
Wrangellia	Hawaii (?)	210
Central Japan	—	280
Kimberlite Provinces		
India		70
Southern Africa		194
Brazil		120
Siberia		148–159

50 km. Andesitic volcanism appears to initiate when a subducting slab has reached a depth of about 70 km and changes to basaltic volcanism when the slab is no longer present or when compressional tectonics changes to extensional tectonics.

Gill (1982) proposed that two conditions are required for voluminous oceanic andesite volcanism: (1) an underlying asthenosphere modified by subduction shortly before the volcanism and (2) compressional stress, or sialic crust greater than about 25 km thick. The latter apparently are required to permit extensive fractionation of the magma prior to eruption.

Although melting of subducted oceanic crust is not necessarily the source of andesitic magmas, the dehydration of subducted sediments and crust may trigger melting in the overlying mantle wedge. Alternatively, the subducted slab may displace material from the MORB source, which then ascends, becomes contaminated, and fractionates prior to eruption. The crust of back-arc basins is more like typical MORB and may represent less impeded ascent of MORB magmas from a deep (>200 km) source.

The overwhelming majority of andesitic volcanoes are located above the intermediate- and deep-focus earthquakes that define the "deep seismic zone." This zone typically dips toward the continent at angles between 15° and near vertical. The dip generally steepens with depth and often bends most noticeably just below the volcanic arc. In most regions the earthquakes cease between about 200 and 300 km. The deep seismic zone is often called the Benioff zone, but it was first described in Japan, and if a name must be applied it should be called the Wadati or Wadati-Benioff zone. Earthquakes occur below 300 km in regions where old (>50 Ma) oceanic lithosphere has subducted. This could mean that old, cold lithosphere has a higher negative buoyancy than young lithosphere, and can therefore sink deeper into the mantle, or that it takes longer for the thicker lithosphere to warm up and lose its brittleness. The crust and depleted peridotite parts of the slab are buoyant relative to the underlying mantle and can only sink when they are very cold. Although the crustal part of the slab can convert to dense eclogite at depths of the order of 50 km, it experiences high temperatures and high stresses as it enters the mantle and may well remain in the shallow mantle. The large number of earthquakes at shallow depths certainly indicates that the top of the slab is extensively fractured before it reaches 50 to 100 km. Even if subducted crust does not contribute directly to the overlying andesites, it probably contributes to the high H₂O and LIL content of the shallow mantle, including the mantle wedge. Oceanic crust becomes extremely altered by hydrothermal activity as it ages and is not a suitable candidate, by remelting, to provide depleted MORB, unless the high temperatures and dehydration experienced during subduction can undo the alteration. If subducted crust can escape melting and converts to eclogite, the relatively small fragments would have diffi-

culty sinking very far into the mantle unless dragged down by the underlying lithosphere.

The mantle immediately below the crust is probably olivine-rich harzburgite or dunite. This is consistent with oceanic P_n velocities and anisotropy and with the mantle section at the base of ophiolites. The thickness of this layer, however, is unknown. In some models the harzburgite is considered the depleted complement to the overlying basalts and is therefore expected to be 25 to 30 km thick for 20 percent melting of a pyrolite source. Since it is depleted in the basaltic component, it is poor in Al₂O₃ and, at high pressure, in garnet. It is therefore buoyant and can only **subduct**, temporarily, if it is very cold. Over the long run, therefore, it stays in the shallow mantle or returns to the shallow mantle after heating.

The seismic data, however, do not require such a thick olivine-rich layer in the oceanic lithosphere. Probably 5–10 km would be thick enough to explain the P_n observations. There are also petrological and geochemical data suggesting that the peridotites in oceanic sections are not genetically related to the overlying basalts. They may, in fact, be the refractory residues from ancient melting processes. Since depleted peridotites are buoyant relative to fertile mantle, and since melting has been occurring over all of Earth history, there should be harzburgites and dunites in the shallow mantle that date back to early Earth history. The cold lithosphere under stable shields may be the best place to find ancient material.

If the oceanic crust represents only part of the molten material that attempts to rise to the surface of the Earth, then there may be basaltic or picritic material elsewhere in the shallow mantle that eventually intrudes or underplates the oceanic lithosphere. Part of the lower oceanic lithosphere may therefore convert to eclogite as it ages. High-velocity regions in the oceanic lithosphere imply a large garnet content, consistent with eclogite. The negative buoyancy of old lithosphere may therefore be due to chemical rather than temperature effects, and the deep subduction of old oceanic lithosphere can be understood.

The most effective way to cool the mantle is to **subduct** old, cold lithosphere. The cooling works in two ways: Hot mantle is replaced or displaced by the cold subducted material, and as the slab warms up the adjacent mantle is cooled. The latter process is very slow because of the long conduction time constant. The overall temperature of the slab and adjacent mantle system stays roughly constant since the heating of one is counterbalanced by the cooling of the other.

Subduction tends to occur in oceanic lithosphere of all ages. There does not seem to be a critical age at which subduction can initiate, and therefore a thermal instability, by itself, is not sufficient for spontaneous subduction. It may be that subduction always initiates at continental boundaries or at regions of thick crust, such as oceanic plateaus, or at fracture zones. A novel possibility is that sub-

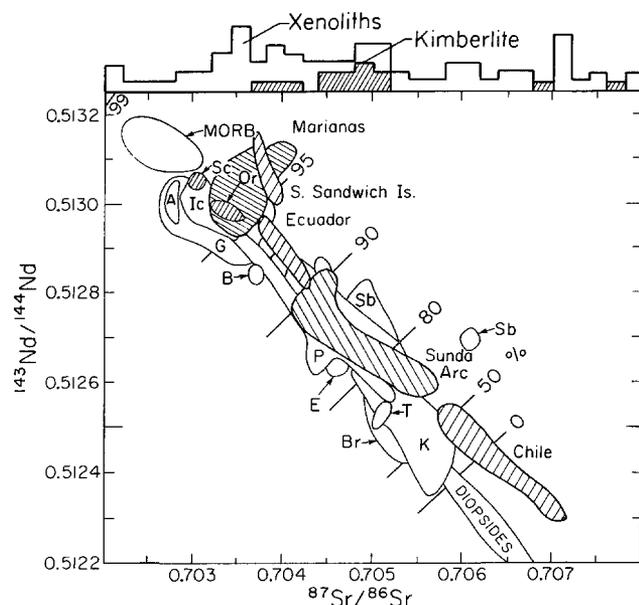


FIGURE 8-12

Neodymium and strontium isotopic ratios for oceanic, ocean-island, continental and island-arc (hatched) basalts and diopside inclusions from kimberlites. The numbers are the percentage of the depleted end-member if this array is taken as a mixing line between depleted and enriched end members. Mixing lines are flat hyperbolas that are approximately straight lines for reasonable choices of parameters. The fields of MORB and CFB correspond to >97 percent and 70–95 percent, respectively, of the depleted end-member. The enriched end-member has been arbitrarily taken as near the enriched end of Kerguelen (K) basalts. The most enriched magmas are from Kerguelen, Tristan da Cunha (T) and Brazil (Br). Other abbreviations are Sc (Scotia Sea), A (Ascension), Ic (Iceland), H (Hawaii), G (Gouch), Or (Oregon), Bo (Bouvet), Sb (Siberia), Co (Columbia River), P (Patagonia) and E (Eiffel). Along top are strontium isotopic data for xenoliths and kimberlites.

duction can occur, not only in normal collision situations, but also when the oceanic lithosphere overrides abnormally hot and light mantle. The Tonga-Fiji arc, for example, seems to be embedded in such a region, with abnormally slow mantle velocities to great depth.

The "mantle wedge" between the base of the crust and the top of the slab has low seismic velocities and high attenuation, consistent with partial melting. Below about 300 km the mantle near subduction zones is faster than average, presumably due to the insertion of cold material.

Figure 8-12 shows that island-arc basalts (hatched) overlap ocean-island and continental basalts in isotopic characteristics. A similar mantle source is indicated.

MARGINAL BASINS

Marginal basins (see Table 8-10) are small oceanic basins, underlain by oceanic crust, that represent trapped and iso-

lated fragments of oceanic crust (Bering Sea, south Fiji Basin, Caribbean) or formed by processes akin to continental rifting (South China Sea, Tasman Basin, Coral Sea) or sea-floor spreading (Gulf of California, Lau Basin, Andaman Sea). Some are related to subduction and are called **back-arc basins**. Some of these are actively spreading (Mariana Trough, Ryukyu Trough, Lau Basin, Scotia Sea), and oth-

TABLE 8-10

Classification and Ages of Trench–Arc–Back–Arc–Marginal Basin Systems

Continental Arc
Peru-Chile Andes Cordilleran
Middle America Cordilleran
North America Basin and Range
Alaska Cordilleran
Sumatra-Java-Sunda Shelf
Trapped Ocean Remnants
Bering Sea–Aleutian Basin (117–132 Ma)
West Philippine Sea? (39–50 Ma)
South Fiji Basin? (25–34 Ma)
Caribbean Plate (100 Ma)
Emerald Basin? (South of New Zealand)
Active Back-Arc Spreading
Mariana Trough (Late Miocene)
Ryukyu Trough
Lau Basin (Tonga) (0–4 Ma)
Scotia Sea (0–9 Ma)
New Hebrides? (41–55 Ma)
Bonin?
Leaky Transform
Gulf of California
Andaman Sea (0–11 Ma)
Bismark Sea (New Guinea) (0–3.5 Ma)
Fiji Plateau (New Hebrides) (0–10 Ma)
Inactive Back-Arc
Grenada Trough (Antilles)
Kamchatka Basin
Kurile Basin
Japan Basin
Tsushima Basin (Japan)
Shikoku Basin? (Izu-Bonin) (17–26 Ma)
Parece Vela Basin (Mariana) (17–30 Ma)
Small Ocean Basin (Continental rift related)
Coral Sea (56–64 Ma)
South China Sea (17–32 Ma)
Tasman Sea (56–77 Ma)
Uncertain Origin
Banda Sea
Caroline Basin (28–36 Ma)
Celebes Basin (42–48 Ma)
New Caledonia Basin
Solomon Basin
West Scotia Sea (16–30 Ma)

Brooks and others (1984).

TABLE 8-11

Composition of Marginal Basin Basalts Compared with Other Basalts

Oxide	MORB	MORB	Mariana Trough	Scotia Sea	Lau Basin	Bonin Rift	North Fiji Plateau	Arc Tholeiites	Island Tholeiite	
SiO ₂	49.21	50.47	50.56	51.69	51.33	51.11	49.5	48.7	51.1	49.4
TiO ₂	1.39	1.58	1.21	1.41	1.67	1.10	1.2	0.63	0.83	2.5
Al ₂ O ₃	15.81	15.31	16.53	16.23	17.22	18.20	15.5	16.5	16.1	13.9
FeO	7.19	10.42	8.26	8.28	5.22	8.18	6.2	8.4	—	—
Fe ₂ O ₃	2.21	—	—	—	2.70	—	3.9	3.4	11.8	12.4
MnO	0.16	—	0.10	0.17	0.15	0.17	0.1	0.29	0.2	—
MgO	8.53	7.46	7.25	6.98	5.23	6.25	6.7	8.2	5.1	8.4
CaO	11.14	11.48	11.59	11.23	9.95	11.18	11.3	12.2	10.8	10.3
Na ₂ O	2.71	2.64	2.86	3.09	3.16	3.12	2.7	1.2	1.96	2.13
K ₂ O	0.26	0.16	0.23	0.27	1.07	0.38	0.3	0.23	0.40	0.38
P ₂ O ₅	0.15	0.13	0.15	0.18	0.39	0.18	0.1	0.10	0.14	—
H ₂ O	—	—	1.6	—	1.18	—	1.4	—	—	—
⁸⁷ Sr/ ⁸⁶ Sr	0.7023	0.7028	0.7028	0.7030	0.7036	—	—	—	—	—

Gill (1976), Hawkins (1977).

ers are inactive (Kamchatka and Kurile Basins, Japan Basin, Shikoku Basin). Some have an uncertain origin (Banda Sea, New Hebrides–Fiji Plateau, west Philippine Sea).

The crustal structure, heat flow and magnetic lineation patterns of back-arc and major ocean basins are similar, implying that the origins are similar. Although back-arc spreading appears to be related to subduction, most arc-trench systems do not have actively spreading back-arc basins. Back-arc basin basalts (BABB) are similar to mid-ocean-ridge basalts but in many respects they are transitional toward island-arc tholeiites and ocean-island basalts (Tables 8-11, 8-12 and 8-13). They are LILE-enriched compared to normal MORB and generally have LREE enrichment. High ³He/⁴He ratios have been found in some BABB, and in this and other respects they are similar to, or gradational toward, ocean-island basalts and other hot-spot magmas.

Back-arc spreading may be initiated above subducted lithosphere, but with time the axis of spreading migrates and no longer lies above the earthquakes of the Wadati-Benioff zone. If present the seismic zone is deeper than 200–300 km below the spreading center. Spreading centers in the Bismarck Sea, east Scotia Sea, Mariana trough and Coriolis trough (New Hebrides) are beyond the deepest seismic extent of the slab.

If the geometric and temporal relationships can be used as a guide, the relatively enriched island-arc tholeiites come from a depth of about 100 km and BABB generally comes from deeper than 200 km. The seismic velocities and seismic *Q*'s are low in the mantle wedge above subducting plates. This probably reflects partial melting.

It is not yet clear whether back-arc basins are formed as a result of global plate motions or as a direct result of subduction. Many back-arc regions exhibit cordilleran-type

TABLE 8-12

Representative Properties of Basalts and Kimberlites

Ratio	Midocean Ridge Basalt	Back-Arc Basin Basalt	Island-Arc Basalt	Ocean-Island Basalt	Alkali-Olivine Basalt	Kimberlite
Rb/Sr	0.007	0.019–0.043	0.01–0.04	0.04	0.05	0.08
K/Rb	800–2000	550–860	300–690	250–750	320–430	130
Sm/Nd	0.34–0.70	0.31	0.28–0.34	0.15–0.35	0.37	0.18
U/Th	0.11–0.71	0.21	0.19–0.77	0.15–0.35	0.27	0.24
K/Ba	110–120	33–97	7.4–36	28–30	13.4	7.5
Ba/Rb	12–24	9–25	10–20	14–20	15	15
La/Yb	0.6	2.5	2–9	7	14	125
Zr/Hf	30–44	43	38–56	44	41	36
Zr/Nb	35–40	4.8	20–30	14.7	2.7	2.3
⁸⁷ Sr/ ⁸⁶ Sr	0.7023	0.7027–0.705	0.703–0.706	0.7032–0.706	—	0.71
²⁰⁶ Pb/ ²⁰⁴ Pb	17.5–18.5	18.2–19.1	18.3–18.8	18.0–20.0	—	17.6–20.0

TABLE 8-13

Representative Values of Large-Ion Lithophile and High-Field-Strength Elements in Basalts (ppm)

Element	Midocean-Ridge Basalt	Back-Arc Basin Basalt	Island-Arc Basalt	Ocean-Island Basalt
Rb	0.2	4.5–7.1	5–32	5
Sr	50	146–195	200	350
Ba	4	40–174	75	100
Pb	0.08	1.6–3	9.3	4
Zr	35	121	74	125
Hf	1.2	2.8	1.7	3
Nb	1	25	2.7	8
Ce	3	32	6.7–32.1	35
Th	0.03	1–1.9	0.79	0.67
U	0.02	0.4	0.19	1.18

tectonics or nonspreading back-arc basins. In the Marianas, back-arc spreading alternates with arc volcanism. Extension, of course, is associated with spreading, whether as cause or effect, and arcs and trenches are basically convergent or compressional phenomena. Arcs with active back-arc spreading do not have the giant earthquakes associated with the cordilleran-style arcs. This is probably a result of the local coupling between plates.

In most respects active back-arc basins appear to be miniature versions of the major oceans, and the spreading process and crustal structure and composition are similar to those occurring at midocean ridges. The age-depth relationships of marginal seas, taken as a group, are indistinguishable from those of the major oceans. To first order, then, the water depth in marginal sea basins is controlled by the cooling, contraction and subsidence of the oceanic lithosphere, and the density of the mantle under the spreading centers is the same. The marginal basins of Southeast Asia, however, tend to be deeper than similar age oceanic crust elsewhere, by one-half to one kilometer, and to deepen faster with age. The presence of cold subducted material underneath the basins may explain why they are deeper than average but not why they sink faster. The low-density material under the ridge axis may be more confined under the major midocean ridges. On average the upper 200–300 km of the mantle in the vicinity of island arcs and marginal basins has slower than average seismic velocities and the deeper mantle is faster than average, probably reflecting the presence of cold subducted material. The depth of back-arc basins is an integrated effect of the thickness of the crust and lithosphere, the low-density shallow mantle and the presumed denser underlying subducted material. It is perhaps surprising then that, on average, the depths of marginal basins are so similar to equivalent age oceans elsewhere. The main difference is the presence of the underlying deep subducting material, which would be expected to depress the seafloor in back-arc basins.

Basalts in marginal basins with a long history of spreading are essentially similar to MORB, while basalts generated in the early stages of back-arc spreading have more LIL-enriched characteristics. A similar sequence is found in continental rifts (Red Sea, Afar) and some oceanic islands, which suggests a vertical zonation in the mantle, with the LIL-rich zone being shallower than the depleted zone. This is relevant to the plume hypothesis, which assumes that enriched magmas rise from deep in the mantle. The early stages of back-arc magmatism are the most LIL-enriched, and this is the stage at which the effect of hypothetical deep mantle plumes would be cut off by the presence of the subducting slab. Continental basalts, such as the Columbia River basalt province, are also most enriched when the presence of a slab in the shallow mantle under western North America is indicated. The similarity of the isotopic and trace-element geochemistry of island-arc basalts, continental flood basalts and ocean-island basalts and the slightly enriched nature of the back-arc basin basalts all suggest that the enrichment occurs at shallow depths, perhaps by contamination of MORB rising from a deeper layer. The high $^3\text{He}/^4\text{He}$ of Lau Basin basalts also suggests a shallow origin for "primitive" gases.

Both back-arc basins and midocean ridges have shallow seafloor, high heat flow and thin sedimentary cover. The upper mantle in both environments has low seismic velocity and high attenuation. The crusts are typically oceanic and basement rocks are tholeiitic. The spreading center in the back-arc basins, however, is much less well defined. Magnetic anomalies are less coherent, seismicity is diffuse and the ridge crests appear to jump around. Although extension is undoubtedly occurring, it may be oblique to the arc and diffuse and have a large shear component. The existence of similar basalts at midocean ridges and back-arc basins and the subtle differences in trace-element and isotopic ratios provide clues as to the composition and depth of the MORB reservoir.

TABLE 8-14
Major-Element Analyses of Komatiites and Other
High-MgO Magmas (percent)

Major Element	Komatiite		Picritic Tholeiite (Baffin Is.)	Bonin Volcanic Province
	Munro Township	Barberton		
SiO ₂	45.8	45.2	44.0	59.2
Al ₂ O ₃	7.30	3.66	8.3	11.25
TiO ₂	0.30	0.20	0.58	0.22
	—	—	2.2	—
FeO	11.2	11.0	8.8	8.86
MnO	0.21	0.22	0.19	0.16
MgO	26.1	32.2	26.0	11.4
CaO	7.64	5.28	7.3	6.55
Na ₂ O	0.69	0.44	0.90	1.78
K ₂ O	0.10	0.17	0.06	0.43
Cr ₂ O ₃	0.24	—	0.36	0.17

BVP (1980).

KOMATIITES

Komatiites are ultrabasic melts that occur mainly in Archean rocks (Table 8-14). The peridotitic variety (MgO > 18 percent) apparently require temperatures of the order of 1450–1500°C and degrees of partial melting greater than 60–70 percent in order to form. Although they have been identified in the Mesozoic, they are extremely rare in the Phanerozoic. Cawthorn and Strong (1974) estimated that they must originate below 300 km if the melting is due to adiabatic ascent, which seems likely.

The great depth of origin implied for high-temperature magmas such as komatiites means that there is little experimental petrological control on their conditions of formation. At low pressure, olivine is the most refractory phase, and melts are commonly much less MgO-rich than the parent. At high pressure it is possible to generate MgO-rich melts with smaller degrees of partial melting. It is even possible that olivine is replaced as the liquidus phase by the high-pressure majorite phase of orthopyroxene, again giving high-MgO melts. Komatiites may therefore represent large degrees of melting of a shallow olivine-rich parent, small degrees of melting of a deep peridotite source (Herzberg, 1984) or melting of a rock under conditions such that olivine is not the liquidus phase. At depths greater than about 200 km, the initial melts may be denser than the residual crystals (Rigden and others, 1984). This may imply that large degrees of partial melting are possible and, in fact, are required before the melts, or the source region, become buoyant enough to rise. The high CaO/Al₂O₃ ratios of some komatiites suggest that garnet has been left behind in their source region or that high-pressure garnet fractionation occurred prior to eruption. Much has been made of

the high temperatures and large degrees of melting implied for komatiite formation, particularly with regard to high upper-mantle temperatures in the Precambrian. Their existence appears to refute the common claim that melt-crystal separation must occur at relatively small degrees of partial melting, about 20–25 percent in ascending diapirs. The rarity of komatiites since Precambrian times could mean that the mantle has cooled, but it could also mean that a suitable peridotite parent no longer exists at about 300 km depth, or that the currently relatively thick lithosphere prevents their ascent to the surface. Diapirs ascending rapidly from about 300 km from an eclogite-rich source region would be almost totally molten by the time they reach shallow depths, and picrite and basaltic magmas would predominate over komatiitic magma. If peridotites only extend to a depth of 200–400 km in the present mantle, then the absence of recent komatiites could be readily understood. Picritic magmas, the currently popular precursor to basalts, are also rare, presumably because they are too dense to rise through the crust. Cooling and olivine fractionation causes the density to decrease, and tholeiites are now generally viewed as fractionated picrites. The MgO content of magmas also increases with depth of melting. Komatiites may therefore be the result of deep melting. By cooling and crystal fractionation komatiitic melts can evolve to less dense picritic and tholeiitic melts. Low-density melts, of course, are more eruptible. It may be that komatiitic melts exist at present, just as they did in the Precambrian, but, because of the colder shallow mantle they can and must cool and fractionate more, prior to eruption. In any case komatiites provide important information about the physics and chemistry of the upper mantle.

The existence of komatiitic melts seems to contradict the petrological prejudice that large amounts of melting are impossible. However, if melts are to drain from a source region, the surrounding matrix must deform to fill the space. Marsh (1984) argued that melting must exceed 45 to 55 volume percent before eruption can occur.

SUMMARY

Most models of mantle chemistry and evolution based on magmas alone tend to relate the various magma types by varying degrees of partial melting or crystal fractionation. Depth of melt-crystal separation is another parameter. The actual depth of the source region, its composition and the amount of partial melting cannot be tightly constrained by observational or experimental petrology. The simplest petrological models tend to view the mantle as homogeneous throughout and capable of providing basalt, for the first time, by partial melting. When the petrological data are combined with isotopic and geophysical data, and with considerations from accretional calculations, a more complex evolution is required. Similarly, simple evolutionary mod-

els have been constructed from isotopic data alone that conflict with the broader data base.

General References

- Anderson, D. L. (1979) The upper mantle transition region: Eclogite? *Geophys. Res. Lett.*, **6**, 433–436.
- Anderson, D. L. (1981) Hotspots, basalts and the evolution of the mantle, *Science*, **213**, 82–89.
- Anderson, D. L. (1982a) Isotopic evolution of the mantle, *Earth Planet. Sci. Lett.*, **57**, 1–24.
- Anderson, D. L. (1982b) Chemical composition and evolution of the mantle. In *High-pressure Research in Geophysics* (S. Akimoto and M. Manghnani, eds.), 301–318, D. Reidel, Dordrecht.
- Anderson, D. L. (1983a) Chemical composition of the mantle, *J. Geophys. Res.*, **88**, B41–B52.
- Anderson, D. L. (1983b) Kimberlite and the evolution of the mantle, *Proc. Third Intl. Kimberlite Conference*, Clermont-Ferrand, France.
- Anderson, D. L. and J. Regan (1983) Uppermantle anisotropy and the oceanic lithosphere, *Geophys. Res. Lett.*, **10**, 841–844.
- Dawson, J. B. (1980) *Kimberlites and their Xenoliths*, Springer-Verlag, Berlin, 252 pp.
- Taylor, S. (1982) Lunar and terrestrial crusts, *Phys. Earth Planet. Inter.*, **29**, 233.
- Wedepohl, K. H. and Y. Muramatsu (1979) The chemical composition of kimberlites compared with the average composition of three basaltic magma types. In *Kimberlite, Diatremes and Diamonds* (F. R. Boyd and H. O. Meyer, eds.), 300–312, American Geophysical Union, Washington, D.C.

References

- Anderson, D. L. (1975) Chemical plumes in the mantle, *Geol. Soc. Am. Bull.*, **86**, 1593–1600.
- Anderson, D. L. (1977) Composition in the mantle and core, *Ann. Rev. Earth Planet. Sci.*, **5**, 179.
- Anderson, D. L. (1980) *Bull. Volc.*, 663.
- Anderson, D. L. (1982a) Isotopic evolution of the mantle, *Earth Planet. Sci. Lett.*, **57**, 1–24.
- Anderson, D. L. (1982b) Chemical composition and evolution of the mantle. In *High-pressure Research in Geophysics* (S. Akimoto and M. H. Manghnani, eds.), 301–318, D. Reidel, Dordrecht.
- Anderson, D. L. (1982c) Hotspots, polar wander, mesozoic convection and the geoid, *Nature*, **297**, 391–393.
- BVP, Basaltic Volcanism Study Project (1980) *Basaltic Volcanism on the Terrestrial Planets*, Pergamon, New York, 1286 pp.
- Bass, J. D. and D. L. Anderson (1984) *Geophys. Res. Lett.*, **11**, 237–240.
- Brooks, D. A., R. L. Carlson, D. L. Harry, P. J. Melia, R. P. Moore, J. E. Rayhorn and S. G. Tubb (1984) Characteristics of back-arc regions, *Tectonophysics*, **102**, 1–16.
- Burdick, L. and D. L. Anderson (1975) Interpretation of velocity profiles of the mantle, *J. Geophys. Res.*, **80**, 1070–1074.
- Butler, R. and D. L. Anderson (1978) Equation of state fits to the lower mantle and outer core, *Phys. Earth Planet. Inter.*, **17**, 147–162.
- Carlson, R. W., G. W. Lugmair and J. D. Macdougall (1980) Crustal influence in the generation of continental flood basalts, *Nature*, **289**, 160–162.
- Cawthorn, R., R. Grant and D. Strong (1974) The petrogenesis of Itomatiites and related rocks as evidence for a layered upper mantle, *Earth Planet. Sci. Lett.*, **23**, 369.
- Cohen, L. and J. Rosenfeld (1979) Diamond; depth of crystallization inferred from compressed included garnet, *J. Geol.*, **87**, 330–340.
- Crough, S. T. (1983) Hotspot swells, *Ann. Rev. Earth Planet. Sci.*, **11**, 165.
- DePaolo, D. J. and G. J. Wasserburg (1979) Neodymium isotopes in flood basalts from the Siberian Platform and inferences about their mantle sources, *Proc. Nat. Acad. Sci.*, **76**, 3056–3060.
- Dosso, L. and V. R. Murthy (1980) A Nd isotopic study of the Kerguelen Islands; inferences on enriched oceanic mantle sources, *Earth Planet. Sci. Lett.*, **48**, 268–276.
- Frey, F. A., D. H. Green and S. D. Roy (1978) Integrated models of basalts petrogenesis: A study of quartz tholeiites to olivine melilitites from southeastern Australia utilizing geochemical and experimental petrological data, *J. Petrol.*, **19**, 463–513.
- Gill, J. (1981) *Orogenic Andesites and Plate Tectonics*, Springer-Verlag, New York, 390 pp.
- Gill, J. B. (1976) Composition and age of Lau basin and ridge volcanic rocks; implications for evolution of an interarc basin and remnant arc, *Geol. Soc. Am. Bull.*, **87**, 1384–1395.
- Green, D. H. and A. E. Ringwood (1967) The stability fields of aluminous pyroxene peridotite and garnet peridotite and their relevance in upper mantle structure, *Earth Planet. Sci. Lett.*, **3**, 151–160.
- Hawkins, J. W. (1977), in *Island Arcs, Deep Sea Trenches, and Back-Arc Basins*, M. Talwani and W. C. Pittman, eds., AGU, Washington, D. C., 367–377.
- Herzberg, C. T. (1984) Chemical stratification in the silicate Earth, *Earth Planet. Sci. Lett.*, **67**, 249–260.
- Jacobsen, S. B., J. E. Quick and G. J. Wasserburg (1984) A Nd and Sr isotopic study of the Trinity Peridotite; implications for mantle evolution, *Earth Planet. Sci. Lett.*, **68**, 361–378.
- Liu, L. (1974) Silicate perovskite from phase transformation of pyrope garnet, *Geophys. Res. Lett.*, **1**, 277–280.
- Maaløe, S. and R. Steel (1980) Mantle composition derived from the composition of Iherzolites, *Nature*, **285**, 321–322.
- Marsh, Bruce D. (1984). In *Explosive Volcanism*. (F. R. Boyd, Jr., ed.) National Academy Press, Washington, D.C.
- McCulloch, M. T., A. Jaques, D. Nelson and J. Lewis (1983) Nd and Sr isotopes in kimberlites and lamproites from Western Australia: An enriched mantle origin, *Nature*, **302**, 400.
- McCulloch, M. T., R. Arculus, B. Anappell and J. Ferguson (1982) Isotopic and geochemical studies of nodules in kimberlites, *Nature*, **300**, 166.

- McKenzie, D. and F. Richter (1981) Parameterized thermal convection in a layered region, *J. Geophys. Res.*, 86, 11,677.
- Morgan, J. W. and E. Anders (1980) Chemical composition of the Earth, Venus and Mercury, *Proc. Natl. Acad. Sci.*, 77, 6973.
- Morgan, W. J. (1971) Convection plumes in the lower mantle, *Nature*, 230, 42–43.
- Morgan, W. J. (1972) Plate motions and deep mantle convection. In *Studies in Earth and Space Sciences* (Geol. Soc. Am. Mem. 132), 7–22.
- Morgan, W. J. (1981) Hotspot tracks and the opening of the Atlantic and Indian Oceans. In *The Oceanic Lithosphere* (C. Emiliani, ed.), 443–387, Wiley, New York.
- Rigden, S. S., T. J. Ahrens and E. M. Stolper (1984) Densities of liquid silicates at high pressures, *Science*, 226, 1071–1074.
- Ringwood, A. E. (1966) Mineralogy of the mantle. In *Advances in Earth Science*, 357–399, MIT Press, Cambridge, Mass.
- Ringwood, A. E. (1975) *Composition and Petrology of the Earth's Mantle*, McGraw-Hill, New York, 618 pp.
- Ringwood, A. E. and S. Kesson (1976) A dynamic model for mare basalt petrogenesis, *Proc. Lunar Sci. Conf.*, 7, 1697–1722.
- Ringwood, A. E. and S. Kesson (1977) Siderophile and volatile elements in Moon, Earth and chondrites, *Moon*, 16, 425.
- Sykes, L. R. (1978) Intraplate seismicity, reactivation of preexisting zones of weakness, alkaline magmatism, and other tectonism postdating continental fragmentation, *Rev. Geophys. Space Phys.*, 16, 621–688.
- Walck, M. (1982) Models of upper mantle compressional velocity under a spreading center, *Eos Trans. AGU*, 63, 1036.
- Warren, P. H. and J. T. Wasson (1979) The origin of KREEP, *Rev. Geophys. Space Phys.*, 17, 73–88.
- Wilson, J. T. (1963) Evidence from islands on the spreading of the ocean floor, *Nature*, 197, 536–538.