Early Proterozoic oceanic crust and the evolution of subcontinental mantle: Elogites and related rocks from southern Africa

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ABSTRACT

Seven eclogitic nodules from kimberlites in southern Africa have been studied in detail for whole-rock, major- and trace-element geochemistry, petrography, and mineral chemistry by electron microprobe; high-purity mineral separates from six of these samples have been analyzed for trace elements and for the isotopic composition of Sr, Nd, and oxygen. Three eclogite groups are recognized: group A eclogites have very high Mg/Fe, low Na in pyroxene, moderate δ18O (4.7 to 5.3), and low 87Sr/86Sr and 143Nd/144Nd. Olivine and enstatite may also be present as accessory phases in this group. Group B eclogites have moderate to high Mg/Fe ratios, high Na in pyroxene, low δ18O (3.0 to 3.4), high 87Sr/86Sr and 143Nd/144Nd ratios, and extremely LREE-depleted HREE-enriched garnets with no Eu anomalies. Group C eclogites have low Mg/Fe ratios, high Na in pyroxene, and variable Sr-, Nd-, and O-isotope compositions. Accessory feldspar is present in one sample that may be of crustal origin. Mineral separates of garnet and pyroxene have positive Eu anomalies.

The group A eclogites are too refractory to represent magma compositions and must have formed as cumulate dikes in the upper mantle that contain a minor trapped liquid component. This is supported by the presence of accessory olivine and enstatite, and by their KSm and KNb, which are similar to empirical pyroxene/garnet partition coefficients. The Group B eclogites are extremely depleted in incompatible elements and have εNd(0) values 10<sup>-1</sup> to 20<sup>-1</sup> MORB. The high 87Sr/86Sr of these eclogites is not consistent with their strongly depleted MORB and high εNd(0), and cannot be primary—it must have been imposed on the protolith. The low δ18O of these eclogites cannot form by mantle fractionation processes and must also be inherited from the protolith. Both the Sr- and oxygen-isotope data are consistent with high-temperature hydrothermal alteration of a basaltic protolith, followed by partial melting to form the refractory compositions observed now. The hydrothermal fluid may have been seawater, but secondary enrichment of the protolith in Rb is required to generate the observed high Sr ratios. The high Na content of the pyroxenes supports a spilitic alteration event. The major- and trace-element characteristics of the group C eclogites are consistent with recrystallization of a cumulate gabbro protolith. One group C rock is probably a garnet granulite derived from the lower crust. The other may represent the plutonic portion of oceanic crust.

A reconstructed whole-rock isochron for the three Bellsbank eclogites yields an age of 2.1 ± 0.1 b.y. This implies that plate-tectonic processes involving the generation and subduction of oceanic crust have been active since the early Proterozoic. The early Proterozoic ocean crust consisted of two components: a plutonic section of gabbro cumulates (group C eclogites) and a volcanic section of basalt (group B eclogites). The volcanic component has undergone substantial hydrothermal alteration and subsequent partial melting to form a refractory residue; the plutonic component is less modified.

INTRODUCTION

Elogites are garnet + clinopyroxene rocks with basaltic bulk compositions that crystallized (or recrystallized) at relatively high pressures in the lower crust or upper mantle. Kyanite, rutile, orthopyroxene, plagioclase, or hornblende are commonly associated as varietal or accessory phases (Coleman and others, 1965). Elogites with omphacitic pyroxene and Ca-, Fe-rich garnet found in blueschist terranes are thought to represent metamorphosed oceanic crust (Coleman and others, 1965). In contrast, most eclogite xenoliths in kimberlites and alkali basalts are more magnesic and are generally interpreted as high-pressure cumulates (garnet pyroxenites) that formed in dikes within the upper mantle (O’Hara and Yoder, 1967; MacGregor and Carter, 1970; Hatton, 1978; Smyth and Caporuscio, 1980). Eclogites that occur in kimberlites vary greatly in texture, mineral compositions, and isotopic characteristics, however, and there is increasing evidence that some of these nodules may also represent metamorphosed oceanic crust (Helmstaedt and Doig, 1975; Helmstaedt and Schulze, 1979; Ater and others, 1984; Jagoutz and others, 1984; MacGregor and Manton, 1986).

In order to distinguish between these contrasting hypotheses, we have selected a suite of seven African eclogites from the collections of Dr. F. R. Boyd for detailed petrologic, trace-element, and isotopic study (Shervais and others, 1985). Five of these eclogites are from kimberlites that intrude the Kaapvaal craton in southern Africa, including four samples from Bellsbank (South Africa) and one sample from Kao (Lesotho). The other two samples, from Deutsche Erde (Namibia) and Vale do Queue (Angola), are from kimberlites that intrude...
Proterozoic mobile belts northwest of the craton. The location of these kimberlites relative to the Kaapvaal craton is shown in Figure 1.

Coleman and others (1965), in a comparative study of eclogites from a variety of localities, recognized three groups of eclogite on the basis of their mode of occurrence: group A eclogites are xenoliths in kimberlite, group B eclogites are found as lenses in high-grade gneiss terranes, and group C eclogites are associated with blueschist-facies, high-pressure, metamorphic terranes. Coleman and co-workers noted that eclogites from each of these groups (defined on mode of occurrence) have characteristic garnet and clinopyroxene compositions. All of the eclogites studied here occur as xenoliths in kimberlite, but the mineral compositions in each are distinct and allow the correlation with the three groups defined by Coleman and others (1965). We will use these groupings to compare and contrast the eclogites studied here.

METHODS

The petrography of all samples was studied using standard optical techniques, and their mineral chemistry analyzed with a MAC 400 S electron microprobe at the University of Tennessee. Whole-rock samples were analyzed for major and trace elements at the University of Tennessee using an energy-dispersive X-ray fluorescence spectrometer and by instrumental neutron-activation analysis (INAA) at Washington University. Whole-rock samples were prepared by crushing ~8–10 grams of each xenolith and splitting for XRF (~5 grams) and INAA (~100 milligrams).

High-purity, hand-picked mineral separates were prepared at the University of California, San Diego (UCSD). These mineral separates were produced by disaggregating portions of each sample in a stainless-steel impact mortar, washing them in distilled water, hand-picking in air under a binocular microscope to produce grain concentrates, and hand-picking the concentrates in alcohol under the binocular microscope using a fiber-optic ring lamp. The ring lamp illuminates internal fractures and facilitates preparation of inclusion-free mineral separates. After each hand-picking, the mineral separates were further cleaned by etching in a solution of HF/HNO₃ (1:3 with distilled water) followed by 4N HCl (2N for plagioclase). Each mineral separate (200–300 mg) was then split into three fractions: one for trace-element analysis by INAA at Washington University, one for Sr- and Nd-isotope analysis at the University of California, San Diego, and one for oxygen-isotope analysis at the University of Chicago. Detailed analytical procedures for each of these labs have been published elsewhere (Clayton and others, 1975; Stosch and others, 1980; Korotev, 1987).

PETROGRAPHY

Ka6

The general geology of the Kao kimberlite has been discussed by Rolfe (1973) and Clement (1973), and the mafic and ultramafic xenoliths from this pipe have been described by Nixon and Boyd (1973), Hornung and Nixon (1973), and MacGregor (1979). In addition to eclogite, other xenolith types found at Kao include spinel lherzolite, garnet lherzolite, harzburgite, ilmenite-bearing lherzolite, and discrete pyroxene nodules.

The Kao eclogite studied here (PHN 1850-1) consists of subangular to rounded, pale orange garnets up to 8 mm in diameter in a matrix of smaller, gray-green clinopyroxenes (~3 mm) that have been extensively altered along their grain boundaries to a fine-grained mass of phlogopite and other secondary phases (Table 1). The garnet grains are outlined by kelyphitic rims of fine-grained pyroxene and spinel that are distinct from the clinopyroxene alteration assemblage. The garnets commonly contain irregularly distributed patches of exsolved rutile needles; similar rutile inclusions in pyroxene are less common. Trace amounts of rutile also occur as small (~0.1 mm) primary grains interstitial to pyroxene.

Angola

The sample from Vale do Queve, Angola (FRB 340) is unique in that it contains about 5%–10% primary feldspar and is more properly considered a garnet-rich granulite (Table 1). It was first described by Boyd and Danchin (1980), who noted the presence of plagioclase (An₇₃) and called it a plagioclase-bearing eclogite. They also noted that the mineral chemistry of this rock corresponds to the type II Roberts

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**TABLE 1: MODAL ANALYSES OF ECLOGITE XENOLITHS FROM SOUTHERN AFRICA**

<table>
<thead>
<tr>
<th>Sample</th>
<th>Qt</th>
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<th>Cps</th>
<th>Qt</th>
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</table>

*Alternites = phlogopite, chlorite, serpentine, green hornblende, amphibole, and calcite.

*Boyd and Danchin (1980) reported 5% feldspar in their thin section.
Victor eclogites of MacGregor and Carter (1970). FRB 340 consists of large rounded garnets (<3 mm diam) in a granoblastic polygonal matrix of clinopyroxene (<0.5 mm) and plagioclase (<2 mm). The distribution of plagioclase suggests that it was originally interstitial to garnet and pyroxene, and that the primary pyroxene had a grain size similar to that of garnet. Rutile is a common minor constituent that forms small ovoid grains 0.05 to 0.1 mm across in plagioclase and is probably primary. Rutile also occurs as oriented needles within garnet and pyroxene. Secondary phases include actinolite, mica, and carbonate, which are largely concentrated along pyroxene grain boundaries and in patches.

One aspect of this eclogite that has not been noted previously is the occurrence of K-feldspar. The K-feldspar forms subhedral to euhedral laths up to 0.2 × 1.5 mm in size that are intergrown with plagioclase. These laths still retain their distinctive shape even though they have been preferentially altered to low-temperature minerals and the surrounding plagioclase has been recrystallized to a granoblastic polygonal texture. Microprobe analyses show that it ranges in composition from Or37 Ab54 to Or57 Ab38.

### Deutsche Erde

The Deutsche Erde kimberlite has been described by Jane (1975). Our eclogite sample (DE-15-1) consists of subequal portions of intergrown pink garnet and bright apple-green clinopyroxene (Table 1). In contrast to the other eclogites, pyroxene grains are generally larger than garnet (<12 mm versus <8 mm). The interfingerings of garnet and pyroxene is similar to an exsolution texture, but the relative abundance of garnet argues against that interpretation. Some garnet does occur as inclusions within pyroxene, both as elongate ooids (0.5 × 1.5 mm) and as lamellae (0.5 × 8 mm). These inclusions are probably exsolved from the pyroxene. The larger-scale interfingerings of these phases described above is probably a primary feature and may result from coticetic crystalization of garnet and clinopyroxene.

Alteration is extensive along grain boundaries and fractures (especially in pyroxene), but the cores of most grains are fresh, and alteration products comprise less than 15% of the mode (Table 1). Secondary minerals include fine-grained pyroxene, amphibole, and minor phlogopite.

### Bellbank

The geology and petrology of kimberlite dikes in the Bellbank region has been described by Bosch (1969), Fesq and others (1975), and Bontor and Boyd (1982). Meyer and McCallister (1984) have described the petrology of two-pyroxyene megacrysts from the Main dike at Bellbank, and eclogites from the nearby Bobbejaan dike have been described by Smyth and others (1980), Smyth and Caporuscio (1980), and Carswell and others (1981).

Two distinct suites of eclogite are found in the Bellbank kimberlite. The first suite, designated by sample numbers beginning FRB-438, consist of blocky or rounded orange-colored garnets 5–10 mm across, surrounded by smaller (2–5 mm) dark-gray-green clinopyroxene grains (Table 1). Garnet and pyroxene are the only primary phases. Secondary minerals include phlogopite, chlorite, green hornblende, fibrous actinolite, carbonate, and serpentine. Phlogopite is an early phase that is found on the walls of veins and fractures and occurs as grains up to 0.5 × 1 mm in size. The occurrence of early phlogopite veins in these xenoliths suggests a metasomatic origin unrelated to the host kimberlite magma.
Elogites of the second Bellsbank suite, designated by sample numbers beginning FRB-437, contain minor olivine and/or enstatite and thus are not true eclogites; a more proper name would be olivine-bearing garnet pyroxenite or garnet websterite (Table 1). These rocks consist of blocky, sub-equitant grains of lavender garnet (8–12 mm across) intergrown with bright apple-green clinopyroxene (3–5 mm), enstatite, and olivine (Fo92.5). Secondary minerals include minor phlogopite (in veins and fractures), chlorite, and serpentine pseudomorphs after olivine.

The garnet pyroxenites have granoblastic textures with relatively common 120° triple boundaries. Primary enstatite and olivine are sparse, forming blocky grains up to 2 mm across. Most enstatite occurs as exsolution lamellae up to 0.2 mm thick in diopside and as exsolved blebs along grain boundaries. Some garnet also formed by exsolution, but most occurs as large primary grains. The lack of undulatory extinction, kink bands, and other strain features suggests that the current textures result from annealing.

MINERAL CHEMISTRY

Major Elements

Major-element compositions of the primary silicate phases in the rocks studied here have been determined by electron-microprobe analysis and are presented in Tables 2 and 3. These phases are extremely homogeneous in all samples except the garnet granulate from Angola (FRB-340), which exhibits moderate variations in the jadeite component of pyroxene and the grossular component of garnet (Boyd and Danchin, 1980). The major-element compositions of the primary silicate phases do exhibit large variations between samples and allow us to correlate these eclogites with the three eclogite groups defined by Coleman and others (1965).

Garnet. Garnets exhibit large variations in the relative proportions of their pyrope, almandine, and grossular components, as well as variations in Cr concentration (Table 2). Elogites 437-1 and 437-2 from Bellsbank contain garnets that are pyrope-rich relative to the other samples studied here, with low grossular and almandine contents (Fig. 2) and high Cr2O3. Garnet from the Deutsche Erde eclogite is also pyrope-rich but has higher grossular and lower Cr than 437-1, 2 (Fig. 2). Garnets from all three plots in the field of “Group A” eclogites of Coleman and others (1965).

Elogites 438-2 and 438-7 from Bellsbank contain garnets that are lower in pyrope and higher in both almandine and grossular than the other Bellsbank eclogites (Fig. 2). The garnet in these samples also has much lower Cr concentration (<0.1 wt% Cr2O3). The pyr-alm-gross ratios of these garnets plot in the field of “Group B” eclogites of Coleman and others (1965). Garnets from the Kao eclogite and Angola garnet granulate have the lowest pyrope and highest almandine of any of the samples studied here (Fig. 2). Their grossular content is slightly higher than the Bellbank garnets, but nearly the same as the Deutsche Erde garnet. Cr is low in both samples. The pyr-alm-gross ratio of the Kao garnet is similar to the “Group C” eclogites of Coleman and others (1965), but the Angola garnet granulate plots just inside the group B field.

Pyroxene. Pyroxenes show limited compositional variation within the pyroxene quadrilateral (Fig. 2). The mole fraction of Wo is relatively constant; most variation is caused by changes in the Mg# (= mol% Mg/(Mg+Fe)) between pyroxenes in different samples. The variation in clinopyroxene Mg# is sympathetic with Mg# variations in the co-existing garnet so that tie-lines between the two phases do not cross in a ternary plot of Mg-Fe-Ca (Fig. 2). Elogite 437-2 also contains orthopyroxene with Mg#s = 93 (the same as coexisting clinopyroxene).

Concentrations of Cr, Al, and Na in the pyroxenes are more variable (Table 3). The eclogites with high Mg# pyroxene (437-1, 437-2, Deutsche Erde) have clinopyroxene relatively low in Na and Al (jadeite component) and high in Cr (Cr2O3 = 0.16 to 1.30). The K2O for Fe/Mg distribution between these pyroxenes and coexisting garnet is low (3.5–5), as shown by the steep slope of the garnet-pyroxene tie lines in the Mg-Fe-Ca ternary (Fig. 2). Pyroxenes from the more Fe-rich Bellsbank eclogites (438-2, 438-7) have higher concentrations of Na and Al (Table 3) and low Cr2O3 (<0.10 wt%). Garnet-pyroxene K2O for these eclogites are also higher (4.8–5.8). Pyroxenes from the Kao eclogite and the Angola garnet granulate have the highest jadeite contents (Na and Al), low Cr2O3 (<0.10 wt%), and high garnet-pyroxene K2O values (5–6).

The variations observed in the minor-element chemistry of these pyroxenes are consistent with the variations discussed by Coleman and others (1965) and correlate with the eclogite groups defined on the basis of garnet compositions. Thus, pyroxenes in group A eclogites (437-1, 437-2, Deutsche Erde) have the lowest jadeite content, pyroxenes in group B eclogites (438-2, 438-7) have higher jadeite, and pyroxene in the group C eclogite (Kao) has the highest jadeite content (Coleman and others, 1965). Although garnet from the Angola garnet granulate plots just inside the group B field, the high jadeite content of the pyroxene suggests a closer relationship to the group C eclogites. In subsequent sections, we shall refer to these "groups" as representative of the samples correlated with each group.

Trace Elements

Hand-picked mineral separates from six of the seven eclogites studied here were analyzed for trace-element concentrations by INAA (Table 4). The compatible transition metals exhibit moderate variations in concentration that can be explained by differences in bulk composition.
and the phase assemblage of the eclogite. For example, Ni concentrations in whole-rock samples increase from group C to group A eclogites. Ni concentrations in pyroxene and garnet also increase from group C to group A, with the exception of the olivine-bearing garnet pyroxenite, 437-1. The Ni in this sample has been partitioned strongly into olivine, and the co-existing garnet and pyroxene have Ni concentrations similar to group C eclogites. The other transition metals exhibit similar behavior.

Chondrite-normalized REE (rare-earth element) patterns for eclogite garnets display the depletion of LREE (light REE) and enrichment of HREE (heavy REE) that is characteristic of this phase (Fig. 3). Nonetheless, garnets from each of the three groups used here are distinct. Group C garnets have flat or slightly positive HREE slopes that turn down at Sm (Yb/Sm = 0.7–2.2 × chondrite) and distinct positive Eu anomalies (Fig. 3). Group B garnets feature extreme depletion of the LREE and enrichment of the HREE (Yb/Sm = 50–60 × chondrite). Group A garnets have intermediate characteristics and feature smooth positive slopes from La and Yb (Yb/Sm = 3–7 × chondrite).

Clinopyroxene separates from all of the samples studied here, except Angola, are LREE-enriched (Fig. 4). Pyroxene from the Angola garnet granulate has a chondrite-normalized REE pattern that is convex upward, which results in part from partitioning of the LREE into plagioclase (Fig. 4). The most LREE-enriched pyroxenes are those in group A, with Ce/Tb ratios of 600–1,200 (30–70 × chondrite). The Angola garnet granulate has a Ce/Tb ratio of 9 (0.5 × chondrite), which is similar to MORB. Pyroxenes from the other group C and group B eclogites have Ce/Tb ratios which range from 20–90 (1.2–5.3 × chondrite).

Pyroxenes from groups C and B exhibit two other unusual features that distinguish them from those of group A. Chondrite-normalized REE patterns for group B pyroxenes have a sigmoidal shape with a minimum at Sm and maxima at La and Tb (Fig. 4). This may be an effect of partitioning between pyroxene and garnet, because the group B garnets are strongly enriched in HREE and depleted in LREE (Fig. 3). The second unusual feature is displayed by the group C pyroxenes, which have small but significant positive Eu anomalies (Fig. 4).

**WHOLE-ROCK CHEMISTRY**

**Major and Trace Elements**

All of the samples studied here are approximately “basaltic” in the sense that they contain 45%–51% SiO₂; however, only the group C
samples could represent a basaltic-liquid composition (Table 5). The other eclogites are too low in TiO₂, and too high in CaO (DE), MgO (437-1, 2), or Mg#. This is especially true of the group A eclogites, which have whole-rock Mg#s of 84-89, 15%-20% MgO, and 300-600 ppm Ni. High Mg/Fe ratios such as these are virtually restricted to ultramafic cumulates and refractory peridotites; even ultramafic komatiites with 30% MgO have Mg#s of only 86. The group B eclogites have bulk-rock compositions similar to basaltic komatiites, with =13% MgO, 200-250 ppm Ni, and Mg#s of 64-69. The group C samples are characterized by high alumina contents (Al₂O₃ = 13-17 wt%), similar to gabbros and high-alumina basalts, and low Mg#s (42-61).

The trace-element data are generally consistent with the data of the major elements: incompatible elements (Rh, Sr, Zr, Nb) are depleted in the eclogites with high Mg#s (group A) and enriched in those with low Mg#s (group C). Compatible transition metals (Ni, Cr) are highest in group A and lowest in group C. Elements that are strongly partitioned into garnet (V and Y) are enriched in the group B eclogites, even though the modal abundance of garnet is similar in all three groups.

Rare-Earth Elements

Rare-earth-element (REE) concentrations have been determined in six samples by INAA. These data are reported in Table 5 and are shown in Figure 5A as chondrite-normalized plots. The Bellsbank samples have U-shaped patterns with both the LREE and HREE enriched relative to the MREE. The Angola (group C) and DE-15 (group A) samples have LREE-enriched patterns with flat or moderately negative HREE slopes, whereas the Kao eclogite (group C) has a pattern similar to MORB—flat HREE with moderate LREE depletion (Fig. 5A). Both group C samples have positive Eu anomalies.

Interpretation of the whole-rock data is difficult because of the common occurrence of secondary minerals in all of these eclogites. These minerals are generally enriched in incompatible elements such as the LREE and alter the chemistry of the original rocks by enriching them in these elements. This problem is illustrated by the two group B eclogites, 438-2 and 438-7. Both have similar primary modes and nearly identical garnet and pyroxene compositions; they differ only in the modal abundance of secondary minerals (=15% in 438-2 versus 34% in 438-7). Whole-rock analyses of these two samples have nearly identical HREE concentrations, but 438-7 has LREE concentrations 5x higher than 438-2 (Fig. 5A). To what extent has this secondary enrichment affected the other eclogites? The primary concentrations of REE in these eclogites can be determined approximately by using estimates of their original modal composition (Table 1) and the REE concentrations of the primary mineral phases (Table 4). Chondrite-normalized REE patterns for reconstructed whole-rock compositions based on these esti-
TABLE 5. RESULTS OF WHOLE ROCK ANALYSES OF AFRICAN EOLITES BY X-RAY FLUORESCENCE (X) AND INSTRUMENTAL NEUTRON ACTIVATION (N)

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<td>0.04</td>
<td>0.02</td>
<td>0.07</td>
<td>0.35</td>
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<tr>
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<td>18.75</td>
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<td>6.58</td>
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<td>0.20</td>
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<td>P₂O₅</td>
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<td>0.08</td>
<td>0.19</td>
<td>0.12</td>
<td>0.15</td>
<td>0.11</td>
<td>0.37</td>
</tr>
</tbody>
</table>

Trace elements (ppm)

- Sr  | N  | 450 | 34.9 | 41.5 | 63.2 | n.a. | 68.3 | 66.4 |
- Y   | X  | 245 | 35.5 | 33.2 | 190 | 186.6 | 400 | 473.6 |
- Ce  | N  | 35.8 | 52.5 | 43.3 | 51.6 | n.a. | 60.3 | 58.6 |
- Ni  | X  | 277 | 196.7 | 171 | 409.5 | 623 | 204 | 248 |
- Cr  | N  | 480 | 230.6 | 620 | n.a. | 240.1 | 315.0 |
- Re  | X  | 4.07 | 27.8 | 12.0 | 20.8 | 13 |
- Ru  | N  | <0.10 | <1.0 | <1.0 | n.a. | <16.0 |
- Sr  | X  | 127 | 105.1 | 254.1 | 160.1 | 198.1 | 61.1 | 198.1 |
- Sm  | X  | 170 | 110.7 | 470.1 | n.a. | 150.0 | 350.0 |
- Y   | X  | 7.0 | 14.3 | 13.8 | 11.6 | 44.1 |
- Zr  | X  | 31.0 | 9.0 | 7.0 | 38.9 | 39.0 | 30.1 | 68.8 |
- Nb   | X  | 1.0 | 18.7 | 7.3 | 6.0 | 14.0 | 22.2 |
- Cs  | N  | 0.37 | <0.15 | 9.55 | 2.64 | 2.19 | 9.48 |
- Ba  | X  | 149 | 85.0 | 640 | 390 | n.a. | 570.0 | 1090 |
- La  | X  | 318 | 163.0 | 124.0 | 154.0 | n.a. | 51.0 | 66.0 |
- Ce  | X  | 90.0 | 4.4 | 27.3 | 33.3 | n.a. | 10.0 | 46.7 |
- Nd  | N  | 5.1 | 3.9 | 15.9 | 9.9 | n.a. | 2.9 | 14.0 |
- Sm  | N  | 0.74 | 1.26 | 3.26 | 3.80 | n.a. | 0.56 | 1.68 |
- Eu   | X  | 0.238 | 0.79 | 1.40 | 0.214 | 0.271 | 0.54 |
- Tb   | N  | 0.098 | 0.32 | 0.63 | 0.10 | 0.70 | 0.05 | 10.20 |
- Dy  | N  | 0.41 | 1.32 | 1.20 | 1.15 | n.a. | 9.4 | 8.15 |
- Ho   | N  | 0.099 | 0.233 | 0.186 | 0.23 | n.a. | 1.48 | 1.25 |
- Er  | N  | 0.14 | 0.57 | 1.69 | 0.81 | 0.48 | 1.86 | 2.57 |
- Tm   | N  | 0.19 | 0.37 | 0.20 | 0.12 | 0.26 | 0.92 | 5.20 |
- Yb   | N  | 0.15 | <0.1 | 0.22 | 0.14 | 0.04 | 0.15 | 0.40 |
| U    | N  | <0.5 | <0.8 | <0.8 | n.a. | <0.5 | 1.8 | 1.5 |

n.a. = not analysed.

The isotopic composition of Sr and Nd provides important constraints on (1) the age of formation of igneous and metamorphic rocks and on (2) the time-averaged enrichments and depletions of these rocks and their source regions. The pervasive effects of metamorphic enrichment prior to, or after entrainment of the xenoliths in the host kimberlite, however, require special measures to distinguish between primary and secondary isotopic characteristics. Customary whole-rock analyses, therefore, would be meaningless because of the dominating overprint caused by late-stage, REE-rich, metamorphic fluids. This problem can be largely avoided through the use of very pure and thoroughly cleaned (etched) mineral separates, as shown by Stosch and others (1980) and Jagoutz and others (1985). Our techniques are described in the Methods section.

The results of our isotopic analyses are presented in Table 6, along with concentration data for Rb, Sr, Sm, and Nd by isotope dilution. Clinopyroxene has extremely low Rb concentrations (87Rb/86Sr < 2 × 10⁻⁴ for all pyroxenes), and so the isotopic composition of Sr in the pyroxenes today is essentially the same as the whole rock when it last equilibrated. The isotopic composition of Sr ranges from 0.7033 to 0.7042 in the group A eclogites and from 0.7094 to 0.7100 in the group B eclogites, based on the 87Sr/86Sr of clinopyroxene. The two group C samples vary significantly in their 87Sr/86Sr (0.7043 versus 0.7074), despite the similarity of their major- and trace-element characteristics. Plagioclase in the garnet granulate from Angola has Sr with the same composition as the coexisting pyroxene (0.70741 versus 0.70738).
Figure 5A. Chondrite-normalized REE concentrations in whole-rock samples of eclogite. The pronounced LREE enrichment in all of the samples except Kao is due largely to secondary mineralization. Note the positive Eu anomalies in the group C eclogites, only one of which (Angola) contains modal plagioclase. See text for further discussion.

Figure 5B. Primary chondrite-normalized REE concentrations for eclogite whose rocks are reconstructed from the primary-phase compositions and the inferred primary mode of each sample. Most of the LREE enrichment noted in Figure 5A is gone, but the positive Eu anomalies in the group C eclogites remain. See text.

Figure 5C. Measured REE concentrations of eclogite whole rocks normalized to the reconstructed primary values shown in Figure 5B. The flat HREE patterns show correspondence between the measured and whole-rock values and imply that the secondary assemblage has low HREE. The variable LREE enrichment is caused by increasing proportions of the secondary assemblage and implies that the metasomatizing fluid (magma?) was LREE enriched. See text.
The group B eclogites contain highly radiogenic Sr, which implies either derivation from Rb-enriched protolith or isotopic exchange with a component rich in radiogenic Sr, or both. In contrast, the group A eclogites have $^{87}$Sr/$^{86}$Sr ratios that are equal to or slightly depleted relative to the present-day bulk earth. The Kao eclogite has $^{87}$Sr/$^{86}$Sr identical to the group A Bellsbank eclogite, whereas the Angola eclogite contains Sr only slightly less radiogenic than the group B eclogites (Table 6). We emphasize that the enriched Sr isotopic compositions of the Angola and group B eclogite pyroxenes cannot result from insufficient leaching, but must be intrinsic (see Methods section).

The isotopic composition of Nd in each of these eclogites can be calculated from the composition of the primary phases using the estimated primary modes (Table 1). Expressed in terms of their relative deviation (in parts per 10$^4$) from the isotopic composition of an undifferentiated chondritic reservoir ($e_{0}$), the Nd in these eclogites ranges in group A from -19 to +6, in group B from +122 to +235, and in group C from +0.3 to +16 (Table 6). $e_{0}$ of the group B eclogite garnets are among the highest known from any rocks; they are exceeded only by garnets from the Roberts Victor eclogites studied by Jagoutz and others (1985). These high $e_{0}$ values result from the extreme fractionation of REE in the group B garnets and imply that this fractionation occurred at least 1 b.y. ago. $e_{0}$ values of minerals in the group C and group A eclogites are closer to current bulk earth, but their limits are still more extreme than generally encountered in terrestrial samples.

A whole-rock isochron diagram can be constructed for the three Bellsbank samples by using the estimated primary mode of each eclogite and the Sm-Nd systematics of the primary minerals. Because an accurate mode is required to properly reconstruct the whole-rock isotopic characteristics, we cross-checked our textural point counts (in which secondary minerals are assigned to the phases with which they are associated) with mixing calculations that use a least-squares mixing routine to reconstruct the whole-rock composition. The mixing calculations assume that the metamagmatic agent was a kimberlitic fluid that can be modeled as a mixture of kimberlite + phlogopite. This assumption is consistent with petrographic evidence of early phlogopite addition (in veins) and the whole-rock mass balance (see Fig. 5C). The results agree well with the point-count data (Table 1) and are very robust; the proportion of garnet to pyroxene does not change much if the composition of the kimberlite component is altered, although the percentage of kimberlite required does change.

The three Bellsbank eclogites define a whole-rock isochron of 2.1 ± 0.1 b.y. (Fig. 6). This Early Proterozoic age is somewhat younger than the Late Archean ages inferred for the Roberts Victor eclogites (Jagoutz and others, 1984) and for clinopyroxenes from xenoliths in South African and Siberian kimberlites (McCulloch, 1982). The largest uncertainty in the age derives from uncertainties in the reconstructed mode, which are estimated at ±3% from the mixing calculations (Table 1). This age coincides with the Early Proterozoic stabilization of the Orange River mobile belt (Fig. 1).

The Bellsbank eclogite isochron implies an initial $^{143}$Nd/$^{144}$Nd of 0.510893, corresponding to an $e_{0}$ of $+19.7$ (Fig. 6). This initial ratio is almost twice as high as MORB today, and points toward large-scale fractionation of the Earth's mantle early in the Archean. Similar results have been reported by McCulloch (1982) and Jagoutz and others (1984). Note that a high, positive $e_{0}$ is required for sample 437-1, whose whole-rock $^{143}$Nd/$^{144}$Nd is controlled by clinopyroxene with a low Sm/Nd ratio (Table 5).
Even if it is argued that these samples are unrelated, therefore, the need for an early mantle fractionation event is sustained.

**Oxygen Isotopes**

The isotopic composition of oxygen in eclogites has received considerable attention since the pioneering work of Garlick and others (1971), who noted the large variations in δ18O between different eclogite suites. Further work by Clayton and others (1975) showed that these variations cannot be explained by igneous fractionation processes within the upper mantle. The range of oxygen-isotope values in eclogites are in the same range observed for hydrothermally altered MORB (Muehlenbachs and Clayton, 1972; MacGregor and Manton, 1986). The effect of hydrothermal alteration on basaltic depends on the temperature at which the isotopic exchange takes place: low-temperature alteration increases δ18O relative to primary mantle values (+5.5), while high-temperature alteration lowers δ18O by 1–3 per mil relative to the mantle (Muehlenbachs and Clayton, 1972; Gregory and Taylor, 1981).

The isotopic composition of oxygen in the eclogites studied here is listed in Table 6. Only the garnet granulite from Angola comes close to a primary mantle ratio. The lowest values of δ18O (+3 to +3.4) are found in the group B eclogites (Table 6). The high-temperature hydrothermal exchange implied by these low values may be connected with their highly radiogenic 87Sr/86Sr ratios (0.709–0.710). Group A eclogites have higher δ18O (4.7–5.3) than the group B eclogites, but are still lower than primary mantle values. Sr isotopic compositions in the group A eclogites are correspondingly lower than those in group B. The group C eclogite from Kao overlaps the range of δ18O in group A eclogites (4.0–4.7) and has a similar low 87Sr/86Sr ratio. If the Angola garnet granulite is excluded from consideration, there is an overall negative correlation between δ18O and 87Sr/86Sr, similar to that observed in oceanic crust (Muehlenbachs and Clayton, 1972) and in the eclogites from Roberts Victor (MacGregor and Manton, 1986).

**DISCUSSION**

The data presented in the preceding sections show that the eclogites and related samples studied here can be correlated with the three eclogite groups defined by Coleman and others (1965). These distinctions, which are based on the major-element composition of garnets and pyroxenes, are supported by trace-element and isotopic data as well (Table 7). Although the use of these groups (which are defined by geologic setting) is not entirely warranted for a xenolith suite from one of these settings, it does provide a convenient framework to examine the relationships between the diverse samples studied here.

The group B eclogites and group C eclogite from Kao correlate well with the Roberts Victor type B and type A eclogites, respectively, described by Jagoutz and others (1985), and appear to correlate with the type II and type I eclogites described by MacGregor and Manton (1986). The group C garnet granulite from Angola does not fit with the other group C (or type I) samples in many regards and may not be related. Nonetheless, it does display some characteristics similar to "typical" group C eclogites (for example, positive Eu anomaly). The group A eclogites (garnet pyroxenites) do not correlate with any of the eclogite groups described at Roberts Victor but are more similar to garnet pyroxenites found in alkali basalts (for example, Beeson and Jackson, 1970; Sershov and others, 1973), orthopyroxene-bearing eclogites from some other kimberlites (Shue and Gurney, 1979), and two-pyroxene megacrysts (granat) in kimberlite (Meyer and McCallister, 1984).

Both Jagoutz and others (1985) and MacGregor and Manton (1986) suggested that the Roberts Victor eclogites formed by metamorphism of subducting oceanic crust. The close similarity of the group B and group C eclogites described here to eclogites from the Roberts Victor kimberlite suggests that they may have formed by the same process. In the discussion that follows, we attempt to integrate this suggestion with the petrologic and geochronal data presented above to establish internally self-consistent models for the origin of each eclogite group.

**Geothermometry**

Garnet-pyroxene temperatures calculated using the geothermometer of Ellis and Green (1979) at an assumed pressure of 30 kbar range from 785 to 1090 °C. These temperatures do not correlate with the compositional groups but do correlate with geographic location. Kao and the four Bellbank eclogites all have temperatures of about 800 to 850 °C; these temperatures are supported by a two-pyroxene temperature of 790 °C for the Bellbank eclogite FRB-437-2 using the geothermometer of Wells (1977). Garnet-pyroxene temperatures for the Angola garnet granulite (940 °C) and the Deutsche Erde eclogite (1090 °C) are higher than for the other eclogites.

The Bellbank and Kao kimberlites are located on the Kaapvaal craton and sample lithosphere that was almost 2.0 b.y. old at the time of kimberlite emplacement. If the lithospheric source of these eclogites had a geothermal gradient normal for an old, thermally equilibrated craton (–0.95 μcal/cm2-sec), pressures on the order of 33 kb to 36 kb are implied. This is consistent with a pressure of 35–40 kb for sample FRB-437-2, based on the alumina content of orthopyroxene in equilibrium with garnet at 800–850 °C (Lane and Ganguly, 1980; Harley, 1984). These pressures coincide with depths of 100–135 km, significantly shallower than the
165- to 190-km depths inferred by MacGregor and Manton (1986) for the Roberts Victor eclogites. A possible explanation for this discrepancy is that Bellsbank and Kao lie near the edge of the Kaapvaal craton where the subcratonic lithosphere is expected to be thinner than at Roberts Victor (MacGregor, 1975).

Two of the eclogites studied here have equilibration temperatures >850 °C: Angola (940 °C) and Deutsche Erde (1090 °C). Both of these eclogites are from kimberlites that erupted through Proterozoic mobile belts northwest of the Kaapvaal craton. These higher temperatures imply either higher geothermal gradients at the time of equilibration or greater depths of origin. The persistence of relic feldspar in the Angola eclogite suggests the former explanation, because feldspar will react to form aluminous pyroxene and spinel at higher pressures. This sample also has the highest normative feldspar content, however, which will expand the field of plagioclase stability to higher pressures (Griffith and others, 1979). A maximum pressure of 22 kb (70 km) may be obtained for the Angola eclogite by comparing its equilibration temperature to the plug-out reaction boundaries of Ringwood (1975). This implies a minimum geothermal gradient corresponding to a surface heat flow of 1.4 μcal/cm²·sec for Angola at the time of kimberlite emplacement. If the Angola garnet granulite is crustal in origin, the crustal thickness in central Africa (≈37 km) implies a geothermal gradient twice that level. Elevated geotherms such as this may be characteristic of mobile-belt lithosphere (MacGregor, 1975).

Sm and Nd Partitioning

The determination of Sm and Nd concentrations by isotope dilution allows us to evaluate the partitioning of these elements between the constituent phases of these eclogites. We can define K_i = C_i^Sm/C_i^Nd where i = Sm or Nd. The magnitude of this partition coefficient varies between the eclogite groups defined here and may be used to distinguish group A eclogites from the eclogites of groups C and B. The change in this partition coefficient between groups also provides important clues on eclogite petrogenesis.

Group B eclogites have partition coefficients for Sm and Nd between pyroxene and garnet that are low compared to “normal”: K_Sm < 1.0 and K_Nd ~ 4. One result of this relative partitioning is that garnet contains a significant fraction of the total REE and exerts a major influence on whole-rock values of Sm/Nd and 143Nd/144Nd.

Group A eclogites are characterized by strong partitioning of both Sm and Nd into clinopyroxene: K_Sm = 2–4, and K_Nd = 11–27. Similar values are derived from studies of partitioning in mantle peridotites (Shimizu, 1975; Ehrenberg, 1982). As a result, calculated whole-rock compositions are dominated by pyroxene (Table 6). The high partition coefficients result from higher concentrations of REE in pyroxene, not from lower concentrations in garnet. This means that deviation from Henry's Law behavior in garnets (Harrison, 1981) is not responsible for the high partition coefficients in the group A eclogites.

The group C samples show contrasting behavior. K_Sm and K_Nd in the Kao eclogite are identical to the group B eclogite values, suggesting that partitioning was controlled by the same process in both. The Angola garnet granulite has K_Sm = 2 (similar to the group A eclogites) and K_Nd = 0.15 (even lower than the group B eclogites). This reversal results from unusually high REE concentrations in the Angola garnet; REE concentrations in the pyroxene are about the same as in the group B pyroxenes. The Angola eclogite is the only sample in this study that retains plagioclase; however, the effect of this additional phase on garnet-pyroxene equilibrium is unclear.

The similarity of empirical and experimental REE partition coefficients to K_Sm and K_Nd determined for the group A eclogites suggests that the distribution of REE in these eclogites formed by equilibrium partitioning between garnet, pyroxene, and magma. This is consistent with the origin that we propose below for these eclogites as crystal cumulates (± trapped liquid) in mantle dikes. The low partition coefficients of the groups B and C eclogites may be inherited from a previous mineral assemblage (for example, garnet amphibolite) during metamorphic recrystallization to eclogite. Shimizu (1975) has suggested a similar origin for peridotite garnets that have negative HREE patterns, which may be inherited from a clinopyroxene precursor.

Hydrothermal Effects

A major problem with the isotopic data presented earlier is the apparent decoupling of the Sr- and Nd-isotope systems. This problem is especially highlighted by the group B eclogites, which are characterized by both highly radiogenic Nd (requiring an old source depleted in LREE) and highly radiogenic Sr (requiring an old source enriched in Rb). As potential source regions are not likely to be both LREE-depleted and Rb-enriched, simple two-stage evolution of an initial chondritic reservoir is precluded.

The high 87Sr/86Sr of the group B eclogites is correlated with extremely low ²⁰⁸O. As discussed above, low ²⁰⁸O cannot form by igneous fractionation processes, but is a common result of high-temperature hydrothermal alteration (Muelhensbachs and Clayton, 1972). This suggests that the high ⁸⁷Sr/⁸⁶Sr of the group B eclogites is not a primary igneous feature but was imposed on the igneous protolith of these rocks after formation and before or during metamorphism to the eclogite facies. Hydrothermal exchange with a fluid reservoir affecting oxygen and Sr but not Nd thus appears to be a required step in the petrogenetic history of the group B eclogites. This process may have affected the progenitors of the group A and group C eclogites as well.

PETROGENESIS

Group A

The group A eclogites are the only eclogites studied here whose geochemical characteristics are consistent with an origin as cumulate dike rocks in the upper mantle. They have high Mg# primary-phase compositions that are similar to

<table>
<thead>
<tr>
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<th>GROUP B</th>
<th>GROUP C</th>
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<tr>
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<tr>
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</table>
the refractory peridotites thought to compose most of the upper mantle, low concentrations of incompatible elements, and Sr-Nd systematics that are similar to ocean-island basalts. Some group A eclogites also contain modal olivine and enstatite, which are common accessories in upper-mantle pyroxenite dikes (Sershervi, 1979b). The reconstructed REE patterns for these eclogites are consistent with the accumulation of pyroxene and garnet in these dikes, along with minor trapped liquid, and the values for $K_{Sm}$ and $K_{Nd}$ are consistent with empirical pyroxene/garnet partition coefficients. The slightly low $\delta^{18}O$ values in these eclogites may be inherited from the source region of the partial melts.

**Group B**

The group B eclogites are characterized by extremely depleted LREE and $e[O]$ values 10$^2$ to 20$^2$ MORB. The high $^{87}Sr/^{86}Sr$ of these eclogites is not consistent with their strongly depleted LREE and high $e[O]$, and cannot be a primary feature of the protolith; it must have been imposed on the protolith prior to metamorphism. This may occur by isotopic exchange with a reservoir enriched in radiogenic Sr or by high Rb in the protolith prior to metamorphism. The extremely low $\delta^{18}O$ of the group B eclogites cannot form by mantle fractionation processes and must also be inherited from the protolith. Both the low $\delta^{18}O$ and high $^{87}Sr/^{86}Sr$ are consistent with high-temperature hydrothermal alteration of a basaltic protolith. These data support the interpretation that the group B eclogites represent metamorphosed, subducted oceanic crust, as proposed by Jagoutz and others (1984) and MacGregor and Manton (1986).

The low concentrations of incompatible trace elements in the whole rocks and the constituent minerals, and the extreme LREE-depleted/HREE-enriched character of the garnets, are most consistent with a refractory origin for the group B eclogites. The volcanic portion of the oceanic crust will be affected more strongly by dehydration and partial melting than the less-hydrated plutonic portion. The isotopic compositions of Sr and oxygen must be inherited from the "spilitized" protolith, along with its high Na$_2$O (now = jadeitic pyroxene). The lower jadeite in pyroxene relative to the group C eclogites probably reflects Na loss during partial melting. Finally, the low values of $K_{Sm}$ and $K_{Nd}$ in the group B eclogites is not consistent with a magmatic origin and may reflect the formation of a garnet amphibolite precursor to these eclogites during subduction, with expulsion of the REE into the partial melt prior to eclogite-facies metamorphism.

Two major differences between the Bellbank group B eclogites and the Roberts Victor group B eclogites of Jagoutz and others (1984) are the isotopic compositions of Sr and oxygen. Jagoutz and others (1984) found extremely low $^{87}Sr/^{86}Sr$ ratios (0.7009-0.7033) in the Roberts Victor eclogites, and only one of the four samples they studied had $\delta^{18}O < 5.7$. In contrast, the type II (= group B) Roberts Victor eclogites studied by MacGregor and Manton (1986) have $\delta^{18}O = 2.2$ to 5.3, and $^{87}Sr/^{86}Sr = 0.7070$ to 0.7112. Unfortunately, their Sr data are not strictly comparable to ours because they analyzed whole-rock samples, not acid-leached mineral separates.

The high $^{87}Sr/^{86}Sr$ ratios of the Bellbank group B eclogites cannot be derived solely by isotopic exchange with sea water enriched in radiogenic Sr. Veizer and Compston (1976) have shown that (1) the low $^{87}Sr/^{86}Sr$ of early Proterozoic carbonates (0.7042-0.7050) requires equally low ratios in contemporary sea water and that (2) highly radiogenic sea water did not evolve until the Late Proterozoic. MacGregor and Manton (1986) have suggested that enrichment of hydrothermally altered oceanic crust in Rb may lead to elevated $^{87}Sr/^{86}Sr$ in the protolith prior to subduction. Subsequent partial melting of the subducted crust strips this Rb from the newly formed eclogite and leaves a component of unsupported radiogenic Sr in its constituent minerals (MacGregor and Manton, 1986). A similar explanation is required to explain the high, unsupported Sr that we observe in the group B eclogites.

As the "group B" eclogites defined by Coleman and others (1965) represent lenses in high-grade gneiss terranes, we must consider whether a similar origin is consistent with the group B kimberlite eclogites described here. Eclogites from the 525-m.y.-old Munchenbarg gneiss have $\delta[O] = -8.5$ and $^{87}Sr/^{86}Sr = 0.7042$ (Stosch and Languir, 1986). These ratios are significantly lower than the group B eclogites from Bellbank. The strongest evidence against a lower continental crust origin for the Bellbank eclogites, however, are the extremely low $\delta^{18}O$ values; high-grade metamorphic rocks in continental crust are typically characterized by $\delta^{18}O$ values greater than mantle values (Shieh and Schwarz, 1974).

**Group C**

The group C samples have whole-rock chemistry similar to high-alumina basalts or gabbros (Table 5). The positive Eu anomaly that is superimposed on the garnet, pyroxene, and whole-rock REE patterns of the group C samples is characteristic of plagioclase accumulation and requires a cumulate gabbro protolith. The reconstructed whole-rock REE patterns are similar to cumulate gabbros, with flat HREE-MREE at $\approx 10^2$ chondrite, slightly depleted LREE, and small positive Eu anomalies (Fig. 5). The source of the parent magma was somewhat less depleted than MORB and possibly slightly enriched, as shown by $e[O]$ that are close to bulk earth today (1.3 to +7.0). The $^{87}Sr/^{86}Sr$ ratios are consistent with the $e[O]$ and lie on or near the trend of ocean-island basalts in a Sr-Nd correlation diagram. The $\delta^{18}O$ of Angola is slightly higher than that of mantle values, suggesting low-temperature exchange with sea water; Kao is depleted in heavy oxygen compared to mantle values, suggesting exchange at elevated temperatures. Both eclogites contain jadeite-rich pyroxene, consistent with a plagioclase-enriched protolith.

The group C eclogite from Kao (PHN-1850) probably represents the plutonic section of ocean crust, as proposed by Jagoutz and others (1985) and MacGregor and Manton (1986) for Roberts Victor type I eclogites. This interpretation is consistent with the isotopic data, and with our bulk-rock analyses, which suggest that gabbro is the most likely protolith. The plutonic gabbro portion of oceanic crust contains less water than does the overlaying volcanic portion and is less affected by dehydration and partial melting than are the overlaying spilitic basalts. This may explain why the group C type I eclogites are less refractory than are the group B eclogites, which are thought to represent the volcanic portion of the crust.

The Angola garnet granulate retains plagioclase in its mode and must have equilibrated at a lower pressure than the chemically similar Kao eclogite. The Angola garnet granulate differs from typical group C eclogites in other ways. The Angola garnet granulate probably formed in the lower continental crust, as suggested by Griffin and others (1979) for similar plagioclase-bearing xenoliths from southern Africa.

**CONCLUSIONS**

The proposal that eclogite xenoliths in kimberlite represent metamorphosed ocean crust, first presented by Helmstaedt and Doig (1975), has become a popular alternative to the commonly proposed mantle dike origin. We propose that both hypotheses are correct, and that eclogites may form by either mechanism. The eclo-
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gites studied here may be correlated with the
tree ecologie groups defined by Coleman
ta (1965). Two of these ecologie groups may
represent metamorphosed ocean crust; the third
group probably formed as cumulative dikes in the
upper mantle. Unlike the ecologie xenoliths
described by Helmsma and Doig (1975) and the
Franciscan ecologies described by Coleman and
others (1965), which formed during Phanero-
zoic subduction, the ocean-crust ecologies
described here are Early Proterozoic in age
and support the idea that plate-tectonic processes
have been active for at least the past 2.1 by.

The group A ecologies that formed as cumu-
lative dikes in the mantle differ substantially
from ecologies that form from metamorphosed ocean
crust. These samples have mineral assemblages
consistent with fractional crystallization of a bas-
altic melt at high pressure, and mineral compo-
ositions that are in equilibrium with the refractory
peridotite that must form the wall rock. Further
studies are required to determine what relationships,
if any, exist, between these mantle dikes rock and
the associated oceanic ecologies.

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