[AMERICAN JOURNAL OF SCIENCE, VOL. 277, SUMMER, 1977, P. 801-832]

PARTIAL MELTING IN THE JOSEPHINE PERIDOTITE I, THE EFFECT ON MINERAL COMPOSITION AND ITS CONSEQUENCE FOR GEOBAROMETRY AND GEOTHERMOMETRY

HENRY J. B. DICK*

Department of Geology and Geophysics, Yale University,
New Haven, Connecticut 06520

ABSTRACT. The Josephine Peridotite is an alpine-type peridotite, 640 km^2 in area, on the western edge of the Klamath Mountains of southwestern Oregon and northern California. It is characterized by a highly restricted lithology and mineralogy typical of many alpine type peridotites. The peridotite is largely harzburgite (>95 percent), with scattered patches of dunite (<5 percent), and irregular seams of orthopyroxenite and olivine websterite (<<1 percent). The average modal composition of the harzburgite (31 samples) is 78 ± 6 percent olivine, 20 ± 5 percent orthopyroxene, 1.3 ± 1.3 percent diopside, and 0.6 ± 0.3 percent chromian spinel. The harzburgite olivine ranges from $Fo_{91.2}$ to $Fo_{89.5}$, while the orthopyroxene ranges from $En_{91.9}$ to $En_{90.3}$. Accessory spinel in the harzburgite, unlike the silicates, has a large range in composition with Cr(Cr+Al) ratios from 0.18 to 0.76.

The only other comparable bodies of crustal ultramafic rocks, the layered intrusions, differ significantly from the Josephine Peridotite. Whereas gabbroic rocks associated with layered intrusions greatly exceed in volume the ultramafic portion, the proportion of gabbroic to ultramafic rock in the vicinity of the Josephine Peridotite is approximately 1:1. No layered intrusion is known to contain a thickness of an ultramafic rock type similar to that of the Josephine Peridotite. Layered intrusions are also characterized by diverse lithologies and mineralogies and typically have large systematic ranges of silicate compositions with a restricted spinel composition.

The composition of coexisting spinel and olivine indicates a large range of crystallization temperatures in layered intrusions, in contrast to a relatively uniform crystallization temperature throughout the Josephine Peridotite. The Josephine Peridotite is therefore not likely to be a product of fractional crystallization but is instead the residue of partial melting in the upper mantle.

The minerals in the Josephine Peridotite show limited systematic variations in composition (Fe, Ca, Al depletion, Mg enrichment) from pyroxene-rich to more refractory pyroxene-poor samples, consistent with a limited variation in the degree of fusion and melt removal. Though the extent of these variations is small in the Josephine or any other single alpine-type peridotite, taken together for many such bodies they define a broad trend consistent with a large range in the degree of partial melting.

Enstatite compositions in the alpine-type peridotites diverge strongly from those in layered intrusions toward Wo_0 En_{94} Fs_6 suggesting that in the former clinopyroxene was not present as a separate phase during the late stages of melting. Accordingly, the small amounts of clinopyroxene now present in most alpine-type peridotites must have precipitated from trapped melt at the end of fusion. This and the large range in the degree of partial fusion implied for different alpine-type peridotites suggest that a significant amount of melt (possibly >6 percent) could not be removed during melting in the upper mantle.

INTRODUCTION

Alpine-type peridotites are bodies of olivine-rich ultrabasic rock found in zones of alpine-type deformation. They range from small lenticular pods and narrow belts following the local tectonic grain to vast sheets covering more than a thousand square kilometers. Alpine-type peridotites occur in groups along continental margins and island arcs, often forming well defined belts sometimes thousands of kilometers long

---

* Present Address: Department of Geology and Geophysics, Woods Hole Oceanographic Institution, Woods Hole, Massachusetts 02543

coinciding with major fault zones. While alpine-type peridotites are composed of the most refractory of igneous minerals, there is generally little or no contact metamorphism at their margins compared to basic intrusions of similar size (Hess, 1938). Though the peridotites are commonly foliated and highly deformed, apparently at high temperatures, surrounding rocks are often of low metamorphic grade and lack the foliation of the peridotites. The contacts between peridotite and country rock are generally faulted, and the bases of the large sheets appear to have been the loci of penecontemporaneous faulting, shearing, and serpentinization as well as low-grade metamorphism of the underlying country rocks (Armstrong and Dick, 1974; Williams and Smyth, 1973). It is evident, then, that many of the large sheets are crystalline nappes and that alpine-type peridotites have had a complex history prior to their final emplacement.

At the present time, many investigators, impressed by the similarity of the world-wide association of alpine-type peridotite, gabbro, intrusive diorite and diabase dikes, and metavolcanics to geophysical reconstructions of ocean lithosphere have proposed that ophiolites represent fragments of ocean crust and upper mantle (that is, Davies, 1971; Coleman, 1971; Moores and Vine, 1971). The discovery of serpentinite, gabbro, and greenstone at the transverse oceanic fracture zones (for example, Miyashiro, Shido, and Ewing, 1969; Cann, 1971; Aumento and Loubat, 1971) appears to support this hypothesis.

Other investigators, impressed by the persistent association of alpine-type peridotite with orogenic andesite, have proposed that the peridotite represents the cumulate or residual portion of andesitic magmas brought to the surface by faulting, uplift, and erosion (Hess, 1955; Green and Ringwood, 1967; Challis, 1969; Osborn, 1969). Ewart and Bryan (1973) have noted that early Tongan arc suites are similar to some ophiolites, and Miyashiro (1973) has suggested that some ophiolites geochemically resemble island-arc volcanics. Dick (1973) also demonstrated that many of the basic dikes associated with some U.S. Cordilleran ophiolites are late Jurassic marginal intrusives. It appears then that ophiolites represent a variety of oceanic and circum-oceanic provenances.

Different origins have been proposed for the peridotite portion of the ophiolite sequence. Some investigators feel that the peridotite is a cumulate formed by crystal settling from a basaltic magma during the formation of ocean crust (Thayer, 1964, 1967, 1969; McTaggert, 1971) or during fractionation of basaltic magma to form andesite (Osborn, 1969; Challis, 1969). Others have suggested that the ultramafics represent residual mantle material, the refractory residue of fusion and magma generation in either an oceanic or circum-oceanic and environment (Hess, 1964; Green and Ringwood, 1967b; Irvine and Findlay, 1972; Boudier and Nicholas, 1972; Dick, 1974, 1975, 1976; Menzies, 1973). In this paper I examine the chemistry of the minerals in the Josephine Peridotite in light of the partial-melting hypothesis; the bulk lithologies will be examined in a later paper.

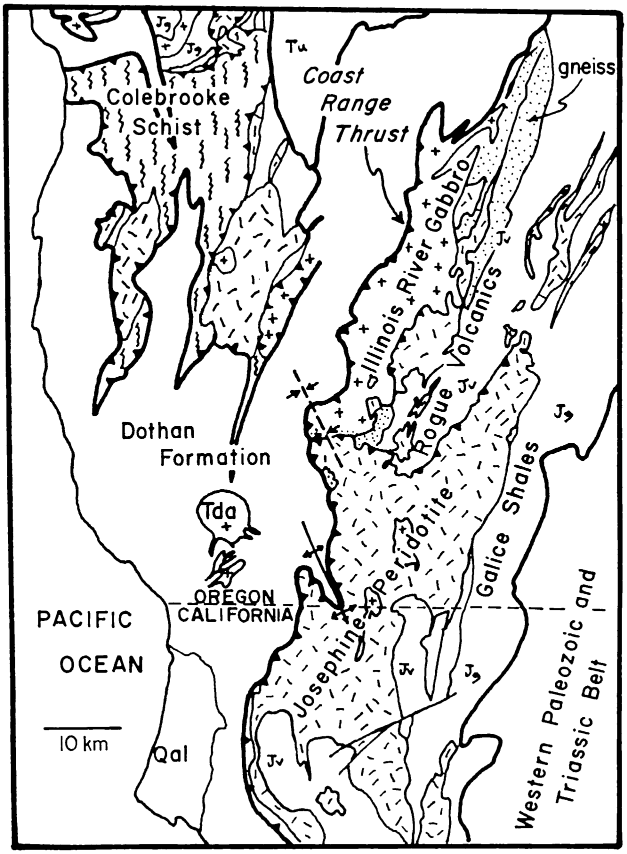


Fig. 1. Generalized tectonic sketch map of the Josephine Peridotite and its vicinity. Data taken from the geologic map of western Oregon published by the Oregon Department of Geology and Mineral Industries, the Weed Sheet, published by the California Division of Mines and Geology, Dott (1971), Coleman (1972), and material gathered by the writer.

REGIONAL GEOLOGY

The Josephine Peridotite is a large east-dipping thrust sheet of alpine-type ultramafic rocks along the western edge of Irwin's (1960) western Jurassic Lithic Belt of the Klamath Mountains (fig. 1). The peridotite covers 640 km^2 and, judging by the exposed relief, is more than one kilometer thick. The numerous outlying tectonic slices of peridotite in the region suggest that it is now only part of a much larger mass which possibly extended over as much as 2000 km^2.

Some of the peridotite is heavily altered to serpentine, particularly at its base and along a number of broad north-northeast-trending shear zones that cross the body. In less altered portions, the peridotite has a well developed internal fabric and folded discontinuous compositional banding both crosscut by Nevadan age faults. The compositional banding appears to have formed during an episode of high-temperature creep which is not evident in the country rocks and therefore must reflect deformation predating the emplacement of the peridotite.

The peridotite was initially emplaced from the east during the late Jurassic Nevadan Orogeny over greenschist facies calc–alkaline volcanics (Rogue Formation) and gabbro, quartz–diorite, and amphibolite of the Illinois River Gabbro Complex. Abundant quartz–diorite and diabase dikes and stocks, similar both in age and chemistry to the gabbro complex, intrude both the Rogue volcanics and the peridotite (Dick, 1973). Many of these dikes appear to have been feeders to the volcanics, are of the same metamorphic grade, and are crosscut by the Nevadan faults. Others, however, have been intruded along the Nevadan faults and into the peridotite. These relations suggest, then, that the peridotite may have been emplaced over an area of active late Jurassic island-arc volcanism represented by the Illinois River Gabbro complex and the Rogue volcanics. The source of the peridotite accordingly may have been a collapsing marginal basin to the east or the tectonically disrupted basement of the arc itself.

Subsequently the peridotite was again thrust to the west as part of a larger plate of Klamath Mountain Nevadan-age basement rocks, overiding Cretaceous and late Jurassic shale, graywacke, and metavolcanics along the Coast Range Thrust. The movement along this fault was intermittent and probably continued well into the Cretaceous (Suppe and Armstrong, 1972). To the east, the peridotite and volcanics are in juxtaposition with metashale and metagraywacke of the late Jurassic Galice Formation along a steep westward-dipping reverse fault. This probably represents a simple late uplift of a prism of crystalline rocks from the western edge of the Nevadan basement. In Tertiary time, the peridotite was involved, along with the rest of the Coast Ranges, in a cycle of low-amplitude folding which produced a series of southeast-plunging synclines and anticlines (fig. 1).

PETROLOGY

Lithology

The Josephine Peridotite is largely an immense mass of harzburgite. Point-count modal analyses of thirty-four samples indicate an average composition and standard deviation of 78 ± 6 percent olivine, 20 ± 5 percent orthopyroxene, 1.3 ± 1.3 percent diopside, and 0.6 ± 0.3 percent spinel. Variations in composition of the harzburgite generally occur gradually over a distance of hundreds of meters. Locally, however, there are large areas showing discontinuous alternating compositional bands of pyroxene-poor and pyroxene-rich harzburgite 2 to 4 cm thick.

Dunite occurs in minor amounts as discontinuous tabular bodies, pods, lenses, and patches from hundreds of meters to centimeters across; it constitutes perhaps less than 5 percent of the peridotite. The dunite apparently formed late and often crosscuts compositional banding in the harzburgite. Chromite is locally concentrated in some dunite bodies as bands, schlieren, and pods. The larger of these deposits commonly contain well developed cumulus textures. Nonetheless, there appears to be a complete gradation from large cumulus ore deposits to centimeter-thick chromite bands in isolated dunite patches.

Volumetrically insignificant amounts of orthopyroxenite and olivine-websterite are found locally as discontinuous seams in the harzburgite. These seams have undergone a complex history and are often folded or may be crosscut and partially replaced by dunite. The orthopyroxenite generally appears to have formed later than the olivine websterite and may locally cut across isolated seams and patches of dunite.

Petrography

Olivine.—Olivine occurs in a range of textures and morphologies typical of plastic deformation (fig. 2), including textures typical of syntectonic recrystallization, annealing recrystallization, and simple plastic deformation of individual grains and perhaps diffusion-controlled dissolution-reprecipitation creep (see discussion). In dunite olivine is commonly much coarser grained (grains 4 or 5 cm long are not uncommon), apparently as the result of the absence of pyroxene which normally acts as a locus for the disruption of olivine during deformation of the harzburgite (Ragan, personal commun.).

Enstatite.—Enstatite grains in the harzburgite are either extremely irregular, lobed, and deeply embayed, or they are rounded, abraded by deformation, and enclosed in a recrystallized matrix of olivine (fig. 2). In thin section there is an apparent irregular variation in the size, number, and shape of clinopyroxene exsolution lamellae within and between specimens of the harzburgite. In general, the degree of clinopyroxene exsolution in enstatite, the amount of separate diopside grains, and the amount of enstatite in the harzburgite decrease together. In some samples, clinopyroxene exsolution in enstatite is closely associated with kinking.

Spinel.—Chromian spinel is an ubiquitous accessory phase in harzburgite and occurs locally as concentrations in dunite although the form of the spinel grains is quite different in the two hosts. Spinel is commonly

TABLE 1

Composition* of olivine from Josephine harzburgite

Oxide	J28-F	J137-8	J46-H	J50-1	J57-7	J120-3
FeO	9.6	9.1	9.0	8.1	8.8	9.5
MgO	49.5	48.4	49.8	51.6	51.1	49.5
SiO_2	39.9	39.7	39.7	40.2	39.6	39.3
NiO	0.4	0.5	0.4	0.4	0.4	0.4
MnO	0.2	0.2	0.2	0.2	0.2	0.2
Total	99.6	97.9	99.1	100.5	100.1	98.9
%Fo	90.1	89.6	90.7	9.17	91.1	90.2

* Electron microprobe analyses made at Yale University using a natural olivine standard. Data here are average of three separate grains from each sample.

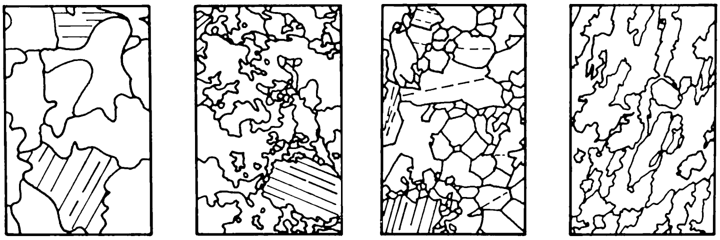


Fig. 2. Synoptic picture of the different textural types found in the Josephine Peridotite. Elements of each of these textures can be found in most thin sections. Deformation mechanisms suggested: (A) dissolution-reprecipitation creep during partial fusion, (B) inter- and intragranular recrystallization: syntectonic recrystallization, (C) polygonal recrystallization: static high temperature annealing recrystallization, and (D) simple plastic deformation and elongation: cold plastic deformation. The textures may have developed sequentially during the upward intrusion of the peridotite, or they may have developed penecontemporaneously due to variable strain rate and/or temperature across the peridotite. Considering strain rate these textures suggest:

(A) low strain rate — high T, melt present, (B) high-intermediate strain rate — high T ± melt, (C) low strain rate — high T, (D) intermediate strain rate — low T.

euhedral or subhedral in dunite. Where concentrated in chromite-rich pods, the chromite grains frequently exhibit relict cumulate textures similar to those shown by the stratiform chromites of layered intrusions. The alpine deposits, however, are smaller and generally appear to have undergone a greater degree of "late magmatic" (post-cumulus) overgrowth and coalescence of grains than those in layered intrusions.

The spinel in harzburgite is commonly anhedral and highly irregular in shape (fig. 3) and only rarely shows euhedral forms. It may be interstitial or intergranular and occasionally even extends from within pyroxene grains into the adjoining olivine groundmass. The spinel is commonly abundant adjacent to or within pyroxene grains in thin section. In volume, it may in some places equal or exceed pyroxene (fig. 3). In some samples, spinel poikilitically encloses brecciated enstatite grains.

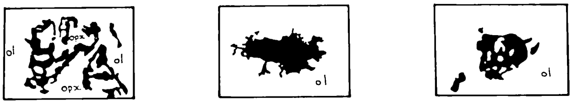
Textural interpretation.—Texturally the harzburgite of the Josephine Peridotite is an ultramafic granulite. The most common textures are those of plastic deformation and recrystallization. Features such as the kink-controlled exsolution of clinopyroxene from enstatite demonstrate that the deformation was, in part, at high temperature (Himmelburg and Loney, 1973, estimate that the temperature for coexisting Josephine clinopyroxene and orthopyroxene was around 800°C). Several lines of evidence, reviewed below, also indicate that deformation and partial fusion of the peridotite occurred together.

Textural type "A" in figure 2 is characterized by smooth curved lobate and cuspate grain boundaries. In metamorphic rocks such features are thought to be characteristic of resorption of mineral grains or dissolution-reprecipitation creep (Sprye, 1969). In the Josephine Peridotite, then, these features might be a partial melting texture. The most significant feature of textural type "A" in this regard is that olivine and pyroxene have much the same grain shape even though they generally

(A) J-120-10
Harzburgite

(B) J-127-11
Harzburgite

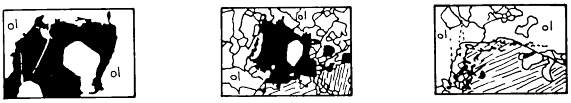
(C) J-127-11
Harzburgite



(D) J-133-8
Harzburgite

(E) J-120-3
Harzburgite

(F) J-124-12
Harzburgite



(G) J-79-6
Dunite

(H) J-31-E
Harzburgite

(I) J-124-5
Harzburgite

Fig. 3. Drawings of accessory spinels in harzburgite. Note the irregular shape; inter- and intragranular character and the fact that some spinel grains and trains cross from pyroxene into groundmass olivine. Field of view: A, D = 1 mm across; the remainder = 3 mm across.

behave differently during mechanical deformation as illustrated by textural types "B" and "C" in figure 2. The features of textural type "A" can be understood if melt was present or forming during deformation; where strain rates were sufficiently low, creep might have occurred by dissolution-reprecipitation with the melt acting as the medium for dissolution at pressure points on impinging grain boundaries, and for reprecipitation in areas where grains do not impinge (a form of diffusion-controlled creep).

Spinel textures in the harzburgite also suggest penecontemporaneous partial fusion and creep of the Josephine Peridotite. Chrome spinel poikilitically encloses brecciated enstatite grains in some thin sections of the harzburgite, suggesting that deformation took place at or above the temperature of spinel formation. The amoeboid, interstitial poikilitic texture of most of the spinel in the harzburgite is unlike that of accessory spinel in either basalts or layered intrusions, where its euhedral shape reflects free growth in a melt. In fact, much of the accessory spinel in the Josephine harzburgite closely resembles that found in metamorphosed peridotite. As the accessory spinel often occurs isolated within the groundmass-olivine or may greatly exceed in volume the associated pyroxene, exsolution from pyroxene or breakdown of garnet does not appear to be a satisfactory explanation for its occurrence. These features, the interstitial nature of most of the accessory spinel, a close spatial association of spinel and pyroxene grains, and the fact that individual spinel grains and trains cross from within pyroxene grains into the adjoining olivine groundmass, suggest that the spinel formed as the result of incongruent melting of pyroxene, its resorption, and the formation of olivine + spinel + melt. Euhedral spinel occurring locally in the harzburgite, then, may be attributed to precipitation from small pockets of melt trapped or moving interstitially through the peridotite.

I believe that the accessory spinel and that in the ore pods formed penecontemporaneously. The cumulate textures in the latter clearly demonstrate that it is magmatic. That the enclosing dunite often cuts across compositional banding in the harzburgite, however, demonstrates that dunite and ores formed late in the history of the peridotite. Apparently, then, the podiform chromitite and dunite bodies formed as lag deposits from relatively large bodies of melt that coalesced from the harzburgite and moved out of, or through, the peridotite during partial melting (see also Dickey and Yoder, 1972; Viljoen and Viljoen, 1969). The large degree of late overgrowth of many spinel concentrates may be explained as growth from additional melt filter pressed through old lag deposits due to prolonged fusion and upward flow of the peridotite from the mantle.

MINERALOGY

Olivine

The average composition of olivine in 19 samples of Josephine harzburgite is $Fo_{90.4}$ ± 0.5. Green (1964) and Challis (1965) compared olivine from alpine-type peridotites (table 2) with that from layered intrusions. Although there is some overlap (Challis, 1965), the olivine from alpine-type peridotites is generally more magnesian and has a much more restricted range of composition. The total range for alpine-type peridotites lies essentially between $Fo_{94.4}$ and $Fo_{87.8}$ (Green, 1964; Challis, 1965; Jackson and Thayer, 1972; Ross, Foster, and Myers, 1954); in contrast olivine from the Stillwater layered intrusive complex alone ranged from Fo_{94} to Fo_{80} (Jackson, 1961). On the other hand olivine from peridotite nodules in basaltic rocks is strikingly similar to that found in the Josephine Peridotite: nine analyses given by Ross, Foster, and Myers (1954) from widespread localities average $Fo_{90.6}$ and have a range of only 1.8 percent.

Enstatite

Analytical results.—Microprobe analyses of 16 enstatites from 8 Josephine harzburgites are plotted in figures 4 and 5. The analyses and analytical procedure may be found in Dick (1976). Clinopyroxene exsolution lamellae were excluded during analysis, and hence the compositions plotted are those of the host grain only. Although the data are limited, the amount of clinopyroxene exsolution in enstatite and separate

TABLE 2

Olivine composition of alpine-type peridotites (tectonites)

Body	Location	Composition	(% Fo)	Source
Lizard	Cornwall	89.1-91.0		Green (1964)
Burro Mtn	California	91.1-91.4		Loney, Himmelburg, and Coleman (1971)
Troodos	Cypress	90.75-91.25		Greenbaum (ms)
Bay of Islands	Newfoundland	87.8-92.2		Irvine and Findlay (1972)
Red Mtn.	California	84.3-90.4	Fo_{90} avg	Himmelburg and Coleman (1968)
New Zealand	New Zealand	89.4-93.8	$Fo_{91.5}$ avg	Challis (1965)
Coast Ranges	Oregon	90.0-90.5		Medaris (1972)
Papua	Papua	91.6-93.6		England and Davies (1973)
Etang de lers	Spain	90.1 avg		Dickey (ms)
Cypress Island	Washington	89-93		Raleigh (1965)
Twin Sisters	Washington	91.8		Ragan (1963)
Mt. Albert	Quebec	89.4-91.6		Irvine and Findlay (1972)
Beni Bouchera	Morocco	90.5-89.1		Kornprobst (1969), Dickey (ms)
Serrania de Ronda	Spain	89.0 avg		Dickey (ms)
New Caledonia	New Caledonia	87-92		Rodgers (1975)
Josephine	Oregon	91.2-89.5	$Fo_{90.7}$ avg	This paper

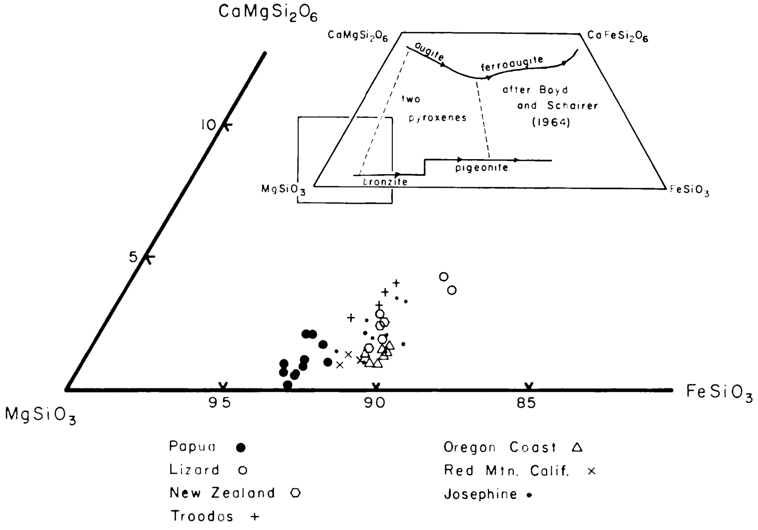


Fig. 4. Alpine enstatite trend. Enstatite corner of the pyroxene quadrilateral showing the systematic compositional variation found collectively for alpine peridotites. The entire pyroxene quadrilateral with the trends for layered intrusions is inset in the upper right corner for comparison. Area in square is that expanded in figure. Note that the alpine trend starts where the layered trend ends and diverges toward Mg-rich compositions on the En–Fs join. The high-temperature Lizard intrusion and low-temperature Papuan sheet plot at opposite ends of the alpine trend suggesting a correlation between the composition of the peridotites and their manner of emplacement. The observed trend with a relatively restricted range for individual bodies is consistent with the interpretation that the peridotites are the residues of partial melting. Source of analyses same as figure 5.

coexisting clinopyroxene grains in thin sections is largest for samples with the most calcium-rich enstatite host grains. In some thin sections, however, there appears to be considerable local variation in pyroxene composition.

The alpine enstatite composition trend.—As seen in figure 4, the enstatite in the Josephine Peridotite, and that in other alpine-type peridotites, has a limited but systematic compositional variation. Collectively, alpine-type harzburgites show a consistent trend across the pyroxene quadrilateral from Wo_4 En_{86} Fs_{10} toward En_{94}. Figure 5 shows an analogous trend of iron and aluminum depletion with magnesium enrichment. In both plots, enstatites from the pyroxene-rich "high-temperature" Lizard intrusion plot at the high-iron end, and those from the pyroxene-poor Papuan thrust sheet plot at the higher magnesium end.

Hess (1941), Brown (1967), and Boyd and Schairer (1964) have shown that pyroxenes from different layered intrusions have a large, systematic chemical variation (fig. 4), and that the enstatites all lie on approximately the same trend parallel to the En–Fs join. Thus variations in the complex natural system only affect the enstatite limb of the two-pyroxene

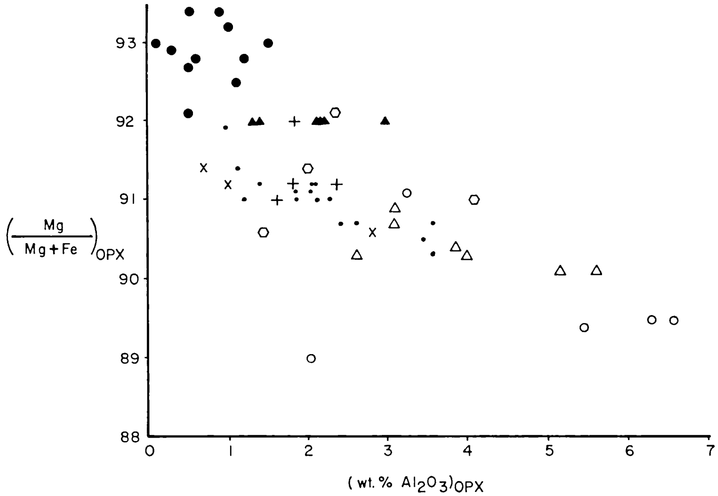


Fig. 5. Plot of Mg/Mg+Fe versus weight percent Al_2O_3 for alpine enstatites. The pattern found in the enstatite quadrilateral is repeated with an overall trend, not entirely obvious in enstatite from any one peridotite, defined collectively by the bodies. As in figure 4, the high-temperature Lizard intrusion and low-temperature Papuan sheet plot at opposite extremes. Closed circle, Papua (England and Davies, 1972); closed triangle, Burro Mtn (Loney, Himmelburg, and Coleman, 1972); plus sign, Troodos (Greenbaum, 1972); hexagon, New Zealand (Challis, 1965); small closed circle, Josephine (Dick, this paper); cross, Red Mtn. (Himmelburg and Coleman, 1968); open triangle, Oregon Coast (Medaris, 1972); open circle, Lizard (Green, 1964).

solvus slightly at magmatic temperatures. Davis and Boyd (1966) suggest that for a given temperature the two-pyroxene field expands when alumina is added to the system, but the most aluminous alpine enstatites are also generally the most calcic. The observed variation of alpine enstatite compositions, therefore, does not appear to reflect changes in the two-pyroxene solvus due to the effect of cations other than Ca, Fe, and Mg.

The expansion of the two-pyroxene solvus with decreasing temperature is well documented and provides a basis for geothermometry in many rocks (Davis and Boyd, 1966; Warner and Luth, 1974). It might therefore account for the alpine enstatite trend, provided that the most Ca-poor enstatites coexist with the most Ca-rich diopsides. From figure 6 it can be seen that Ca-poor Coast Range and Red Mountain enstatites from the western United States do coexist with relatively Ca-rich diopsides, and the variation of coexisting Lizard pyroxene compositions are consistent with a range of equilibration temperatures. On the other hand, in the Papuan and New Zealand Peridotites, many of the most Ca-poor diopsides and enstatites occur together. In fact, it is apparent from figures 4, 5, and 6 that overall decreasing diopside content in enstatite correlates best with increasingly refractory (Mg-rich) Fe- and Al-poor enstatite composition rather than a more Ca-rich coexisting

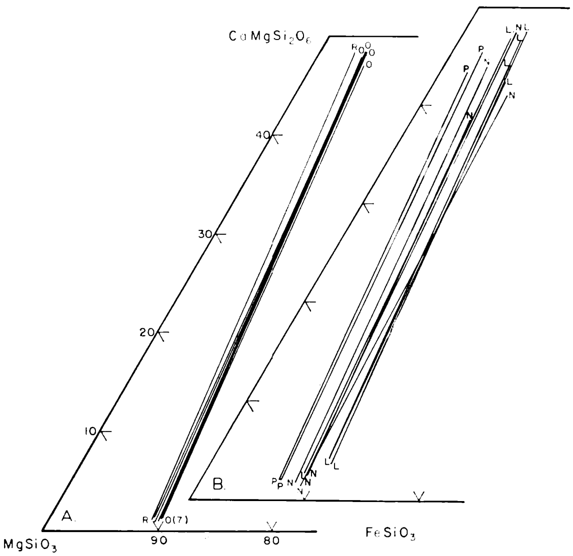


Fig. 6. Plots of coexisting diopside and enstatite in alpine-type peridotites. Note that some of the least diopsidic New Zealand and Papuan enstatites appear to coexist with the most enstatitic diopsides, contrary to what would be expected from the two-pyroxene geothermometer. Source of analyses same as in figure 5.

diopside. Thus, the theoretically lowest temperature pyroxenes are the most refractory and occur in the most refractory, pyroxene-poor rock— a very strong contradiction. Thus, while variation in equilibration temperature may account for some differences in the composition of coexisting pyroxenes within individual peridotites, it is not an adequate explanation for the overall trend of alpine enstatite compositions.

The observed trend of alpine enstatite compositions is that expected, however, for variable degrees of partial fusion and depletion of the mantle in magmaphile elements, perhaps the best explanation for the alpine enstatite trend.

This explanation leads to the very important hypothesis that alpine-type peridotites were generally undersaturated with respect to diopside, at least during the late stages of melting. This appears to be the consequence of the departure of the alpine enstatite trend from the trend for layered intrusions (which presumably reflects the saturation surface under a range of magmatic conditions). The small amount of diopside now present in many harzburgites probably formed at a late stage, either exsolving from enstatite or precipitating from trapped melt.

Pyroxene geobarometry.—Considerable attention has been given to the possibility that the alumina content of enstatite is a function of pressure (for example, MacGregor, 1974, 1967, 1965; O'Hara, 1967). MacGregor's (1974) experimental data on the system MgO–Al_2O_3–SiO_2 in the stability field of spinel peridotite indicate that the alumina content of orthopyroxene decreases with an increase in pressure at constant temperature in the presence of spinel ($MgAl_2O_4$) and olivine.

To date, attempts to utilize the solubility of Al_2O_3 in enstatite in combination with the Ca/(Ca+Mg) ratio of coexisting clino- and orthopyroxene have suggested a wide range of P and T in alpine peridotites (MacGregor, 1974; Wilshire and Jackson, 1975). To some extent, the textural as well as chemical evidence points to more than one generation of pyroxene within individual peridotites (for example, MacGregor, 1974; Challis, 1965; Loney, Himmelburg, and Coleman, 1971; Himmelburg and Loney, 1973; Medaris, 1972). MacGregor (1974) has proposed that these temperatures and pressures record the migration path of the intrusions to the surface and so define a paleo-geotherm. Such a hypothesis, however, suggests that preservation of gross disequilibrium throughout the peridotite, particularly as the spread of pressure indicated within individual alpine-type peridotites vastly exceeds that anticipated from their present thickness (Wilshire and Jackson, 1975). Such gross disequilibrium seems unlikely, considering the long slow cooling of such a large body as the Josephine Peridotite. Wilshire and Jackson, in fact, have shown that large apparent pressure differentials (up to 14 kb) calculated for pyroxenes within individual xenoliths by MacGregor's method actually correspond to bulk composition gradients within the nodules.

It should be noted that exsolution of enstatite from diopside affects the Ca/(Ca+Mg) ratio of the host grain more than the exsolution of diopside affects the Al_2O_3 content of enstatite. As a practical matter it is nearly impossible to get an accurate microprobe analysis of an entire pyroxene grain that contains exsolution lamellae. Most quoted microprobe pyroxene analyses are accordingly those of the host pyroxenes, excluding any late exsolution. If microprobe analyses of pyroxene are used, then the temperature found from the Ca/(Ca+Mg) ratio of the diopside host may not represent that of the recorded pressure of equilibrium found from the alumina content of the enstatite host. Thus, the resulting line of P–T points found for coexisting diopside–enstatite pairs may bear no relation to the actual paleogeotherm.

The application of the enstatite geobarometer to harzburgite also assumes that the pyroxene coexists with a pure $MgAl_2O_4$ spinel. Yet spinel in alpine-type peridotite is generally a chrome spinel with a wide range in composition. Himmelberg and Loney (1973) and I have found that the Al_2O_3 content of pyroxene and spinel vary sympathetically in the Josephine Peridotite (fig. 7), demonstrating that neither spinel nor pyroxene is saturated with $MgAl_2O_4$. Thus without taking into account the diluting effect of Cr_2O_3 on the activity of aluminum in the peridotite, application of the pyroxene geobarometer will give incorrect results (Dick, 1975).

As shown in figure 5, the Al_2O_3 content of enstatite in the Josephine Peridotite decreases with increasing Mg/(Mg+Fe). As discussed in the

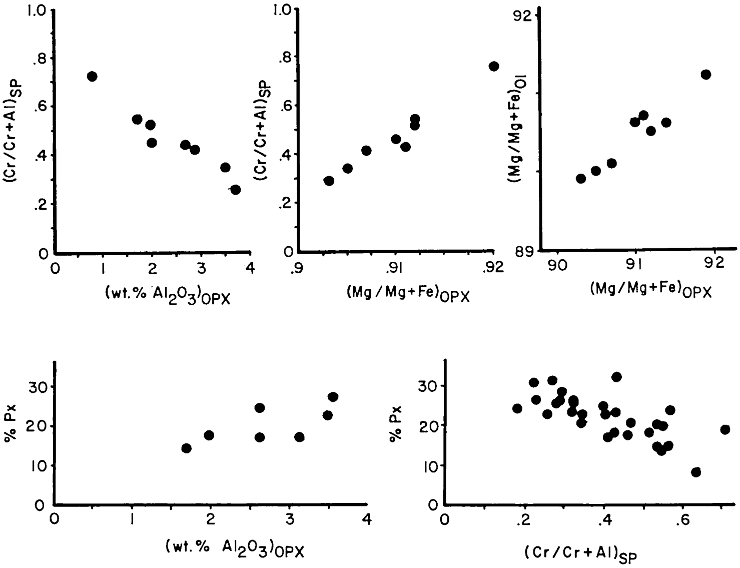


Fig. 7. Covariance of spinel, enstatite, and modal compositions for samples of Josephine harzburgite. (A) Cr/Cr+Al ratio of spinel versus alumina content and Mg/Mg+Fe ratio of the coexisting enstatite and Mg/Mg+Fe ratio of olivine versus Mg/Mg+Fe ratio of enstatite. (B) Modal composition of harzburgite versus alumina content of enstatite and Cr/Cr+Al ratio of spinel (note that olivine plus pyroxene are equal to 99 percent of the harzburgite modal analysis).

next section, the alumina content of spinel decreases with the amount of pyroxene in the rock, and the correlation of the Al content of spinel and pyroxene (fig. 7) suggests the same for enstatite. Thus the first-order effect on the concentration of alumina in alpine enstatites is bulk composition — not pressure. Recently, Stroh (1976) has taken the effect of chromium into account and has estimated that many alpine-type peridotites have recrystallized at low to moderate pressure ($\leq$ 15 kb).

Chrome Spinel

Analytical results.—Compositions of chrome spinel from 62 harzburgites and 48 ores are plotted in figure 8. As Cr, Al, Mg, and Fe^{2+} generally constitute more than 95 percent of the analysis of Josephine chrome spinel, this figure fairly represents most of the major element variation. All the analyses were done using the electron microprobe at Yale University with natural chromite standards. Three grains were analyzed separately and then averaged for each sample. Differences between grains in individual hand specimens were small and generally within the range of analytical precision. Ferric and ferrous iron were calculated from total iron on the basis of spinel stoichiometry. Details and the analyses may be found in Dick (1976).

Josephine chrome spinel is similar to that found in other alpine-type peridotites with a large nearly reciprocal range of Cr and Al contents, a

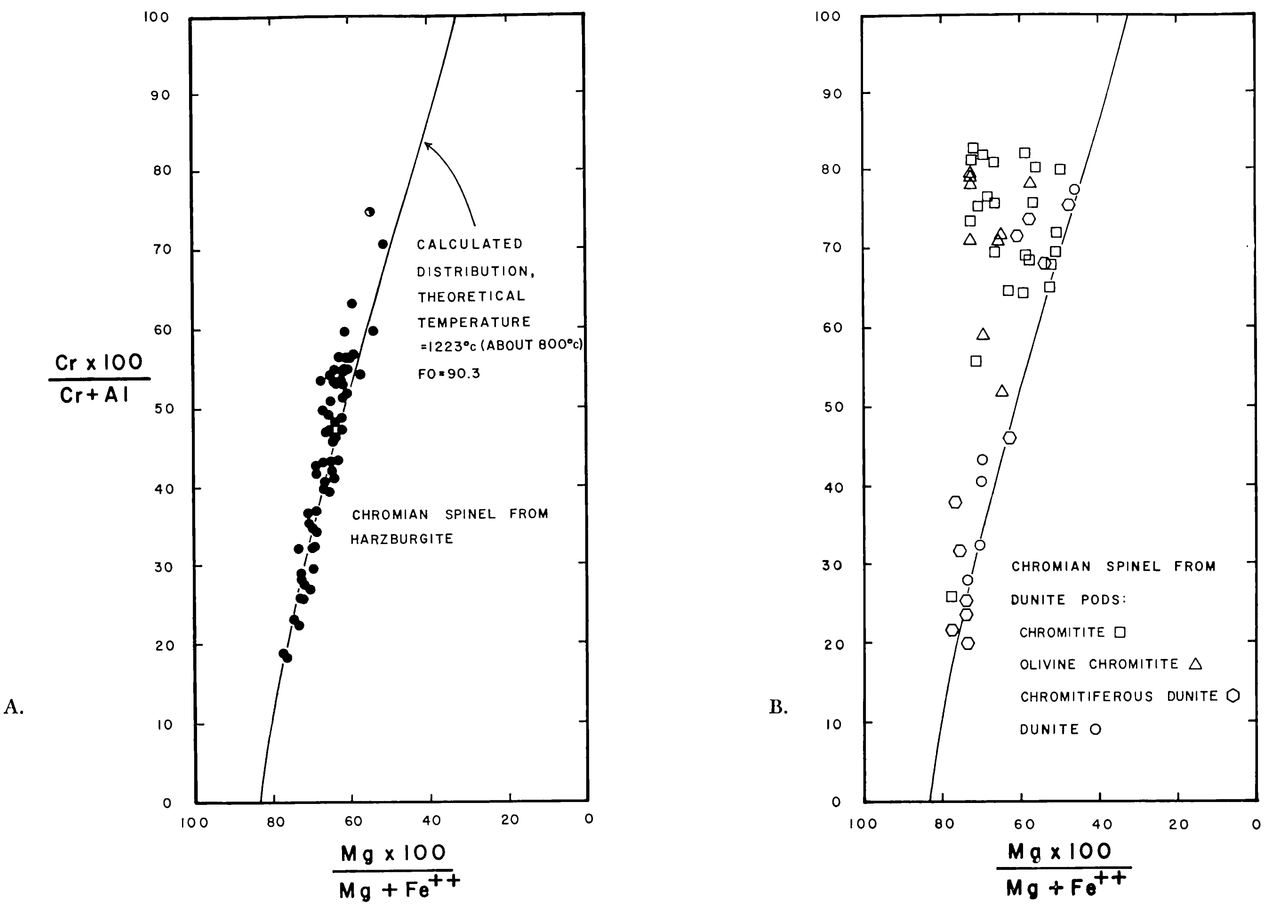


Fig. 8. Plots of chrome spinel from the Josephine Peridotite showing the variation of major components: (A) accessory chrome spinel from the harzburgite country rock, (B) chrome spinel from concentrations in dunite pods (symbol refers to rock type). Fine lines represent the theoretical composition of all chrome spinels in equilibrium with the average Josephine olivine at a nominal temperature of 1223°C. The nominal temperature, calculated using the Jackson equation for Mg–Fe^{+2} partitioning and the average compositions of Josephine spinel and olivine, probably corresponds to a real temperature of about 800°C.

restricted variation of Mg/(Mg+Fe), and a positive correlation between these two ratios (Thayer, 1946; Irvine, 1967). The range in composition found in the Josephine Peridotite is equal to that found previously for all alpine-type peridotites. Analytical data for other alpine-type peridotites has largely been from ore bodies; as shown in figure 8, however, while the range in composition is similar, there are significant differences between spinel from ore and that from harzburgite. In particular there is a much better correlation between Mg/($Mg+Fe^{2+}$) and Cr/(Cr+Al) for the accessory spinel, and, for a given value of Cr/(Cr+Al), the Mg/($Mg+Fe^{2+}$) ratio of the accessory spinel lies at the lower bounding limit for the ore spinel.

Unlike the silicates, chrome spinel has a larger range of solid-solution in alpine-type peridotites than in any other terrestrial rock type. In comparison the composition of chrome spinel from the layered intrusions is relatively restricted.

Partitioning of Mg and Fe^{2+} between olivine and spinel.—Irvine (1965) has shown that while the effect of pressure over moderate ranges may be negligible, the relative distribution of ferrous iron and magnesium between olivine and spinel may be expected to vary significantly with temperature. Jackson (1969), using a thermodynamic treatment similar to that used by Irvine and the available thermodynamic data, has produced an equation relating the composition of coexisting olivine and spinel:

$$T = \frac{5580\,\infty + 1018\beta - 1720\lambda + 2400}{0.90\,\infty + 2.56\beta - 3.08\lambda - 1.47 + \ln K_D} \qquad (1)$$

where

$$K_D = \frac{(X_{Mg}{}^{01})\,(X_{Fe^{2+}}{}^{Sp})}{(X_{Fe^{2+}}{}^{01})(X_{Mg}{}^{Sp})}$$

and ∞, β, and λ are the respective fractions of Cr, Al, and Fe^{3+} in chromian spinel.

Assuming ideal solution behavior, this expression can be used to calculate the theoretical composition of all spinels coexisting with an olivine of fixed composition at a given temperature by substituting different values for ∞, β, λ. Thus we can contour figure 9 for different olivine compositions as shown in figure 10. Irvine (1965) termed these contours $Mg\text{–}Fe^{2+}$ equipotential lines.

Using Jackson's formula and the average olivine and spinel compositions from the Josephine harzburgite yields a theoretical equilibration temperature of 1223°C. Considering the narrow range of forsterite content of the olivine in the Josephine harzburgite (2.2 percent Fo), it is instructive to calculate the composition of all chrome spinels in equilibrium with the average harzburgite olivine ($Fo_{90.3}$) at 1223°C (theoretical) and compare it to the actual distribution in figure 8. As may be seen, the natural assemblage closely approximates the theoretical isothermal distribution curve. A better estimate of the theoretical isothermal spinel distribution curve, however, is obtained using figure 9 and taking into account the small change in olivine composition in the harzburgite (which becomes increasingly forsteritic as the coexisting spinel becomes more chromian). The estimated curve (fig. 9) is the same as that of the natural assemblage within analytical error, suggesting near isothermal crystallization of the harzburgite spinel.

The ore spinels, on the other hand, all lie above the theoretical 1223°C distribution curve calculated for the harzburgite spinels in figure 8. In addition, the olivine in these ores has a much broader range of Mg/(Mg+Fe) ratios than that in the harzburgite. Irvine (1967) has suggested that thermal reequilibration may strongly affect the partitioning of Fe^{+2} and Mg between spinel and olivine. Reequilibration of

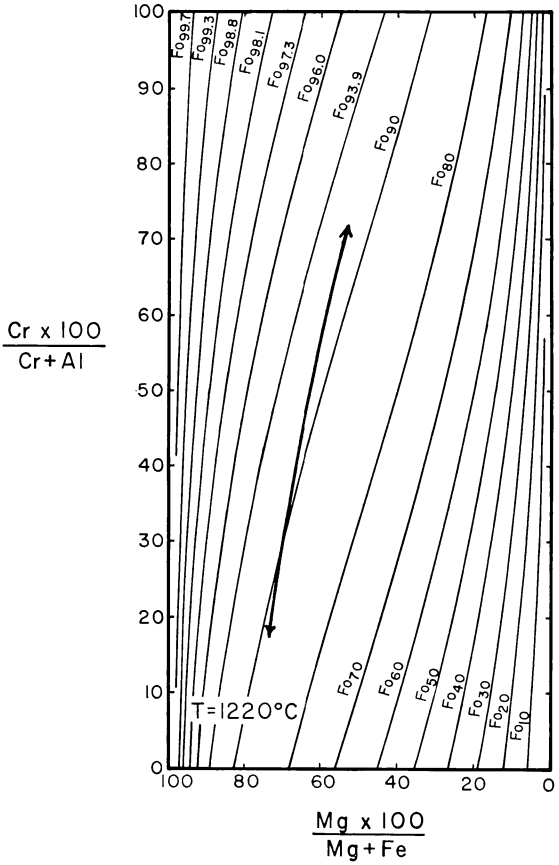


Fig. 9. Mg–Fe^{+2} equipotential plot for spinels in equilibrium with olivines of specific compositions at fixed temperature. Each contour corresponds to the composition of all spinels, for varying activities of Cr and Al, in equilibrium with an olivine of constant composition. T = 1220°C is a nominal temperature predicted by the Jackson spinel geothermometer equation.

chrome spinel with olivine at lower temperatures should shift its composition to the right in figure 8, toward a lower magnesium content, because the change in olivine and spinel compositions, over a given temperature interval, is dependent on the relative proportions in the rock. The shift in composition would be greatest for accessory spinel and almost negligible for olivine in the harzburgite, whereas the reverse would be true for a massive chrome ore with little olivine. This effect is illustrated graphically in figure 10. Thus the larger range of Mg/(Mg+Fe^{+2}) ratios in spinel and olivine from Josephine ores probably reflects their highly variable proportions of spinel and olivine and reequilibration to lower temperatures from an originally uniform high temperature of formation. As anticipated, the largest compositional shift has occurred for accessory spinels, which, because of their effectively uniform proportion relative to coexisting olivine, retained their original isothermal distribution.

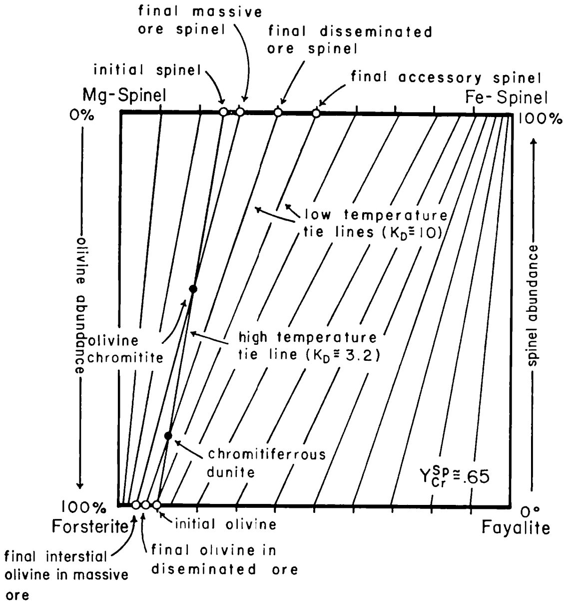


Fig. 10. Effect of bulk composition during equilibration of olivine–spinel mixtures. Tie lines connect coexisting spinel and olivine compositions at low temperatures ($K_D = 10$) and high temperatures ($K_D = 3.2$). $K_D = 3.2$ is measured from coexisting olivine and spinel in Hawaiian basalts crystallized at about 1200°C (Evans and Wright, 1972), and $K_D = 10$ is from the slowly cooled Josephine Peridotite at $Y^{Cr}_{Sp} = 0.65$. The final spinel and olivine compositions are dependent on both their relative proportions in the rock during reequilibration and their initial composition. Thus spinel in a pure chromitite and olivine in a pure dunite will be unaffected by reequilibration. The maximum shift in spinel composition will be for accessory spinel and will be equal to the range of compositions produced for ores consisting of variable proportions of olivine and spinel. Thus the pattern of ore and accessory spinels in the Josephine Peridotite (fig. 8) is exactly that predicted for reequilibrated magmatic spinels initially crystallized near 1200°C.

Irvine (1967) suggests that the original temperature of crystallization can be approximated by using spinel and olivine from adjacent layers formed during the same general stage of paragenesis where they are at their maximum concentration. Using Jackson's formula, olivine–spinel pairs from three isolated pockets of chromite-bearing dunite give excessive temperatures (up to 3059°C) that increase systematically with the spinel concentration (table 3) suggesting that Jackson's formula underestimates the change in composition due to reequilibration.

Evans and Wright (1972) have applied Jackson's formula to spinel–olivine pairs from recent Hawaiian eruptions. Although they found the predicted direction of falling temperature to be correct, the temperatures calculated are again greatly in excess of the actual eruption temperatures. Using the inverse relationship between Ln K_D and temperature with data from the Hawaiian spinels and from metamorphic assemblages, Evans and Frost (1975) have produced a calibrated spinel geothermometer. This geothermometer suggests a final equilibration temperature in the range of 800°C for the Josephine harzburgite spinels, rather than the calculated 1223° temperature.

The Mg–Fe distribution coefficient measured by Evans and Wright (1972) for coexisting olivine and spinel from rapidly cooled basalt ($K_D = 3.2$) can be used to estimate the composition of the harzburgite spinel at magmatic temperatures (fig. 10). The predicted shift in composition for the accessory spinel in figure 10 is the same as the observed spread of ore analyses to the left of the accessory spinel compositions in figure 8 suggesting that the chrome spinel in the Josephine Peridotite crystallized at magmatic temperatures (around 1200°C) and has since reequilibrated down to about 800°C.

Cr and Al partitioning.—It is of interest to consider the wide variation of Cr and Al in Josephine spinel in terms of a possible coexisting melt. The concentration of Cr and Al in spinel should reflect the relative activities of these components in the melt from which it precipitates. Irvine (1967) has contoured the diagram in figure 8 for spinels in equilibrium with melts of different Cr/(Cr+Al) ratios in a manner analogous to that used for partitioning of Mg and Fe between olivine and spinel

TABLE 3

Theoretical equilibration temperature of Josephine olivine–spinel pairs based on Fe–Mg partitioning

Sample No	X_{Mg}^{01}	X_{Mg}^{Sp}*	T°K	% SP
J120-9	90.0	70.1	1323°C	2%
J30-F	90.0	77.3	1511°C	20%
J46-G	91.2	60.5	1663°C	15%
J32-F	90.3**	72.6	2042°C	30%
J112-2	90.3**	75.4	3059°C	99.5%

* $X_{Mg}^{Sp} = N_{Mg}^{Sp}/(N_{Mg}^{Sp} + N_{Fe^{2+}}^{Sp})$, $X_{Mg}^{01} = N_{Mg}^{01}/(N_{Mg}^{01} + N_{Fe^{2+}}^{01}$ where N_i^a = mole fraction of i in a.

** Olivine composition used is that of olivine in average Josephine harzburgite.

in the previous section (fig. 11). Although the diagram is at present only qualitative, it can be used in conjunction with the Mg–Fe contours to interpret the chemical trends of spinel assemblages.

In a hypothetically "normal" magmatic environment, any change in melt composition is associated with a gain or loss of heat. Thus in a differentiated sequence, such as the Stillwater Complex, a systematic change in composition of the principal phases has accompanied progressive crystallization of the melt. Compositions of chromite from the G and H chromitite zones of the Stillwater Complex taken from Jackson

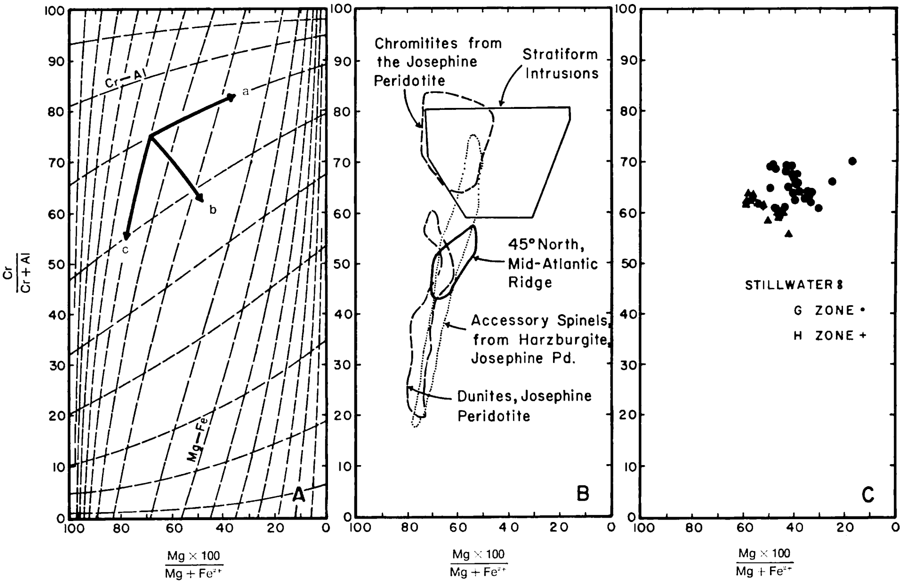


Fig. 11. Compositional variation of spinel in different parageneses.
A. Theoretical trends:

Activity ratios

Trend	Mg/Mg+Fe	Cr/Cr+Al	T°
a	Variable	Constant	Constant
	Constant	Constant	Variable
b	Variable	Variable	Constant
	Constant	Variable	Variable
c	Constant	Variable	Constant

B. Composition fields for Josephine dunite, chromitite and harzburgite spinel, stratiform intrusions and 45° MAR spinels from Aumento and Loubat, 1971 (45°N data plotted in mole percent rather than wt percent as in Aumento and Loubat, 1971).

C. Compositional variation for spinels from G and H zones of the Stillwater Complex (Jackson, 1969). The pattern for the Stillwater spinels, trend B, is consistent with fractional crystallization while the constrasting pattern for Josephine spinels trend C, suggests near isothermal crystallization which is consistent with an origin due to partial melting in the peridotite.

(1969) are shown in figure 11C. As can be seen by comparison with figure 11A, the trend of spinel composition, as pointed out by Jackson, is consistent with a progressive decrease of Mg/(Mg+Fe) and Cr/(Cr+Al) in the melt and falling temperature during two cycles of fractional crystallization.

The present isothermal pattern of Josephine accessory spinel compositions seen in figures 8 and 11B should arise only if the original temperature of formation were also nearly uniform. Reequilibration of accessory spinels with a range of olivine compositions, as in the layered intrusions, would have greatly increased the scatter of accessory spinel compositions in figures 8 and 11B. It is unlikely that spinels of such varied Cr and Al contents could have precipitated from a fractionally crystallizing melt with olivine of nearly uniform composition; while the composition of the spinel indicates a radical variation in the melt composition, during fractional crystallization, that of the olivine indicates almost none.

MnO and TiO_2 content of chrome spinel.—In figure 12, MnO and TiO_2 are plotted against Cr_2O_3 for Josephine spinels. As might be expected, Mn^{2+} appears to follow Fe^{2+} in the spinels with similar positive correlations with chrome in the ores and accessory harzburgite spinels. TiO_2, however, is uniformly low in harzburgite spinels, whereas the ore spinels have larger variable amounts. In neither case do TiO_2 and Cr_2O_3 correlate.

As shown in figure 13, chromian spinel in alpine-type peridotites is generally depleted in TiO_2 relative to that in layered intrusions and in Hawaiian basalts but is similar to that in sea floor basalts and some metamorphic rocks. In particular, spinel from Josephine and Bay of Islands harzburgites is very poor in TiO_2 as compared with spinel in most terrestrial rocks.

The geochemical behavior of Ti^{4+} is different from that of elements such as Cr^{3+} and should be enriched in the melt relative to the solid phases during crystallization or partial melting (Burns, 1973). It has been demonstrated in lunar basalts that titaniferous spinels form late in crystallization (Cameron, 1970), and Reid (1971) has shown that there is probably a continuous simple solid-solution series from chromian to ulvö–spinel. Accordingly, the titanium content should be useful in placing a spinel in a differentiation sequence. Evans and Wright (1972) found that TiO_2 content increases systematically with FeO and Fe_2O_3 contents in spinel from Hawaiian basalts with decreasing amounts of Cr_2O_3. In figure 14, it can be seen that there is also a rough negative correlation between the contents of Cr_2O_3 and TiO_2 of Great Dyke spinels. The contrasting behavior of TiO_2 in Josephine spinels suggests, therefore, that they did not form in a normal differentiation sequence.

The low uniform TiO_2 content of the spinels is consistent with a TiO_2-depleted residue of partial melting. The higher, but variable, concentrations of TiO_2 in the ore spinels could be explained as due to

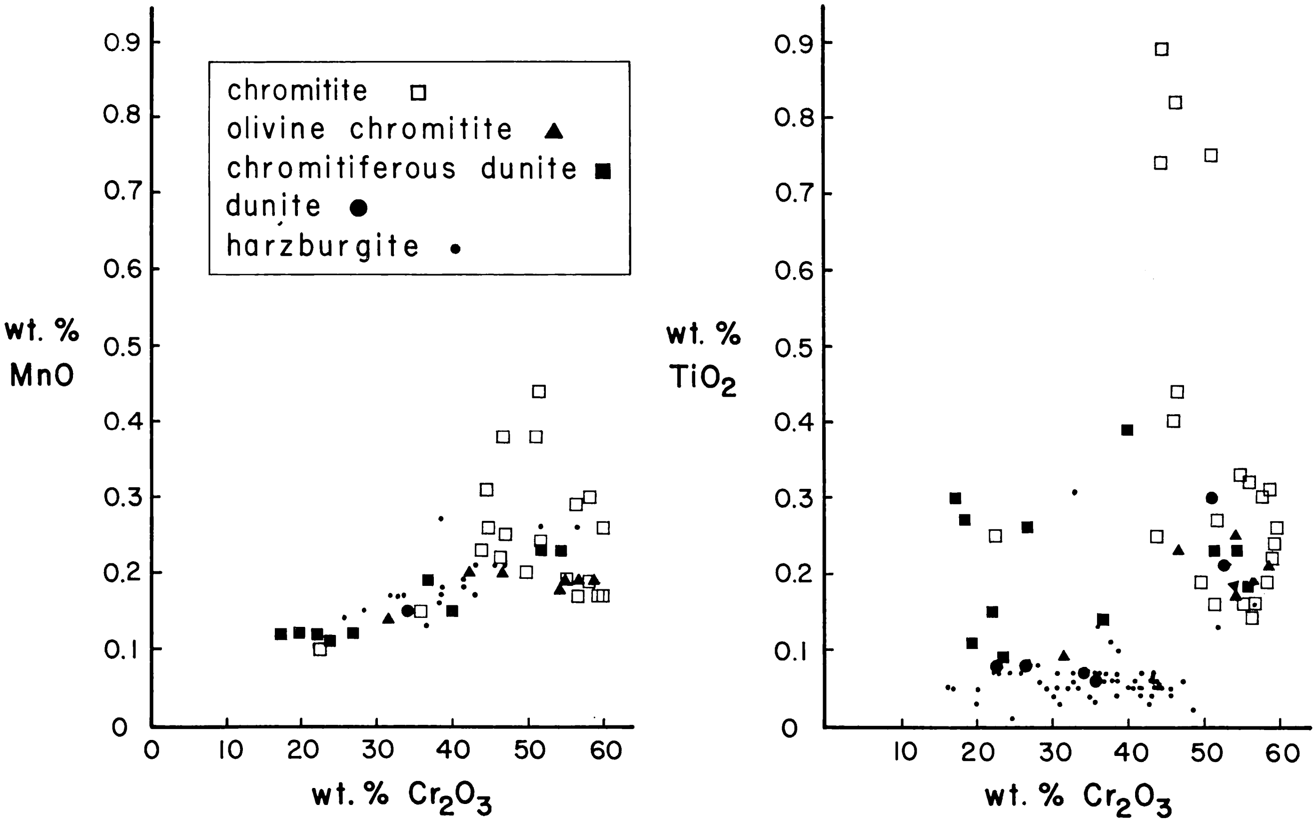


Fig. 12. (A) Plot of MnO versus Cr_2O_3 of Josephine spinels. The behavior of MnO in the spinels is similar to that of iron; at a given concentration of Cr_2O_3 ore spinels have a broader range of MnO than accessory spinel from the harzburgite. (B) Plot of TiO_2 versus Cr_2O_3 for Josephine spinels. Symbols refer to the host rock of the spinel.

direct crystallization from pockets of TiO_2-enriched melt moving out of the peridotite. The degree of enrichment of MnO in basalt and andesite relative to MnO in peridotite is similar to that of FeO but much less than that of TiO_2. The difference in the behavior of TiO_2 and MnO in Josephine spinels is to be expected, therefore, if they formed during melting in the harzburgite.

Chrome spinel as an oxygen fugacity indicator.—Experimental and theoretical work has shown that the composition and stability of chrome spinel is dependent on F_{O_2} (Irvine, 1965, 1967; Ulmer, 1969; Hill and Roeder, 1974). Using Irvine's (1965) treatment, we may relate F_{O_2} and the composition of coexisting spinel, olivine, and pyroxene in the following manner:

since:

$$6Fe_2SiO_4 + O_2 = 2Fe_3O_4 + 6FeSiO_3 \tag{2}$$

then:

$$F_{O_2} = (a_{Mt}{}^{Sp})\,(a_{Fs}{}^{Opx})^6/K_D\,(a_{Fa}{}^{Ol})^6 \tag{3}$$

Since the compositions of olivine, pyroxene, and spinel may all be related, F_{O_2} may be determined on the basis of spinel composition alone if the other phases are present. Thus Irvine (1965), estimating K_D to be about $10^{-5.4}$ at 1250°C, has been able to contour the spinel composition prism

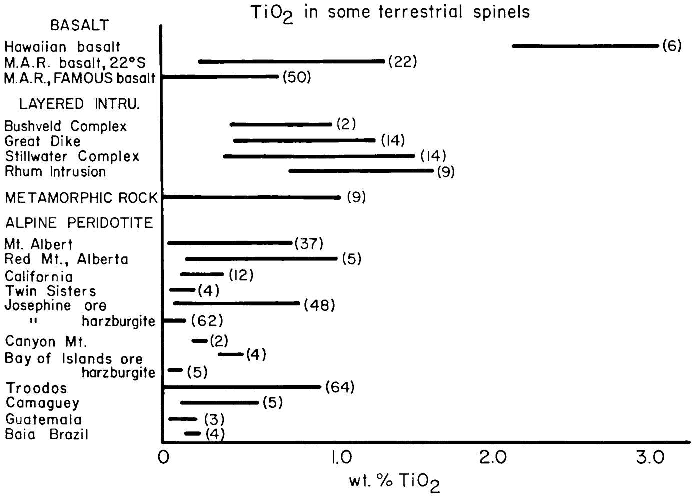


Fig. 13. TiO_2 contents of some terrestrial spinels: Hawaiian basalts, Evans and Wright (1972); mid-Atlantic ridge basalt: 22°S, Carol (unpub. data), FAMOUS basalt, Dick (in preparation); Bushveld Complex, Cameron and Desborough (1969); Great Dike, Bichan (1969); Stillwater Complex, Samson (1969), Stevens (1944); Rhum Intrusion, Henderson (1975); Metamorphic rocks, Evans and Frost (1975); Mt. Albert, MacGregor and Smith (1963); Red Mt. (Kenai Peninsula), California Alpine-type peridotites, Twin Sisters Dunite, Canyon Mt. (Oregon), Camaguey (Cuba), Guatemala, Baia Brazil, Stevens (1944); Josephine ores and harzburgite, Dick (1976); Bay of Islands ore and harzburgite, Malpas and Strong (1975); Troodos, Greenbaum (ms). Unless specified spinels may be either accessory or in ore concentrations.

with a set of surfaces representing the possible compositions for different oxygen fugacities at constant P and T. The intersection of these surfaces (F_{O_2} isobars) with the chrome- and aluminum-free faces of the spinel composition prism and the compositions of Josephine spinels are plotted in the projection of figure 15.

Despite possible errors in the calculation of ferric iron, the harzburgite spinels plot along a relatively well defined trend in figure 15 that resembles an oxygen-fugacity isobar. This trend might be interpreted, therefore, as the result of a restricted low range of oxygen fugacities reflecting general equilibrium over a large volume of rock. Extrapolation of the intermediate Josephine spinel compositions between the suggested positions of the Cr–free and Al–free F_{O_2} isobars, however, actually indicates a significant increase in F_{O_2} from Al–rich to Cr–rich ends of the Josephine trend (or vice-versa).

Though the dunite and chromitite spinels contain more Fe_2O_3 than the harzburgite spinels and are similar to those from layered intrusions, they plot like the harzburgite spinels in figure 15. Spinel in the dunite and chromitite, therefore, appears to record higher oxygen fugacities than the harzburgite spinel.

The chromitite and dunite both lack pyroxene, however, and a_{Fs} may be equal to or lower than in a similar assemblage with pyroxene, so that the value of F_{O_2} determined from figure 15 represents only a maximum. As the dunite and chromitite appear to have formed late in the pyroxene bearing harzburgite, the a_{Fs} during initial spinel crystallization was probably the same as in the harzburgite. The oxygen fugacity is a function of temperature and would normally decrease during cooling of the harzburgite. Thus reequilibration at lower temperatures should shift reaction (2) to the left, consuming $FeSiO_3$ from the pyroxene and Fe_3O_4 from the spinel in the harzburgite. Reequilibration would only occur in the dunite and chromitite to the extent that spinel could react with pyroxene in the adjoining harzburgite. Thus the high ferric iron content of dunite and chromitite spinels may have been essentially "locked in", in the absence of pyroxene, since initial high temperature

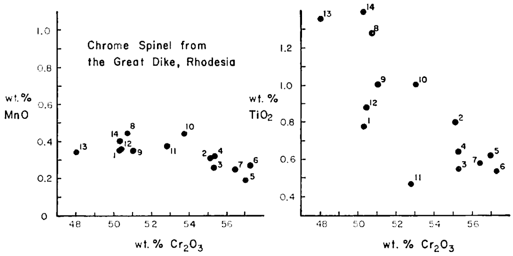


Fig. 14. (A) Plot of MnO versus Cr_2O_3 for Great Dike spinels. Unlike the Josephine spinels in figure 12, there is a rough correlation. (B) Plot of MnO versus Cr_2O_3 for Great Dike spinels. Trend has opposite slope to that in figure 12 for Josephine spinels. Numbers refer to sequential stratigraphic position in section going from bottom to top.

crystallization, while the lower ferric iron content of the harzburgite spinels reflects late low-temperature reequilibration.

If the preceding argument is correct, then F_{O_2} during initial spinel crystallization in the Josephine Peridotite may have been similar to that in the layered intrusions. This seems to me to be the best explanation. There are, however, alternate explanations for the observed pattern of spinel compositions. For example, dunite containing accessory and ore concentrations of spinel is often heavily serpentinized relative to the harzburgite, and the spinels may have oxidized late during serpentinization. On the other hand, with their generally chrome-rich, iron-poor compositions, at a given F_{O_2} the dunite and chromitite spinels should plot below the harzburgite spinels in figure 15. If the effect of reequilibration is not large and if water content controlled F_{O_2} in the peridotite, the pattern might reflect crystallization of spinel in H_2O enriched pockets of melt and H_2O depleted residual mantle.

DISCUSSION

A body of ultramafic rock the size of the Josephine Peridotite has only two likely sources: the mantle, which has the appropriate geophysical parameters, or cumulates formed by fractional crystallization of a basic magma beneath an ocean ridge or in some other large intrusion. The enormous thickness of a single rock type, harzburgite, found in the

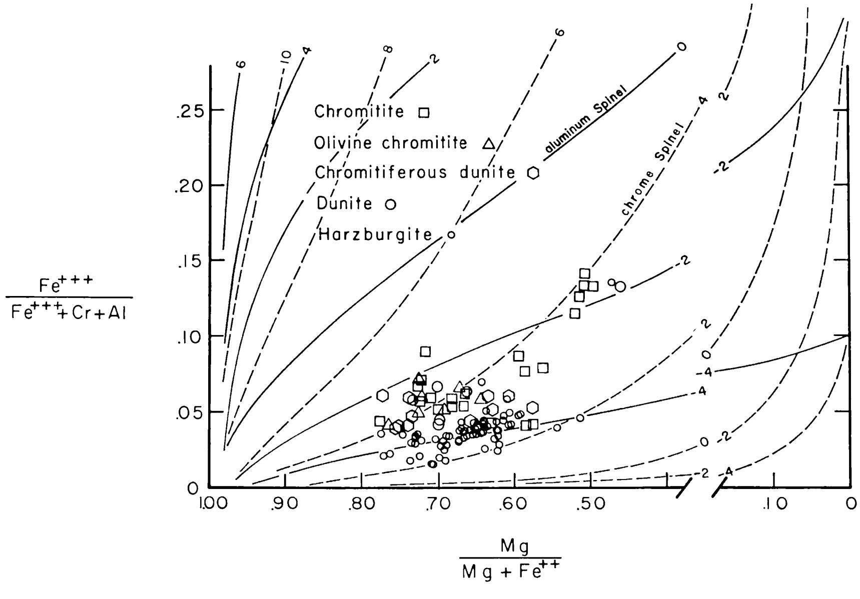


Fig. 15. $Fe^{+3}/(Fe^{+3}+Cr+Al)$ versus $Mg/(Mg+Fe^{2+})$ projection of Stevens' (1944) spinel composition prism contoured with theoretical constant oxygen fugacity isobars. Dashed lines are for aluminum free system, solid lines for chrome free system. To interpret natural assemblages with variable Cr:Al ratios one must interpolate between the two sets of isobars.

Josephine Peridotite, however, has no known parallel in a differentiated intrusion (see Irvine and Findlay, 1972), a major problem for any fractional-crystallization model for the Josephine Peridotite. The proportion of gabbroic to ultramafic rocks exposed in the vicinity of the Josephine Peridotite does not greatly exceed 1:1. The fractional crystallization of basaltic melts, on the other hand, is thought to yield gabbroic and ultramafic fractions in the approximate proportions of 10:1. In the Stillwater and Bushveld Complexes the proportions of exposed gabbro and ultramafics, despite erosion, are about 4:1 and 6:1 respectively (Jackson, 1961). Given the very restricted compositions of the silicates in the ultramafic, compared to those in layered intrusions, the proportion of gabbroic differentiates produced by fractional crystallization of the Josephine Peridotite must greatly exceed that for layered intrusions. The present thickness of the Josephine Peridotite, for example, would require basic differentiates much thicker than the present-day ocean crust.

A variety of recent experimental, theoretical, and petrographic studies (for example, Kushiro, 1968, 1975; Burns, 1973; Green and Ringwood, 1967a and b; Irvine and Findlay, 1972; Ito and Kennedy, 1967; Presnall, 1969; Carter, 1970; Menzies, 1973; Menzies and Allan, 1974; Mysen and Boettcher, 1975; and many others) provides a basis on which to distinguish the residue of partial melting.

Partial melting is likely to produce residues of restricted lithology and mineralogy, such as the Josephine Peridotite, that should stand in sharp contrast to the varied lithology and mineralogy produced by fractional crystallization. During fractional crystallization there is a continuous change in melt composition, with different phases appearing successively on the liquidus. Thus the precipitates are characterized by sharp, often cyclic, changes in lithology and by solid solutions with large systematic compositional variations. In contrast partial melting produces a continuous change in the composition of the residue, successively eliminating crystalline phases as it proceeds. Though, in theory, simple melting alone may drive the composition of rocks with different initial compositions apart, the mechanical effect of the collection and removal of melt during fractional fusion is probably an efficient means for eliminating such differences. Reequilibration due to the shift in the bulk composition of the system (rock + melt) as melt passes through an area will cause a shift in the composition of the new host rock toward that of the magma source — in essense a process similar to zone refining (Yaroshevskii, 1967).

During fractional fusion, the residue remains a chemically open system until melt removal stops. Thus if equilibrium between melt and crystal fractions is maintained, the temperature of formation of the mineral residue is that of the end of melting. Melting in the mantle is probably a regional event without large lateral temperature gradients. Thus in the course of melting, the composition of temperature-dependent solid solutions such as olivine should be driven toward restricted refractory compositions over a large geographic area. During fractional crystallization, on the other hand, the residues are isolated by continued sedimentation from the magma. Thus the precipitates in a differentiated intrusion, such as spinel, form over a temperature range reflecting the entire cooling history of the magma.

The residue of partial melting should also be distinguishable from the products of fractional crystallization by its refractory composition. The extraction of melt will leave a residue depleted in fusible components like Na, Al, K, and Ti and enriched refractory elements like Cr and Ni. Although monomineralic refractory precipitates are common in layered intrusions, they contain up to 40 percent interprecipitate liquid. This liquid is enriched in fusible components relative to the precipitate and often forms a complex intercumulus with a mineralogy unlike the restricted mineralogy to be expected in a melting residue. Refractory monomineralic orthocumulates are found in layered intrusions, but they are generally rather local as their formation requires special conditions with low sedimentation rates.

I consider, then, that the Josephine Peridotite is the residue of partial melting. The restricted composition, lithology, mineralogy, silicate mineral compositions, and granulite texture all point to this interpretation. The harzburgite also has little potential basaltic mesostatis and, as expected, is depleted in fusibles (that is, Al, Ca, Na, K, Si, Fe) and enriched in refractory elements (that is, Mg, Cr, Ni). In addition, olivine–spinel mineral pairs reflect isothermal crystallization over a large area, as predicted for a residue of mantle melting.

There is also a close relationship in the Josephine Peridotite between the composition of chrome spinel, that of the coexisting silicate phases, and the modal composition of the harzburgite. As the rock becomes pyroxene-poor and olivine-rich (more refractory), the spinel is enriched in Cr and depleted in Al while at the same time the coexisting pyroxene is depleted in Al and Fe and enriched in Mg (fig. 7). This direct relationship between modal proportions and phase composition, not identified in layered intrusions, is expected in the residue of partial fusion where there would be a continuous change in mineral composition and proportions as melting advanced.

Despite the restricted composition of the silicate minerals, a detailed examination of the composition of spinel and orthopyroxene in the Josephine Peridotite demonstrates systematic variations unlike any in the known products of fractional crystallization. These trends reflect progressive depletion of the residual minerals in fusibles as the peridotite goes to a refractory olivine rich composition. Taken together alpine peridotites define much larger trends, that apparently record varying degrees of fusion and depletion for different peridotites. This variation may be the result of different melting histories and mantle environments or of initial compositional heterogeneity in the mantle.

The data in this paper have a number of important implications for models of magma genesis. First, minerals like enstatite, which are the principal reservoir for fusible components, do not have fixed compositions during melting, particularly with respect to magmaphile or incompatible elements. Alumina, for example, decreases in concentration in residual enstatite from 6.8 to 0.1 wt percent while the Mg/(Mg+Fe) ratio changes by only 4 percent (fig. 7). Thus the assumption of constant, or near constant, mineral composition during partial fusion in models of magma genesis is likely to be wrong.

Second, since the Josephine Peridotite was undersaturated with respect to diopside during melting, the diopside now present in the harzburgite must have formed at the end of melting either from trapped melt or exsolved from enstatite. Accordingly we can use the volume of separate diopside grains in the harzburgite to calculate the *minimum* amount of melt trapped in the peridotite. Various assumed melt compositions produce estimates of around 6 to 7 percent trapped melt. Since the peridotite appears to have been undergoing plastic deformation during melting and was therefore at its yield point, melt extraction must have been highly efficient. Thus at relatively shallow mantle depths it may be extremely difficult, if not impossible, to tap off a significant volume of melt representing much less than a 6 or 7 percent partial melt. On the other hand, since the effectiveness of melt extraction is dependent on magma viscosity which is in turn dependent on pressure (Waff, 1975; Kushiro, Yoder, and Mysen, 1976), this observation is not inconsistent with the derivation of such melts at much greater mantle depths, and it may explain the rarity of alkaline melts at ocean ridges, where large volumes of tholeiitic basalt appear to be separating from the mantle at relatively shallow depths.

ACKNOWLEDGMENTS

This paper represents work done, in large part, in fulfillment of the requirements for the doctoral degree at Yale University. Writing and editing of successive versions of the manuscript were done while in residence at the Woods Hole Oceanographic Institution as a post-doctoral investigator. I would like to extend my thanks to Dr. Richard Armstrong, my principal advisor, and to Dr. Horace Winchell, Dr. Philip Orville, and Dr. John Rodgers of Yale University without whom the work would neither have been initiated nor completed.

I am grateful to Dr. Neil Irvine of the Geophysical Laboratory for his excellent review and seemingly tireless efforts which have produced a much improved manuscript. Reviews of this paper by Dr. G. Thompson, Dr. W. Bryan of the Woods Hole Oceanographic Institution, and Dr. Martin Menzies of UCLA were also greatly appreciated. Mr. Wallace Phelps of Yale University provided considerable technical assistance during sample preparation.

This work has been supported by grants from the National Science Foundation, The Oregon Department of Geology and Mineral Industries, and Sigma Xi.

REFERENCES

Armstrong, R. L. A., and Dick, H. J. B., 1974, A model for the development of thin overthrust sheets of crystalline rock: Geology, v. 2, p. 35-40.

Aumento, F., and Loubat, H., 1971, The mid-Atlantic Ridge near 46°N, XVI, serpentinized ultramafic intrusions: Canadian Jour. Earth Sci., v. 8, p. 631-663.

Bichan, R., 1969, Chromite seams in the Hartley Complex of the Great Dike of Rhodesia: Econ. Geology Mon. 4, p. 95-113.

Boudier, F., and Nicholas, A., 1972, Fusion partielle gabbroique dans la lherzolite de Lanzo (Alpes piemontaises): Bull. Suisse. Mineralogie Petrographic, v 52, p. 39-56.

Boyd, F. R., and Schairer, J. F., 1964, The system $MgSiO_3$–$CaMgSi_2O_6$: Jour. Petrology, v. 5, p. 545-560.

Brown, G. M., 1967, Mineralogy of basaltic rocks, *in* Hess, H. H. and Poldervaart, Arie, eds., Basalts: New York, Intersci. Pub., p. 102-162.

Burns, R., 1973, The partitioning of trace transition elements in crystal structures: a provocative review with applications to mantle geochemistry: Geochim. et Cosmochim. Acta, v. 57, p. 2395-2403.

Cameron, E. N., 1970, Opaque minerals in certain lunar rocks from Apollo 11, *in* Proc. Apollo 11 Lunar Sci. Conf.: Geochim. et Cosmochim. Acta, Suppl. 1, v. 1, p. 221-245.

Cameron, E. N., and Desborough, G. A., 1969, Occurrence and characteristics of chromite deposits — Eastern Bushveld complex: Econ. Geology Mon. 4, p. 23-40.

Cann, J. R., 1971, Petrology of basement rocks from Palmer Ridge, N.E. Atlantic: Royal Soc. (London) Philos. Trans. v. 268, p. 605-617.

Carter, J. L., 1970, Mineralogy and chemistry of the earth's upper mantle based on the partial fusion — partial crystallization model: Geol. Soc. America Bull., v. 81, p. 2021-2034.

Challis, G. A., 1965, The origin of New Zealand ultramafic intrusions: Jour. Petrology, v. 6, p. 322-364.

——— 1969, Discussion of the paper "The origin of ultramafic and ultrabasic rocks" by P. J. Wyllie: Tectonophysics, v. 7, p. 495-505.

Coleman, R. G., 1971, Plate tectonic emplacement of upper mantle peridotites along continental edges: Jour. Geophys. Research, v. 76, p. 1212-1221.

——— 1972, The Colebrooke schist of southwestern Oregon and its relation to the tectonic evolution of the region: U.S. Geol. Survey Bull. 1139, 61 p.

Davies, H. L., 1971, Peridotite-gabbro-basalt complex in eastern Papua: an overthrust plate of oceanic mantle and crust: Australian Bur. Mineral Resources, Bull. 128, p. 48.

Davis, B. T. C., and Boyd, F. R., 1966, The join Mg_2SiO_2–$CaMgSi_2O_6$ at 30 kilobars pressure and its application to pyroxenes from kimberlites: Jour. Geophys. Research, v. 71, p. 3567-3576.

Dick, H. J. B., 1973, K-Ar dating of intrusive rocks in the Josephine Peridotite and Rogue Formations west of Cave Junction, southwestern Oregon: Geol. Soc. America Abs. with Programs, v. 5, p. 33-34.

——— 1974, The Josephine Peridotite, a refractory residue of the generation of andesite: Am. Geophys. Union Trans., v. 56, p. 464.

——— 1975, Partial melting in the Josephine Peridotite and its consequence for pyroxene geobarometry and thermometry, *in* Boettcher, A. L., ed., Extended Abstracts: Internat. Conference on Geothermometry and Geobarometry, Pennsylvania State University.

——— 1976, The origin and emplacement of the Josephine Peridotite: Ph.D. thesis, Yale Univ., New Haven, Ct. (Univ. Microfilms, Ann Arbor, Michigan).

Dickey, J. S., ms, 1969, Partial fusion products in alpine-type peridotites: Serrania de la Ronda and other examples: Ph.D. dissert., Princeton Univ., Princeton, N. J.

Dickey, J. S., and Yoder, H. S., 1972, Partitioning of Cr and Al between clinopyroxene and spinel: Carnegie Inst. Washington Year Book 71, p. 384-392.

Dott, R. H., Jr., 1971, Geology of the southwestern Oregon coast west of the 124th meridian: Oregon Dept. Geology and Mineral Industries Bull. 69, 63 p.

England, R. N., and Davies, H. L., 1973, Mineralogy of ultramafic cumulates and tectonites from eastern Papua: Earth Planetary Sci. Letters, v. 17, p. 416-425.

Evans, B. R., and Frost, B. R., 1975, Chrome spinel in progressive metamorphism, a preliminary analysis: Geochim. et Cosmochim. Acta, v. 39, p. 955-972.

Evans, B. R., and Wright, T. L., 1972, Composition of liquidus chromite from the 1959 and 1965 eruptions of Kilauea Volcano Hawaii: Am. Mineralogist, v. 57, p. 217-230.

Ewart, A., and Bryan, W. B., 1973, Petrography and geochemistry of the igneous rocks from Eua, Tongan Islands: Geol. Soc. America Bull., v. 83, p. 3281-3298.

Green, D. H., 1964, The petrogenesis of the high temperature alpine-type peridotite at the Lizard, Cornwall: Jour. Petrology, v. 5, p. 134-188.

Green, D. H., and Ringwood, A. E., 1967, The genesis of basaltic magmas: Contr. Mineralogy Petrology, v. 15, p. 103-190.

——— 1967, The stability fields of aluminous pyroxene peridotite and garnet peridotite and their relevance in upper mantle structure: Earth Planetary Sci. Letters, v. 3, p. 151.

Greenbaum, D., ms, 1972, The geology and evolution of the Troodos plutonic complex and associated chromite deposits, Cyprus: Ph.D. thesis, Univ. Leeds, England.

Henderson, P., 1976, Reaction trends shown by chrome spinels of the Rhum Layered intrusion: Geochim. et Cosmochim. Acta, v. 39, p. 1035-1044.

Hess, H. H., 1938, A primary peridotite magma: Am. Jour. Sci., 5th ser. v. 35, p. 321-344.

——— 1941, Pyroxenes of common mafic magmas: Am. Mineralogist, v. 26, p. 515-535.

——— 1955, Serpentines, orogeny and epeirogeny: Geol. Soc. America Spec. Paper 62, p. 391-408.

——— 1964, The oceanic rust, the upper mantle, and the Mayaguez serpentinized peridotite: Natl. Acad. Sci. Natl. Research Council Pub. 1118, p. 169-175.

Hill, R., and Roeder, P., 1974, The crystallization of spinel from basaltic liquid as a function of oxygen fugacity: Jour. Geology, v. 82, p. 709-729.

Himmelburg, G. R., and Coleman, R. C., 1968, Chemistry of primary minerals and rocks from the Red Mountain — Del Puerto ultramafic mass, California: U.S. Geol. Survey Prof. Paper, 600C, p. 18-26.

Himmelburg, G. R., and Loney, R. A., 1973, Petrology of the Vulcan Peak alpine-type peridotite, southwestern Oregon: Geol. Soc. America Bull., v. 84, p. 1585-1600.

Irvine, T. N., 1965, Chromian spinel as a petrogenetic indicator; part 1, theory: Canadian Jour. Earth Sci., v. 2, p. 648-671.

——— 1967, Chromian spinel as a petrogenetic indicator; part 2, petrologic applications: Canadian Jour. Earth Sci., v. 4, p. 71-103.

Irvine, T. N., and Findlay, T. C., 1972, Alpine peridotite with particular reference to the Bay of Islands Complex, *in* The Ancient Ocean Lithosphere: Ottawa, Canada, Dept. Energy, Mines and Resources, Earth Physics Branch, v. 43, p. 97-126.

Irwin, W. P., 1960, Geologic reconnaissance of the northern coast ranges and Klamath Mountains, California: Calif. Div. of Mines Bull. 179, 80 p.

Ito, K., and Kennedy, G. C., 1967, Melting and phase relations in a natural peridotite to 40 kb: Am. Jour. Sci., v. 265, p. 519-538.

Jackson, E. D., 1961, Primary textures and mineral associations in the ultramafic zones of the Stillwater Complex, Montana: U.S. Geol. Survey Prof. Paper 358, 106 p.

——— 1969, Chemical variation in coexisting chromite and olivine in chromite zones of the Stillwater Complex: Econ. Geology Mon. 4, p. 41-71.

Jackson, E. D., and Thayer, T. P., 1972, Some criteria for distinguishing between stratiform, concentric and alpine peridotite-gabbro complexes: Internat. Geol. Cong., 24th, Montreal 1972, sec. 2, p. 289-296.

Kornprobst, J., 1969, Le massif ultrabasique des Beni Bouchera (Rif Interne, Maroc) etude des peridotites de haute temperature et de haute pression et det pyroxenolites, a grenat ou sans grenat, qui leur sont associes: Contr. Mineralogy Petrology, v. 23, p. 283-322.

Kushiro, I., 1968, Composition of magmas formed by partial zone melting of the earth's upper mantle: Jour. Geophys. Research, v. 73, p. 619-634.

——— 1975, On the nature of silicate melt and its significance in magma genesis: regularities in the shift of the liquidus boundaries involving olivine, pyroxene and silica minerals: Am. Jour. Sci., v. 275, p. 411-431.

Kushiro, I., Yoder, H. S. and Mysen, B. O., 1976, Viscosity of basalt and andesite magmas at high pressure: Am. Geophys. Union Trans., v. 57, p. 354.

Loney, R. A., Himmelburg, G. R., and Coleman, R. G., 1971, Structure and petrology of the alpine-type peridotite at Burro Mountain, California, U.S.A.: Jour. Petrology, v. 12, p. 245-309.

MacGregor, I. D., 1965, Aluminous diopsides in the three phase assemblage diopside solid-solution + forsterite + spinel: Carnegie Inst. Washington Year Book 64, p. 134-135.

———— 1967, Mineralogy of model mantle compositions, *in* Wyllie, P. J., ed., Ultramafic and Related Rocks: New York, John Wiley & Sons, p. 382-392.

———— 1970, The effect of CaO, Cr_2O_3, Fe_2O_3 and Al_2O_3 on the stability of spinel and garnet peridotites: Physics Earth and Planetary Interiors, v. 3, p. 372-377.

———— 1974, The system Mg–Al_2O_3–SiO_2: solubility of Al_2O_3 in enstatite for spinel and garnet peridotite compositions: Am. Mineralogist, v. 59, p. 110-119.

MacGregor, I. D., and Smith, C. H., 1963, The use of chrome spinels in petrogenetic studies of ultramafic intrusions: Canadian Mineralogist, v. 7, p. 403.

Malpas, J., and Strong, D. F., 1975, A comparison of chrome spinels in ophiolites and mantle diapirs of Newfoundland: Geochim. et Cosmochim. Acta, v. 39, p. 1045-1060.

McTaggart, K. C., 1971, On the origin of ultramafic rocks: Geol. Soc. America Bull., v. 82, p. 23-42.

Medaris, L. G., 1972, High pressure peridotites in southwestern Oregon: Geol. Soc. America Bull., v. 83, p. 41-58.

Menzies, M., 1973, Mineralogy and partial melt textures within an ultramafic-mafic body, Greece: Contr. Mineralogy Petrology, v. 42, p. 273-385.

Menzies, M., and Allan, C., 1974, Plagioclase lherzolite-residual mantle relationships within two eastern Mediterranean ophiolites: Contr. Mineralogy Petrology, v. 45, p. 197-213.

Miyashiro, A., 1973, The Troodos Complex was probably formed in an island arc: Earth Planetary Sci. Letters, v. 19, p. 218-224.

Miyashiro, A., Shido, F., and Ewing, M., 1969, Composition and origin of serpentinites from the mid-Atlantic ridge near 24° and 30° north latitude: Contr. Mineralogy Petrology, v. 23, p. 117-127.

Moores, E. M., and Vine, F. J., 1971, The Troodos Massif, Cyprus and other ophiolites as oceanic crust: evaluation and implications: Royal Soc. (London) Philos. Trans., ser. A, v. 268, p. 443-466.

Mysen, B. O., and Boettcher, A. L., 1975a, Melting of a hydrous mantle: I. Phase relations of natural peridotite at high pressures and temperatures with controlled activities of water, carbon dioxide and hydrogen: Jour. Petrology, v. 16, p. 520-548.

———— 1975b, Melting of a hydrous mantle: II. Geochemistry of crystals and liquids formed by anatexis of mantle peridotite at high pressures and temperatures as a function of controlled activities of water, hydrogen and carbon dioxide: Jour. Petrology, v. 16, p. 549-593.

O'Hara, M. J., 1963, The join diopside–pyrope at 30 kb pressure: Carnegie Inst. Washington Year Book 62, p. 116-118.

———— 1967, Garnetiferous ultrabasic rocks of orogenic regions, *in* Wyllie, P. J., ed., Ultramafic and Related Rocks: New York, John Wiley & Sons, p. 167-172.

Osborn, E. F., 1969, The complementariness of orogenic andesite and peridotite: Geochim. et Cosmochim. Acta, v. 33, p. 307-324.

Presnall, D. C., 1969, The geometrical analysis of partial fusion: Am. Jour. Sci., v. 267, p. 1178-1194.

Reid, J. B., Jr., 1971, Apollo 12 spinels and petrogenetic indicators: Earth Planetary Sci. Letters, v. 10, p. 351-356.

Rodgers, K. A., 1975, A comparison of the Papuan and New Caledonian ultramafic belts: Jour. Geology, v. 83, p. 47-60.

Ross, C. S., Foster, M. D., and Myers, A. T., 1954, Origin of dunites and of olivine-rich inclusions in basaltic rocks: Am. Mineralogist, v. 39, p. 693-737.

Samson, E., 1969, Discussion — Stillwater, Montana chromite deposits: Econ. Geology Mon. 4, p. 72-75.

Sprye, A., 1969, Metamorphic textures: New York, Pergamon Press, 350 p.

Stevens, R. E., 1944, Composition of some chromites of the western hemisphere: Am. Mineralogist, v. 29, p. 1-34.

Stroh, J. M., 1976, Solubility of alumina in orthopyroxene plus spinel as a geobarometer in complex systems. Applications to spinel-bearing alpine-type peridotites: Contr. Mineralogy Petrology, v. 54, p. 173-187.

Suppe, John, and Armstrong, R. L., 1972, Potassium-argon dating of Franciscan metamorphic rocks: Am. Jour. Sci., v. 272, p. 197-216.

Thayer, T. P., 1946, Preliminary chemical correlation of chromite with the containing rocks: Econ. Geology, v. 41, p. 202-217.

———— 1964, Principle features and origin of podiform chromite deposits, and some observations on the Canleman-Soridag District, Turkey: Econ. Geology, v. 59, p. 1497-1524.

———— 1967, Chemical and structural relations of ultramafic and feldspathinc rocks in alpine intrusive complexes: *in* Wyllie, P. J., ed., Ultramafic and Related Rocks: New York, John Wiley & Sons, p. 222-239.

———— 1969, Gravity differentiation and magmatic re-emplacement of podiform chromite deposits, *in* Magmatic Ore Deposits: Econ. Geology Mon. 4, p. 132-146.

———— 1969, Alpine type sensu-strictu (ophiolitic) peridotites: refractory residue from partial melting or igneous sediments? A contribution to the discussion of the paper "The Origin of ultramafic and ultrabasic rocks" by P. J. Wyllie: Tectonophysics, v. 7, p. 511-516.

Ulmer, G. C., 1969, Experimental investigations of chromite spinels, *in* Magmatic Ore Deposits: Econ. Geology Mon. 4, p. 132-146.

Viljoen, R. P., and Viljoen, J. J., 1969, Evidence for the composition of the primitive mantle and its products of partial melting from a study of the mafic and ultramafic rocks of the Barberton Mountain Land: Geol. Soc. South Africa, Upper Mantle Symp., p. 276-295.

Waff, Harve, 1975, Pressure induced coordination changes in magmatic liquids: Geophys. Research Letters, v. 2, p. 193-196.

Warner, R. D., and euth, W. C., 1974, The diopside-orthoenstatite tow phase region in the system $CaMgSi_2O_6$–$Mg_2Si_2O_6$: Am. Mineralogist, v. 59, p. 98-109.

Williams, H., and Smyth, W. R., 1973, Metamorphic aureoles beneath ophiolite suites and alpine peridotites: tectonic implications with west Newfoundland examples: Am. Jour. Sci., v. 273, p. 594-621.

Wilshire, H. G., and Jackson, E. D., 1975, Problems in determining mantle geotherms from pyroxene compositions of ultramafic rocks: Jour. Geology, v. 83, p. 313-329.

Yaroshevskii, A. A., 1967, The principle of zone melting and its application to certain geochemical problems, *in* Vinogradov, A. P., ed., Chemistry of the Earth's Crust: Jerusalem, Israel, Program for Sci. Translations, v. 2, p. 54-75.