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OLIVINE AND CHROMIAN SPINEL IN PRIMITIVE CALC-ALKALINE AND THOLEIITIC LAVAS FROM THE SOUTHERNMOST CASCADE RANGE, CALIFORNIA: A REFLECTION OF RELATIVE FERTILITY OF THE SOURCE

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ABSTRACT

Chromian spinel and coexisting olivine phenocrysts from a geochemically diverse suite of primitive tholeiitic and calc-alkaline basalts and magnesian andesites from the Lassen region, in the southernmost Cascade Range, in California, show that the sub-arc mantle is zoned. Depleted calc-alkaline basalts and magnesian andesites erupt in the forearc region, and calc-alkaline basalts contain increasing abundances of incompatible elements toward the backarc. High-alumina olivine tholeiites erupt from the arc and backarc areas. Olivine from all these lavas displays a limited compositional range, from Fo_{86} to Fo_{91}, and crystallized at high temperature, generally 1225–1275°C. Chromian spinel trapped in the olivine phenocrysts displays a large range of composition: Cr# values span the range 9–76. Excess Al in the spinel relative to that in 1-atm spinel suggests that it crystallized at elevated pressure. The phenocrysts in these lavas are in equilibrium with their host liquids. The full range of Cr# of the spinel compositions cannot be explained by differentiation or variable pressure, variations in $f(O_2)$, subsolidus equilibration or variations in degree of partial melting of a single peridotitic source. Rather, the systematic compositional differences among phenocrysts in these primitive lavas result from bulk chemical variability in their mantle sources. Correlations between spinel and host-rock compositions support the assertion that the geochemical diversity of Lassen basalts reflects the relative fertility of their mantle sources.

Keywords: chromian spinel, olivine, arc basalt, mantle fertility, Lassen, Cascade arc, California.

SOMMAIRE

Les phénocristaux de spinelle chromifère et d'olivine coexistante provenant d'une suite géochimiquement diversifiée de basaltes tholéiitiques et calco-alcalins et d'andésites magnésiennes provenant de la région de Lassen, dans le secteur sud de la chaîne des Cascades, dans le nord de la Californie, montrent que le manteau sous cet arc est zoné. Les basaltes calco-alcalins stériles et les andésites magnésiennes se trouvent dans l'avant-arc, et les basaltes calco-alcalins davantage enrichis en éléments incompatibles se trouvent plutôt vers l'arrière-arc. Les tholéiites à olivine riches en Al affleurent aussi bien dans la zone axiale de l'arc que dans l'arrière-arc. L'olivine de toutes ces variétés de laves basiques fait preuve d'un intervalle restreint de composition, entre Fo_{86} et Fo_{91}, et a cristallisé à une température relativement élevée, entre 1225° et 1275°C. Le spinelle chromifère piégé dans les phénocristaux d'olivine, en revanche, montre un intervalle important de composition, comme en témoigne l'intervalle des valeurs de Cr#, entre 9 et 76. Un excédent d'aluminium dans le spinelle par rapport aux compositions équilibrées à un atmosphère fait penser que sa cristallisation a eu lieu à une pression élevée. Les phénocristaux dans ces laves semblent équilibrés avec la composition probable du liquide coexistant. L'intervalle complet en Cr# du spinelle des laves ne peut résulter ni d'une différentiation ou d'une pression variable de cristallisation, ni de variations en fugacité d'oxygène ou de taux de fusion partielle d'un seul socle péridotitique, ni de ré-équilibrage subsolidus. Nous croyons plutôt que les variations systématiques en composition impliquant les phénocristaux des laves primitives illustrent une variabilité dans la composition de leurs sources dans le manteau. D'après les corrélations entre la composition du spinelle et celle des roches hôtes, la diversité géochimique parmi les basaltes de la suite de Lassen dépendrait surtout de différences dans la fertilité de leurs sources mantelliques.

(Traduit par la Rédaction)

Mots-clés: spinelle chromifère, olivine, basalte d'arc, fertilité du manteau, Lassen, arc des Cascades, Californie.

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INTRODUCTION

The large range of solid solution in chromian spinel and its common occurrence in basalts and peridotites have promoted its use as a petrogenetic indicator of environment of crystallization. The composition of chromian spinel in igneous rocks exhibits a significant range of variation (Allan *et al.* 1988), and correlates with the major-element composition of coexisting primitive liquids (Irvine 1967, Eales 1979, Dick & Bullen 1984). This observation, combined with the fact that chromian spinel is the liquidus phase in many melts of basaltic composition, suggests that it potentially records information about the most primitive aspects of melt composition. On this basis, spinel compositions have been used to constrain differentiation processes such as fractional crystallization, magma mixing, and magma-chamber recharge (*e.g.*, Natland *et al.* 1983, Allan *et al.* 1988). The strong correlation between the Cr# [100Cr/(Cr + Al)] of spinel and the modal mineralogy of the host peridotite has led several authors to suggest that the composition of spinel in lavas may be used as an indicator of the degree of partial melting (Dick & Bullen 1984, Dick 1989) and source fertility (Fujii 1989). But the composition of chromian spinel also is dependent on intensive variables such as pressure and $f(O_2)$ (Sigurdsson & Schilling 1976, Fisk & Bence 1980, Roeder & Reynolds 1991), and on the extent of melt differentiation (Arai & Takahashi 1987, Arai 1994a). Therefore, before chromian spinel can be applied as a petrogenetic indicator, compositional variation resulting from these factors must be assessed.

Primitive arc lavas from the Lassen region of northern California contain chromian spinel that exhibits a large compositional range. Here, we use the compositions of coexisting chromian spinel and olivine to infer the relative depletion in the peridotite sources of the suite of primitive Lassen lavas. These lavas provide an excellent test of the usefulness of chromian spinel as a petrogenetic indicator because both the degree of partial melting and relative fertility of the source peridotite have been estimated on the basis of major, trace element, and isotope geochemistry (Clynne 1993, Borg *et al.* 1997). We review the compositional dependence of chromian spinel on the intensive variables pressure and $f(O_2)$, as well as on extent of fractional crystallization and partial melting. Subsequently, correlations between the compositions of coexisting spinel and of olivine and geochemical indices of differentiation, partial melting, and source composition in the Lassen lava suite are discussed.

GEOLOGICAL SETTING

The subduction-related magmatic regime of the southernmost Cascade arc was described by Guffanti *et al.* (1990), and summaries can be found in Borg *et al.* (1997) and Bacon *et al.* (1997). Eruption of lavas from hundreds of coalescing volcanoes has built a broad mafic platform in the Lassen area. Individual volcanoes range from monogenetic cinder cones of basalt and basaltic andesite to larger lava cones and shields of basaltic to andesitic composition (Clynne 1990). Linear alignments of vents demonstrate that the locations are related to north- to north–northwest-striking normal faults that reflect extension resulting from overlap of Basin and Range tectonics on the volcanic arc (Guffanti *et al.* 1990). Although primitive lavas are sparse, they are widely distributed in the region, and as a result of crustal extension, are more common than in other parts of the Cascade Range (Bacon *et al.* 1997). The majority of primitive basalts discussed by Clynne (1993) and Borg *et al.* (1997) are less than 0.5 Ma in age, and thus are broadly contemporaneous. Sample locations can be found in Clynne (1993) and Borg *et al.* (1997).

GEOCHEMISTRY AND PETROGENETIC MODELS FOR THE ORIGIN OF LASSEN MAGMAS

This paper focuses on the constraints that can be imposed on models of the petrogenesis of primitive lavas from the Lassen region by the compositions of their phenocrysts. We use the term *primitive* to denote magnesian lavas with simple assemblages of phenocrysts and low contents of phenocrysts (generally <5%), that have high contents of compatible elements (*e.g.*, Ni > 100 ppm, Cr > 200 ppm), and FeO*/MgO values and compositions of ferromagnesian minerals in equilibrium with mantle assemblages (Mg# > 86). Generally, these lavas have ≥8 wt% MgO and <53 wt% SiO_2, although some samples of magnesian andesite also meet these criteria. Lavas containing phenocrysts with reverse zoning or anomalous abundances of one or more compatible elements, or lavas containing multiple populations of phenocrysts, or other evidence suggestive of a complex history, are common in the area, but are not considered here.

Lavas from the Lassen area can be divided into two types; the most magnesian of each type meet the criteria above. The first group is similar to circumpacific calc-alkaline basalt found worldwide, and is called high-alumina basalt by many investigators (*e.g.*, Kuno 1960, Waters 1962, Ewart & LeMaitre 1980, Walker 1981, Kay *et al.* 1982). Calc-alkaline magnesian andesites with primitive mineralogy that are chemically similar to boninites and rocks found in Japan (Tatsumi 1981, 1982) and the Mt. Shasta area (Baker *et al.* 1994) also are present in the Lassen area. The second group is similar to low-K, high-Al olivine tholeiite from the Medicine Lake Highland (Donnelly-Nolan *et al.* 1991, Baker *et al.* 1991) and the Devils Garden area of northeastern California (McKee *et al.* 1983), and is found throughout the northwestern Great Basin (Hart 1985, Hart *et al.* 1984). These lavas were called high-Al olivine tholeiite or HAOT by the above investigators and

Bacon *et al.* (1997). HAOT vents are found from the arc axis to the backarc, but are absent from the forearc.

Calc-alkaline lavas (CA)

Calc-alkaline (CA) lavas display a considerable continuum of compositional diversity (Fig. 1). At one end of the continuum are magmas with low contents of incompatible elements, high ratios of concentrations of large-ion lithophile element to high field-strength elements (*LILE*/*HFSE*), and Sr and Nd isotopic ratios approaching those of mid-ocean-ridge basalt. The high abundance of Sr and high Sr/*LREE* (light rare-earth element) and Sr/P values relative to primitive mantle led Borg at al. (1997) to characterize this end member as having a high $(Sr/P)_N$ value. The other end of the continuum is characterized by magmas with higher abundances of incompatible elements, lower *LILE*/*HFSE*, and Sr, Nd, and Pb isotopic ratios that are like those of ocean-island basalts. Relative enrichment of *LREE* over Sr and low Sr/P led Borg *et al.* (1997) to characterize this end member as having a low $(Sr/P)_N$ value. Clynne (1993) classified mafic lavas of the Lassen area by their K_2O content, using an arbitrary division between lower- and higher-K_2O lavas. Since $(Sr/P)_N$ correlates inversely with the concentration of the *LILE* and particularly K, classifications of CA lavas based on K_2O and Sr/P are essentially equivalent. Low-$(Sr/P)_N$ lavas are equivalent to higher-K basaltic lavas, and high-$(Sr/P)_N$ lavas are equivalent to lower-K basaltic lavas. The highest-$(Sr/P)_N$ lavas are mostly primitive magnesian andesites. Below, we adopt the $(Sr/P)_N$ terminology, and use $(Sr/P)_N = 3.3$ to distinguish low- and high-$(Sr/P)_N$ lavas. The magnesian andesites form a subgroup of high-$(Sr/P)_N$ lavas.

Clynne (1993) and Borg *et al.* (1997) discussed across-arc variation in the geochemistry of primitive lavas. Primitive CA lavas are found from the forearc to the backarc. Although exceptions exist, low-$(Sr/P)_N$ basalts characterize the arc axis and backarc, whereas high-$(Sr/P)_N$ basalts and magnesian andesites characterize the forearc. Using incompatible trace-element abundances, Borg *et al.* (1997) modeled the compositional continuum as the result of the production of arc magmas through the interaction of mantle peridotite with a decreasing proportion of a slab-derived fluid component from the forearc to the backarc. Their models suggest that the degree of melting of the mantle ranges from a maximum of about 10% for high-$(Sr/P)_N$ basalts and magnesian andesites erupted in the forearc to a few percent for the low-$(Sr/P)_N$ basalts erupted in the backarc.

Low-K, high-Al olivine tholeiite (HAOT)

The most striking feature of HAOT from the Lassen region (and in general) is the limited range of major- and trace-element variability displayed by these lavas. Representative results of major- and trace-element analyses of HAOT are given in Bacon *et al.* (1997) and Clynne (1993). Distinctive features of primitive HAOT are low SiO_2 (48 wt%), FeO*/MgO of 0.9, high Al_2O_3 (18 wt%), and low alkali content, especially K_2O (around 0.2 wt% or less). In comparison to calc-alkaline lavas, HAOT have lower concentrations of *LILE*, higher concentrations of *HREE*, lower *LILE*/*HFSE*, and lower *LREE*/*HREE* (Fig. 1). HAOT display weak enrichments of Ba, Sr, and Pb relative to *LILE* or *LREE*. Sr and Nd isotopic systematics of HAOT are distinct from those of CA lavas, primarily in having higher $^{144}Nd/^{143}Nd$ at equivalent $^{87}Sr/^{86}Sr$ (Bullen & Clynne 1989, Borg *et al.* 1997, Bacon *et al.* 1997).

Bartels *et al.* (1991) have experimentally equilibrated a magnesian HAOT composition with a spinel

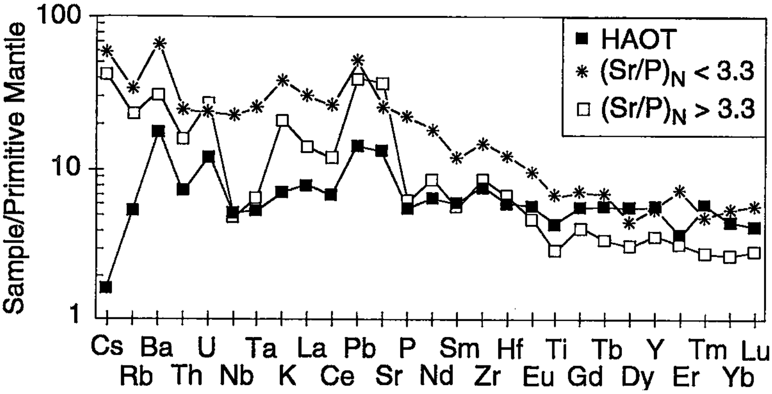


FIG. 1. Spider diagram showing the differences between the types of Lassen basalts. Average trace-element concentrations for the groups of lavas, taken from data reported by Clynne (1993), Borg *et al.* (1997), and Bacon *et al.* (1997), are normalized to the primitive mantle values of Sun & McDonough (1989).

lherzolite assemblage at a temperature of about 1290°C and pressure of ~11 kbar. Clynne (1993) suggested that HAOT were derived by approximately 10% partial melting of depleted subcontinental mantle subjected to Fe and Al enrichment by addition of pyroxene through metasomatism. Baker *et al.* (1994) concluded that HAOT at Mt. Shasta represents a nearly anhydrous 6–10% partial melt of subcontinental mantle that last equilibrated near the base of the crust. Bacon *et al.* (1997) attributed their weak subduction-related geochemical signature (elevated Sr, Ba, Pb, and Sr/P) to passage of ancient (probably Mesozoic) arc magmas through their mantle source-region. Clynne (1993) proposed and Bacon *et al.* (1997) agreed that the mantle source-regions of HAOT and CA lavas are compositionally distinct.

ANALYTICAL TECHNIQUES

Minerals were analyzed with an automated nine-spectrometer ARL SEMQ electron microprobe at the U.S. Geological Survey in Menlo Park. We used the matrix-correction procedures of Bence & Albee (1968) and the modifications of Albee & Ray (1970). Corrections were made for interferences of Cr*K*β on Mn*K*α, and Ti*K*β on V*K*α for spinel analyses. Synthetic and natural minerals were used as standards. The proportion of Fe^{3+} in spinel was calculated by charge balance and stoichiometry (fixing 8 positive charges and 3 cations per formula unit). The Tiebaghi chromite (USBM 117075, Jarosewich *et al.* 1980) was used as an internal standard for analyses of chromian spinel. Precision and accuracy associated with analyses of the Tiebaghi chromite are reported in Clynne (1993), and compare favorably with those reported by Ballhaus *et al.* (1991) and Forsythe & Fisk (1994).

MINERALOGY

CA lavas in the Lassen region, like arc basalts and andesites worldwide, tend to be porphyritic, but we

TABLE 1. REPRESENTATIVE COMPOSITIONS OF OLIVINE IN HAOT AND CALC-ALKALINE LAVAS

	LC84-569	LC86-1046	LC88-1311	LC86-951	LC88-1398	LC82-970	LC85-735	LC84-514	LC83-347	LC85-677	LC88-1314	LM87-1333	LC86-831	LM92-2443	LC85-671	LC82-905
SiO_2 (wt%)	39.94	40.22	39.91	39.84	40.52	39.91	39.88	40.34	39.74	40.00	39.37	39.59	40.00	40.38	39.99	39.80
MgO	47.49	47.12	47.20	47.12	46.84	47.25	46.40	46.96	44.82	46.45	46.42	46.21	46.24	47.93	46.44	46.97
FeO	12.18	11.72	11.67	11.71	12.10	12.34	12.73	13.67	15.21	13.08	12.93	13.72	13.43	11.62	12.54	12.20
MnO	0.20	0.21	0.21	0.19	0.20	0.22	0.22	0.20	0.25	0.23	0.18	0.23	0.23	0.16	0.19	0.16
NiO	0.30	0.26	0.27	0.34	0.29	0.30	0.28	0.32	0.22	0.30	0.37	0.24	0.37	0.32	0.37	0.31
CaO	0.21	0.26	0.28	0.21	0.24	0.29	0.24	0.22	0.17	0.18	0.13	0.17	0.15	0.18	0.16	0.18
Total	100.31	99.79	99.55	99.41	100.19	100.31	99.74	100.71	100.41	100.24	99.41	100.16	100.41	100.59	99.69	99.62
Si	0.989	0.994	0.983	0.980	1.003	0.990	0.990	0.997	0.996	0.995	0.975	0.989	0.995	0.995	0.990	0.984
Mg	1.753	1.735	1.734	1.728	1.728	1.746	1.716	1.729	1.675	1.722	1.713	1.721	1.714	1.760	1.715	1.731
Fe	0.252	0.242	0.241	0.241	0.250	0.256	0.264	0.261	0.319	0.272	0.268	0.287	0.279	0.239	0.260	0.252
Mn	0.004	0.004	0.004	0.004	0.004	0.005	0.005	0.004	0.005	0.005	0.004	0.005	0.005	0.003	0.004	0.003
Ni	0.006	0.005	0.005	0.007	0.006	0.006	0.006	0.006	0.005	0.006	0.007	0.005	0.007	0.006	0.007	0.006
Ca	0.005	0.007	0.007	0.006	0.006	0.008	0.006	0.006	0.005	0.005	0.004	0.005	0.004	0.005	0.004	0.005
Fo	87.4	87.8	87.8	87.8	87.4	87.2	86.7	86.9	84.0	86.4	86.5	85.7	86.0	88.0	86.8	87.2
Ni (ppm)	2318	2035	2122	2672	2255	2381	2200	2483	1760	2389	2939	1847	2892	2538	2892	2420
$(Sr/P)_N$*	HAOT	HAOT	HAOT	HAOT	HAOT	HAOT	HAOT	HAOT	1.25	1.45	1.52	1.52	1.53	1.57	1.67	1.69

	LC86-834	LC86-1006	LC88-1408	LC88-1371	LC88-1305	LC86-1005	LC88-829	LM91-2116	LM92-2697	LC88-1303	LC86-1056	LC88-1312	LC83-255	LC88-1310	LC88-1308	LC86-1009
SiO_2 (wt%)	40.81	40.73	40.69	40.01	39.49	40.13	40.38	40.99	41.16	40.72	40.21	41.51	40.59	41.09	40.72	40.51
MgO	48.69	50.58	49.83	47.48	43.10	45.24	49.16	49.54	50.16	49.63	47.42	48.80	46.45	48.93	47.91	48.94
FeO	10.89	8.49	9.23	12.80	16.68	14.16	10.31	10.32	9.65	10.07	11.36	9.91	11.57	9.14	10.67	9.54
MnO	0.14	0.14	0.17	0.21	0.24	0.23	0.17	0.14	0.16	0.16	0.19	0.19	0.16	0.15	0.19	0.13
NiO	0.32	0.38	0.38	0.19	0.31	0.23	0.33	0.34	0.32	0.25	0.10	0.15	0.41	0.35	0.27	0.73
CaO	0.14	0.13	0.14	0.21	0.11	0.12	0.17	0.16	0.19	0.13	0.22	0.21	0.12	0.11	0.20	0.08
Total	100.99	100.44	100.46	100.89	99.94	100.10	100.53	101.49	101.52	100.96	99.50	100.78	99.31	99.77	99.96	99.93
Si	0.994	0.991	0.993	0.988	1.000	1.003	0.990	0.993	0.992	0.992	0.987	1.009	0.994	1.001	1.002	0.994
Mg	1.778	1.834	1.813	1.749	1.626	1.685	1.797	1.789	1.807	1.802	1.735	1.768	1.696	1.777	1.758	1.790
Fe	0.223	0.173	0.188	0.264	0.353	0.296	0.211	0.209	0.195	0.205	0.233	0.201	0.237	0.186	0.220	0.196
Mn	0.003	0.003	0.004	0.004	0.005	0.005	0.004	0.003	0.003	0.003	0.004	0.004	0.003	0.003	0.004	0.003
Ni	0.006	0.008	0.008	0.004	0.006	0.005	0.007	0.007	0.006	0.005	0.002	0.003	0.008	0.007	0.005	0.014
Ca	0.004	0.004	0.004	0.005	0.003	0.003	0.005	0.004	0.005	0.004	0.006	0.005	0.003	0.003	0.005	0.002
Fo	88.9	91.4	90.6	86.9	82.2	85.1	89.5	89.5	90.3	89.8	88.2	89.8	87.7	90.5	88.9	90.1
Ni (ppm)	2538	3002	3010	1477	2467	1776	2593	2672	2538	1941	802	1202	3245	2758	2153	4420
$(Sr/P)_N$*	1.91	2.42	2.50	3.04	3.31	3.79	3.89	3.91	4.66	4.79	5.10	5.35	5.47	5.88	5.96	6.59

Compositions derived from microprobe data. See Table 2 for host rock SiO_2 and MgO; * calc-alkaline lavas show value for $(Sr/P)_N$

emphasize that the majority of lavas selected for detailed study are primitive basalts and magnesian andesites that have simple assemblages of minerals (many of them contain olivine as the only phenocryst phase) and a low content of phenocrysts (<5%). Although titaniferous magnetite is the stable oxide phase in the groundmass of CA lavas, the most magnesian lavas usually contain chromian spinel preserved as inclusions in olivine phenocrysts. Crystals in primitive CA lavas are small, usually less than 1 mm in maximum dimension, and unzoned, except for a thin, normally zoned quench-induced rim. The groundmass ranges from being holocrystalline to holohyaline, and is composed of plagioclase, clinopyroxene, and titaniferous magnetite, although olivine also is present in some of the more magnesian CA lavas.

HAOT lavas are distinctive rocks, typically aphyric or with sparse olivine or plagioclase (or both) in a holocrystalline, usually diktytaxitic groundmass (as defined by Williams *et al.* 1954) composed of plagioclase, clinopyroxene or olivine (or both), and titaniferous magnetite. Most HAOT lavas contain chromian spinel as inclusions in olivine phenocrysts, and a few also contain chromian spinel as microphenocrysts. Slightly evolved HAOT lavas may contain glomeroporphyritic clots composed of olivine and plagioclase, accompanied rarely by clinopyroxene.

About 30% of the primitive CA lavas reported by Clynne (1993) and Borg *et al.* (1997) contain clinopyroxene phenocrysts. Clinopyroxene compositions fall in a small range around $Wo_{45}En_{50}Fs_5$, contain high Cr contents (generally 0.5–1.0 wt% Cr_2O_3), and correspond to chromian diopside or chromian Mg-rich augite. They have Mg# [100Mg(Mg + Fe^{2+})] values in the range 88–92 and are in equilibrium with the host-rock composition if one assumes a K_D of 0.23 (Grove *et al.* 1982). The compositions of the clinopyroxene in CA lavas are similar in all respects to those found in primitive lavas in other arcs (*e.g.*, BVSP 1981, Tatsumi & Ishizaka 1982, Kay & Kay 1985, Luhr & Carmichael 1985, Bailey *et al.* 1989), and are similar in major components to those found in many lherzolites (Dick & Fisher 1984, Wilshire *et al.* 1988).

Plagioclase phenocrysts are sparsely present in about 25% of the primitive CA lavas and nearly 50% of HAOT lavas described in Clynne (1993) and Borg *et al.* (1997). Plagioclase phenocrysts in CA lavas are generally small, up to a few mm in size and unzoned, but occasionally display weak normal or reverse zoning toward their margin. The composition of plagioclase in CA lavas ranges from An_{60} to An_{90}. The composition of plagioclase phenocrysts in HAOT is in the range An_{72-83}; the phenocrysts are unzoned or display weak normal zoning of a few mol.% An. The An content of plagioclase from HAOT and of low- to high-$(Sr/P)_N$ CA lavas plot as distinct parallel linear arrays *versus* whole-rock CaO/Na_2O, and have Sr and K_2O contents appropriate for equilibrium with their hosts (Clynne 1993).

Olivine

Olivine is a ubiquitous phenocryst phase in both CA and HAOT lavas. Most of the primitive lavas in this study contain less than 5% olivine, usually as small, <1 mm, unzoned to weakly zoned, euhedral crystals. Primitive CA lavas have core compositions in the range Fo_{86-91}, with 2500–3500 ppm Ni (Table 1, Fig. 2A). In general, the most magnesian forsterite (Fo_{89-91}) occurs in CA lavas with high $(Sr/P)_N$. The Ca content of forsterite from CA lavas ranges from 0.10 to 0.22 wt% CaO (Fig. 2B).

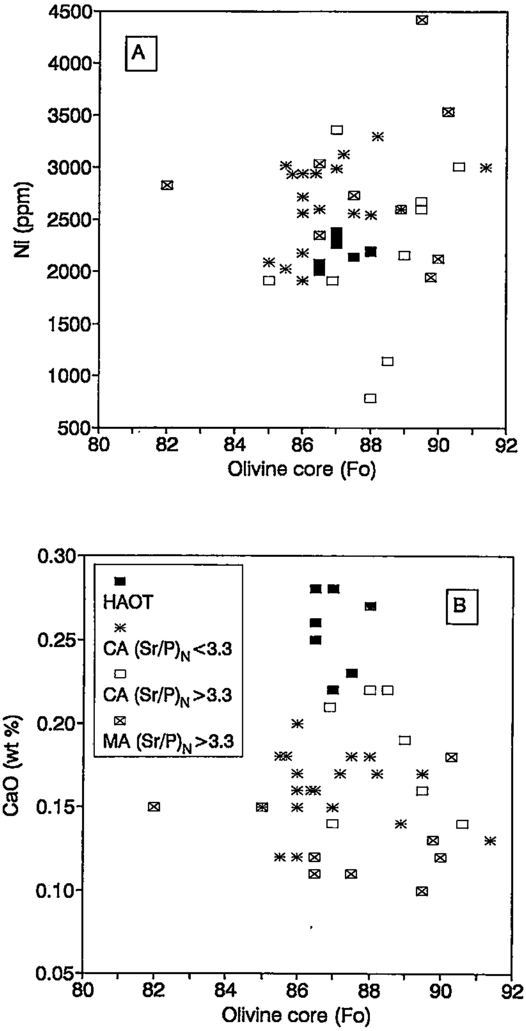


FIG. 2. Plot of abundances of (A) NiO and (B) CaO *versus* Fo of olivine phenocrysts from HAOT and CA basalt and magnesian andesite (MA) lavas from the Lassen region. Olivine in HAOT lavas has consistently higher levels of Ca, and generally lower levels of Ni, than olivine in CA lavas.

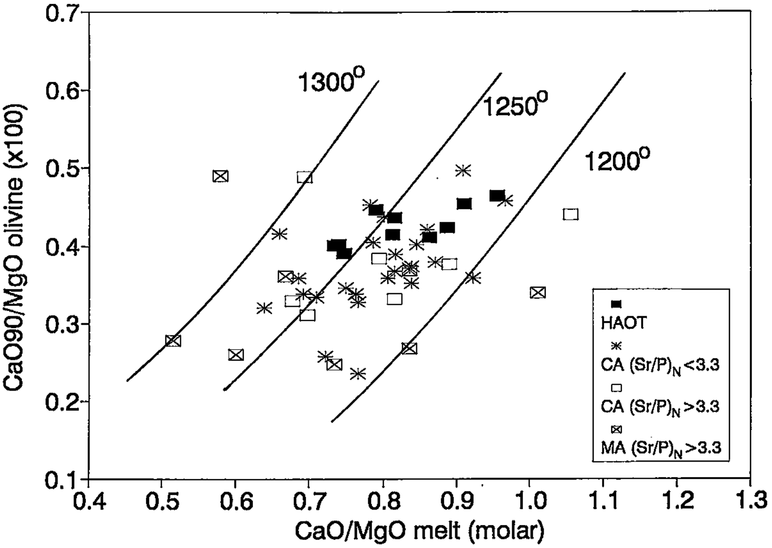


FIG. 3. CaO_{90}/MgO (×100) in olivine *versus* molar CaO/MgO in coexisting whole rock (melt), where CaO_{90} is the Ca content of olivine normalized to Fo_{90} olivine (Jurewicz & Watson 1988). Temperature contours from Jurewicz & Watson (1988) have an uncertainty of ±25°C.

Olivine phenocrysts in sparsely phyric HAOT lavas are euhedral and generally unzoned or display weak normal zoning of a few mol.% Fo. The range of core compositions is quite restricted, generally Fo_{86-88}, with Ni content in the range 2000–2400 ppm, and Ca content in the range 0.22–0.28 wt% CaO (Table 1, Figs. 2A, B). Compared to the sparsely phyric lavas, the olivine phenocrysts in porphyritic HAOT lavas shows a slightly wider and more evolved compositional range (Fo_{82-86}), with generally lower Ni content, in the range 1500–2000 ppm, and 0.22 wt% CaO. Olivine phenocrysts in HAOT contain a greater amount of Ca than phenocrysts in CA lavas having a similar Fo content (Fig. 2B).

There is considerable disagreement on the factors that influence the partitioning of Ca between magnesian olivine and melt; temperature, pressure, concentrations of Ca and Fe in the melt have all been cited as controlling factors by various investigators. A synopsis and critical evaluation of the role of each of these factors are given by Jurewicz (1986), and these factors were investigated experimentally by Jurewicz & Watson (1988). Their results indicate that the Ca content of olivine depends on the concentration of Ca in the melt and the relative activity of iron, as indicated by the Fe/Mg of the olivine; pressure and $f(O_2)$ have no direct influence at conditions relevant to crystallization of olivine in Lassen magmas. Jurewicz & Watson (1988) reported a partition coefficient for Ca in olivine of 0.021 ± 0.015, and presented an empirical equation to evaluate the effect of iron on the equilibrium Ca content of magmatic olivine. The measured partition-coefficient [normalized to Fo_{90} using the procedure outlined by Jurewicz & Watson (1988), and assuming that bulk-rock Ca equals the amount of Ca in the melt] for olivine in the Lassen samples is 0.021 ± 0.008, identical to that of Jurewicz & Watson (1988). Crystallization temperatures based on amount of Ca in olivine (Jurewicz & Watson 1988) range from about 1275 to about 1225°C for the majority of HAOT and CA lavas (Fig. 3). Temperatures calculated for HAOT using an empirical function based on whole-rock SiO_2 and MgO content (Albarède 1992) are mostly in the range from 1250 to 1315°C, whereas temperatures calculated for CA lavas are mostly lower, but in the range 1200–1300°C.

It is important to demonstrate that in terms of trace elements, the phenocrysts are in equilibrium with whole-rock (wr) compositions. At equivalent Fo, olivine phenocrysts in most primitive CA lavas contain higher Ni than olivine phenocrysts in HAOT, despite similar to lower Ni content in the host lava (Fig. 2A). Measured D_{Ni}(ol/wr) are plotted against whole-rock MgO content in Figure 4. Most fall close to the D_{Ni}(ol/wr) reported by Hart & Davis (1978). The exceptions are mostly samples of high-$(Sr/P)_N$ magnesian andesite, and their apparently higher D_{Ni} is probably due to the increase of D_{Ni} of olivine with increasing SiO_2 of the melt and with decreasing temperature of crystallization (Kinzler *et al.* 1990). Olivine is less abundant in the magnesian andesites in which D_{Ni} is high; instead, these rocks also

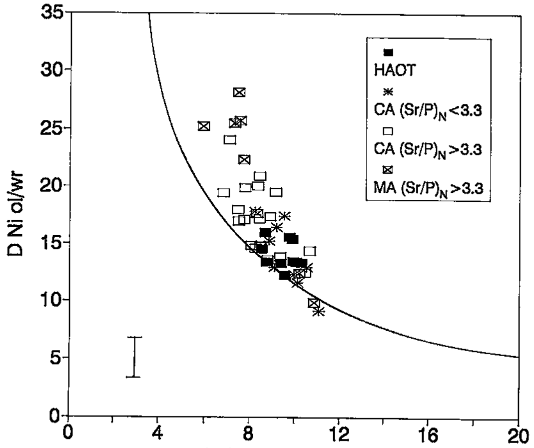


FIG. 4. Distribution coefficient for Ni in olivine *versus* MgO in whole rock. Solid line denotes the equilibrium curve of Hart & Davis (1978) for olivine in equilibrium with liquid. Error bar approximates the error in the measurement of Ni in whole rock and olivine (Clynne 1993). Hart & Davis (1978) did not report an uncertainty for the distribution coefficient.

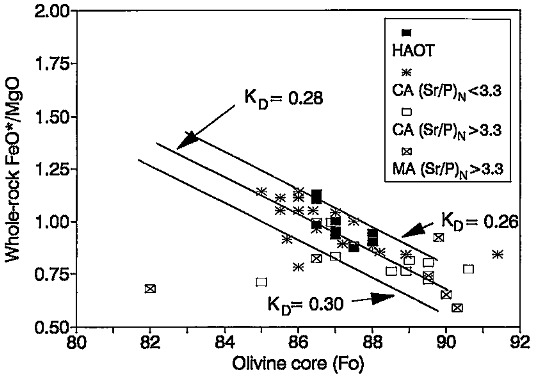


FIG. 5. Plot of whole-rock FeO*/MgO *versus* composition of the core of forsterite phenocrysts. Solid lines display contours of olivine–liquid K_D. The majority of HAOT and CA lavas plot between K_D values of 0.29 and 0.26, and thus are in equilibrium with whole-rock FeO*/MgO. See text for discussion of the exceptions.

contain clinopyroxene phenocrysts. Consequently, they may have undergone some combination of olivine accumulation or fractionation, which may well have modified D_{Ni}(ol/wr). However, the presence of chromian spinel and high-Ni olivine at low whole-rock FeO*/MgO values indicates that crystal fractionation has not significantly modified their bulk compositions, whereas relatively low whole-rock Ni content and olivine – whole rock equilibrium suggest that olivine accumulation is likewise not an important factor in the evolution of the magnesian andesites. Consequently, the higher SiO_2 of the high-$(Sr/P)_N$ magnesian andesites is probably the most important factor in the origin of their high D_{Ni}.

Olivine compositions of HAOT and CA lavas are plotted as a function of whole-rock total iron as FeO *versus* MgO in Figure 5. The majority of HAOT and CA lavas form a coherent trend lying between the K_D = 0.26 and 0.28 contours, slightly lower than the K_D = 0.29 determined for Medicine Lake Volcano basalt at the quartz – fayalite – magnetite (QFM) buffer by Grove *et al.* (1982). Thus the data suggest that HAOT and CA lavas exhibit equilibrium between olivine phenocrysts and host lava in terms of FeO*/MgO, probably at a $f(O_2)$ higher than QFM. A few samples of magnesian andesite contain olivine with lower Fo content than expected for their FeO*/MgO, yet are sparsely phyric and contain high levels of Ni. These lavas also contain chromian spinel with high Fe^{2+}/Fe^{3+}, and may have crystallized under more reducing conditions than the majority of CA lavas. Conversely, a few samples show evidence for a small amount of olivine accumulation.

Chromian spinel

The majority of primitive lavas of the Lassen area contain chromian spinel trapped as inclusions in forsterite phenocrysts. Spinel crystals trapped sequentially in growing phenocrysts may record changes in composition due to crystal fractionation, decompression or destabilization of chromian spinel. The ranges of composition of chromian spinel in many Lassen lavas are similar to those reported by Luhr & Carmichael (1985), Arai & Takahashi (1987), and Umino *et al.* (1991). The compositions reported here are from the core of euhedral grains located in the core of olivine phenocrysts, and are representative of those with lowest Cr#, highest Mg#, and lowest proportion of the titaniferous magnetite component taken from the array of analyzed chromian spinel crystals in each rock examined (Clynne 1993). Thus the effects of fractional crystallization and reaction with host liquid (see below) are minimized. Further discussion is limited to this specially selected group of chromian spinel compositions, which are reported in Table 2, along with the forsterite content of the host olivine phenocryst monitored directly adjacent to the chromian spinel.

Chromian spinel is found included in olivine phenocrysts, but does not occur in the groundmass of CA lavas. In the CA lavas, the grains are smaller (10–50 μm in diameter, 20 μm is typical), but more abundant (there are generally ten to one hundred crystals per cross section of an olivine phenocryst) than in HAOT lavas. The crystals range in color from golden brown through

TABLE 2. REPRESENTATIVE COMPOSITIONS OF CHROMIAN SPINEL IN HAOT AND CALC-ALKALINE LAVAS

	LC84-569	LC86-1046	LC88-1311	LC86-951	LC88-1398	LC82-970	LC85-735	LC84-514	LC83-347	LC85-677	LC88-1314	LM87-1333	LC86-831	LM92-2443	LC86-855	LC85-671
SiO_2 (wt%)	0.00	0.00	0.00	0.00	0.00	0.11	0.06	0.05	0.00	0.00	0.00	0.00	0.00	0.08	0.00	0.00
TiO_2	0.32	0.15	0.18	0.33	0.26	0.18	0.18	0.23	0.98	0.66	0.96	0.87	0.98	0.49	0.62	0.68
Al_2O_3	41.88	58.16	57.11	45.93	47.72	56.52	52.44	49.63	36.87	39.83	25.15	36.39	30.40	22.89	38.02	28.30
Cr_2O_3	21.88	8.50	10.01	18.58	18.79	9.61	12.01	14.89	22.46	23.10	33.53	24.37	28.72	39.91	21.55	31.20
FeO	12.56	10.44	10.85	11.41	11.77	11.80	10.03	11.83	16.61	13.96	16.87	14.65	16.17	14.42	16.45	15.20
Fe_2O_3	6.48	3.31	3.35	6.21	3.83	2.62	5.20	5.64	8.40	5.93	10.26	8.23	9.04	7.48	9.77	10.06
MnO	0.15	0.10	0.11	0.14	0.14	0.12	0.13	0.16	0.20	0.13	0.20	0.13	0.20	0.23	0.19	0.17
MgO	17.23	20.31	20.16	18.53	18.53	19.22	19.89	18.62	14.47	16.29	12.97	15.74	13.91	13.94	14.45	14.25
NiO	0.25	0.29	0.27	0.28	0.22	0.27	0.23	0.30	0.17	0.24	0.26	0.22	0.30	0.08	0.18	0.28
CaO	0.00	0.02	0.00	0.01	0.01	0.03	0.00	0.00	0.01	0.01	0.02	0.00	0.00	—	0.01	0.00
V_2O_3	0.09	0.08	0.11	0.09	0.13	0.10	0.12	0.11	0.18	0.14	0.14	0.17	0.18	0.16	0.14	0.17
Total	100.84	101.37	102.15	101.50	101.39	100.58	100.29	101.47	100.36	100.29	100.36	100.78	99.90	99.68	101.38	100.31
Si	0.000	0.000	0.000	0.000	0.000	0.003	0.001	0.001	0.000	0.000	0.000	0.000	0.000	0.002	0.000	0.000
Ti	0.007	0.003	0.003	0.007	0.005	0.003	0.004	0.005	0.021	0.014	0.022	0.019	0.022	0.011	0.013	0.015
Al	1.366	1.753	1.719	1.458	1.505	1.732	1.628	1.554	1.250	1.321	0.901	1.223	1.064	0.826	1.275	0.994
Cr	0.481	0.173	0.203	0.397	0.399	0.198	0.251	0.314	0.513	0.516	0.809	0.552	0.677	0.970	0.487	0.738
Fe^{2+}	0.291	0.223	0.232	0.257	0.263	0.256	0.221	0.263	0.400	0.328	0.429	0.349	0.402	0.369	0.391	0.379
Fe^{3+}	0.135	0.064	0.064	0.126	0.007	0.051	0.103	0.113	0.182	0.126	0.235	0.176	0.202	0.172	0.209	0.226
Mn	0.003	0.002	0.002	0.003	0.003	0.003	0.003	0.003	0.005	0.003	0.005	0.003	0.005	0.006	0.005	0.004
Mg	0.711	0.774	0.768	0.744	0.739	0.745	0.781	0.738	0.621	0.683	0.588	0.669	0.616	0.637	0.613	0.633
Ni	0.006	0.006	0.006	0.006	0.005	0.006	0.005	0.006	0.004	0.006	0.006	0.005	0.007	0.002	0.004	0.007
Ca	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.001	0.000	0.000	—	0.000	0.000
V	0.002	0.002	0.002	0.002	0.003	0.002	0.003	0.002	0.004	0.003	0.003	0.004	0.004	0.004	0.003	0.004
Mg#	71.0	77.6	76.8	74.3	73.7	74.4	78.0	73.7	60.8	67.6	57.8	65.7	60.6	63.3	61.0	62.6
Cr#	26.0	9.0	10.6	21.4	21.0	10.3	13.4	16.8	29.1	28.1	47.3	31.1	38.9	54.0	27.6	42.6
Fe#	6.8	3.2	3.2	6.3	3.9	2.6	5.2	5.7	19.3	6.4	12.1	9.0	10.4	8.7	10.6	11.5
ol Fo*	87.4	87.8	87.8	87.8	87.4	87.3	86.7	87.0	84.0	86.4	86.5	85.7	86.0	88.0	84.2	86.8
host SiO_2	49.58	48.12	48.37	48.84	49.50	48.33	48.61	48.58	51.23	51.17	50.82	50.71	52.81	51.60	49.94	51.35
host MgO	8.71	10.08	9.41	9.56	10.30	9.78	8.76	9.90	6.83	7.53	8.43	7.74	7.09	9.34	8.26	8.05
$(Sr/P)_N$.**	HAOT	HAOT	HAOT	HAOT	HAOT	HAOT	HAOT	HAOT	1.25	1.45	1.52	1.52	1.53	1.57	1.66	1.67

	LC82-905	LC86-834	LC88-1408	LC88-1371	LC88-1305	LC86-1005	LC86-829	LM91-2116	LM92-2697	LC86-1056	LC88-1312	LC83-255	LC85-743	LC88-1310	LC88-1308	LC86-1009
SiO_2 (wt%)	0.04	0.13	0.09	0.09	0.00	0.00	0.09	0.11	0.10	0.00	0.00	0.00	0.00	0.00	0.00	0.00
TiO_2	0.51	0.31	0.30	0.43	0.23	0.48	0.34	0.34	0.35	0.42	0.29	0.91	0.71	0.27	0.27	0.50
Al_2O_3	35.60	45.33	25.35	32.37	17.84	23.33	29.32	28.91	19.46	32.95	31.49	14.15	18.30	10.70	28.21	13.61
Cr_2O_3	26.85	18.14	39.50	29.36	43.74	37.22	34.32	35.39	45.28	31.72	35.05	39.74	41.35	51.92	36.17	49.32
FeO	13.59	9.95	11.45	14.77	22.32	19.00	12.36	11.82	15.14	13.49	12.31	19.89	17.52	17.26	13.63	16.23
Fe_2O_3	7.31	6.57	6.90	7.71	7.21	8.85	7.87	6.95	5.93	5.30	3.42	15.30	10.45	7.20	6.55	7.40
MnO	0.17	0.13	0.17	0.22	0.30	0.23	0.13	0.20	0.29	0.18	0.13	0.20	0.17	0.23	0.21	0.21
MgO	16.05	19.15	16.11	14.74	8.19	11.37	16.08	16.35	13.04	15.94	16.20	9.84	11.88	10.50	15.15	12.04
NiO	0.23	0.21	0.21	0.02	0.11	0.05	0.22	0.07	0.10	0.06	0.13	0.22	0.13	0.11	0.16	0.13
CaO	0.00	—	—	—	0.00	0.05	—	—	—	0.01	0.03	0.02	0.01	0.03	0.03	0.01
V_2O_3	0.14	0.11	0.12	0.18	0.10	0.20	0.15	0.12	0.10	0.22	0.18	0.16	0.15	0.13	0.17	0.25
Total	100.49	100.03	100.20	99.89	100.03	100.79	100.88	100.26	99.79	100.29	99.22	100.43	100.66	98.36	100.55	99.70
Si	0.001	0.004	0.003	0.003	0.000	0.000	0.003	0.003	0.003	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Ti	0.011	0.006	0.007	0.010	0.005	0.011	0.007	0.008	0.008	0.009	0.006	0.022	0.017	0.007	0.006	0.012
Al	1.200	1.452	0.890	1.119	0.681	0.847	1.009	0.999	0.715	1.123	1.086	0.543	0.677	0.421	0.982	0.517
Cr	0.609	0.391	0.934	0.683	1.124	0.910	0.795	0.823	1.121	0.728	0.814	1.026	1.031	1.374	0.848	1.261
Fe^{2+}	0.325	0.226	0.285	0.362	0.604	0.489	0.302	0.290	0.395	0.326	0.301	0.541	0.460	0.481	0.337	0.437
Fe^{3+}	0.157	0.134	0.154	0.170	0.176	0.205	0.173	0.153	0.139	0.115	0.075	0.374	0.247	0.181	0.146	0.179
Mn	0.004	0.003	0.004	0.005	0.008	0.006	0.003	0.005	0.008	0.004	0.003	0.006	0.005	0.006	0.005	0.006
Mg	0.684	0.776	0.715	0.644	0.396	0.523	0.700	0.715	0.606	0.687	0.706	0.478	0.556	0.522	0.667	0.578
Ni	0.005	0.005	0.005	0.000	0.003	0.001	0.005	0.002	0.003	0.001	0.003	0.006	0.003	0.003	0.004	0.003
Ca	0.000	—	—	—	0.000	0.002	—	—	—	0.000	0.001	0.001	0.000	0.001	0.001	0.000
V	0.003	0.002	0.003	0.004	0.003	0.005	0.003	0.003	0.003	0.005	0.004	0.004	0.004	0.004	0.004	0.006
Mg#	67.8	77.5	71.5	64.0	36.9	51.6	69.9	71.2	60.6	67.8	70.1	46.9	54.7	52.0	66.5	57.0
Cr#	33.7	21.2	51.2	37.9	62.3	51.8	44.1	45.2	61.0	39.3	42.8	65.4	60.3	76.6	46.3	70.9
Fe#	8.0	6.8	7.8	8.6	8.9	10.4	8.7	7.8	7.0	5.9	3.8	19.3	12.6	9.1	7.4	9.2
ol Fo	87.3	88.9	90.6	86.9	80.6	85.1	89.5	89.5	90.3	88.0	89.8	86.0	86.7	91.0	88.9	89.0
host SiO_2	50.36	49.38	54.08	51.99	53.68	51.13	51.51	51.17	51.54	50.57	50.83	54.47	54.53	56.14	51.47	58.18
host MgO	8.43	10.46	8.89	8.42	10.86	11.06	9.20	10.15	13.21	8.24	9.98	7.73	7.55	10.20	9.09	7.57
$(Sr/P)_N$.**	1.69	1.91	2.50	3.04	3.29	3.79	3.89	3.91	4.66	5.10	5.35	5.47	5.83	5.88	5.96	6.59

Compositions derived from microprobe data. * composition of host olivine immediately adjacent to analyzed spinel; Mg#= $(100Mg/(Mg+Fe^{2+})$, Cr#= (100Cr)/(Cr+Al), Fe#= $(100Fe^{3+})/(Fe^{3+}+Cr+Al)$, ** $(Sr/P)_N$ of host rock for calc-alkaline lavas shades of reddish brown that become increasingly dark with increasing Cr and Fe content. Chromian spinel with a significant proportion of the titaniferous magnetite component (more than 1–2 wt% TiO_2) is abundant in slightly to moderately evolved lavas, and can be readily identified optically because it is opaque (Sigurdsson 1977).

Primitive CA lavas from the Lassen region contain chromian spinel with a wide range of Cr# at high Mg# (Fig. 6); this range is similar to that reported from all other arcs combined (Arai 1992). The composition of chromian spinel in CA lavas correlates with $(Sr/P)_N$ and Al content of the host lava. Low-$(Sr/P)_N$ basalts contain a chromian spinel with a Cr# in the range 20–50, slightly to significantly less aluminous that the spinel in HAOT lavas. High-$(Sr/P)_N$ basalts and magnesian andesites contain a chromian spinel with a large range of Cr#, from 40 to 77, but most are in the range 50–70. There is a good negative correlation between Cr# and Mg# in the collection of chromian spinel compositions in all Lassen basalts and magnesian andesites, and a good positive correlation between Cr# and $(Sr/P)_N$ for CA lavas (Fig. 6).

Chromian spinel is occasionally found in HAOT lavas as euhedral phenocrysts up to 1 mm in size, but usually occurs as sparse (a few crystals per olivine phenocryst), 10- to 100-μm euhedral crystals in olivine phenocrysts; 40- to 60-μm diameters are typical, and they are greenish brown to light yellow-brown in color.

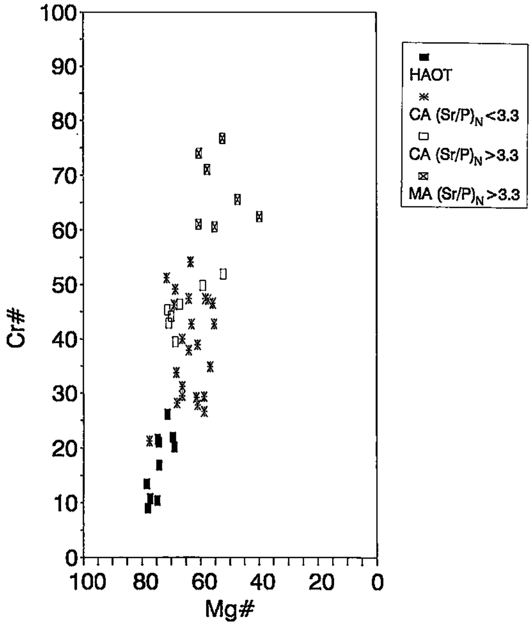


FIG. 6. Cr# and Mg# of chromian spinel found in HAOT lavas and the array of low- to high-$(Sr/P)_N$ CA lavas from the Lassen region.

Grains of chromian spinel in HAOT contain high levels of Al and Mg, and low levels of Cr and Fe, have Cr# in the range 9–26, and Mg# in the range 80–65 (Fig. 6), and are similar to aluminous spinel found in MORB and abyssal peridotite (Dick & Bullen 1984). Although there is considerable overlap, grains of chromian spinel in HAOT generally contain lower amounts of Fe^{3+}, Ti, and V, and higher amounts of Ni than chromian spinel in CA lavas.

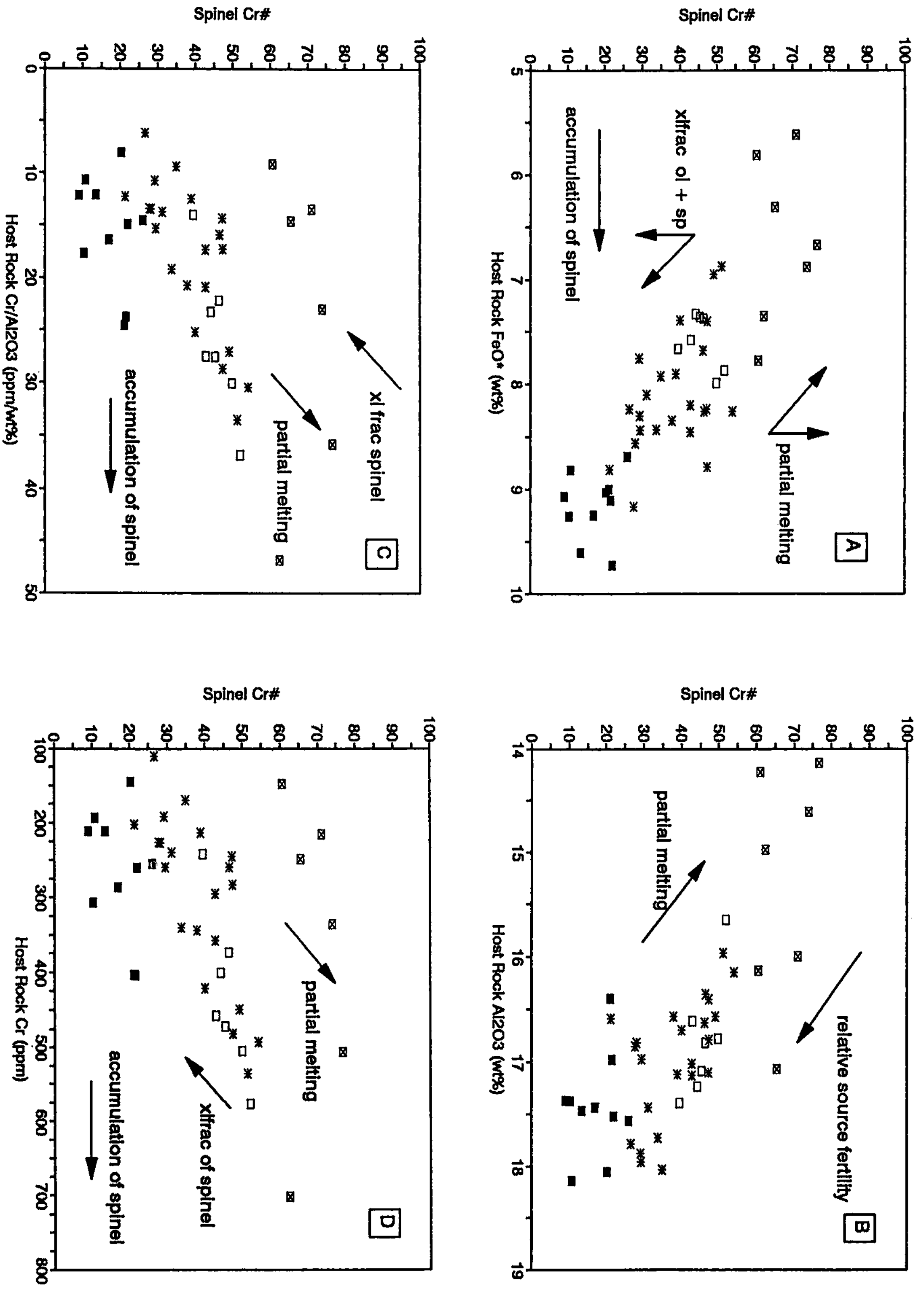
Spinel compositions plotted against host FeO* and Al_2O_3 of primitive lavas from the Lassen area reveal systematic correlations. The Cr# of spinel shows a good negative correlation with FeO* and Al_2O_3 of the host lava for all Lassen samples of basalt (Figs. 7A, B). The entire range of spinel Cr# is present at constant host-lava Cr/Al_2O_3 and Cr in Lassen lavas (Figs. 7C, D).

FACTORS THAT AFFECT THE COMPOSITION OF SPINEL IN VOLCANIC ROCKS

The composition of the chromian spinel crystallizing as the liquidus phase from a primitive magma should be close to that of the residual spinel in the source region. Differences between residual and phenocryst spinels reflect differences between the physical conditions at the sites of melt formation and crystallization. The analysis of chromian spinel crystallized in natural and experimental systems demonstrates that its composition is affected by extent of fractional crystallization, pressure, $f(O_2)$, and degree of partial melting. However, the only significant factor affecting the composition of chromian spinel crystallizing at liquidus temperature is change in melt composition (Allan *et al.* 1988). Below, we review the factors that affect the composition of spinel in volcanic rocks and evaluate their relevance in connection with the origin of the compositional array of chromian spinel found in the Lassen rocks.

Effect of fractional crystallization on spinel composition

Chromian spinel is typically the liquidus phase in mafic magmas, but crystallizes over only a narrow temperature-interval (Hill & Roeder 1974, Fisk & Bence 1980, Luhr & Carmichael 1985, Arai & Takahashi 1987), so that its occurrence is typically restricted to minimally fractionated lavas (Sigurdsson & Schilling 1976). The composition of chromian spinel closely reflects the composition of its host melt, and is therefore dependent on differentiation processes (Irvine 1965, 1976, Allan *et al.* 1988, Sack & Ghiorso 1991a). Chromian spinel reacts with its host liquid either during fractional crystallization, or once it ceases to be a stable phase (Ridley 1977, Tatsumi & Ishizaka 1982, Luhr & Carmichael 1985). Preservation of chromian spinel with a near-liquidus composition requires that the magma be rapidly decompressed, erupted, and quenched, or that olivine enter the crystallizing assemblage simultaneously with or


A
Spinel Cr#
Host Rock FeO* (wt%)
partial melting
xlfrac ol + sp
accumulation of spinel
B
Spinel Cr#
Host Rock Al2O3 (wt%)
relative source fertility
partial melting
C
Spinel Cr#
Host Rock Cr/Al2O3 (ppm/wt%)
xl frac spinel
partial melting
accumulation of spinel
D
Spinel Cr#
Host Rock Cr (ppm)
partial melting
xlfrac of spinel
accumulation of spinel

shortly following chromian spinel to encapsulate and preserve spinel crystals (Maaløe & Hansen 1982).

The Cr# of spinel may increase, decrease or remain constant as the Fo of coexisting olivine decreases during crystal fractionation in magmas (Arai 1992). Crystallization of calcic plagioclase raises the Cr# of the chromian spinel by lowering the Al content of the liquid, but has minimal effect on the Mg# of coexisting silicate phases. Crystallization of olivine lowers the Mg# of the liquid, coexisting phases, and spinel, but lowers the Cr# of spinel only slightly (Dick & Bullen 1984, Arai 1994b). Clinopyroxene and chromian spinel are normally not stably coexisting phases in mafic magmas (Irvine 1965, 1967). Since chromian spinel remains on the liquidus for only a small interval of temperature, and plagioclase is not an important phase in primitive Lassen lavas, the equilibrium compositions of spinel are not expected to be strongly affected by small amounts of olivine or plagioclase fractionation. The phenocrysts of forsterite with high Ni content confirm that the Lassen lavas considered here are little fractionated.

The primary way in which differentiation of the host basalt affects the composition of the chromian spinel occurs upon cessation of its crystallization as a liquidus phase. The spinel component ($MgAl_2O_4$) is rapidly replaced by magnetite (Fe_3O_4) and ulvöspinel (Fe_2TiO_4) components, converting chromian spinel sequentially to chromian titaniferous magnetite and titaniferous magnetite (Luhr & Carmichael 1985). Clynne (1993) observed that spinel grains located on cracks in olivine phenocrysts also may re-equilibrate with melt. Consequently, we avoided analyzing grains located along cracks in their olivine hosts. The TiO_2 content, Fe^{2+}/Fe^{3+} values, and Fe^{3+} content, as measured by Fe^{3+}# [$100Fe^{3+}/(Cr + Al + Fe^{3+})$] are used to assess the extent of solid solution toward magnetite and ulvöspinel (Arai 1992). Small amounts of the magnetite and ulvöspinel components do not significantly affect the Cr# of spinel. Here, chromian spinel is considered to be unaffected by solid solution toward magnetite and ulvöspinel if Fe^{2+}/Fe^{3+} exceeds 1.5, TiO_2 is less than 1.5 wt%, and Fe^{3+}# is less than or equal to 10. Most spinel compositions showing such solid solution were eliminated from the data set.

Because all grains of chromian spinel in CA lavas are included in olivine, accumulation of chromian spinel requires accumulation of olivine. The observation that FeO* of the host magma decreases with increasing Cr# of spinel prohibits the accumulation of olivine or chromian spinel from playing any role in the origin of the compositional array (Fig. 7A). We conclude that the systematic differences in Cr# and Mg# among spinel populations in the Lassen lavas are not the result of fractional crystallization (or assimilation combined with fractional crystallization) or of accumulation of chromian spinel with olivine (Arai 1992, Luhr & Carmichael 1985). However, it is possible that some of scatter in Figures 7A–D may be due to small amounts of accumulation of chromian spinel (+ olivine).

FIG. 7. (A) Cr# of spinel *versus* FeO* of host lava for primitive Lassen lavas. Arrows schematically illustrate the direction of evolution for various magmatic processes. Symbols as in Figure 6. Accumulation of spinel has a negligible effect, but accumulation of olivine plus spinel would have a significant effect on the position of data points on this plot, driving them to higher FeO*. Crystal fractionation may contribute to the scatter of the array, but can be seen not to have a significant influence on these parameters. Partial melting could produce the array from a single composition of peridotite composition by varying the degree of melting over a large range. If so, the degree of melting must increase in the order HAOT → lower-$(Sr/P)_N$ CA → higher-$(Sr/P)_N$ CA → magnesian andesite. However, that order is inconsistent with trace-element systematics [Fig. 1; Clynne (1993), Borg *et al.* (1997), Bacon *et al.* (1997)]. (B) Cr# of spinel *versus* Al_2O_3 content of host lava. Variable fertility of the source and variable degree of partial melting are consistent with the array for Lassen lavas, but partial melting is ruled out as in (A). (C) Cr# of spinel *versus* Cr/Al_2O_3 content of the host lava. Accumulation or crystal fractionation of olivine has a negligible effect on Cr/Al_2O_3 of the host lava. Accumulation of chromian spinel could explain some of the variation of Cr/Al_2O_3, especially in CA lavas with high Cr. Increased degree of partial melting could explain some of the increase in Cr# for the array of low- to high-$(Sr/P)_N$ CA lavas, but cannot relate HAOT and CA lavas. (D) Cr# of spinel *versus* Cr content of host lava. Accumulation of chromian spinel could explain some of the variation in Cr content. Increased degree of partial melting is consistent with the array for the groups of CA lavas, but cannot relate HAOT and CA lavas. In conjunction with Figures 7A–C and trace-element systematics, variation in source composition is the only explanation consistent with the spinel data.

Effect of pressure and $f(O_2)$ on spinel composition

Pressure affects the Al-content and Al/Cr ratio of chromian spinel: at equal temperature and melt composition, spinel crystallized at a higher pressure is expected to have higher Al content and Al/Cr value than spinel crystallized at low pressure (Dick & Bullen 1984, Green *et al.* 1971, Ballhaus *et al.* 1991). Roeder & Reynolds (1991) reported a decrease in the Cr# of spinel of about 1 per kbar in experiments on basalt between 1 atm and 10 kbar. Comparison of experimental results on lherzolite by Jaques & Green (1980) at 2–15 kbar indicates a decrease of Cr# of spinel of about 1.5 per kbar. Thus, extreme and systematic variation in pressure would be necessary if pressure variation was the major cause of Cr# variation in spinels in Lassen lavas.

Spinel–liquid equilibria were calculated for spinel-bearing HAOT and CA lavas using an algorithm by T.D Bullen (written comm.) who adapted the method developed by Allan *et al.* (1988) to account for the Ti^{4+} component in addition to Cr^{3+}, Al^{3+}, and Fe^{3+} in the spinel's composition. The results are plotted along with the compositions of coexisting spinel–olivine crystallized at low pressure over a significant interval of temperature (1200–1300°C) and range of $f(O_2)$ (quartz – fayalite – magnetite to nickel – nickel oxide) from a wide variety of natural lavas, MORB glasses, and experimental charges (Fig. 8). Chromian spinel – olivine pairs from Lassen basalts are consistently displaced from the 1-atm equilibrium as a result of consistently elevated Al in the spinel. The wide ranges in composition and conditions used to construct the 1-atm array make it is unlikely that differences in temperature, $f(O_2)$ or bulk composition contribute significantly to displacement of spinel – olivine pairs from Lassen from the 1-atm trend.

The compositions of spinel – olivine pairs from peridotite and synthetic compositions in high-pressure experiments are plotted along with spinel – olivine pairs

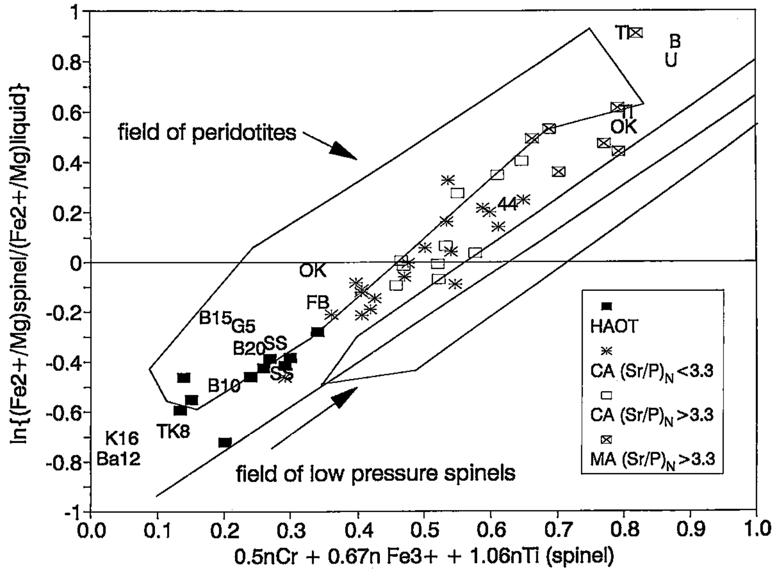


FIG. 8. Comparison of composition of chromian spinel from the suite of Lassen basalts with spinel from low- and high-pressure experiments, MORB glasses, primitive basalts, and spinel peridotites. The comparison is based on an adaptation of the approach of Allan *et al.* (1988), which takes into account the Ti component (coefficient from Sack & Ghiorso 1991b) in addition to Cr, Al, and Fe^{3+} in the spinel's composition. However, for most samples, the position on the abscissa is primarily a measure of the Cr/Al of the spinel. The figure uses data for spinel and contiguous olivine. Liquid Fe^{2+}/Mg is calculated from measured olivine Fe^{2+}/Mg and a K_D of 0.30 for peridotites, 0.275 for MORB (Allan 1994), and for Lassen and arc lavas. The solid line denotes a linear regression (r^2 = 0.88, standard error of Y estimate = 0.11) of the data in the field of low-pressure natural (Allan *et al.* 1988, Allan 1994, Fisk & Bence 1980, Sigurdsson & Schilling 1976) and experimentally produced (Green *et al.* 1971, Sack *et al.* 1987, Thy *et al.* 1991) spinel–olivine pairs. Spinel in the Lassen basalts contain excess Al relative to low-pressure spinel. It is similar to spinel in volcanic rocks interpreted to have crystallized at high pressure and to spinel in a variety of mantle nodules. Data for spinel peridotites from Brey *et al.* (1990), Frey & Prinz (1978), Galer & O'Nions (1989), Griffin *et al.* (1984), Press *et al.* (1986), Sigurdsson (1977), Stolz & Davies (1988), Umino & Yoshizawa (1996), and Wilshire *et al.* (1988). The composition of high-pressure spinel from volcanic rocks is shown by FB (MORB: Fisk & Bence 1980), SS (MORB picrite: Sigurdsson & Schilling 1976), 44 (Jor 44 primitive CA basalt from Jorullo volcano, Mexico: Luhr & Carmichael 1985), TI (Shodo-shima, Japan: Tatsumi & Ishizaka 1982), B (primitive ankaramite from Epi, Vanuatu: Barsdell & Berry 1990), OK (primitive CA basalts from Okmok, Aleutian arc: Nye & Reid 1986), and U (boninite from Bonin Islands: Umino 1986). Data points of the form B10 indicate the pressure in kbar of spinel–olivine equilibria in experiments on natural lherzolitic compositions; B: Brey *et al.* (1990), G: Green *et al.* (1972), Ba: Bartels *et al.* (1991), K: Kinzler & Grove (1992), TK: Takahashi & Kushiro (1983).

equilibrated at high pressure from primitive basalt, and spinel lherzolite and harzburgite xenoliths (Fig. 8). Here, the spinel also exhibits an elevated Al content relative to 1-atm natural and experimental results, which supports an interpretation of high-pressure crystallization of the spinel in all the Lassen basalts. Some of this difference might be due to re-equilibration of spinel and olivine (see discussion below), and spinel and pyroxene, in peridotites at subsolidus temperatures (Roeder 1994), but quickly cooled high-pressure experiments on peridotites also plot in the field of natural peridotites. The agreement between the 1-atm line and MORB suggests that re-equilibration is not important in quickly cooled primitive lavas (Allan *et al.* 1988, Allan 1994). A few HAOT lavas contain chromian spinel enclosed in olivine and in the groundmass. The spinel grains in the groundmass have a symplectitic texture, with a core of spinel surrounded by plagioclase and liquid. The core zone of the symplectitically intergrown grains of spinel and the spinel in the olivine phenocrysts have the same composition and support the interpretation that spinel in Lassen basalts crystallized at high pressure.

HAOT lavas contain a spinel with the lowest Cr# of the suite of Lassen lavas, but these probably equilibrated at a depth equivalent to a lower pressure (~11 kbar: Bartels *et al.* 1991) than the CA lavas. Therefore, the difference in composition of spinel in Lassen lavas cannot be primarily the result of variable pressure of crystallization.

Experiments by Roeder & Reynolds (1991) on basalts equilibrated at various oxygen fugacities indicate that the relative content of trivalent cations (Al, Fe^{3+}, and Cr) in spinel is affected by $f(O_2)$. As $f(O_2)$ increases, Cr_2O_3, Al_2O_3 and Fe^{2+}/Fe^{3+} decrease as the proportion of Fe_2O_3 increases. Fe^{3+} substitutes for Cr in preference over Al in the spinel structure, so that increasing $f(O_2)$ results in a decrease of approximately 1 to 2 in Cr# per log unit $f(O_2)$. The Ti^{4+}, Fe^{2+}, and Mg contents of the spinel are not strongly affected by $f(O_2)$. As $f(O_2)$ increases, Fe^{2+} slightly decreases and Mg slightly increases, resulting in a small increase of Mg# with $f(O_2)$. Thus, the array of spinel compositions in Lassen lavas cannot be explained by any plausible variation in $f(O_2)$.

Re-equilibration of spinel and olivine

Chromian spinel may exchange Fe^{2+} and Mg with its olivine host or adjacent crystals during cooling. In intrusive and metamorphic rocks, this exchange has been used as a geothermometer (Sack & Ghiorso 1991b). The Cr# of spinel is not affected by exchange with olivine because of the lack of sites for Cr and Al in the olivine structure. Scowen *et al.* (1991) reported the re-equilibration of spinel and olivine in slowly cooled basalt in the Kilauea Iki lava lake as compared to quickly cooled scoria. Ozawa (1984) evaluated the extent of re-equilibration of chromian spinel with olivine as rocks cool. Above about 850°C, slowly cooled olivine and spinel maintain equilibrium, and below about 600°C, the rate of diffusion becomes too small to affect compositions on a relevant time-scale. At cooling rates appropriate for volcanic rocks, Fe^{2+} and Mg exchange is minimal, amounting to a maximum decrease of about 5 mol.% Mg in the core of spinel grains 10–20 μm in diameter. The core of spinel grains 50 μm in diameter is unaffected. Because of the large reservoir of olivine relative to spinel, the effect on the olivine host is a few tenths of a mol.% at most. Similar Fe^{2+}/Mg values of spinel trapped in olivine and in the groundmass in some samples of HAOT, similar compositions of small and large grains of spinel in the same rock, and identical compositions for spinel trapped in olivine and clinopyroxene in one sample (Clynne 1993) suggest that Fe^{2+}–Mg exchange between spinel and its olivine host is not important in determining its composition in primitive Lassen basalts. Nevertheless, some of the scatter on Figures 6 and 8 may be attributed to re-equilibration of olivine and spinel.

Partial melting

Experiments by Jaques & Green (1980) and Baker & Stolper (1994) demonstrate that the composition of spinel in restite assemblages changes with increasing degree of partial melting. As melting proceeds, the Al and Mg contents of spinel decrease, and the Cr and Fe^{2+} contents increase, as expected from the coupled substitution of (Cr + Fe^{2+}) for (Al + Mg). Thus the Cr# of spinel increases and the Mg# of spinel decreases with increasing degree of partial melting. The change in spinel Cr# resulting from increase in melting is estimated to be approximately 1.5 Cr# per 1% increase in partial melting (Fig. 9), on the basis of anhydrous experiments on fertile peridotite (Baker & Stolper 1994). If these experiments are applicable to magmas with moderate H_2O contents, generation of the range of spinel compositions observed in Lassen basalts from a single source would require 40–45% variation in the degree of melting. However, geochemical modeling of the CA lavas indicates that they can be explained by a small range of melting, from a maximum of about 10% in the forearc to a few percent in the backarc (Borg *et al.* 1997). HAOT lavas can be generated by probably no more than 10% melting (Bartels *et al.* 1991). Thus, to produce the observed variation in the chromian spinel, unrealistic amounts of partial melting are required, especially in light of the geochemical modeling.

Variations in source fertility

The olivine–spinel mantle array, as defined by Arai (1987, 1990), is the result of compositional covariation observed in forsterite (Fo) content of olivine and the Cr# of coexisting spinel from xenoliths and mantle-derived peridotites (Fig. 9). The Cr# of spinel varies from near 0 to over 80, compared with a 10 mol.% variation

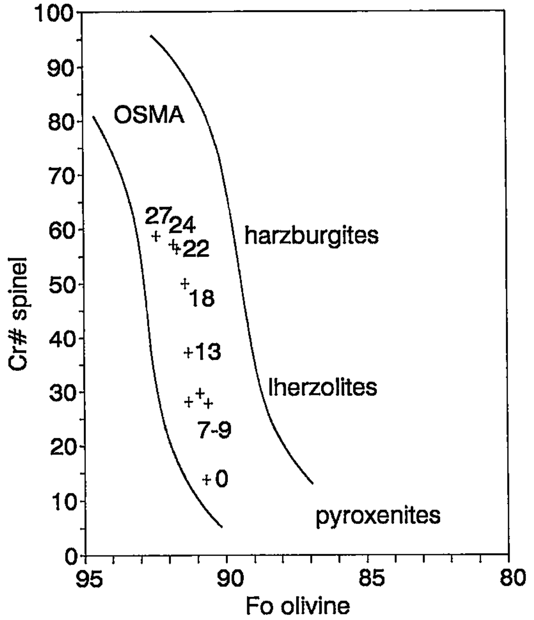


FIG. 9. The olivine–spinel mantle array (OSMA) of Arai (1987). Points give the composition of olivine–spinel pairs from results of progressive melting experiments on a fertile spinel lherzolite, reported by Baker & Stolper (1994). Numbers indicate the degree of melting (%) of the lherzolite.

in Fo. Fo and Cr# are positively correlated, whereas the Fo and Mg# of spinel correlate negatively. The composition of the chromian spinel correlates with the modal mineralogy of the host peridotite, such that Cr# increases as the peridotite becomes more refractory (Dick & Fisher 1984). Below, we use the term *fertility* to indicate the relative amount of a mafic melt that can be extracted from a peridotite at a given temperature, pressure, and volatile concentration. The relative fertility of peridotites is indicated in several ways. Compositionally, the increasing relative fertility is indicated by higher concentration of the easily fusible major-element components (especially Fe, Al, Ti, Na, and H_2O). Mineralogically, increasing relative fertility is indicated by the increasing Al and Na content and decreasing Mg/Fe^{2+} of silicate minerals, and by decreasing Cr# of spinel. Modally, increasing relative fertility is indicated by decreasing olivine:pyroxene and orthopyroxene:clinopyroxene ratios, and by increasing abundance of spinel and hydrous minerals. Thus, fertile spinel lherzolites and pyroxenites contain a lower-Cr# spinel and lower-Fo olivine, and less fertile harzburgites contain a higher-Cr# spinel and higher-Fo olivine (Dick & Fisher 1984).

In the following discussion, we refer to source fertility as an important control on the compositions of chromian spinel in Lassen basalts. We realize that magmas are in equilibrium with restites, not initial source-compositions. However, because Lassen basalts were derived by relatively small degrees of melting (Borg *et al.* 1997), the initial compositions of spinel crystallizing from the magmas are not significantly different from the spinel compositions in the source.

The wide range of spinel Cr# at a low Cr content of the host lava suggests that spinel Cr# is not controlled simply by Cr concentration of host-magma (Fig. 7D). We suggest that the Cr# of spinel in Lassen lavas is in part a function of the spinel Cr# in the mantle source of those lavas, and that there is considerable variability of the Cr# of spinel in the sources. Compared to high-$(Sr/P)_N$ CA lavas, the more aluminous compositions of spinel and whole rocks, higher FeO*/ MgO, less Ni-rich olivine, and lower SiO_2 of HAOT are consistent with the hypothesis that they are derived from a more fertile source than high-$(Sr/P)_N$ CA basalts, especially if HAOT is generated by a similar degree of melting. Low-$(Sr/P)_N$ CA lavas contain a less aluminous spinel than HAOT, have lower FeO*/MgO, and contain olivine with similar Fo as in HAOT, but have higher SiO_2. In general, they contain a more aluminous spinel, and have a higher FeO*/MgO and less magnesian olivine than the majority of the high-$(Sr/P)_N$ CA lavas. Hence, their characteristics suggest a source with fertility between HAOT and high-$(Sr/P)_N$ CA lavas. Of all the high-$(Sr/P)_N$ CA lavas, the relatively rare magnesian andesites contain the most Cr-rich spinel and most Mg-rich olivine, have the lowest FeO*/MgO, and highest SiO_2. These features suggest that they are derived from an even less fertile source than the majority of the high-$(Sr/P)_N$ CA lavas.

To further test the relationship of source fertility and basalt geochemistry, we make use of trace-element correlations between spinel composition and host-lava geochemistry in primitive Lassen lavas. Consider the relationship between Cr# of spinel and the Yb content of the host lava. The Yb content is perhaps the best geochemical measure, and Cr# of spinel, the best mineralogical measure, of the history of melting and incompatible-element depletion of peridotites. Specifically, decreasing whole-rock Yb and increasing Cr# of spinel reflect decreasing fertility of the peridotite (McDonough & Frey 1989, Dick & Fisher 1984). Consequently, if the Yb content of the primitive magmas directly reflects the relative fertility of their sources, then Cr# of liquidus spinel should increase as Yb content of the host lava decreases. The data for the primitive Lassen region lavas display a linear trend that satisfies this constraint (Fig. 10A).

The K and La contents of HAOT lavas increase with increasing Cr# of spinel (Fig. 10B). This correlation indicates that the range of HAOT cannot be related simply by increasing degree of partial melting. Instead, we suggest that the incompatible element content of HAOT magmas is a function of source fertility. That is, melting proceeds until the easily fusible component of

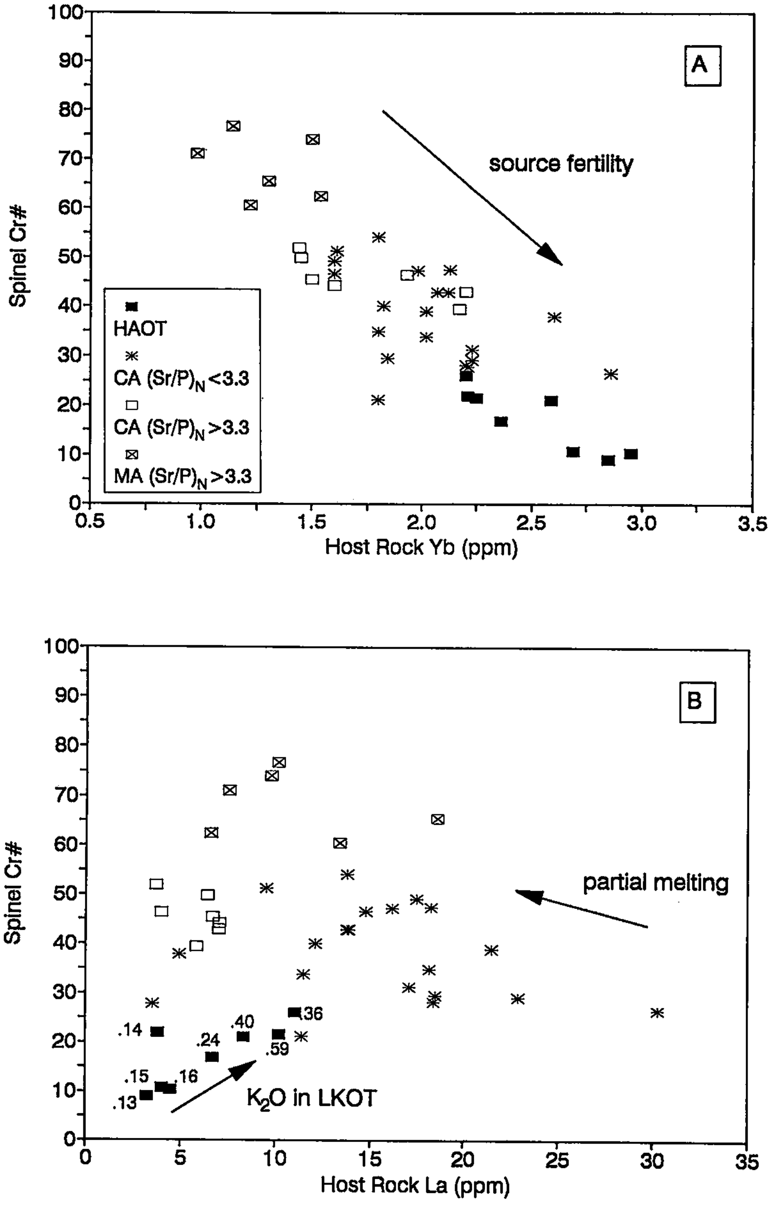


FIG. 10. (A) Host-rock Yb content (Clynne 1993, Borg *et al.* 1997) *versus* Cr# of spinel for primitive Lassen basalts. The clear correlation between decreasing Yb content and increasing Cr# of spinel links source mineralogy and geochemistry, and reflects the decreasing fertility of the CA source relative to the HAOT source. Arrows indicating changes with degree of melting are schematic. (B) Host-rock La content *versus* Cr# of spinel for primitive Lassen basalts. Note the increasing La and K_2O content with increasing Cr# for HAOT. This correlation is inconsistent with derivation of the HAOT array by increasing the degree of melting, and suggests that source fertility or incompatible-element content of the source peridotite or both control the incompatible-element content of HAOT lavas. The low- to high-$(Sr/P)_N$ CA basalts could be related by variation in degree of melting, but the source of the high-$(Sr/P)_N$ magnesian andesites must have had different initial *LREE* contents and spinel composition.

the source is exhausted, and the degree of melting is primarily a function of relative fertility. Thus, the most fertile source melts the most extensively.

The opposite correlation exists for CA basalts: in general, the abundances of the incompatible elements (*e.g.*, La) decrease with increasing Cr# of spinel (Fig. 10B) and $(Sr/P)_N$ (Borg *et al.* 1997). The correlation between Cr# of spinel and Cr content of host lava for low- and high-$(Sr/P)_N$ CA lavas is partially explained by a small increase in Cr# of residual spinel as partial melting proceeds, but a variably fertile source is probably still required (Figs. 7C–D). No correlation between Cr# of spinel and incompatible element abundance exists for the magnesian andesites. Furthermore, the relatively refractory source-compositions indicated by their Cr-rich spinel probably cannot melt to a high degree. Thus the composition of magnesian andesites is probably not controlled by a variable degree of partial melting. Baker *et al.* (1994) estimated the H_2O content of magnesian basaltic andesites at Mt. Shasta to approach 6%, and Borg *et al.* (1997) showed that these lavas have a larger proportion of slab component than other CA lavas. Thus, high-$(Sr/P)_N$ magnesian andesites owe their origin to small degrees of melting of refractory peridotite fluxed by a H_2O-rich slab component.

Figures 7A–D suggest that a variable degree of partial melting or relative fertility of the source peridotite determines the composition of primitive Lassen lavas. Because increased degree of partial melting in effect reduces source fertility, it will be difficult to distinguish between the two processes. However, it is apparent that neither process alone or in conjunction with the other can explain production of the entire range of Lassen basalts from a single source.

Borg *et al.* (1997) proposed a model to explain the major- and trace-element systematics of Lassen CA basalts that is consistent with the spinel data presented here. In their model, relatively fertile and incompatible-element-rich peridotite in the mantle wedge is progressively depleted by melting as it moves beneath the backarc to the forearc region. Melting of relatively infertile, depleted peridotite is facilitated by an increasing component of slab-derived fluid toward the forearc, which also produces the enhanced arc geochemical signature of the magmas toward the forearc.

CONCLUSIONS

Two important conclusions can be made from the composition of the phenocrysts carried in HAOT and primitive CA lavas in the Lassen region.

1. The phenocryst compositions are primitive and in equilibrium with peridotite. The simple assemblages of minerals, high contents of compatible elements and low Fe/Mg values of olivine and clinopyroxene, and the presence of chromian spinel, indicate that both HAOT and primitive CA lavas are mantle-derived and little fractionated, and the lavas have not significantly accumulated phenocrysts. In terms of Fe/Mg, olivine and clinopyroxene are in equilibrium with the whole rock. The Al/Cr ratio of the chromian spinel suggests that the phenocrysts of chromian spinel (and olivine) in both HAOT and primitive CA lavas crystallized at elevated pressure. The level of Ca in olivine from both HAOT and primitive CA lavas indicates crystallization in the range 1225–1275°C, probably near the base of the crust.

2. The full range of Cr# of spinel compositions in the Lassen lavas cannot be explained by differentiation or variable pressure, $f(O_2)$ or degree of partial melting. Rather, the systematic compositional differences between phenocrysts in HAOT and primitive CA lavas, and between the low-$(Sr/P)_N$ and high-$(Sr/P)_N$ groups of CA lavas, result from bulk chemical variability in their mantle sources. Correlations between spinel and host-rock compositions support the assertion that the geochemical diversity of Lassen basalts reflects the relative fertility of their mantle sources. The relative fertility of the mantle sources of Lassen lavas decreases in the order HAOT lavas – low-$(Sr/P)_N$ CA basalt – high-$(Sr/P)_N$ CA basalt – high-$(Sr/P)_N$ magnesian andesite.

The observation that compositions of chromian spinel included in olivine reflect the relative fertility of mantle sources of basaltic magmas is especially useful in studies of arc petrogenesis, where the geochemical behavior of incompatible elements can be complicated by mixed contributions from several sources. However, caution is necessary in applying patterns of spinel compositions to the petrogenesis of their host lavas. To obtain useful information about the source from chromian spinel, it is essential to study primitive lavas with simple assemblages of primary minerals, and even then, it is usually necessary to look through the complex evolutionary history, clues of which are preserved in the population of spinel in the volcanic rocks.

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